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QUATERNARY GEOLOGY OF CANADA

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XII. Quaternary Geology of Canada

INTRODUCTION

The term Quaternary refers to about the last million years of the earth's history and is set apart from the Tertiary period by reason of the climatic changes that gave rise to successive glaciations of vast regions, and to a general lowering of snowlines throughout the world. Some recent oceanographical and paleontological data suggest that the climatic changes actually may have begun as early as 3 million years ago.

The geology of the glacial and non-glacial deposits that mantle the bedrock and various aspects of the landscape that are attributable to Quaternary events are discussed in this chapter. The term Quaternary in the writer's opinion, may be interchanged with Pleistocene as applied to Canada. This term is frequently used in Canada as it more readily conveys the glacial and climatic connotations, and accordingly is generally used here. The Pleistocene also includes the present, very short, non-glacial interval which is commonly termed Recent. In Canada, Recent may be regarded as comprising the last 7,000 years—the period following dissipation of the major part of the last mainland ice sheet.

At present, about 10 per cent of the earth's land surface is covered by glacier ice whereas during former glaciations as much as 30 per cent was under ice and permanent snowfields. About 97 per cent of Canada has been glaciated, and hence this country contains more glaciated terrain than any other; at present about one per cent remains under glacial cover—in the Queen Elizabeth Islands, Baffin Island, and in the mountains of western Canada. An area of about 70,000 square miles in the western Yukon, in the shadow of the coastal mountains, escaped glaciation altogether. Two elongate areas along the mountain front west of Mackenzie River in Northwest Territories, comprising about 4,000 square miles, are also thought to be partly unglaciated, but have been little studied. A small area in the Foothills of southwestern Alberta stood higher than the adjacent interior ice sheet, but harboured elongate valley glaciers from the mountain ice on its western side. Parts of the southern Interior Plains close to the International Boundary and near the southern terminus of the interior ice sheet apparently stood higher than the surrounding glaciers and escaped glaciation.

The organic remains preserved in the surficial deposits are mainly those of modern species. Study of their stratigraphic position and geographic location relative to their present distribution reveals information on former migration of plants and animals. Extinction and evolution of species occurred in the Quaternary, but not to the same degree as in earlier periods.

Man made his appearance during the Quaternary and his development has been controlled to a large degree by the climatic conditions so characteristic of the period. It is generally believed that it was the lowering of sea level, consequent upon the amount of water incorporated in continental ice, that allowed Asiatic tribes to migrate to

XII Quaternary Geology of Canada

V.K. Prest

Viking Ice Cap and glaciers, Ellesmere Island, Northwest Territories.

North America, mainly by land, via the Bering Sea land-bridge (north of Aleutian Islands chain). Later, as the glaciers waned, these people left the unglaciated *refugia* in Alaska and Yukon Territory, migrating eastward along the arctic coast and southward into the interior of the continent. Some of these people—the early North American Indians—camped along glacial lakes and spillways that have long since disappeared.

The Quaternary includes four major glaciations each of which occupied a period of about 100,000 years. From oldest to youngest these are known as Nebraskan, Kansan, Illinoian, and Wisconsin. They were separated by longer-term, interglacial intervals—the Aftonian, Yarmouth, and Sangamon—when climates were as warm or warmer than now and when the continent must have been largely ice free. Under such conditions a rich and varied flora and fauna must have occupied most parts of Canada. Few deposits, however, can be assigned with certainty to these interglacial intervals though all are probably represented in some known occurrences. Interglacial sediments buried by glacial drift have been reported from the Maritimes, Great Lakes, Interior Plains, Cordillera, and Arctic; all the intervals must be represented in the stratigraphy of Porcupine Plain in northern Yukon Territory, though this has as yet been little studied.

Each of the four major glaciations was interrupted by relatively short-term non-glacial intervals or interstades. These were times when the climate ameliorated and the glaciers receded extensively from peripheral zones but presumably continued to occupy part of the mainland. Plants and animals tolerant of a cool climate migrated into the newly deglaciated areas, only to have their progeny displaced by the re-advancing glaciers at a later date. In Canada such interstades have been recognized only within the Wisconsin Glaciation. The best known occurrences are in Lake Erie region of Ontario and around Strait of Georgia, British Columbia.

There is no generally accepted classification scheme for Wisconsin age sediments, but they may be subdivided on the basis of glacial and non-glacial deposits. By means of radiocarbon (C^{14}) dates of organic materials contained in some deposits, reliable correlation between scattered sites dating back to about 45,000 years have been made, but the detailed geological and biological studies necessary to support the age determinations are usually lacking. Reliable age-datings are generally restricted to the last 25,000 or 30,000 years B.P.[1]. Investigations in the Lake Erie region indicate a mid-Wisconsin interstade occupying a period of about 20,000 years. This has enabled subdivision of the Wisconsin Glaciation into two main stages—early and late. The latter is commonly termed the 'Classical' Wisconsin. Fluctuations of the ice fronts during the mid-Wisconsin interstade complicate the stratigraphic record. This interstade was brought to a close by a major ice-frontal advance

[1] B.P., "Before Present", refers to the year 1950. This is to be understood in connection with all datings that follow, though it will not always be repeated.

during which the glaciers returned to the vicinity of the limit reached during the earlier Wisconsin Glaciation. The late or Classical Wisconsin climax occurred about 20,000 years ago, and as mentioned earlier most ice had disappeared from the mainland by 7,000 years B.P. The time transgressive stratigraphic relationships recognized

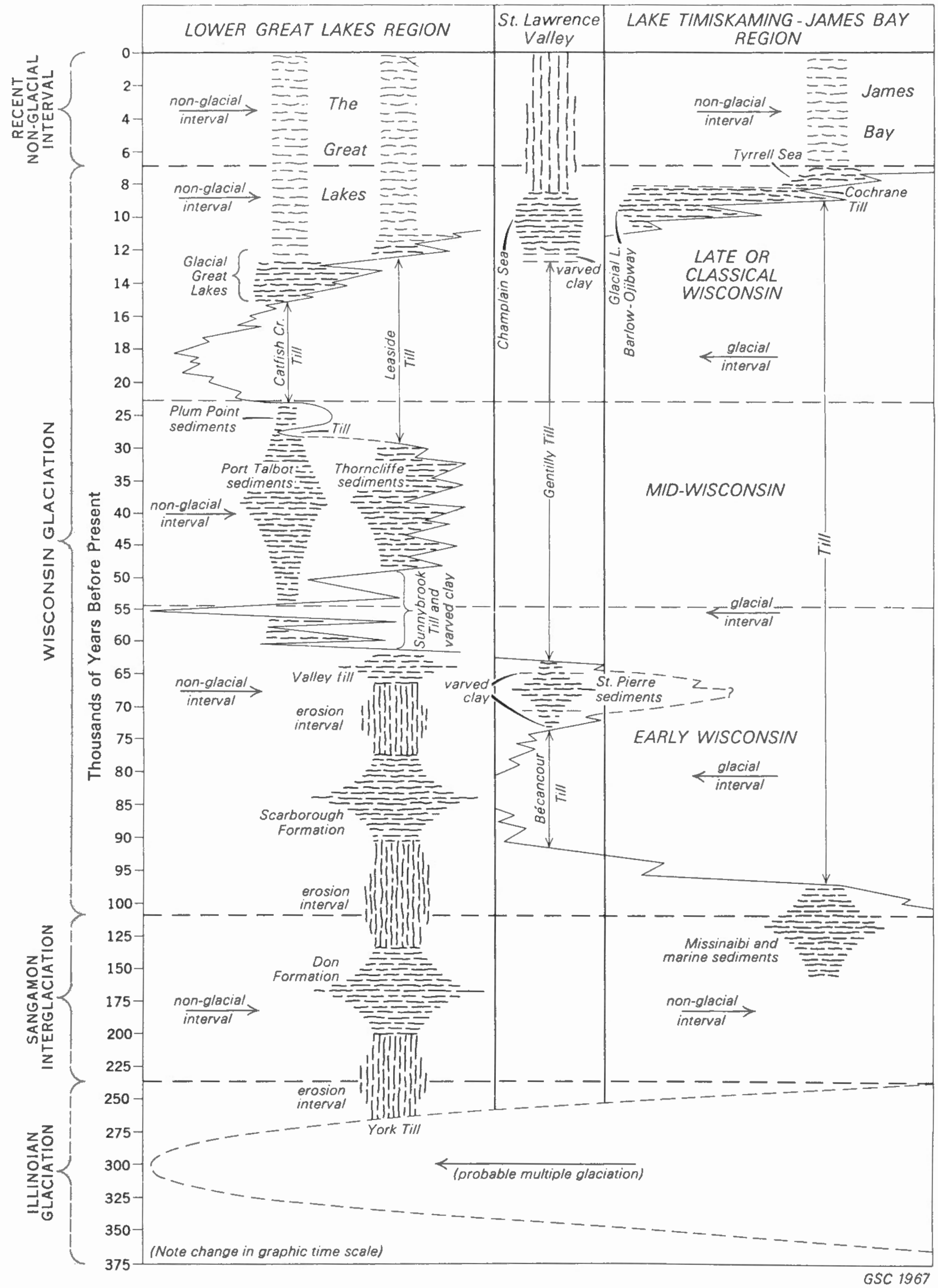


FIGURE XII-1
Time-stratigraphic relationships of late Pleistocene in east-central Canada.

in east-central Canada during the late Quaternary are illustrated diagrammatically in Figure XII-1; similar relationships are no doubt applicable to the whole of the Quaternary and to all parts of Canada.

Because there are uncertainties regarding Wisconsin events and their implications in different parts of Canada, and because many samples submitted for radiocarbon analyses have been given "greater than" datings, it is not yet possible to discuss interglacial deposits separately from Wisconsin, or other interstadial deposits. Furthermore, because of the general lack of paleontological or other data sufficiently accurate to identify specific interglacial deposits, the associated till sheets cannot be identified. Thus all deposits that predate the climax of the late, main, or Classical Wisconsin are discussed together, followed by an account of the last glaciation, and of deglaciation. In some parts of the Arctic and the Cordillera, glacial conditions have prevailed to the present and are manifested in the form of ice caps and mountain glaciers; for convenience, fluctuations of these ice masses are discussed along with Wisconsin deglacial events though strictly speaking they are post-Wisconsin.

THE OLDER GLACIAL AND NON-GLACIAL RECORD

This section deals with the Canadian record of the Quaternary prior to the climax of the Classical Wisconsin Glaciation; it includes glacial, interglacial, and interstadial deposits and the events that caused them or otherwise modified the landscape. Locally the deposits may be preglacial and represent events occurring in the earliest Quaternary.

Organic remains form a small but vitally important part of the sedimentary record, and are given considerable attention here. As a result of recognition of the environment and age of the non-glacial sediments the glacial sediments above or below them may be tentatively assigned to a particular glaciation. The deposits themselves and pertinent aspects of the physiography are treated on a regional rather than temporal basis.

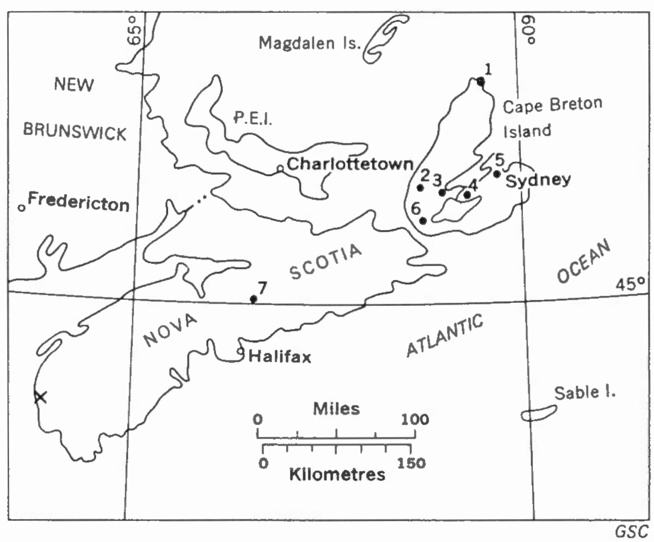
Appalachian Region

The physiography of the Appalachian Region reflects some aspects of the older glacial and non-glacial intervals but these are, in general, effectively masked by the more obvious fresh landforms and deposits of the last glaciation. Numerous estuaries represent the drowned parts of former river systems that may have been operative during times of lower sea level in Quaternary time. R. H. MacNeill (Acadia Univ.) believes many of the streams in Nova Scotia are re-excavated preglacial channels, and also that some till-mantled coastal terraces were probably cut during preglacial times. A grey sandstone regolith is found in parts of eastern New Brunswick both at the surface and beneath a mantle of red, sandy till. It is formed of soft Pennsylvanian sandstone and siltstone, up to 25 feet thick, and argues against extensive modification of much of the topography at this time. The highlands of New Brunswick show surprisingly little evidence of glaciation; erratics appear to be absent over large areas and the bedrock is soft and deeply weathered.

Interglacial and interstadial deposits together with associated glacial deposits that predate the Classical Wisconsin have been recognized on Cape Breton Island, and one deposit is known on the Nova Scotia mainland; none has been recognized in Newfoundland, Prince Edward Island, New Brunswick, or the Gaspé.

Buried Organic Deposits (Fig. XII-2)

Bay St. Lawrence. Unconsolidated materials on the northeastern end of Cape Breton Island form a seacliff up to 150 feet high and comprise stratified sediments, with some organic materials, between stony till-like layers. Towards the southwest, bevelled bedrock is exposed beneath the drift. Up to 20 feet of hard-packed sand-gravel 'till' occurs at the base of the drift section. This is overlain by as much as 4 feet of interbedded fine gravel and silt, or by a boulder layer believed to be a washed-surface of the till. The stratified unit and boulder layer is overlain by a dense organic layer, a few inches to a foot thick, followed by 2 feet of stratified silt and as much as 30 feet of interbedded sand and gravel grading upward into cobble gravel. The silt beneath the organic layer was



1. Bay St. Lawrence 2. Hillsborough 3. Whycocomagh 4. Benacadie 5. Leitches Creek 6. Inhabitants River 7. Milford
× Old' shell site: shells in till, Cape St. Mary

FIGURE XII-2. Location of buried organic deposits in Nova Scotia.

found to contain spores derived solely from Mississippian rocks. The organic bed represents a detrital sediment and contains pollen of alder, birch, black and white spruce, jack pine, balsam fir; a trace of juniper, willow, blue beech and walnut; and a variety of shrubs, herbs, grasses, ferns, mosses, and fungi (Mott and Prest, 1967). A piece of wood identified as tamarack has yielded a radiocarbon dating of >38,300 years (GSC-283)[1]. The immediately overlying silt beds contain similar pollen and spores but higher strata are barren. Nearby, E. H. Muller noted a lens of tan, silty clay, 12 feet thick at its maximum and 200 feet long. Overlying this lens and the cobble gravel (reported above) is a sandy boulder till some 30 to 90 feet thick, which becomes increasingly coarser upward; boulders as much as 8 by 5 feet occur near the surface.

In the clay lens are fragments of *Megayoldia thracaeformis,* a marine mollusc inhabiting waters colder than those surrounding Cape Breton today. The clay also contains pollen of alder and birch indicative of a cool climate but with pine, some oak, and a trace of basswood; otherwise the assemblage is similar to that of the detrital organic layer. Dinoflagellate cysts of Quaternary age and spores from the nearby Mississippian rocks are also present. The fossil record and radiocarbon date on the buried organic materials clearly indicate a pre-Classical Wisconsin cool period; this may be very early Wisconsin or perhaps late Sangamon.

Hillsborough. In southwestern Cape Breton Island between Mabou and Hillsborough from 8 to 12 feet of dull red, compact, clayey till overlies some 5 feet of stratified silty to clayey sediments that contain streaks and thin lenticular beds of carbonaceous material, and a basal layer of peat, silt, and wood a few inches to two or more feet thick. The organic layer rests on up to 18 inches of silt which in turn rests on an uneven ortstein layer developed on the surface of a highly oxidized sand and gravel more than 11 feet thick. The base of the exposed section at road level is about 10 feet above the level of the river and the sea.

The silty sediments and the peat bed (Mott and Prest, 1967) contain five pollen zones with a different assemblage predominating in each. The whole assemblage indicates a forest cover similar to that of the Boreal Forest Region rather than that of Cape Breton today. Wood from the base of the peat layer is dated at >51,000 years (GSC-370). An early Wisconsin or other interstadial interval is indicated but correlation with the Bay St. Lawrence site is not justifiable.

Whycocomagh. Buried organic beds in a highway-cut in the village of Whycocomagh, Cape Breton Island, were re-examined by the writer following an anomalous radiocarbon dating obtained on wood from this site. Gravelly till-like material 10 feet thick overlies 5 feet of stratified sediments ranging from fine gravel to silt and including a few inches of silty peat with scattered wood. The organic layers rest on 15 inches of partially oxidized clay and silt overlying 10 to 16 feet or more of stony, clay till that rests on an irregular bedrock surface. The stratified sediments extend for more than 50 feet, pinching out to the west as both bedrock and till mantle rise towards the surface. Pollen in the organic layer (Mott and Prest, 1967) is characterized by an assemblage dominated by alder, birch, spruce, and pine, and similar to part of the Hillsborough pollen diagram. Wood from the organic layer has an age of >44,000 years (GSC-290). The sites are only 15 miles apart at opposite ends of a through-going valley with a low drainage-divide; both sites lie close to sea level. They are believed correlative.

Benacadie. Intertill stratified sediments occur in a shore-cliff south of Benacadie at the entrance to East Bay, Bras d'Or Lake. Some 50 feet of reddish, stony, clay till overlies 5 feet of well-bedded clayey silts and a lower 20 feet of sand-gravel till. The clay-silt beds have a high percentage of pollen of pine (probably jack pine) and sedge, with lesser amounts of birch, alder, and grass (Mott and Prest, 1967). The assemblage has more affinities with the Bay St. Lawrence pollen diagrams than with those of the nearer Whycocomagh and Hillsborough sites. Another occurrence of intertill stratified sediments was found by Terasmae and Mott near Derby Point 4 miles northwest of Benacadie. Beneath the level of the shore road, 2 feet of reddish, clayey till overlies a few feet of silty clay, sand with plant detritus, and sandy gravel that rest on about 10 feet of slumped till down to sea level. The intertill silty clay did not carry pollen.

Leitches Creek. Part of a drill-log and some samples from a borehole at Leitches Creek, west of Sydney, that penetrated some 190 feet of overburden, indicate the presence of two layers of organic-bearing sediments each underlain and overlain by till. Preliminary pollen studies suggest that the lower organic layer is of interglacial age and that the upper layer correlates with the Hillsborough interstadial interval.

Inhabitants River. Buried organic deposits were reported in 1868 by J. W. Dawson from Inhabitants River, Cape Breton Island. A hard, peaty bed with roots and branches of coniferous trees rests on grey clays and underlies about 20 feet of till. The site has not been relocated.

Milford. A unique deposit of buried organic material found as overburden was removed at the gypsum quarry 2½ miles south of Milford Station, Nova Scotia. Beech nut *(Fagus* sp.), hickory nut *(Carya* sp., cf. *C. aquatica),* bayberry seed *(Myrica pennsylvanica),* and a beaver-cut stick were collected from a sinkhole in the gypsum. The

[1] These letters refer to the radiocarbon laboratory responsible for the age dating and the number is the sample reference number. GSC, Geological Survey of Canada; Gx, Geochron. Laboratories: Gro and GrN, Groningen, Netherlands: I, Isotopes Inc.: L, Lamont Geological Observatory; S, University of Saskatchewan; Y, Yale University; W, U.S. Geological Survey.

PLATE XII-1
Karst topography on gypsum at site of interglacial deposits, Milford, Nova Scotia. The sinkholes contained a mixture of glacial and non-glacial sediments including organic materials dated at >38,000 years B.P. The gypsum surface is ice scoured and overlain by a complex of tills and fossiliferous sediments visible in bluff in background.

wood was dated at >33,800 years (GSC-33). The following notes on the quarry site and its environs are based on observations by W. Take (Nova Scotia Museum).

The surface of the gypsum displays a karst topography (Pl. XII-1). In the base of depressions there is usually a mixture of slumped till and glaciofluvial materials, overlain by unfossiliferous brown to grey clay and sphagnum peat containing gastropods and patches of slumped till. Overlying the peat unit is highly fossiliferous grey clay and sandy clay. The thin basal part of the clay is characterized by abundant macrofossils of white pine and rare hemlock; the latter increases upward and is evenly distributed. Logs, beaver-sharpened sticks, cones, insects, mollusca, amphibian and mammalian remains were collected from the stratified clays. An erosional unconformity separates the clay from plant-bearing sands—mainly spruce, fir, and rare white pine cones.

The sinkhole deposits and the glacially scoured surface of the bedrock are evenly truncated and overlain by a compact, grey, gypsiferous till. In places this till carries woody detritus; in two places it has a boulder pavement developed on it. The till is generally overlain by sediments grading from unfossiliferous grey clay through fossil-bearing silty clay, into highly fossiliferous grey sand with abundant twigs of coniferous trees as well as cones of spruce and rarely of fir. Overlying these sediments are two very similar tills, separated in valley bottoms by orange-brown gravelly sands. The basal part of each till, and especially of the lower one, is gypsiferous. The tills change from grey near the base to grey-brown at the top. In upland sections the combined till is a uniform brown colour, does not contain gypsum but is deeply oxidized. At one locality it has a well-developed soil-profile.

Lying on a grey-brown till about 35 feet above sea level in some valleys is an orange-brown to red-brown laminated clay with vertical rootlets in its upper part suggestive of a marsh environment. At somewhat higher elevations, a black clay carries mollusc remains, white pine cones, hemlock bark and cones, and red oak acorns. A distinctive red-brown clay till overlies the stratified sediments in low areas, and the grey-brown tills on higher ground. This is followed by a complex of red clays, red till, minor glaciofluvial sand and gravel, and a thin sand-gravel till unit, the last strictly confined to the valleys. Glaciofluvial sand and gravel, and stratified clay and silty clay, both with the modern soil profile or a cultivated surface, complete the sequence of deposits.

Take considers the Milford deposits to represent parts of the Kansan, Illinoian, and Wisconsin Glaciations, as well as the Yarmouth and Sangamon interglacial intervals. Certainly interglacial as well as interstadial fossil assemblages are indicated, but the implications of the karst topography have precluded definite correlations.

St. Lawrence Lowlands

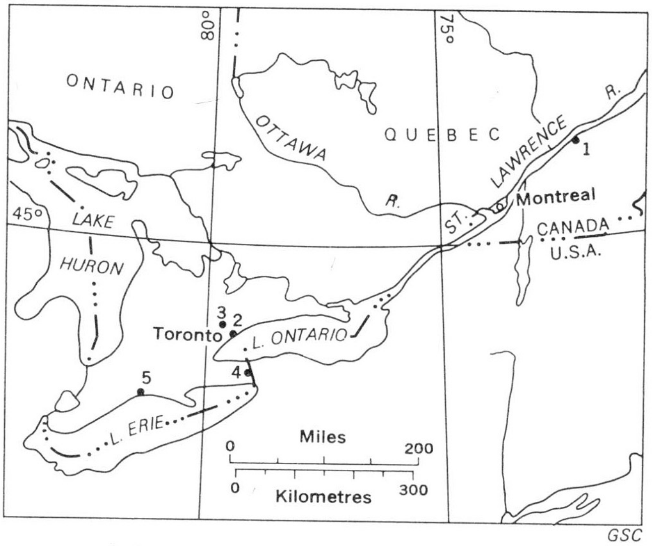
Major features of the physiography of the St. Lawrence Lowlands, such as Niagara Escarpment and the bordering highlands of the Canadian Shield and Appalachian Region, predate the Quaternary and, except for local over-deepening of valleys and some glacial scour, were little changed by glaciation. The great morainal belts in southwestern Ontario bear testimony to the direct effects of glaciation. Drift thickness in the Oak Ridges interlobate moraine reaches over 1,000 feet, and there is some evidence of a buried drainage system north of Lake Ontario beneath what is now a major ridge overlooking the lake. The St. Lawrence and Ottawa River valleys are filled by 100 to 200 feet of marine sediments, and in an area of high bedrock knobs north of Oka-sur-le-Lac a channel containing about 400 feet of drift rests on rotted bedrock some 175 feet below present sea level. Thus the flat valley bottoms bear little resemblance to the more mature drainage systems of interglacial and preglacial times.

In the St. Lawrence Lowlands the only undoubted interglacial deposits are those of the Toronto area where the Don Formation is assigned to the Sangamon interglacial interval, and the basal York Till is thought to represent an Illinoian glaciation (Karrow, 1967). Near Trois-Rivières, Quebec, and Toronto and London, Ontario, numerous occurrences of buried organic materials appear to represent non-glacial intervals within the span of the Wisconsin glacial period.

Buried Organic Deposits (Fig. XII-3)

Trois-Rivières. The Quebec part of the St. Lawrence Lowlands near Trois-Rivières has been studied in some detail by Gadd (1960, *in press*). The St. Lawrence Valley drainage system appears to have been blocked in very early Wisconsin time, giving rise to a glacial lake and deposition of reddish varved clays (Table XII-1). These clays were later overridden by the glacier and the Bécancour Till, a brick-red, somewhat sandy, clay till, was deposited. This till is known over a wide area south of the St. Lawrence, but the southern limit reached by the glacier is unknown. The till derives its rich red colour from the Ordovician, Queenston, red shales. Most stones in the till, however, are Precambrian types and hence the glacier that deposited the till came from the Canadian Shield, presumably from the Laurentian Highlands. However, neither the red till nor a correlative grey till has been recognized north of St. Lawrence River.

The type section of the St. Pierre sediments is near St. Pierre les Becquets, Quebec (Gadd, 1960). The name 'St. Pierre' is given to the intertill sediments and to the interval during which they were deposited. The sediments overlie the red Bécancour Till and consist of sand and silty sand with lenticular and discontinuous beds of highly compressed peat and some disseminated organic matter. The sand unit has a maximum thickness of 25 feet, but at the type section is only 13 feet thick and includes three layers of peat. The upper and thickest peat bed where observed along a ravine near St. Pierre les Becquets was 1.75 feet thick and was traced for over 200 yards. Each peat layer begins with a gyttja and grades upward through *Carex* peat into *Sphagnum* peat with abundant tree and



1. Trois Rivières area
2. Toronto
3. Woodbridge
4. Niagara Falls
5. Port Talbot

FIGURE XII-3. Location of buried organic deposits in St. Lawrence Lowlands.

other plant remains, the whole suggestive of periodic flooding of the lowlands.

The St. Pierre sediments occur for about 50 miles along the south side of St. Lawrence River from Pierreville to Deschaillons and inland for about 20 miles to Ste. Brigitte (Pl. XII-2). They have also been found on the north side of the river, at Les Veilles Forges, about 6 miles northwest of Trois-Rivières (Gadd and Karrow, 1960). According to P. F. Karrow, it is very likely that the organic-bearing sediments reported by A. P. Coleman at Donnacona 50 miles to the east are the equivalent of the St. Pierre sediments. Some peaty sediments of uncertain stratigraphic position and non-organic sediments between

PLATE XII-2
Section of postglacial sediments, Gentilly Till, and St. Pierre sediments on Bécancour River, north of Aston Junction, Quebec. Sediments are probably underlain by red Bécancour till exposed downstream. (A) St. Pierre sand and pebbly sand >60 feet, (B) Gentilly Till 15 feet, (C) Champlain Sea clay 24 feet, (D) alluvial sand 6 feet.

two tills elsewhere in the region suggest that the St. Pierre sediments are even more widespread. The whole appears to represent an old fluvial system that is slowly being exposed by the St. Lawrence River and its tributaries.

Palynological studies of St. Pierre sediments reveal that spruce, pine, birch, and alder are the main tree types present, and that oak, beech, maple, elm, ash, and hickory together form only 2 to 5 per cent of the tree pollen in the middle and warmest part of the sequence. Hemlock is noticeably absent. The pollen spectrum is similar to that of the early postglacial assemblages in the St. Lawrence Lowlands, except for the absence of hemlock, and it is similar to that of the boreal forest today (Terasmae, 1958, p. 20). Beetle wings are fairly common in the peat beds and a few ostracods have been noted. Terasmae concludes that the climate remained fairly constant throughout most of the St. Pierre interval with sub-arctic conditions prevailing near the bottom and top, and that a relatively short span of time is involved.

Samples from Pierreville and St. Pierre were dated at Groningen by an isotopic enrichment method at 67,000 ± 2,000 years and 64,000 ± 1,000 years, and a sample from Donnacona by conventional techniques at >44,470 years (Y-463). Thus, St. Pierre sediments are inferred to represent an early Wisconsin non-glacial interval. The St. Pierre sediments are overlain by the Lake Deschaillons varved clay, 70 feet thick, deposited in a proglacial lake; these in turn are overlain by Gentilly Till deposited by the advancing ice sheet. This ice cover persisted throughout the greater part of Wisconsin time until final recession of the ice sheet and incursion of the sea into St. Lawrence

TABLE XII-1 *Composite section of Pleistocene deposits, St. Lawrence Lowlands (by N. R. Gadd)*

	(Max. known thickness, feet)
Bog deposits	20
Low terrace sands, el. ca. 100′	10
High terrace sands, el. ca. 300′	10
Champlain Sea sand	25
St. Narcisse Till	15+
Champlain Sea clay	100
Gentilly Till	15
Lake Deschaillons varved clay	70
St. Pierre sediments	25
Red-banded varved clay	5
Bécancour Till	55
Red varved clay	5
Bedrock (Queenston red shale)	—

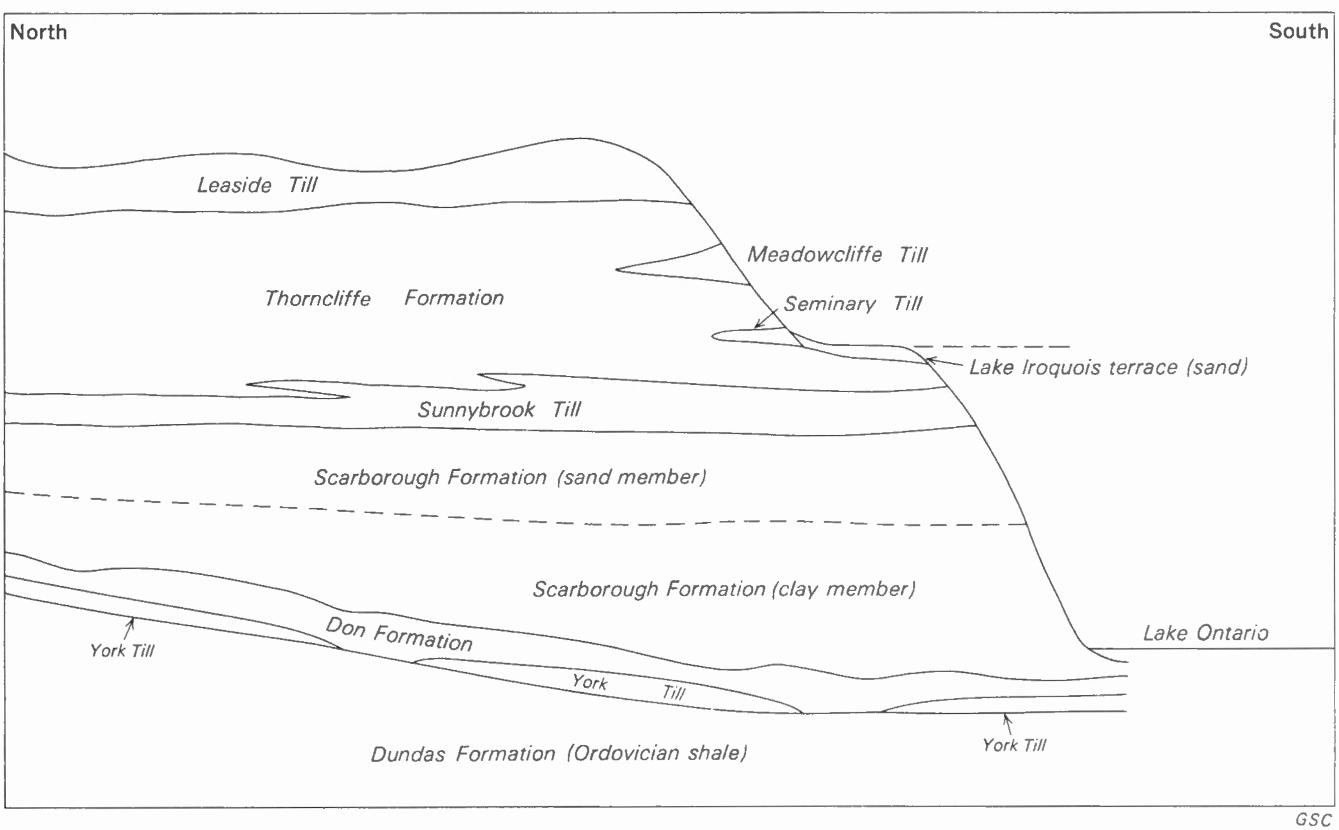


FIGURE XII-4. Schematic diagram showing stratigraphic relationships of Pleistocene deposits at Scarborough Bluffs, Toronto, Ontario (after Karrow, 1964).

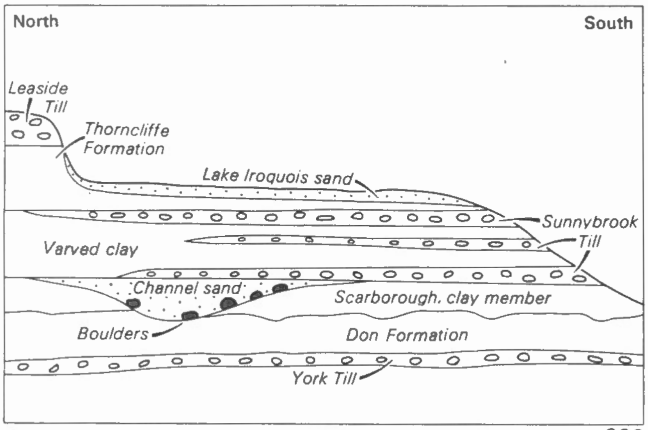


FIGURE XII-5. Schematic diagram showing stratigraphic relationships of Pleistocene deposits in Don Valley brickyard, Toronto, Ontario (after Karrow, 1964).

Lowlands almost 12,000 years ago. Subsequently, differential uplift of the land caused regression of the sea, and the present drainage system was established.

Toronto. The most famous buried, non-glacial Pleistocene deposits in Canada are those of the Toronto area. They have been known for over a century and have been under more or less continual observation in the Don Valley brickyard operations since 1889. Recent studies have done much to clarify the chronology recorded by the complex deposits. The non-glacial organic-bearing beds have been encountered in the extensive subway excavations and other construction projects of the last decade. The sequence of non-glacial deposits in the region is shown schematically in Figures XII-4 and 5.

In the Don Valley brickyard the basal York Till is one to four feet thick and rests on Ordovician strata. It is overlain by fossil-bearing stratified sediments of the Don Formation (Karrow, 1964, 1967). An appreciable hiatus occurred between the deposition of the York Till and the Don Formation as the basal parts of the latter contain pollen of a warm flora (Terasmae, 1960, p. 33).

The Don Formation consists of up to 25 feet of generally well stratified clay and sand displaying some crossbeds and cut-and-fill structures, and containing scattered remains of a wide variety of plants and animals. Noteworthy among the plants, in that they do not range so far north today, are southern white cedar, blue ash, osage orange, iron oak, chestnut oak, and black locust. In all, some 44 taxa of plants have been identified from macroscopic remains. Most of these have been recognized also as pollen, plus an additional 28 taxa, including pollen of sweet gum, that are not found in the Toronto area today (Terasmae, 1960). Terasmae also identified some 20 species of diatoms, and noted the presence of freshwater sponge spicules from the middle part of the Don Formation, which indicate lake, stream, and bog habitats. Animal remains recovered include the shells of some 40 species of pelecypods and gastropods (including a few land snails), part of a catfish, and bones of groundhog, deer, bison, bear, and giant beaver. Terasmae states: "An ecological-climatological interpretation of all evidence supplied by the fossils from the Don beds suggests that the annual mean temperature at the time of their deposition reached a maximum probably 5°F warmer than the present." In recent years investigations by Karrow and others have shown that wood is not common and leaves and vertebrate remains are very rare.

The Don Formation occurs some 60 feet above the level of Lake Ontario (el. 246 feet) in the brickyard but its upper surface is below lake level at Scarborough Bluffs. The lower part was deposited at the mouth of a river as it entered a lake in the Ontario basin. Lowering of the lake level is recorded by the character of the diatom assemblage in the middle part of the beds and by the sandy nature of the upper part of the formation. The upper sandy beds furthermore had been leached and weathered (Terasmae, 1960; Karrow, 1964) prior to the deposition of the overlying cool-climate Scarborough Formation. In the Don Valley brickyard the Don Formation is locally separated from the overlying Scarborough Formation by a layer of hard, compact, non-calcareous sand with some pebbles and cobbles. The formational contact was also encountered in borings at Scarborough Bluffs 15 feet below lake level. The Don Formation clearly represents part of an interglacial interval. It is generally assigned to the Sangamon but exact dispositon must await further work.

The Scarborough Formation (Karrow, 1964) is best known from the exposures on Scarborough Bluffs where it comprises a lower clayey-silt unit about 100 feet thick and an upper sandy unit about 50 feet thick, but it is also present in the Don Valley brickyard. There, the Scarborough clay unit is less than 25 feet thick and seemingly devoid of fossils although a few plant detrital seams have been noted between the thin clay layers. This deposit is very finely bedded and is regarded as a deep water deposit. The clay unit at Scarborough Bluffs includes peaty layers in places half an inch or more thick. These have yielded many small fossils including diatoms, and the leaves, seeds or spores of some 15 plants and, notably, the wing covers and other chitinous parts of some 72 species of beetle (Coleman, 1933). The beetles are reported as mostly extinct species but a re-study of the beetle content of the Scarborough plant detrital layers is needed. Terasmae has identified 41 plants from the clay unit, which indicates a boreal forest cover. The overlying sandy unit also contains plant detrital layers, sparse ostracods, and molluscs. Terasmae reports a pollen assemblage of boreal forest species similar to those found in the lower clay unit. He concludes that the climate during Scarborough times was perhaps 10°F cooler than now. The sands were deposited by southeast-flowing waters and presumably represent a delta formed in a lake (Lake Scarborough) that stood some 200 feet higher than the present lake. A piece of wood from near the top of the Scarborough Formation at Scarborough was dated at >52,000 years (Gro-2555), hence the precise age of the beds remains unknown.

The writer considers the Scarborough beds to be older than the St. Pierre sediments. Lake Scarborough formed after a lengthy erosional interval during which the Don beds were scoured and weathered. Lake Scarborough may be attributed to plugging of the drainage system by an advancing glacier, and the biotic record seems to support this view. The St. Pierre sediments represent an old river system with associated flood plain deposits, which required a through-flowing St. Lawrence Valley drainage system. Thus the Scarborough Formation has to be either younger or older than the St. Pierre sediments, and based on other evidence noted below it is considered to be older.

Organic matter is also present in post-Scarborough deposits. During and following the lowering of the Scarborough lake, deep valleys were cut into the Scarborough Formation, and in places into the Don Formation. Rising lake levels again brought about deposition of sandy sediments within these valleys. Boulders and cobbles, and occasional balls of till occur in the bottom of a channel in the Don Valley brickyard, which may indicate a significant time gap. The organic materials found in these sediments have been in part re-deposited from the older beds but some are indigenous to this stage of sedimentation. Of special interest is the first occurrence noted in Canada of the distinctive gastropod *Hendersonia occulta*. Coleman (1933) records water-worn "vertebrae of bison, part of a lower jaw of a bear, and a horn of an extinct deer, *Cervalces borealis*," and bits of ivory from mammoth or mastodon; these are from the Christie Street sand pits which P. F. Karrow correlates with the above-mentioned valley-fillings.

Subsequent to a mid-Wisconsin glacial invasion into the Lake Ontario basin and deposition of the Sunnybrook Till and varved clays, another period of non-glacial sedimentation occurred. The sediments formed at this time contain sparse plant remains and comprise the Thorncliffe Formation (Karrow, 1964), which consists of stratified clay, silt, and sand believed to have been deposited in both lakes and streams. Near Lake Ontario two wedges of till interfinger with the Thorncliffe Formation as a result of short-term expansions of the westward-moving ice. Both the sparse plant fragments and the till lenses (Seminary and Meadowcliffe Tills) indicate a rather cold climate. Jack pine and spruce were the dominant tree types with minor tamarack, oak, and birch; non-arboreal pollen was rather abundant. A small sample of plant material from the Thorncliffe beds has been dated at 38,900±1,300 years (GSC-271). Glaciation, foreshadowed by the interfingering tills, then engulfed the region during the classical Wisconsin with deposition of the Leaside Till. Vegetal and animal life did not return until late glacial and early postglacial times.

Markham and Woodbridge. A few miles north of Toronto old organic materials have been observed beneath till. At Markham, a peat ball found in a gravel pit and dated at >34,000 years (W-194) is thought to have been derived from Thorncliffe beds or the Scarborough Formation. At Woodbridge, five discrete tills and a few feet of bedded silt, the whole comprising up to 35 feet, rest on some 25 feet of clayey silt with streaks and lenses of

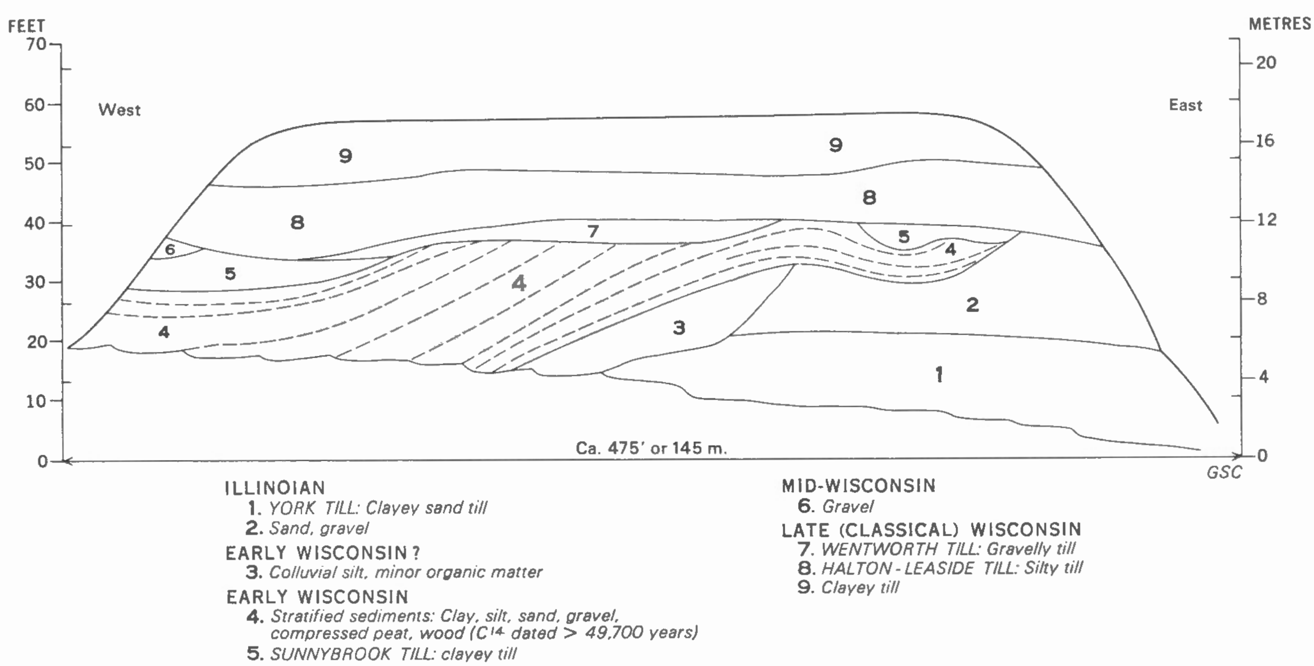


FIGURE XII-6. Wisconsin and Illinoian drift in railway-cut, Woodbridge, Ontario (after Karrow, 1965).

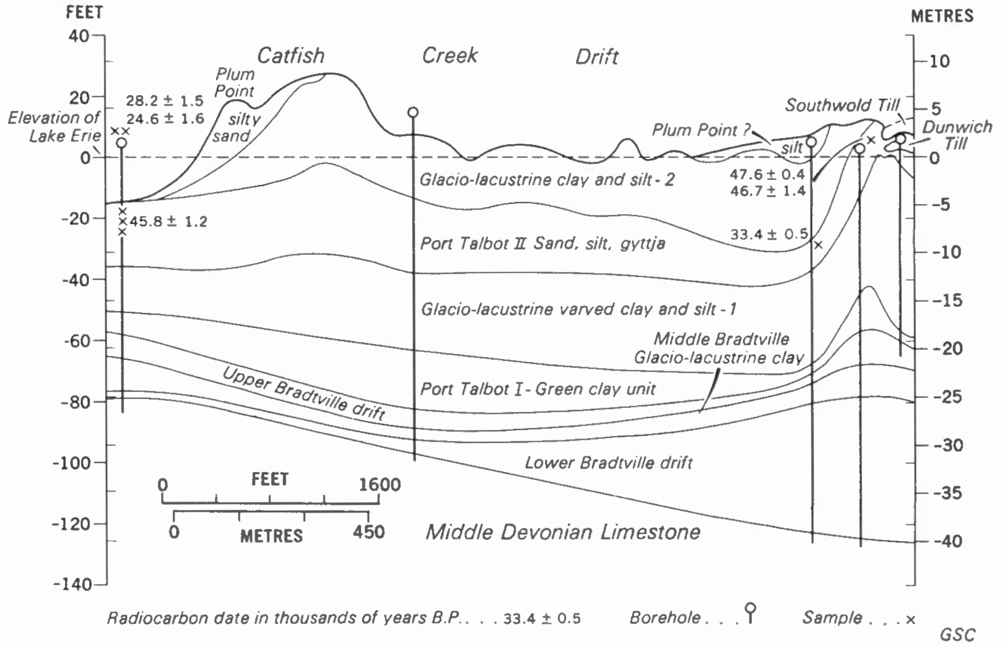


FIGURE XII-7
Generalized profile through Pleistocene deposits near Port Talbot, Ontario (after Dreimanis, et al., 1966).

peat and wood (Fig. XII-6). Another 14 feet of sand and gravel and 9 feet of till of Illinoian age are exposed at the base of the cut (Karrow, 1965). Some peat balls have been seen in the clayey till that overlies the peaty sediments. Palynological studies reveal that a northern boreal forest grew in the area at that time, with jack pine, black and white spruce, and birch the main tree types. This assemblage and a radiocarbon dating of >49,700 years (GSC-203) suggest a correlation with the Scarborough Formation, though a post-Scarborough age is also possible.

Niagara Falls. Silt in the buried St. David's channel near Niagara Falls has yielded a spore and pollen assemblage that is indicative of the northern boreal forest zone. Pollen studies by Terasmae substantiate the early report of fragments of spruce wood, from a boring over the channel, at a depth of 186 feet. Correlation of these silt deposits with other non-glacial beds in Ontario is not yet conclusive, but they may well be mid-Wisconsin.

Port Talbot. Buried organic deposits indicative of non-glacial conditions were reported by A. Dreimanis in 1951 from the Lake Erie shore south of London. As investigations progressed both along the shore and by drilling a complex sequence of deposits was indicated (Fig. XII-7). Radiocarbon datings on organic materials appeared to fall into main groupings, namely, about 48,000 to 44,000 years B.P. and about 28,000 to 23,000 years B.P. The indicated non-glacial intervals were thought to be separated by a glacial period represented by a till sheet (Southwold) found overlying the 'older' sediments in several places. The two intervals were named Port Talbot and Plum Point by Dreimanis (1957, 1958).

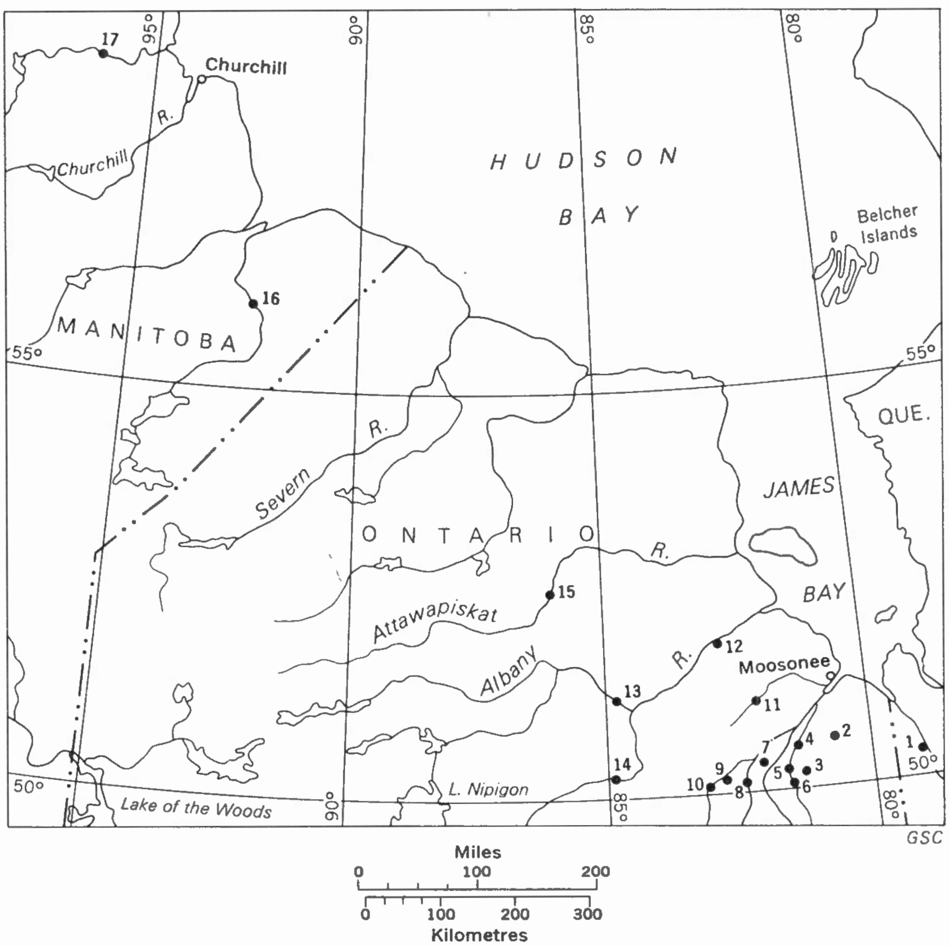
The main organic-bearing deposit at Port Talbot is exposed at the base of a 100-foot bluff overlooking Lake Erie. Contorted beds of silt, clay, calcareous gyttja, and scattered peat balls are overlain by glaciolacustrine clay and silt, organic-bearing silt and silty sand of Plum Point unit, and an uppermost drift complex of the main or Classical Wisconsin Glaciation. The gyttja contains larch, spruce, and water plants (*Potamogeton, Menyanthes,* and *Najas*) (Dreimanis, *et al.*, 1966). Part of a mastodon tusk was found in a clayey bouldery gravel some 600 feet east of the gyttja site at the same stratigraphic horizon. Some 17 species of ostracods have been recorded from the clayey silts both above and below the gyttja horizon. Mollusc shells from both the silts and the gyttja appear to have been crushed by the glacial action that contorted the beds but three genera have been recognized. The pollen assemblage from the gyttja horizon indicates that pine (mostly jack pine), spruce, larch, and birch were the dominant trees in the region. The base of the gyttja includes also large pollen grains of either white or red pine. The organic record indicates that the climate was cooler than that at present along the Lake Erie shore.

In boreholes at Port Talbot, beneath stratified sediments dating 47,700 ± 1,200 years (GSC-217), up to about 30 feet of brownish buff varved silt and clay with a sparse pollen content overlies a few feet of greenish clay and silt with abundant pollen of jack pine, spruce, oak, and non-arboreal pollen. Mineralogical characteristics of the green clay and silt suggest a gap in the record after deposition of the basal, reddish (Bradtville) till, which rests on bedrock some 130 feet below surface of Lake Erie. Thus the Port Talbot non-glacial interval may extend back 50,000 years B.P. The lower Bradtville Till has been subdivided into three units of very similar lithology with the lower two separated by a glaciolacustrine clay. The initial advance of Wisconsin ice into Erie basin, from the east, appears to have removed all evidence of interglacial or older glacial deposits in Port Talbot area. The initial advance was followed by a retreat stage during which a glacial lake occupied the basin. Two later encroachments of the ice formed the middle and upper Bradtville Till sheets. The time occupied by these early Wisconsin events is unknown but possibly 5,000 to 10,000 years.

Recent radiocarbon dates of 33,400±500 and 38,000±1,500 (GrN-4238, 4272) indicate that the time gap between the Port Talbot and Plum Point non-glacial intervals was short. It now appears that one long-continued interval, a mid-Wisconsin interstade, may be represented by the intertill stratified sediments and their contained organic matter. Southwold Till, formerly considered to intervene, is now correlated with the younger Catfish Creek Till which overlies the Plum Point sediments (Dreimanis, *et al.,* 1966). Thus, non-glacial conditions may have prevailed in the Port Talbot area from about 50,000 years to about 24,000 years ago. The ice front was, however, not too far away during this long period of sedimentation; the eastern end of Erie basin was blocked by ice on two occasions during which glacial lake clays were deposited and also ice of northern derivation was nearby when the Dunwich Till was deposited at some time during the Port Talbot interval.

Hudson Bay Lowland

The history of Pleistocene events in Hudson Bay Lowland has been dealt with by Lee (1968a). Important deposits bearing on the early Pleistocene record have recently been obtained by Craig and McDonald (1968).



1. Harricanaw River
2. Nettogami River
3. Little Abitibi River
4. 5. 6. Abitibi River
7. Campbell Lake
8. 9. 10. Missinaibi and Opasatika River
11. Kwataboahegan River
12. 13. Albany River
14. Kenogami River
15. Attawapiskat River
16. Gods River
17. Seal River

FIGURE XII-8
Location of buried organic deposits in Hudson Bay Lowland.

Buried Organic Deposits (Fig. XII-8)

In the Hudson Bay Lowland there are numerous records of organic deposits lying below or between glacial tills. The published records are confusing in that the term lignite has been applied not only to Lower Cretaceous lignite deposits but also to Quaternary compressed peat and to sands, of both glacial and non-glacial affinities, containing detrital lignite. Eliminating the occurrences that are probably Cretaceous and the sands with detrital lignite grains of Quaternary age, a great many occurrences remain which, in the writer's opinion, record Quaternary non-glacial intervals in the Lowland. The occurrences are commonly cited as of interglacial age and indeed their wide distribution lends credence to this concept, but with present knowledge of Wisconsin interstades, a younger setting must be kept in mind where there is no strong evidence to the contrary. The more important sites exclusive of new ones found by Craig and McDonald are described below from east to west:

Harricanaw River. Two miles below Seven Mile Island, rhythmically bedded silt and sand with some thin vegetal layers underlie a marine clay beneath river sand and gravel. The vegetal layers were dated at >42,000 years (Y-1165).

Nettogami River. On the Nettogami River the lignite reported by Bell (1904) is most probably Pleistocene compressed peat. He refers (p. 161) to thick beds of black clay-shale with a great many very thin seams of a rather peaty lignite and to boulder clay containing striated pebbles both above and below the organic beds.

Little Abitibi River. Sections along the lower part of Little Abitibi River and along the adjacent Abitibi River reveal dark grey, compact and jointed silty clay beneath till and Tyrrell Sea deposits. Thin peaty laminae yielded pollen of black spruce, jack pine, and birch, together with varied non-arboreal pollen and spores of ferns and mosses. Farther up Little Abitibi River, silty and carbonaceous beds in the lower part of a 70-foot section of sand yielded much the same pollen assemblage. Wood from the lower part of the sand unit gave an age of >43,600 years (GSC-435). These deposits probably correlate with the Missinaibi beds on the Missinaibi River.

Abitibi River. The occurrence of buried Pleistocene organic deposits in the Moose River basin, including sites along Abitibi River (Bell, 1904; Wilson, 1906), may appear doubtful because of the Lower Cretaceous lignite found along Abitibi, Mattagami, and Missinaibi Rivers. Study of borehole data and excavations at Otter Rapids, however, revealed plant detritus in intertill sediments and indicated correlation with Pleistocene beds along Missinaibi River (Terasmae and Hughes, 1960). Furthermore at the Onakawana lignite workings, close to the river sections observed by Bell and Wilson, buried organic deposits of Pleistocene age were encountered in a shaft and in some 116 exploratory drillholes (Martison, 1953). Two till sheets separated by interglacial or interstadial sand, gravel, and clay are clearly present.

On Abitibi River 8 miles below Otter Rapids the writer has observed contorted beds of compact, dark grey, stony clay, containing broken or fractured marine shells, beneath younger drift. Eleven species of foraminifers, three of ostracods, and two of pelecypods were identified along with indeterminate species of forams, a pelecypod, and sponge spines, and spicules. Most of the species live in Hudson Bay at present. The stony marine clay overlies an oxidized quartzose sand, with a bed of differentially rotted igneous boulders. The marine deposits and the oxidized sand are considered to represent an interglacial interval. They are overlain by a Wisconsin clayey till and Tyrrell Sea deposits. Half a mile downstream the till rests on several feet of gravel containing limestone and igneous rocks; although the gravel appears to overlie the interglacial stony clay it is presumed to be much younger. Coleman (1941) also reported peaty clay and marine shells from interglacial sediments along Abitibi River, as well as shells from crumpled clay beds beneath till at Moose Factory.

Campbell Lake. A drillhole at Campbell Lake (Wawa Lakes) intersected 725 feet of drift resting on Paleozoic rocks (Hogg, *et al.*, 1953). J. Satterly reports that microfauna were recovered from samples taken at intervals between 262 and 645 feet—a section that includes two tills and two or three stratified clayey units. The fauna are of Pleistocene age and, as they were largely marine foraminifers, it is probable that the beds are interglacial. The great depth of overburden, in a section that includes both tills and marine beds, appears to indicate the presence of both preglacial and interglacial valleys of the Mattagami River. Microfauna identified in research laboratories, Shell Oil Co., are as follows:

Nonion grateloupi (d'Orbigny)
Elphidium gunteri (Cole)
Cibicides concentricus (Cushman)
Quinqueloculina lamarckiana (d'Orbigny)
Quinqueloculina seminulum (Linnaeus)
Discorbis sp.
Elphidium discoidale (d'Orbigny)
Siphonina cf. *pulchra* Cushman
echinoid spines
fragments gastropods and pelecypods
unidentified ostracod and plant spores
Cibicides pseudoungeriana (Cushman)
Globigerina bulloides d'Orbigny
Discorbis orbicularis (Terquem)

Missinaibi and Opasatika Rivers. Along Missinaibi River, above its junction with Opasatika River, lignite visible at low water was first noted by Bell (1879, p. 4C). Some of the lignite is undoubtedly the equivalent of the lignite of the Lower Cretaceous Mattagami Formation, but Bell's description suggests that at least some of the occurrences are Pleistocene intertill compressed peat and wood. A section 9 miles above the mouth of Opasatika River is as follows:

Feet	
0 - 10	Hard, drab clay with striated pebbles and small boulders, and holding rather large valves of *Saxicava rugosa, Macoma calcarea,* and *Mya truncata.*
10 - 15	Hard, lead-coloured clay with yellow seams and spots, and red, grey, drab, and buff layers.
15 - 21	Lignite, made up of laminae of moss and sticks.
21 - 22	Clay with spots of lignite.
22 - 62	Unstratified drift full of small pebbles.
62 - 65	Yellowish stratified sand and gravel.

Three miles farther up the river Bell records:

0 - 45	Blue clay with pebbles, some striated.
45 - 47½	Lignite, made up principally of sticks and rushes.
47 - 127	Yellow weathering grey clay with pebbles, some striated.

Important observations were made also by J. M. Bell (1904) and J. Keele (1921), but the occurrence of both Cretaceous and Pleistocene deposits remained unproven until the area was restudied and pollen analyses were made by R. Auer (McLearn, 1927).

Terasmae and Hughes (1960) recognize five main Pleistocene units: (1) a lower drift; (2) a middle drift consisting of till and glaciofluvial sand and gravel; (3) layers of peat, organic silt, and clay, termed the Missinaibi beds; (4) an upper drift consisting mainly of till; and (5) marine clay, sand, and silt. Locally this sequence is eroded and overlain by the fluviatile deposits of the terraces along the rivers. Wood from Missinaibi beds gave a radiocarbon age of >53,000 years (Gro-1435). Palynological studies by Terasmae (1958) indicate a climate similar to the present or slightly cooler and possible correlation with the St. Pierre beds of St. Lawrence Lowlands. The inferred climate is unlike that represented by the Toronto Don beds, even allowing for the difference in latitude between the two areas, and hence the Missinaibi beds, if not representative of a late part of the Sangamon Interglaciation, must have been deposited during an early part of the Wisconsin glacial interval.

Kwataboahegan River. Wilson (1906) found solid peaty material in the bed of this river some 65 miles above its mouth:

> The mass where examined was six feet thick and it can be traced along the river for 430 feet. It is a dark brown colour and breaks off into lumps two to three feet thick. It burned slowly in the camp fire but left a large quantity of ash. Thin layers of the same material are exposed in the bank intercalated with the clay for several miles up the river.

A section, 60 miles above the mouth, is described by J. M. Bell (1904, p. 168) as follows:

> The seam, which has a maximum width of two feet six inches, outcrops almost continuously along the edge of the river for 450 feet in a bank 40 feet high. Though compact and hard it is never pure and is for the most part mixed with clay. Above it lies about 25 feet of hard, blue clay surmounted by six feet of shell-bearing post-glacial material. Below the seam is a hard stony clay containing many shells. This is of great scientific interest as it is the only point in the Moose Basin where interglacial shells are known to occur. The lignite itself is both arenaceous and argillaceous. It consists of thin layers of indurated moss with partings of clay and sand. It burned with considerable difficulty in the camp fire leaving a large residuum of clay and sand.

Though Martison (1953) did not examine the Kwataboahegan organic deposits, he considered them of Pleistocene age, and noted that high ash content is not characteristic of the Cretaceous lignites at Onakawana.

Albany River. Here till overlies a few feet of blue and brown clay (reported as leached) with two 2-inch beds of 'lignite.' Williams (1921) reported one bed to be "composed mostly of moss" and the other as "containing compressed roots." Nearby in a 90-foot section on the north bank of the river the till cover is 50 feet thick and rests on similar clay though no 'lignite' was observed. The Pleistocene age of these materials was corroborated by Terasmae and Hughes (1960). They conclude that the spore and pollen assemblages are entirely unlike those of the lignite of the Cretaceous Mattagami Formation, known on Mattagami, Abitibi, Missinaibi, and Opasatika Rivers.

Farther down Albany River 10 feet of stratified sand rests on 20 feet of pebble clay, presumably till, that in turn rests on 20 feet of thin-bedded clay, peat and moss, and 2 feet varved clay and pebble clay beneath which are 8 feet of pebble clay possibly a till. The bedded clay, peat, and moss sequence was considered by Martison to be interglacial.

Kenogami River, a tributary of Albany River, intersects a preglacial river channel cut into Silurian limestone. R. Bell (1887, p. 38) reported a basal till overlain by a 6- to 8-foot bed of soft lignite, containing many flattened stems of small trees and succeeded by 30 to 40 feet of rudely stratified red and grey drift with rounded boulders and many pebbles. This is most probably interglacial.

Attawapiskat River. Well-stratified clay with silt and sand laminae and sparse plant fragments occurs beneath 6 to 10 feet of sand and gravel and 12 to 15 feet of clayey till. The plant fragments are dated at >35,800 years (GSC-83). It is likely that the organic-bearing sediments are related to the Missinaibi beds.

Gods River. On Gods River, formerly called the Shamattawa, Tyrrell (1913) reports that intertill sand and gravel and the basal part of the overlying till contain moss and wood, partly altered to lignite. He recognized two tills over a wide area, in places separated by a striated boulder pavement. He noted the intertill sand and gravel also along Hayes River but could find no organic remains.

Seal River. A buried non-glacial deposit carrying organic materials was reported by Taylor (1961). This occurrence is on the Precambrian Shield, 85 miles west of the Paleo-

PLATE XII-3. Deep meltwater channel between unglaciated Cypress Hills and Wisconsin end moraine complex, southwestern Saskatchewan. Vertical airphoto of meltwater channel, about 700 feet deep, now occupied by Adams Creek. Road leads north to Maple Creek, Saskatchewan. Scale 1 inch to 3,000 feet.

zoic rocks, very close to the limit reached by the Tyrrell Sea. Beneath sandy till typical of Shield areas, is a fluviatile bouldery gravel with a matrix of goethite. Two 6-inch layers in the gravel contain casts and impressions of leaves and twigs partly replaced by, and set in a matrix of, goethite. The plant-bearing layers represent the replacement of a woody peat composed of moss, sedge, grass, and shrubs including herbaceous plants such as leather-leaf, dwarf lamel, and bear-berry. The plant assemblage is similar to that of the area today, and is believed to have been deposited during an interglacial interval.

Interior Plains

Early Pleistocene Events

Pediment surfaces. In late Tertiary the climate of the Interior Plains is inferred to have been arid to semiarid, and a series of pediplains was developed along the mountain front and around smaller uplands to the east (Gravenor and Bayrock, 1961; Parizek, 1964; Barton, *et al.*, 1965). With the change to a cooler and moister climate at the close of the Tertiary and in earliest Quaternary extensive valley erosion, locally accompanied by alluviation, left the present uplands as remnants of the pediplain system. Some of these uplands that escaped glacial modification are Cypress Hills (Pl. XII-3) and Wood Mountain near the International Boundary. Other uplands such as Missouri Coteau in southern Saskatchewan and Hand Hills in east-central Alberta have been extensively modified. There is some disagreement as to the degree of the effects of glaciation on the gross topography of the Plains. Contour maps of the bedrock surface give a general picture of the preglacial topography but as much as 1,000 feet of drift is known in some valleys in Saskatchewan, indicating that considerable modification of the topography has taken place.

Drainage systems. At the end of the Tertiary period a mature, dendritic drainage system existed on the Interior Plains (Fig. XII-9). A good account of early work on buried valleys dating back to that of Dawson in 1885 is given by Stalker (1961). He reports also that the glaciers responsible for the numerous till sheets recognized on the southern plains had the over-all effect of forcing rivers to follow more southerly courses. In a general way the major elements of the present-day drainage reflect the trend of the old valleys, but there are many divergences as a result of disruption and modification of the preglacial rivers. For instance, near the town of Peace River Hen-

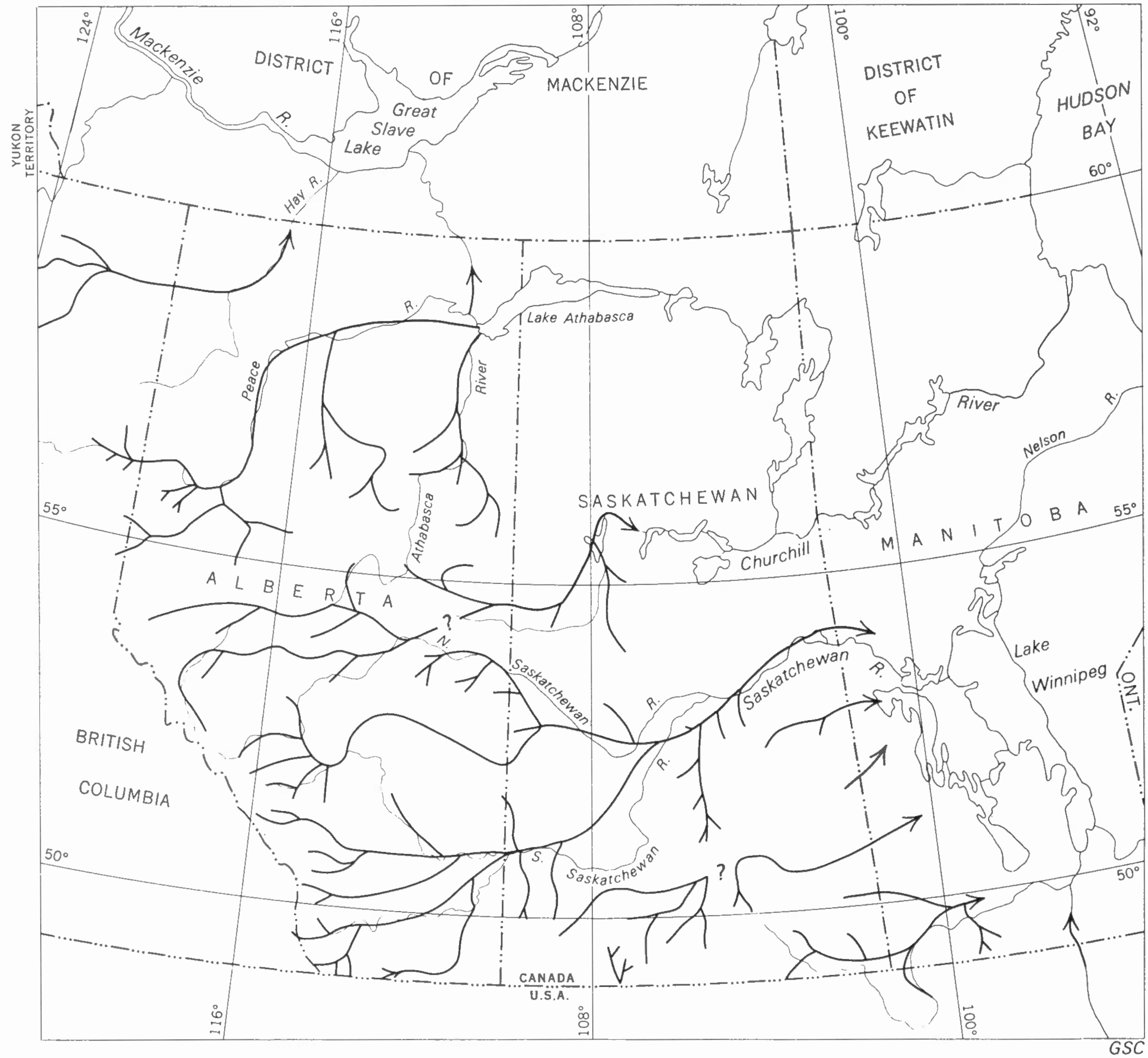


FIGURE XII-9. Inferred preglacial drainage systems of the Interior Plains (modified from available sources).

derson (1959a) records 800 feet of drift in a preglacial channel 3 miles southeast of the present river, and also shows that the preglacial Smoky River joined Peace River some 20 miles southwest of the present junction. In central Alberta the preglacial Red Deer River followed a markedly different course from that of the present, north of Red Deer; this preglacial channel is now occupied in places by Battle River. Similarly the preglacial South Saskatchewan River system in both Alberta and Saskatchewan was quite different from that of the present (Stalker, 1961; Christiansen, 1967). The preglacial Milk River in southern Alberta was probably a tributary of the South Saskatchewan rather than Missouri River, and farther east this latter river may have flowed across southeastern Saskatchewan into Manitoba and thence presumably northward to Hudson Bay rather than joining the Mississippi as at present (Meneley, 1957). The course of the preglacial Qu'Appelle River in south-central Saskatchewan is in doubt (Kupsch, 1964).

On the basis of shape and contained sediments Stalker has identified both preglacial and interglacial valleys in southern Alberta. He has noted a very dark grey till in the lower parts of all preglacial valleys. In the bottom of their channels the preglacial valleys are characterized by the presence of the Saskatchewan gravels, free of stones derived from the Canadian Shield. Gravel is uncommon in the interglacial valleys, and where present contains some Canadian Shield stones, although they may be very rare. The preglacial valleys in southern Saskatchewan are generally 4 to 10 miles wide and have gently sloping sides, whereas interglacial valleys are a mile to 2 miles wide with steeply sloping walls (Christiansen, 1967). Christiansen records that the preglacial valleys are filled with 50 to 1,000 feet of drift. They may be apparent at the surface where the drift is thin, but are completely obscured where it is thick.

Saskatchewan gravels. The occurrence of preglacial Saskatchewan gravels on the Interior Plains is of great interest and their age has long been controversial (Westgate, 1965). These buried gravels are recognized in Alberta, Saskatchewan, and Manitoba, occurring as alluvial terraces and benches below the Tertiary (Miocene–Pliocene) pediment surfaces and as lower-level channels or valley fills. They are very extensive beneath the drift mantle whether this be a few feet or more than 1,000 feet thick; generally they are a few feet to a few tens of feet thick, varying laterally from sand to sand mixed with coarse gravel. Henderson (1959a), however, reports more than 100 feet of gravel in Peace River area. Westgate found that 98 per cent of the gravel-sized material in southeastern Alberta is made up of quartzite, argillite, and chert, the remainder being arkose, limestone, a green porphyry, and local bedrock. Their source is the Cordilleran Region to the west, also earlier deposited Cordilleran gravels that capped the pediment surfaces of the Interior Plains. Over most of the Plains they rest unconformably on Cretaceous rocks. Frost-action structures have been observed beneath Saskatchewan gravels by J. A. Westgate in southern Alberta, and by Westgate and Bayrock (1964) in central Alberta. In places the gravels are undisturbed by the frost structures. A periglacial environment prevailed during deposition of at least the early part of the unit. Westgate notes the occurrence of bones of a woolly mammoth *Mammuthus primigenius,* and of a horse (*Equus* sp.) somewhat smaller than the present-day horse. He considers that the bulk of the Saskatchewan gravels probably range in age from early to late Pleistocene. Stalker, however, considers them to be early Pleistocene.

Proglacial deposits. Fine-grained sediments including varved sediments have been noted resting on Saskatchewan gravels in major preglacial valleys and directly on bedrock in their tributaries. Thicknesses up to 150 feet have been recorded. They have formerly been included with the Saskatchewan gravels as there is no evidence of a prolonged break in sedimentation prior to their deposition. Westgate (1965), however, treats them as a separate stratigraphic unit—the Wolf Island sediments. They were deposited in cold, quiet, proglacial lakes rather than in swift-flowing

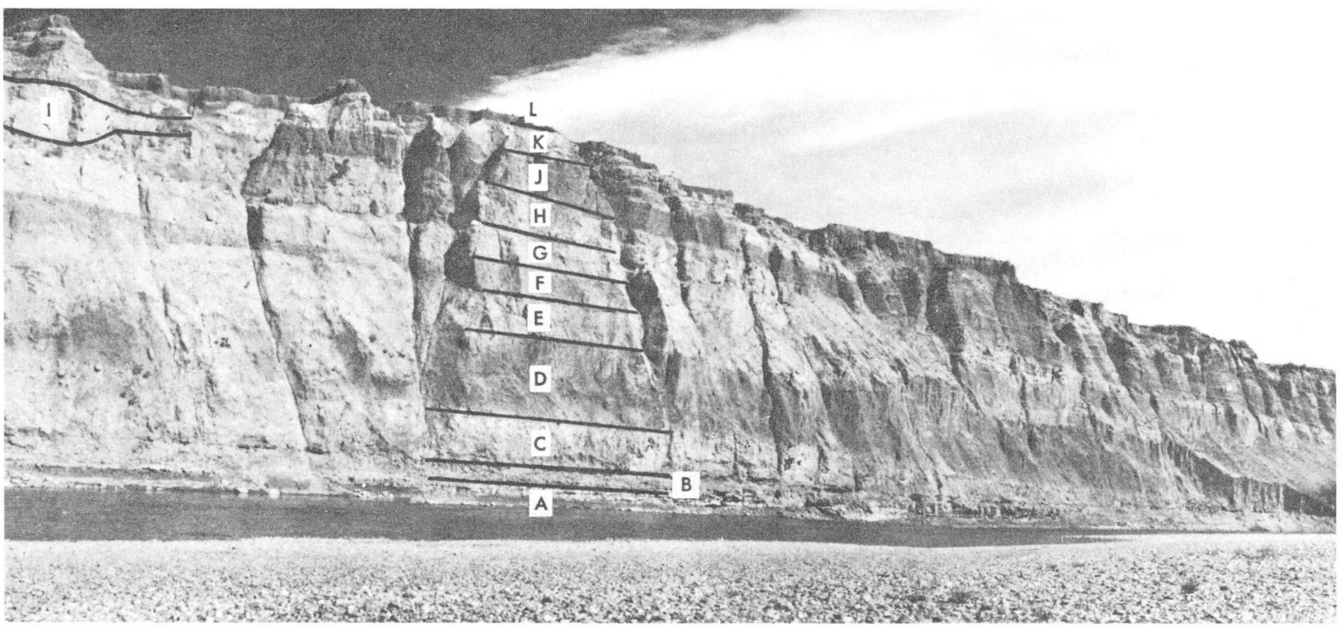


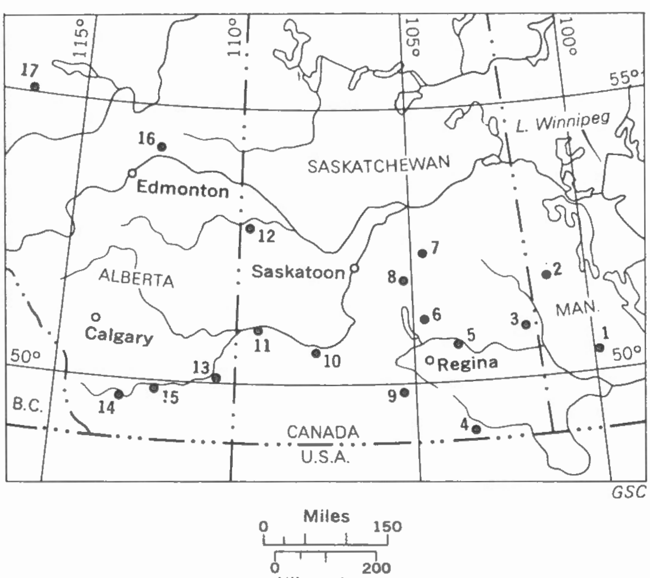
PLATE XII-4. Section of non-glacial and glacial sediments on Oldman River at Brocket, Alberta.

L. Sandy loam; modern soil.
K. Sand; lake and outwash.
J. Varved clay, silt and fine sand; glacial lake.
I. Till (Buffalo Lake); Laurentide.
H. Sand, silt, clay; lacustrine or alluvial; possibly mid-Wisconsin.
G. Till (unnamed), medium light brown; Laurentide?
F. Till (Brocket), dark brown; Laurentide.
E. Till (Maunsell), light bluish grey; Laurentide.
D. Till (Labuma), dark brown to black; Laurentide.
C. Till (Albertan), light grey; Cordilleran.
B. Gravel (no Canadian Shield stones); mostly outwash.
A. Bedrock (Willow Creek Formation); Paleocene.

streams, and appear to reflect a blockage of the northeast-flowing rivers by encroachment of Laurentide ice.

Early glaciations. Very little is known as to the course or extent of pre-Classical Wisconsin glaciers. The generally flat character of the Plains limits observations of Pleistocene stratigraphy to the relatively few sections along entrenched rivers. In Saskatchewan great use has been made of drilling rigs and side-hole samplers to establish the stratigraphy; this practice gives great promise of the establishment of a firm chronology in a region that might otherwise remain an enigma. In many places multiple till sheets, with and without intervening stratified sediments, have been recognized (Pl. XII-4). There is as yet little agreement as to the precise ages of the tills or sediments. Opinions vary from those who believe that all stages of the Pleistocene are represented to those believing in only a Wisconsin glaciation. Some writers place the limit of the Classical Wisconsin drift south of the 49th parallel while others contend that it lies well north of Cypress Hills. Only recently have radiocarbon datings proved the existence of Pleistocene age organic-bearing sediments that predate the Classical Wisconsin. Extensive oxidized zones are recognized in some thick till sections; these appear to represent periods of near-surface exposure during both interglacial and intraglacial times.

In the vicinity of Del Bonita Hills and Cypress Hills an old drift topography occurs at higher altitudes than the surrounding more youthful hummocky terrain. Westgate (1965) terms the higher and older drift the Elkwater drift and considers it correlative with the most extensive and southernmost drift sheet in Montana. He recognizes five Laurentide drift sheets in the Foremost–Cypress Hills area of southeastern Alberta and attributes end moraines to each. The ice sheets generally advanced southeast but with some variations due to lobing in the marginal zones. He considers the oldest drift sheet to be post-Sangamon. Stalker (1963), however, believes that most, if not all, the pre-Wisconsin glacial periods are represented by tills in southern Alberta.



1. Riding Mountain
2. Duck Mountain
3. Churchbridge
4. Outram
5. Fort Qu'Appelle
6. Gregherd
7. Spalding
8. Lanigan
9. Spring Valley
10. Swift Current
11. Prelate Ferry
12. Marsden
13. Medicine Hat
14. Lethbridge area
15. Taber
16. Smoky Lake
17. Goose River

FIGURE XII-10. Location of buried organic deposits in Interior Plains.

Buried Organic Deposits (Fig. XII-10)

Organic materials in stratified deposits beneath one or more till sheets are known from many places on the Interior Plains.

Riding Mountain. Buried organic materials were found in a highway-cut on the north side of Minnedosa River valley (Klassen, *et al.,* 1967). Three till units are separated by silt and sand layers, the whole comprising 61 feet (Pl. XII-5). This complex overlies a 5-foot silt layer that contains plant remains and rodent bones. The silt unit rests on a foot of limonite-stained gravel with small limestone pebbles weathered to a powder. Beneath is 16 feet of dark grey, shale-rich till, the surface of which also appears to be weathered in places. The sediments are almost devoid of pollen, presumably due to oxidation. Bone fragments are of arctic ground squirrel *(Citellus undulatus)* and a large vole (*Microtus* sp.). Plant fragments gave an age of >31,300 years (GSC-297). Several boreholes on the south side of Riding Mountain encountered stratified sediments beneath two tills and overlying a third. Wood chips from a hole near Inglis, farther northwest, were encountered in clay at a depth of 196 to 212 feet and were dated at >30,000 years (GSC-218). Bedded silts 75 feet thick occur beneath the clay unit and rest directly on shale bedrock. It may be that the Riding Mountain buried deposits are interglacial.

Duck Mountain. In 1964 R. W. Klassen re-examined the locale of interglacial sediments reported by Tyrrell (1892, p. 116). The section on Rolling (Roaring) River was heavily slumped but the Heart Hill section was clear. Two or more tills totalling 54 feet overlie 8.5 feet of silty beds containing small pelecypods, gastropods, ostracods, plant fragments, and seeds. The silt beds overlie 10 feet sand and gravel, and 15 feet sand; the lower 58 feet down to river level was heavily slumped. The upper 1.5 feet of organic-bearing silt is brown rather than grey, suggestive of a weathered zone. The contact with the overlying till is gradational over some 6 inches. The basal inch or two of the silt unit is a clayey sand, cemented with limonite. It contains pelecypods and much carbonaceous material including plant fragments. Some of the latter gave a radiocarbon dating of >38,000 years (GSC-284).

The silt unit was sampled at 2½-inch intervals and examined for both shells and pollen. Seven species of

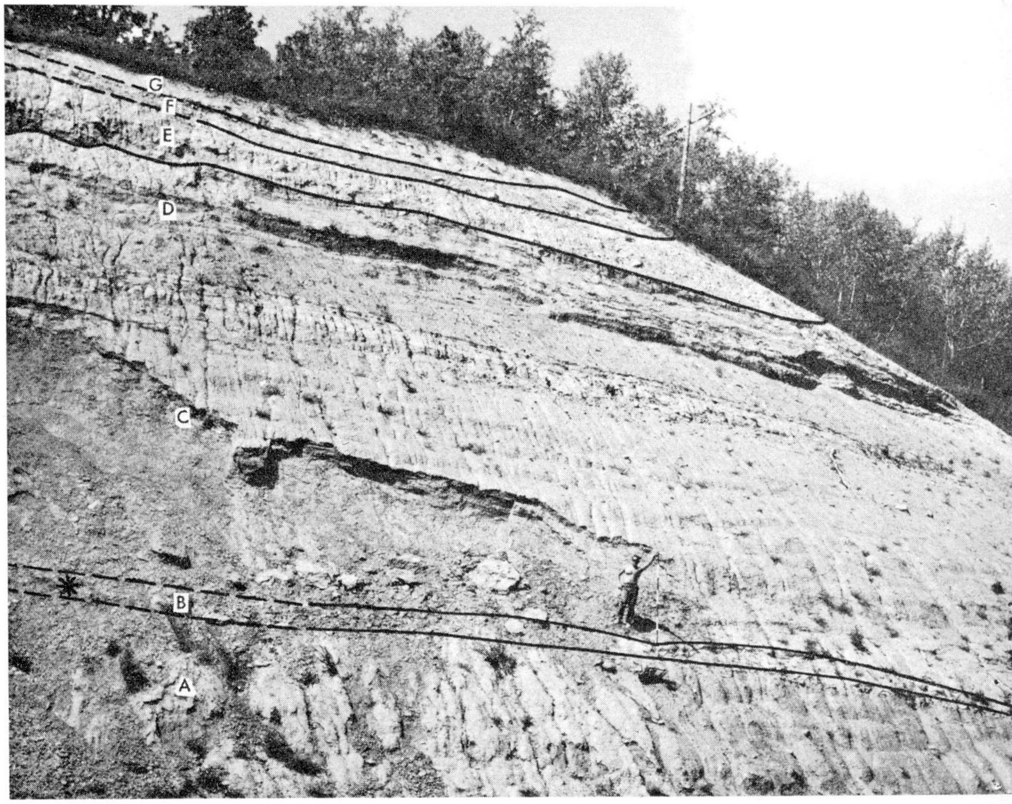


PLATE XII-5
Section of glacial and interglacial deposits on Minnedosa River, north of Minnedosa, Manitoba. A, till; B, silt with bones (collection site); C, till; D, silt, sand, gravel; E, till; F, sand and gravel; G, till.

pelecypods, three gastropods, and ten ostracods have been identified. The ostracods and the pollen indicate a moist climate about as warm as the present, which supports the interglacial interpretation of Tyrrell. The fossil assemblage also indicates a cool–warm–cool sequence; this sequence was terminated by glaciation and deposition of till.

Churchbridge. Tyrrell (1892, p. 142E) recorded the log of a well at Churchbridge that appears to indicate some 32 feet of drift, mainly till, overlying more than 235 feet of clayey and sandy sediments. A piece of wood was found at a depth of 200 feet and identified as a species of larch, *Larix churchbridgensis.* Elsewhere in the area as much as 165 feet of till is reported overlying sand or gravel.

Outram. Christiansen and Parizek (1961) report wood chips from probable lacustrine sediments at a depth of 170 feet from a test-hole in southeastern Saskatchewan. The section from the top down is, light, olive-brown calcareous till, 10 feet; grey, calcareous till with two clayey beds, 92 feet; stratified, oxidized and leached sand rich in organic matter, 8 feet; grey, calcareous till, 20 feet; grey, interbedded sand, silt and clay (from which wood chips were obtained at depth 170 feet), 71 feet; grey, calcareous till, 15 feet; stratified sediments, 144 feet; light brownish grey, calcareous till, 8 feet; gravel, 3 feet; pale yellow, kaolinitic, calcareous till, 4 feet; sand, 2 feet; and light brownish grey, calcareous till, 1 foot +. The wood chips were dated at 27,750 ± 1,200 years (S-96), which suggests correlation with the Prelate Ferry interval.

Fort Qu'Appelle. Fossiliferous sediments beneath 200 feet of drift were reported by Christiansen (1960). He reports that upstream from Fort Qu'Appelle a till overlies 10 to 25 feet sand and 20 to 40 feet of exposed gravel; down the valley towards Lebret, terraced valley-fill is at least 120 feet thick. Vertebrate fossils in the sediments, including bones and teeth of bison, mammoth, horse, wolf, and bear, are tentatively assigned to the Sangamon by L. S. Russell. Christiansen thinks that there is an older drift beneath these sediments.

Gregherd. Wood chips from sand beneath till, in a well at a depth of 300 feet, were dated at >30,000 years (S-111). The overlying till is considered late Wisconsin but the age of the non-glacial sediments is unknown.

Spalding. A boring has yielded wood chips from the lowest of three tills at a depth of 221 feet. The wood was dated at >34,000 years (S-127). It probably represents an interglacial deposit.

Lanigan. Wood from a depth of 540 feet in a drift-filled valley and only 20 feet above bedrock was encountered during sinking of a mine shaft. It was dated at >42,000 years (GSC-632), and is believed to be of Pleistocene age.

Spring Valley. Layers and lenses of peat associated with silty-clay beds occur beneath a till mantle in the hummocky ice-thrust moraine of the Dirt Hills. The peat was dated >38,000 years (GSC-790) and pollen indicates Pleistocene age.

Swift Current Creek. Wickenden (1931) reported a multiple till section, with intervening sediments and organic matter, in Swift Current Creek valley near its junction with South Saskatchewan River. From the surface downward are thin lacustrine clays; 20 to 70 feet of till; 35 feet white sand and gravel with some poorly preserved plant material; 3 to 45 feet dark grey till; and up to 225 feet sand and gravel with some poorly preserved peat and plant detritus. The last sediments rest either directly on bedrock or on a brown to yellowish till that locally displays a weathered zone, 8 feet thick, with poorly preserved plant materials and rootlets. On the basis of the great thickness of the lower gravels, the presence of plant materials, and the weathered zone on the basal till, Wickenden regarded the lower gravels as deposited during a long, moist interglacial interval. Christiansen (1959) has linked the two upper tills to end moraines in the area and assigns the upper unit of sand and gravel to a proglacial, very late part of the Wisconsin. He regards the lower stratified sediments as earlier proglacial Wisconsin deposits.

Prelate Ferry. David (1966) reported a buried soil beneath some 120 feet of drift including three tills from South Saskatchewan River valley north of Prelate. From the surface downward are lacustrine silt and clay, 16 feet; oxidized, calcareous till, 13 feet; stratified sand, 26 feet; mainly oxidized very calcareous till, 39 feet; stratified sand, silt, and a thin basal marl-like layer, 24 to 25 feet; paleosol, 3 to 4 feet; oxidized, very calcareous, sandy clay-loam till, 7 feet; stratified and crossbedded silt, sand, and gravel, 8 feet; and a basal, oxidized, calcareous till, 72 feet. The marl-like layer, only 6 to 8 inches thick, provides a widespread and useful horizon marker for locating the paleosol. The buried soil has been dated at 20,000 ± 850 years (S-176). David considers the paleosol to have developed during a non-glacial interval that he named the Prelate Ferry interval. Its beginning is unknown but it ended about 20,000 years ago when advancing Classical Wisconsin glaciers led to burial of the soil, first by stratified sediments and later by till.

Marsden. Christiansen (1965) reports a soil, buried beneath a few feet of till, about 125 miles north of the Prelate Ferry paleosol exposures. The dating of 21,000 ± 800 years (S-228) suggests that this soil was formed also during the Prelate Ferry interval.

Medicine Hat. Four miles north of Medicine Hat, in South Saskatchewan River valley, Stalker records about 100 feet of intertill sediments exposed in a small buried valley cut in till. Wood fragments and dark carbonaceous layers occur, dated at 24,490 ± 200 and 28,630 ± 800 years (GSC-205, 578). Two tills of differing lithologies overlie the stratified sediments and are considered to represent the Classical Wisconsin. Beneath the sediments in the buried valley are other tills and intervening sediments with abundant wood, bones, and shells which Stalker considers to be Sangamon or older. Wood from near river level at a site 3 miles north of town and beneath 220 feet of drift including several tills was dated >46,700 years (GSC-543). He considers the containing sediments to be of Yarmouth age.

Lethbridge. Multiple till sections have been described by Stalker (1963) on Oldman River. At one site 10 miles west of Lethbridge wood and cones of black spruce were found in the basal part of intertill sand, silt, and gravel. The wood has been dated as >54,500 years B.P. (GSC-237). Two tills occur beneath these sediments at several places along the river and at one place the basal till shows evidence of weathering. Stalker considers these stratified sediments to be present beneath an upper till as far north as Edmonton and to be of Sangamon age. The two lower tills are equally extensive.

At the Kipp section on Oldman River about 7 miles west of Lethbridge, the humic part of a wood sample from the intertill sediments is dated at >37,000 years (L-455A). Some 24 feet of intertill sediments includes a basal, consolidated, white sand that forms a prominent horizon marker in the area. The sediments are overlain by 32 feet of dark, compact till of unknown age and this in turn by some 163 feet of alluvial sediments within which wood fragments have been found. Another 45 feet of sediments overlying the alluvium includes varved clays and minor till lenses and appears to be glacial deposits. These sediments are overlain by 50 feet of sediments carrying some snail shells that may have been deposited relatively close to the ice front. Similar intertill sediments occur near Taber 30 miles east of Lethbridge where wood fragments have been dated at >32,000 years (S-65).

Smoky Lake. A log of spruce was found in till at a depth of 24 feet near Smoky Lake some 60 miles northeast of Edmonton. This wood was dated at >31,000 years (S-92).

Goose River. The northernmost occurrence of buried wood in the Interior Plains is from Goose River northwest of Edmonton. There alluvium and till overlie a crossbedded sand unit in which wood was found that dated >42,500 years (GSC-501).

Mainland Arctic Coastal Plain

A 10-to-20-mile-wide belt of older glaciated terrain on the eastern side of the Cordillera west of Peel River is continuous northward onto the Yukon Coastal Plain. This belt extends westward from Mackenzie Delta to the

Alaskan boundary (Fyles, 1966) and has subdued hummocky topography as far west as Herschel Island, but beyond this area, where the zone narrows to only a few miles, glacial landforms are extremely rare. East from Kay Point, 20 miles southeast of Herschel Island, the older glaciated terrain is bordered on the northeast by a narrow zone of Wisconsin drift. Beneath these deposits interglacial sediments have been seen on headlands between Kay Point and Mackenzie Delta. The sediments grade, in general, from silt in the west to gravel in the east; Fyles considers them to represent a single stratigraphic unit.

In the Mackenzie Delta region, Mackay (1963) found evidence for glaciation of the main and near-shore islands, and possibly also the off-shore islands. Tuktoyaktuk Peninsula was partly glaciated; evidence of glacial action is lacking for the northeastern part. Farther east on Cape Bathurst and Baillie Island, an extensive low-lying plain is underlain by silt and sand beneath which is a marine clay with rare wood (Fyles, 1966). No clear distinction between postglacial and interglacial sediments could be made. This area and part of Tuktoyaktuk Peninsula are shown as outwash (including alluvium) on the Glacial Map of Canada (1253A, *in folio*) pending clarification of the age of the deposits. Possibly older glacial as well as interglacial deposits are present along the northern edge of the Mackenzie Delta complex.

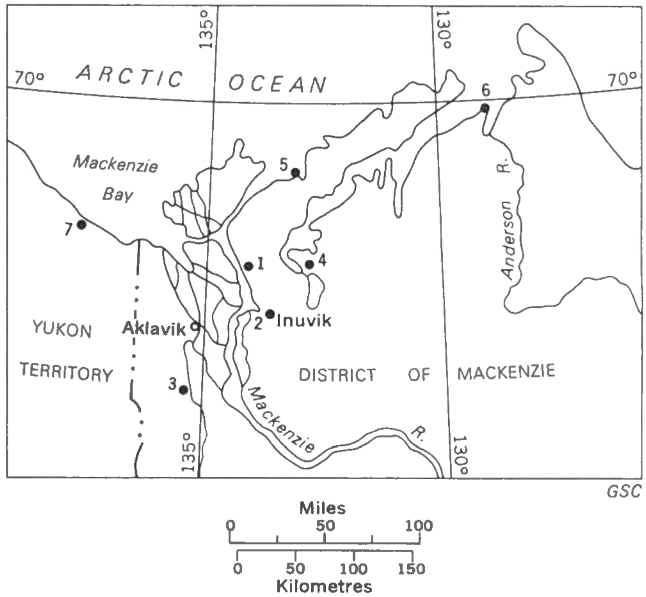


FIGURE XII-11. Location of buried organic deposits of the mainland Arctic Coastal Plain.

Buried Organic Deposits (Fig. XII-11)

Near Reindeer Depot on Mackenzie River 10 feet of peat and silt underlies some 200 feet of drift (Porsild, 1938). Cones of larch were found although the site is 50 miles beyond its present northern limit. Pollen includes abundant birch and alder, much white and black spruce, jack pine, tamarack, and willow, and some ericaceous shrubs, small herbaceous plants, grasses, sedges, and ferns. J. Terasmae considers the pollen assemblage to represent an interglacial deposit, probably the Sangamon. The peat gave an age of >42,000 years (L-522A).

Wood from a delta-kame complex at Inuvik gave an age of >39,000 (GSC-29) and was evidently derived from older non-glacial deposits. Some 60 miles southwest of Inuvik on Rat River, a beaver-chewed stick near the base of a 40-foot section of silt with organic layers and overlying an older till was dated >38,600 years (GSC-120). Evidence elsewhere along the river suggests that Laurentide ice, possibly in Classical Wisconsin time, overrode the site and extended several miles to the west.

Buried organic materials, dated at >50,900 years (GSC-329), occur southwest of Eskimo Lakes where 7 feet of modern peat overlies 76 feet of gravel that rests on at least 46 feet of organic silt with two thin peaty layers in its upper part. O. L. Hughes regards the gravel as outwash in front of a moraine left by the last major ice sheet. According to Terasmae, the pollen content of the peat is unlike that of the present and probably represents an interglacial interval. Fyles (1966) reports a great variety of materials in the southwest part of Eskimo Lakes area where organic matter was noted in three different lithologies.

A non-glacial interval is indicated by the occurrence of abraded wood from a pingo at Tuktoyaktuk dated at >33,000 years (L-300A). Also flattened wood from 20-foot depth in contorted sands, believed to be ice shoved, at the north end of Nicholson Peninsula gave a date of >35,200 years (GSC-34). O'Neill (1924) observed thin peat beds in silty sediments along the arctic coast east of the mouth of Mackenzie River, and he mentions an earlier report by Sir John Franklin of "poor lignite" in similar sediments on nearby Pullen Island.

The interglacial deposits along the arctic coast west of Mackenzie Delta contain a great variety of organic materials. Fyles (1966) reports wood up to a foot in diameter from gravels near the delta, and smaller pieces of wood, peat, freshwater shells, rare bones, and tusks from the silts farther west. Thin beds of marine clay and a few ice-wedge casts complicate the succession. An important site along this coast is at King Point 4 miles east of a moraine believed by O. L. Hughes to mark the western limit of the Classical Wisconsin ice sheet. There up to 3 feet of peat overlies 8 to 15 feet of silt and sand that rests on 20 to 30 feet of Wisconsin till. Beneath this till a stony clay, containing marine shells, grades downward into organic-bearing silts. Plant materials collected from these silts, 2 feet above the base of the seacliff, gave a radiocarbon dating of >51,100 years (GSC-151-2).

The change from plant-bearing silt to shell-bearing stony clay probably reflects a period of marine inundation as a result of the encroaching Wisconsin ice sheet.

Canadian Shield

Mainland

The mainland Shield region has been subjected to intense scouring action by a number of glaciations, but the gross physiography has probably changed little since the end of the Tertiary period. Stream valleys along rugged parts of the Quebec and Labrador coasts, however, have been markedly modified by valley glaciers stemming, in the main, from the interior ice sheet. During interglacial intervals stream erosion probably contributed to valley deepening along these same coasts but elsewhere was relatively ineffective.

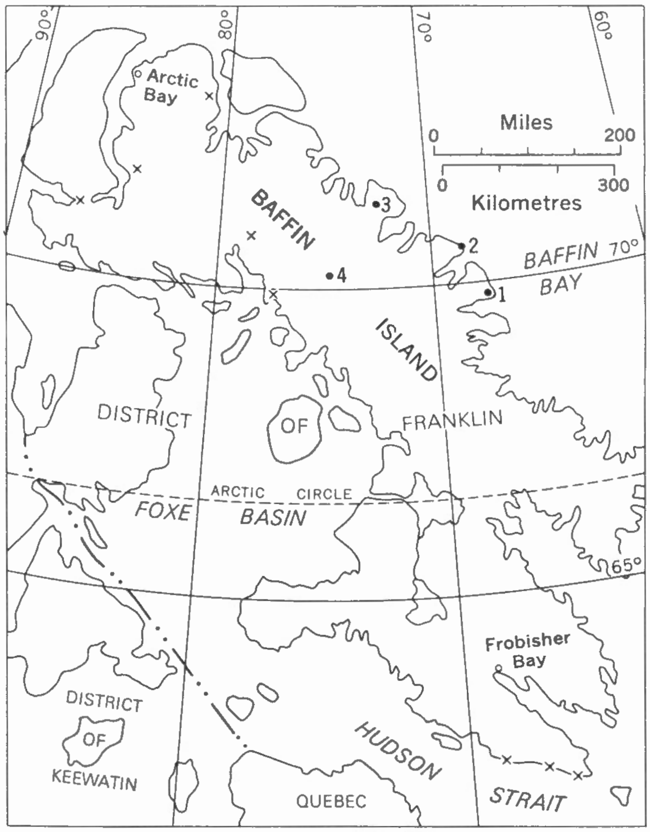
In the Torngat and Kaumajet-Kiglapait Mountains of northern Labrador, an early Torngat Glaciation is believed responsible for widely scattered erratics on the highest peaks, and the later Koroksoak Glaciation for high-level lateral moraines, kame terraces, and a felsenmere trimline well below the mountain tops (Wheeler, 1958; Ives, 1958a, b, 1960a; Tomlinson, 1959; Løken, 1962). A still lower trimline, referred to the Saglek Glaciation is regarded by Andrews (1963) as the probable upper limit of the Classical Wisconsin ice sheet. At that time large parts of the mountains projected above the ice.

The last major glaciation was generally effective in removing the deposits of older glacial and interglacial intervals; buried organic deposits are rare in the Canadian Shield. They are, however, more likely to be found in peripheral areas than in the interior, as for example the Seal River occurrence described under Hudson Bay Lowland. Old deposits may occur along some Labrador coastal lowlands where the adjacent rugged terrain shielded them from the glaciers issuing along the fiords, as on Baffin Island.

Baffin Island

The gross physiography of Baffin Island throughout the Pleistocene was probably much the same as now. In Tertiary time an old erosion surface in central and southern Baffin Island, preserved as concordant hill-tops, was tilted, northeast side up (Goldthwait, 1950). A system of dendritic and parallel valleys resulted, and the drainage divide gradually migrated westward. Glaciers stemming from the interior developed deep outlet valleys and speeded the westward migration of the drainage-divide. The resulting broad strip of mountains along the northeast coast rise to elevations of about 6,000 feet in the Penny Highlands. Bird (1954) reports that lower erosional platforms can be recognized in western Baffin Island down to a 600-foot surface cut in relatively weak rocks. He regards the widespread horizontal surface developed on rolling Precambrian rocks as probably exhumed from beneath a cover of Paleozoic rocks of which remnants survive around Foxe Basin; stripping of this mantle may have been accomplished mainly by glaciers. He states that on Brodeur Peninsula, in northwestern Baffin Island, a well-preserved surface at 1,000 to 1,200 feet surrounds a central upland at about 2,000 feet. These surfaces had reached a late-mature to old-age stage before being elevated. These high erosional surfaces and the main tilted plateau of Baffin Island have persisted throughout the Pleistocene with but little modification by successive glaciations.

Old organic deposits (Fig. XII-12). Information on the pre-Classical Wisconsin history is limited though the occurrences of old organic materials are widespread. Information on earlier Pleistocene events has been obtained from terrain along the coast that was not recently glaciated and from organic deposits beneath drift in the interior. Baffin Coastal Lowland is several miles wide east of Barnes Ice Cap and does not appear to have been covered by ice in the last 50,000 years, though there is some evidence of an older glaciation (Løken, 1966). The last ice sheet supported active glaciers in the main fiords



1. Cape Aston
2. Cape Christian
3. Bruce Mountains
4. Isortoq River

'Old' shell sites: shells winnowed from pre-Classical Wisconsin deposits . . . . ×

FIGURE XII-12. Occurrences of 'old' organic materials on Baffin Island.

of the northeast coast and in a few places formed piedmont glaciers on the coastal lowland. Near Cape Aston, south of Clyde, a delta was built at an altitude of 262 feet, while outlet glaciers filled a nearby fiord and meltwater channels spilled over towards Cape Aston. Marine shells in the deltaic sediments, at elevation 200 feet, are indicative of a cold water environment. They were dated at >54,000 years (Y-1703). Beneath the delta, 25 feet of gravel overlies stony, shell-bearing materials that represent marine deposits ploughed-up by a glacier. Old non-glacial deposits also occur interbedded with glacial deposits in the coastal cliffs near Cape Christian, again in a protected position. Marine shells from a near-shore deposit lying between the two uppermost glacial deposits have a radiocarbon age of >50,000 years (Y-1702). In Bruce Mountains northeast of Barnes Ice Cap flattened pieces of willow and associated moss were found on a low moraine ridge; these appeared to have been winnowed from the till. They are dated at >39,600 years (GSC-209) and may be interglacial.

In the interior of Baffin Island, pre-Classical Wisconsin organic deposits have been found in several places. Fine plant detritus >40,700 years old (GSC-427) occurs in several layers of contorted fluvial sands along Isortoq River only 10 miles west of northern Barnes Ice Cap (Terasmae, *et al.*, 1966). Leaves of *Dryas, Vaccinum,* and *Ledum* suggest a climate at least as warm as the present and thus presumably interglacial. Andrews reports that the beds were contorted by ice flowing westward towards Foxe Basin. Marine shells from western Baffin Island, 100 feet above the postglacial marine limit, were dated at 30,320±820 years (GSC-528); this date may prove to be minimal.

In northern Baffin Island shells indicative of pre-Classical Wisconsin marine events were obtained from varied materials both above and below the postglacial marine limit: finite dates of about 34,200 and 35,400 years (GSC-184, 188) were obtained from Borden Peninsula, and a dating of >30,580 years on shells (GSC-189) from southern Brodeur Peninsula.

In southern Baffin Island shells found in till have given finite dates of about 30,200 and 34,800 years (GSC-414, 426). Blake (1966) regards these as minimum dates and concludes that Hudson Strait was open during the last interglacial or some early Wisconsin interstadial, and that subsequently, during an ice advance, material was scraped from the sea bottom and deposited along the south coast of Baffin Island. Old shells, washed from till, have been found also in postglacial beach deposits (GSC-468).

Some finite radiocarbon age-datings from Baffin Island may indicate an interstadial interval the lower limit of which is unknown. This is not yet proven, but neither do the numerous infinite age-datings prove that the relevant deposits are interglacial. It appears that both types of interval may be represented by the organic-bearing deposits of Baffin Island. However, undoubted glacial till overlying a fossiliferous deposit has yet to be found.

Arctic Islands

The nature of glaciation and the physiography of the arctic islands differ greatly from one region to another. Except for part of Banks Island, the southern islands were overridden by the Laurentide Ice Sheet so that the early Pleistocene record was removed or lies buried by the younger drift and fresh glacial features characterize the landscape. Some of the western islands beyond the northwestern limit of the Wisconsin Laurentide Ice Sheet were covered by an older Laurentide ice sheet. The western part of the Queen Elizabeth Islands is a lowland and includes areas of old and young glacial deposits, which may be related to both regional and local glaciers, and also some areas lacking obvious evidence of former glacial activity (Fyles and Craig, 1965). The main, eastern part is a region of mountains and plateaux that developed their own ice cover and have been extensively scoured. There is, nevertheless, some evidence of the older Pleistocene record.

In the eastern Queen Elizabeth Islands the mountain ranges, plateaux, and lowlands had assumed their present form by the beginning of the Quaternary. High-level terraces and pediment-like surfaces, with associated wood-bearing alluvium on Ellesmere and Axel Heiberg Islands are regarded as remnants of a system of mature preglacial valleys that may perhaps be early Pleistocene. Subsequent fluvial and glacial erosion have carved fiords and inner valleys as much as 1,800 feet below the old valley levels and in some lowland areas underlain by soft rocks have locally reduced the preglacial deposits to hilltop remnants. In the western islands, however, the landscape was reduced to a northwest-sloping surface of low relief when the latest Tertiary and early Quaternary sediments of the Beaufort Formation accumulated on Arctic Coastal Plain. Islands and straits (formerly valleys) comprising the western part of the archipelago are believed to be younger than the Beaufort Formation (Craig and Fyles, 1960) and to be the result of faulting and repeated cycles of fluvial and glacial erosion (Morley and Fortier, 1956; Thorsteinsson and Tozer, 1961).

Preglacial Deposits

Unconsolidated, wood-bearing alluvial sediments of the Beaufort Formation occupy most of the Arctic Coastal Plain from Banks Island to Meighen Island and are believed to extend seaward beyond the present coast; they occur also as isolated patches on hilltops to the east. These beds are tentatively regarded as late Tertiary to early Quaternary. Fyles (1965) has suggested that certain high-level terrace deposits on Axel Heiberg and Ellesmere Islands are also Beaufort equivalents (Pl. XII-6). The Beaufort Formation on Ellef Ringnes and Borden Islands consists of a lower unit, mainly brown silt with

PLATE XII-6
High-level terraces with Beaufort-type sediments overlying Cretaceous bedrock, eastern Axel Heiberg Island, Northwest Territories. View southward from Buchanan Lake. Only the uppermost light coloured material, up to 40 feet thick, is considered Beaufort type. Surface is strewn with a few glacial erratics.

peaty layers, and an upper unit of sand or fine gravel with driftwood. Only the upper unit has been observed on Prince Patrick and Banks Islands where logs of wood up to 2 feet in diameter have been observed.

The Beaufort Formation characteristically contains pollen of spruce, pine, birch, alder, various herbaceous plants such as the ericaceae, and spores of ferns and moss (Fyles and Craig, 1965). Many samples also contain small amounts of pollen of hemlock and of the temperate climate hardwoods such as hazel, beech, elm, hornbeam, and oak. It is thought that much of the hardwood pollen may be secondary, derived from nearby Tertiary strata, but some is definitely not. The pollen from many species other than the hardwoods comprise an imposing array of plants that grew in what is now a treeless, barren country. The assemblage indicates a climate not only much warmer than the present but also warmer than might be expected in any of the interglacial intervals and may be representative, at least in part, of preglacial earliest Quaternary.

On Meighen Island, which has an ice cap at present, the wood-bearing sands of the Beaufort Formation overlie unconsolidated, marine clay and silt (Thorsteinsson, 1961). On the basis of pollen content, which indicates a warm climate, the marine unit is Tertiary, but the sand unit may be early Quaternary. The sand underlies a boulder veneer presumed to be of glacial origin (Fyles and Craig, 1965). The Beaufort Formation on Arctic Coastal Plain of Borden Island is particularly interesting in that hardwood pollen is abundant and the remoteness of known Tertiary rocks suggests a primary rather than a secondary plant assemblage. The wood-bearing sands on Ellef Ringnes Island are believed to be inland, high-level parts of the Beaufort Formation (St. Onge, 1965). The organic-bearing sediments which occur as high-level terrace deposits over large parts of Ellesmere and Axel Heiberg Islands are probably the temporal equivalents of the Beaufort Formation of the coastal plain as they have a similar pollen assemblage. These deposits, characteristically gravel, sand, and silt, have thicknesses up to 200 feet and commonly lie 300 to 2,000 feet above the valleys. They appear to be remnants of broad mature valleys. They commonly contain wood and locally exhibit buried soil profiles, and enclose layers of moss or peat, and peaty beaver-pond deposits with gnawed wood (Fyles and Craig, 1965).

On northwestern Banks Island the Beaufort Formation lies at the surface (Fyles and Craig, 1965). In the higher, eastern part of the island it is only a few feet thick and rests on what appears to be an eroded Tertiary sand. In the lower central part the deposits are thicker and their base is not exposed, as on Ballast Brook where 300 feet of sand and gravel with much wood is underlain by 10 feet of peat which in turn rests on sand and silt containing plant detritus.

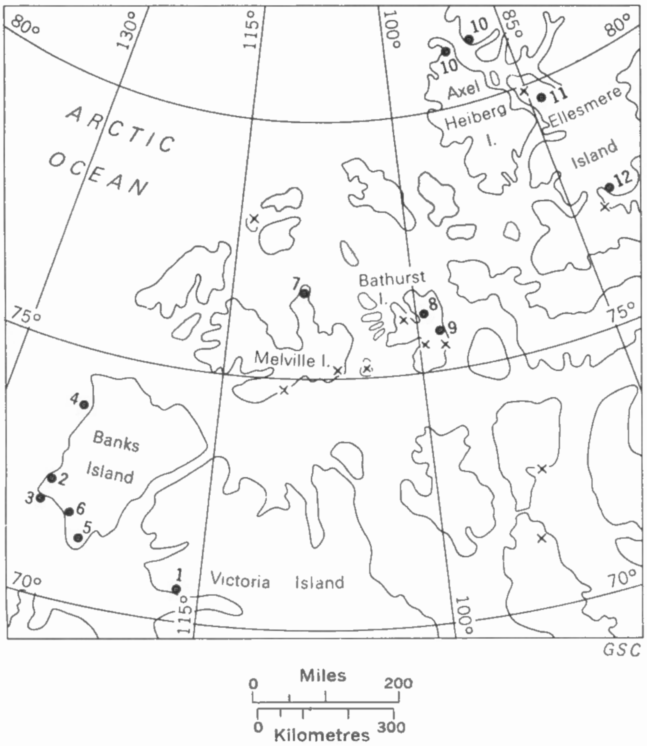
Early Glacial and Non-glacial Events

Terraces and plains within the areas underlain by Beaufort Formation consist of reworked sand and gravel. On Prince Patrick Island, most if not all of these terraces are younger than the northward-trending faults cutting the Beaufort Formation (Fyles, 1965). They indicate alluviation during some pre-Classical Wisconsin non-glacial period.

Parts of Prince Patrick, Eglinton, and Melville Islands, and most of northern and western Banks Island, bear deposits of an older glaciation, and particularly of far-travelled erratics from the Canadian Shield. Study of buried organic deposits has shown that both interglacial and interstadial intervals are probably represented.

Buried Organic Deposits and 'Old' Shell Occurrences (Fig. XII-13)

Victoria and Banks Islands. Fyles (1963a) reports sub-till organic-bearing sediments from the north side of Prince Albert Sound in western Victoria Island. Locally gravels with knob-and-kettle topography rest on 30 feet of till overlying 150 feet of stratified sediments that in turn rest on some 50 feet of dense till. In the lower part of the intertill sediments, 20 feet of thin-bedded silt contains closely packed mats of leaves and other remains of small plants which were radiocarbon dated at about 28,000 years (I-GSC-30). One of the plant layers contains



1. Prince Albert Sound, Victoria Island
2. Worth Point, Banks Island
3. Duck Hawk Bluff, Banks Island
4. Bernard Island, off Banks Island
5. Nelson Head, Banks Island
6. Masik Valley, Banks Island
7. Barrow Dome, Melville Island
8. Stuart River, Bathurst Island
9. Goodsir Inlet, Bathurst Island
10. Nansen Sound, Ellesmere and Axel Heiberg Islands
11. Slidre River, Ellesmere Island
12. Makinson Inlet, Ellesmere Island

'Old' shell occurrences; various islands. . . . .×

FIGURE XII-13. Location of buried organic sites and 'old' shell occurrences on the arctic islands.

sparse pollen of grasses, sedges, and other herbaceous plants. Fyles considers these intertill deposits to have been deposited by rivers and in lakes prior to the last glacial invasion under climatic conditions not greatly different from the present. Similar sediments, lacking organic remains, lie beneath till throughout large parts of northwest Victoria Island and the adjoining coastal region of Banks Island and may be correlative.

At Worth Point, on western Banks Island, beyond the inferred limit of the Classical Wisconsin Laurentide Ice Sheet, an uncompressed peat at the top of a 100-foot exposure of till and stratified sediments has yielded a date of >49,000 years (GSC-367). The peat contains pollen of birch as well as of plants now growing in the area. A stony layer beneath the peat is considered to be colluvium, for beneath this layer are pond silts with layers of only slightly compressed peat and moss. The peat contains wood of small trees, and some beaver-gnawed sticks. The wood gave an infinite radiocarbon age (I-GSC-19). These deposits are believed to be interglacial. At Duck Hawk Bluff, near the top of a 125-foot shore cliff, peat and willow stems occur in silts beneath colluvium or possibly till; the organics also gave an infinite age (GSC-238). The peat contains pollen of alder, willow, spruce, and herbaceous plants and is considered to be an interglacial deposit. The silts rest on an older till that overlies the Beaufort Formation. Another very similar section occurs on Bernard Island, farther north off the west coast of Banks Island. There moss peat from lacustrine deposits contained pollen of conifers as well as of tundra vegetation like that now found in the area. The peat gave an infinite radiocarbon age (I-GSC-28) and the deposit is believed to be interglacial interval.

On southern Banks Island, near Nelson Head, willow wood was found in a sand, gravel, and silt unit beneath 150 feet of till and glaciofluvial gravel and sand. This was dated >41,600 years (GSC-222). The overlying drift dates from the last major glaciation but the organic deposits may be either interstadial or interglacial. In Masik River valley, beneath till, Fyles also has observed spruce wood and layers of moss and peat associated with silt, sand, and gravel. He tentatively considers these organic deposits to date from an interglacial interval.

Melville Island. On northern Melville Island, where little vegetation grows, 2.5 feet of moss and sand was found 6 feet below the surface of a flat-topped hill, at about elevation 300 feet. The deposit appears to predate both glaciation and dissection of the lowland but to post-date the formation of the higher level terraces tentatively considered to be of Beaufort age. The moss, dated at >38,600 years (GSC-422), is probably interglacial.

Shell fragments from raised beach deposits in south-eastern Melville Island that were expected to be post-glacial yielded a radiocarbon date of about 34,050 years (GSC-154). The date is probably minimal and relates to a pre-Classical Wisconsin non-glacial interval.

Also on adjacent Byam Martin Island a shell sample from an outwash deposit was presumably derived from older materials as it yielded an infinite age-dating (GSC-357).

Bathurst Island. On Stuart River, in northern Bathurst Island, postglacial marine silty-clay overlies a terrace within which 4 feet of peat overlies 20 feet of gravel of local derivation. This rests on the eroded surface of a remotely derived white sand possibly of Tertiary age. The gravels are presumed to represent a period of aggradation during some pre-Wisconsin interval during which the sea level was higher than at present. This was followed by a relatively long period of stable conditions during which 4 feet of peat formed in the valley. The peat has been dated at >36,000 years (GSC-165).

Near Goodsir Inlet in east-central Bathurst Island, peat dated at >35,000 years (GSC-178) was found beneath till and as pods within it. Blake (1964) regards the peat to be interglacial although a Wisconsin interstadial is possible. He reports that the till is similar to that from which marine shells have been collected at sites above the postglacial marine limit. Finite datings of about 35,900 and 33,940 years (GSC-212, 378) on shells from sites in central part of island are thought to be minimal.

Ellesmere and Axel Heiberg Islands. J. G. Fyles has observed terrace deposits that are at lower levels than the Beaufort terraces, and he tentatively considers these to be interglacial. In the vicinity of Slidre Fiord on west-central Ellesmere Island he found, associated with a gravel terrace, sedge and moss peat that have been dated at >41,200 years (GSC-268). Boulders on the gravel terrace are probably an indication that it has been glaciated. Glaciation is believed responsible also for the dissemination of old marine shells on similar upland surfaces south of Slidre Fiord and elsewhere on Ellesmere and Axel Heiberg Islands both above and below the postglacial marine limit; these have yielded radiocarbon dates in the 20,000 to 40,000 year range, which, due to possible contamination by old shells, are considered minimal. They provide evidence, however, of marine events prior to the last major glaciation.

Along Nansen Sound, on both Ellesmere and Axel Heiberg Islands, fossiliferous marine sediments, considered to be interglacial, occur well above the highest postglacial marine features and locally they appear to be overlain by glacial drift (Fyles and Craig, 1965). The shells have yielded radiocarbon datings ranging between 35,000 and 40,000 years (GSC-65, 113, 149), which are considered minimal.

Sandy moss peat exposed in the base of a meltwater channel near the head of Makinson Inlet in southern Ellesmere Island overlies bouldery gravel and is overlain by boulders. The peat was dated at >36,400 years (GSC-140), and along with the gravel and boulders is believed to be a remnant of a high-level terrace that may be of early glacial or interglacial age.

Shells from the sandy ground surface at the marine limit on Swinnerton Peninsula, on the southwest side of the inlet, were radiocarbon dated at 29,800±200 years (GSC-134). These shells are believed to record a marine episode prior to the last glaciation.

Cordilleran Region

The Pleistocene history of the Cordillera is somewhat more varied than that of the continental interior in that parts remained unglaciated and other parts were glaciated during only one or more periods prior to the last major glaciation, presumed to be the Classical Wisconsin. The latter, though not everywhere as extensive as older glaciations, did cover the greater part of the Cordillera. Geomorphological studies are limited but the presence of buried valleys and thick sequences of buried glacial and non-glacial sediments do bear testimony to major drainage changes resulting from successive glacial and non-glacial intervals. The record of older Pleistocene events may be preserved in its entirety in the unglaciated parts of the Yukon or in adjoining areas that were covered only by the older glaciers. At present data are sparse. The older Pleistocene record is also fragmentary in the region covered by the last major glacier complex, but locally there is much data bearing on the older events.

Unglaciated Areas

Unglaciated areas in the Cordilleran Region are shown on the Glacial Map of Canada (Map 1253A). A small area in southwestern Alberta lies above and beyond the limit of Cordilleran valley glaciers from the west and Laurentide Ice Sheet from the east. Cordilleran glaciers were restricted in their development on the dry, eastern flank of the Rocky Mountains. The rugged terrain of this unglaciated area is unsuited to the preservation of non-glacial sediments that record early Pleistocene history and climate. Unglaciated areas of very irregular outline also occur in parts of Mackenzie Mountains and Liard Plateau. These existed between the Cordilleran and Laurentide glaciers because of a combination of climatic and topographic factors. There is no information on the Pleistocene stratigraphy of these regions.

By far the largest unglaciated area in Canada is in western Yukon. This area remained unglaciated throughout Pleistocene time as it lies in a dry belt east of the high St. Elias Mountains where most of the moisture in Pacific air masses is precipitated. During glacial intervals western Yukon was both dry and cold and sustained flora and fauna less varied than that of the dry and warm interglacial intervals or that of the present. The record of glacial, interglacial, and interstadial intervals must be sought therefore in the non-glacial sedimentary sequence. To date there is little specific information.

O. L. Hughes states that on Porcupine River, in Old Crow Basin, modern peat mantles 29 feet of silt and clay which he refers to a period of meltwater discharge across Richardson Mountains from a Laurentide ice sheet. Beneath this unit is some 144 feet of brown sand and gravel from which wood was radiocarbon dated at >41,300 years (GSC-199). The older unit also contains pollen indicative of an interglacial interval. This sedimentary sequence records the present non-glacial interval, an older glacial interval, and a still older interglacial interval, but the precise chronology remains unknown.

Throughout most of the unglaciated Klondike area O. L. Hughes has found that the base of the modern peat ranges in age from about 9,000 to 11,000 years B.P., whereas the subsurface organic materials associated with silt and gravel, encountered in placer operations are generally too old to date by radiocarbon methods. At one site on Hunker Creek, however, wood from the base of 20 feet of silty peat was dated 9,520 ± 130 years (GSC-73), whereas wood from 4 feet below this in frozen silt gave a finite date of ca. 30,800 years (GSC-88). Sediments of this mid-Wisconsin interstadial interval were no doubt deposited in many parts of the unglaciated Yukon.

Older Glaciated Areas

In many parts of the Cordillera there is evidence that one or more early glaciations reached higher altitudes than the last major glaciation or extended beyond its outer limits (Map 1253A). Fyles (1963b) reports that elevation of the uplands of southern Vancouver Island and consequent development of narrow stream channels took place prior to deposition of any of the surficial deposits found there at present. These materials and the glacial features are believed to date from mid-Pleistocene time. Pleistocene glaciers overrode the region and extensively modified the existing valleys and elsewhere smoothed and rounded the major topographic features. The area bears evidence of at least two separate glaciations. In the Okanagan Range of Cascade Mountains in south-central British Columbia, ice at one time overrode mountains up to 8,500 feet whereas the last ice sheet did not reach over 7,500 feet. In the Foothills of southwestern Alberta an area of older glaciated terrain surrounding the unglaciated area lies about 500 to 1,000 feet above the Wisconsin limit of both Cordilleran and Laurentide ice.

In Mackenzie and Selwyn Mountains most of an elongate, partly unglaciated region bears evidence of one or more early Pleistocene Cordilleran as well as Wisconsin Cordilleran glaciations. The early glaciers originating in western Mackenzie Mountains and eastern Selwyn Mountains apparently filled the intermontane valleys to higher levels than did the Wisconsin glaciers. The fringe zone of older glaciations is thus very complex and at present remains undifferentiated (Map 1253A). The region lies above and beyond the western limits reached by Laurentide Ice Sheet. The Cordilleran glaciers, however, on flowing northward towards the Arctic Red and Peel Rivers encountered the older, southward-moving, Laurentide Ice Sheet. In this area a zone of older glaciated terrain is both extensive and distinct, but the interrelations of the Cordilleran and Laurentide ice masses are uncertain.

A zone of older glaciated terrain, some 10 to 20 miles wide, lies west of Peel River between the unglaciated part of the Yukon and the western limit of Wisconsin Laurentide Glaciation. Glacial landforms are relatively uncommon in this area, but areas of subdued hummocky terrain occur.

In southwestern Yukon, north of St. Elias Mountains, an area of greatly variable width separates unglaciated terrain from areas to the east and south that were glaciated during the Classical Wisconsin. Several glaciations have been recognized in this fringe zone (Denton and Stuiver, 1967; Bostock, 1966). It appears also that older glacier complexes in the interior mountainous parts of the Yukon attained greater thickness than the Wisconsin ice. Some peaks in the dry central part of the Yukon that protruded above the Wisconsin glacier surface were covered by one or more older glaciers and bear a fringe of older glaciated terrain. Near Kluane Lake on the northeast side of St. Elias Mountains, on the other hand, two older glaciations, the Shakwak and Icefield, were not so extensive as the Kluane or Classical Wisconsin Glaciation. The older glaciers extended 70 to 75 miles northeast of the present Pacific–Interior ice divide, and reached the eastern part of Kluane Lake basin. The last major ice sheet filled Kluane basin and extended 90 to 100 miles northwest along the valley to the vicinity of Snag, Yukon, with an ice tongue extending 20 miles northwards down White River (Bostock, 1952). There, an older glacier was responsible for drift extending another 20 miles down the river. Correlation between the older drifts of Kluane area and that of White River valley has not been made, but apparently the latter is older than either of the former and probably is pre-Illinoian.

In central Yukon two clearly defined glacial limits are evident and may be readily correlated and traced (Bostock, 1966). They are considered to mark the limits of the last and an earlier glacier advance. In addition, older glacial drift, including till and erratics, and modified glacial landforms have been found beyond the limit of the earlier advance and may record two still earlier glaciations. In Stewart Valley the McConnell was the last major glaciation. The youngest of the older drift units, the Reid extends westward down the valley at least 40 miles beyond the McConnell limit; it displays distinct glacial features, including an end moraine. West of the Reid End Moraine, in lower Stewart Valley, glacial landforms are generally lacking but grey, silty to clayey till with mostly fresh stones and scattered erratics represents the older Klaza Glaciation.

The Klaza glaciers extended at least 25 miles beyond the Reid limit. In some other valleys also, subdued glacial landforms related to the Klaza glaciers have been observed. In lower Stewart Valley the oldest glacier extended about 25 miles beyond Klaza glacial limit and reached to within

20 miles of Yukon River; it is referred to as the Nansen Glaciation. It is expressed by a great thickness of drift that has a hummocky to undulating surface or is locally terraced. This drift blocked the channel of a creek, which then cut a canyon 300 feet deep in bedrock. In some valleys Bostock noted that the Nansen drift is extensively weathered; the stones in the drift are rotted, and usually both till and gravels are brown. Till beneath a lava flow, 3 miles downstream from Fort Selkirk on Yukon River, is believed to be related to the Nansen Glaciation but may be older. On the north side of Stewart River valley, the Nansen glacier appears to have reached the altitude of a low pass in the valley side so that melt-waters flowed northward depositing a distinctive gravel train along the valley of Australia Creek and Indian River as far as their junction with Yukon River about 40 miles below the mouth of Stewart River.

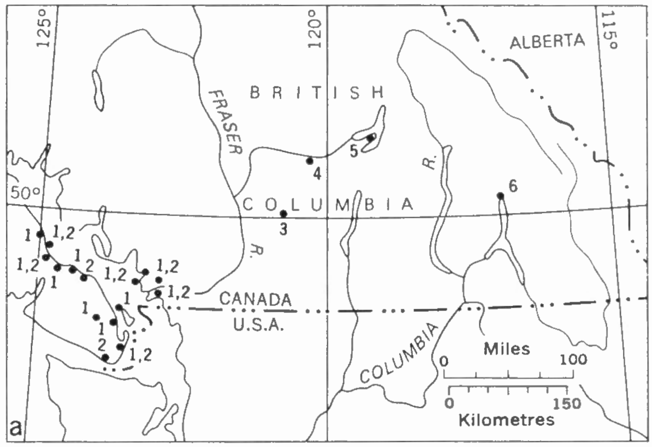
Bostock states that in upper Stewart Valley the profiles of the ice surfaces of the major glacier advances rise eastward gently and converge in elevation so that the moraines of the McConnell advance are so close to those of the earlier advances that they are virtually indistinguishable. On Talbot Plateau, southeast of Mayo, the McConnell and Reid Moraines reach altitudes of 4,000 and 4,400 feet respectively, and scattered erratics believed to be remnants of the Klaza Glaciation occur up to an elevation of 4,700 feet. Two small buttes rising above 4,700 feet elevation may have been nunataks. Other moraines are present in the area some of which may reflect oscillations of the ice fronts during the older glaciations.

In west–central Yukon, Vernon and Hughes (1965) found scattered evidence of one or more old glaciations beyond or above a clearly defined glaciated terrain of an intermediate glaciation which, in turn, lay beyond or above the deposits of the last major glaciation. Little is known of the oldest glaciations other than the occurrence of scattered erratics in otherwise apparently unglaciated areas. The trace of large transection glaciers, formed during the intermediate glaciation, have been delineated in the eastern part of the region and their gradients over long stretches have been determined as from 19 to 35 feet to the mile.

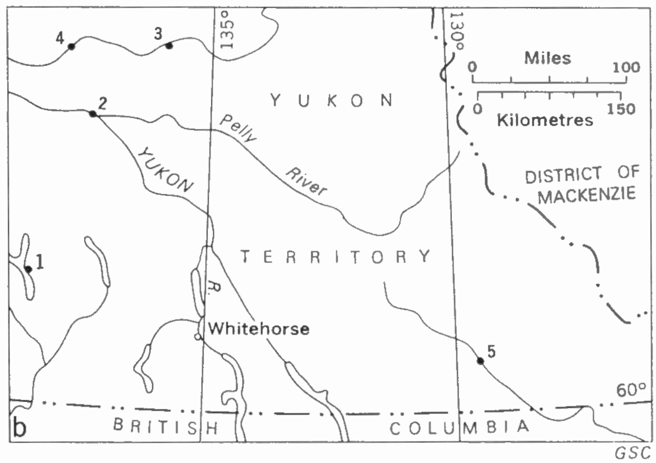
Buried Organic Deposits of the Main Glaciated Areas

Both interglacial and interstadial deposits are recognized in southern British Columbia and perhaps are the equivalents of the Sangamon and mid-Wisconsin intervals. In Yukon the extent of several Pleistocene ice sheets have been delimited but only in one area has it been possible, tentatively, to assign the buried organic deposits—and hence their associated glacial deposits—to any particular Pleistocene period. The location of buried organic deposits in southern British Columbia and the Yukon Territory are shown on Figures XII-14 (a) and (b) respectively.

Straits of Georgia and Juan de Fuca. In the Strait of Georgia region subtill organic-bearing deposits, known as the Quadra sediments, have been assigned to the Olympia Interglaciation which preceded the last major or Fraser Glaciation (Armstrong, *et al.*, 1965). For comparative purposes, these may equate, on the basis of radiocarbon datings, with the mid-Wisconsin and Classical Wisconsin of the mid-continent and Great Lakes region. Recent investigations by J. G. Fyles and E. C. Halstead have shown the wide distribution of Quadra sediments on southeastern Vancouver Island. According to Fyles (1963b), the Quadra sediments comprise a lower unit of clayey silt with stones and marine shells, a middle unit of plant-bearing silt, gravel, and sand, and an upper unit of white sand with local beds of gravel and plant-bearing silt. Maximum thickness of these units is about 80, 40, and 250 feet.



1. Vancouver Island and Fraser Lowland; Quadra sediments
2. Vancouver Island and Fraser Lowland; pre Quadra sediments
3-5. Thompson River valley and tributaries; younger and older intertill sediments
6. Kootenay Lake; palaeosol and intertill sediments



1. Kluane Lake; younger and older intertill sediments
2. Fort Selkirk area; sediments beneath striated lava
3. Mayo Landing area
4. McQuesten area
5. Liard River

FIGURE XII-14. Location of buried organic deposits in the Cordilleran Region: (a) Southern British Columbia; (b) Glaciated parts of Yukon Territory.

The lower unit is considered to represent a transition from glaciomarine to marine conditions. The middle, plant-bearing unit appears to have originated in a swampy coastal lowland during, and also probably following, a regression of the sea in which the underlying marine clay had been deposited. The upper unit is a fluvial-plain deposit characterized by cut-and-fill structures; it contains beds of gravel and plant-bearing silt composed largely of debris from the Coast Mountains of the mainland. Fyles suggests that the present Strait of Georgia may have been completely filled with sediments prior to the Fraser Glaciation.

Pollen and marine molluscs in the Quadra deposits record a climate cooler than the present and somewhat comparable to the present climate of the Gulf of Alaska. The vegetation differs from that on Vancouver Island today in rarity of Douglas fir and abundance of spruce. Radiocarbon dates range from about 20,000 to perhaps 50,000 years B.P. Thus, either the lower limit of a very lengthy non-glacial interval remains to be established or two separate intervals are represented by deposits currently assigned to the Quadra on Vancouver Island.

Quadra sediments on Vancouver Island in places overlie a till and associated sediments, the Dashwood drift, which in turn rest on the non-glacial Mapleguard sediments. At Icarus Point, E. C. Halstead states that peat in the basal part of Quadra-type sediments overlies a till that rests on peat, silt, and silty sand. The lower peats were dated at >37,600 and >36,650 years (GSC-155, 191). These older sediments overlie some 10 feet of stony marine clay, which overlies a third till exposed near beach level. Pre-Quadra sediments containing organic matter also occur beneath till at Cordova Bay, Victoria. According to J. G. Fyles, these sediments contain a pollen assemblage that is believed to represent true interglacial conditions, perhaps the Sangamon Interval. Equivalent deposits that include peat and wood-bearing strata occur in coastal exposures west of Sooke in Juan de Fuca Strait and have been dated at >40,300 years (GSC-358).

Both Quadra and pre-Quadra organic-bearing sediments have been noted also by J. E. Armstrong near Lynn Creek in Fraser Lowland. Quadra sediments, dated 36,200±500 years (GSC-92-2), overlie a till that rests on an older sand, gravel, and peat unit from which wood has been dated >52,300 years (GSC-555). These older sediments overlie an older basal till.

The best-documented and most complete Pleistocene section in Fraser Lowland, if not on the whole west coast, is that at Point Grey, Vancouver (Table XII-2), where some 130 boreholes and a long tunnel have added immeasurably to data available from study of the seacliffs.

Quadra and pre-Quadra sediments occur respectively also in Coquitlam and Surrey municipalities east and southeast of Vancouver.

TABLE XII-2 *Composite section of Pleistocene deposits, Fraser Lowland (by J. E. Armstrong)*

	(Max. known thickness, feet)
Stratified sediments (postglacial)—stream and marine	70
Till and associated sediments (Fraser Glaciation)—glacial, glaciofluvial, glaciolacustrine, glaciomarine	215
Non-glacial (Quadra) sediments (Olympia Interglaciation)—swamp, flood plain, channel and estuarine; peat in basal part dated >36,800 years (GSC-81)	200
Till and associated sediments (Semiamu Glaciation)—glacial, glaciomarine, marine	130
Non-glacial sediments (interglaciation?)—swamp, lacustrine, flood plain, channel, marine; includes some peat	155
Till and stratified deposits (glaciation)—glacial, glaciomarine; includes shells	15
Non-glacial sediments (intertill) channel and flood plain	15+
Till?	—
Bedrock (Eocene)	—

Southern interior, British Columbia. Fulton (1965) cites two distinct intertill sequences in the valley of Thompson River and its tributaries. The lower consists of oxidized sand, silt, and gravel containing volcanic ash, together with wood and freshwater shells. Shells from near Merritt gave a dating of >37,200 years (GSC-258). Wood and shells from Kamloops gave ages of >32,700 and >35,500 years (GSC-275, 413). The upper sequence, also sand, silt, and gravel but unoxidized, is correlated with deposits that at Salmon Arm, Shuswap Lake, contain wood dated at 20,230±120 years (GSC-194). An erosional unconformity separates the two sequences. Till and glaciolacustrine deposits, locally exposed beneath the older oxidized sediments, indicate that at least one glaciation preceded the older non-glacial interval. The intertill sediments were deposited prior to the Fraser Glaciation during a non-glacial interval which comprised at least two periods of aggradation separated by an interval of oxidation and soil formation.

Fulton (1968) indicated rather similar geological events in Purcell Trench, north of Kootenay Lake. He has demonstrated the occurrence of two glacial sequences separated by non-glacial sediments that contain organic matter and a paleosol. An unconformity separates the older till and varied associated deposits from the younger sediments. The paleosol, which includes A, B, and C horizons, is developed on some of the older materials. A finite age of 41,900±600 years (GSC-733) was obtained on roots embedded in the A horizon. The younger non-glacial sediments represent deposits of an aggrading flood plain. Two radiocarbon dates on materials successively nearer the top of the flood plain unit were 43,800±800 and 42,300±700 years (GSC-740, 720). An age of 41,800±600 years (GSC-716), furthermore, was obtained from a stump growing on a slope facies of the paleosol but embedded in the younger gravels. Silty sediments that intertongue with the gravel yielded wood that gave dates of 33,700±330 and 32,710±800 (GSC-542, 493) from successively higher positions. A coarse gravel overlying the silty sediments contained wood in its upper part that dated 25,840±320 (GSC-715). The gravel is capped by till. Fulton relates the upper till to the Fraser Glaciation which, judging by dates from Kamloops region, did not reach the Kootenay area until after about 20,000 years B.P. He relates the non-glacial sediments to the Olympia Interglaciation of the west coast. The paleosol was developed following the older glaciation and was successively buried by the interglacial sediments until perhaps 41,800 years ago.

Central and northern interior, British Columbia. In central and northern British Columbia interglacial sediments have been reported but their interpretation is doubtful. In places non-glacial intervals are indicated by buried river channels cut into older tills or bedrock and by the occurrence of buried placer deposits. Lignite or peat occurs in gravel beneath till along Stikine and Tuya Rivers, but the origin of the deposits is unknown. The indicated non-glacial intervals must predate the last major glaciation but whether earlier Wisconsin, pre-Wisconsin, or pre-Pleistocene has not been determined. In many parts of central British Columbia the Tertiary deposits are unconsolidated and not readily distinguishable from Pleistocene sediments. The occurrence of lava flows or sills in unconsolidated sediments is more characteristic of the Tertiary, but as lavas overlie till sheets in some places, volcanic activity must have continued into the Pleistocene. Lava flows associated with the tills or other Pleistocene sediments may serve as a means of dating some events and provide information on the older Pleistocene.

Interior Yukon Territory. Within the glaciated part of the Cordilleran Region in the Yukon few occurrences of buried organic materials have been reported, but undoubtedly many more will be found as the region is studied.

In Kluane Lake area on the northern side of St. Elias Mountains organic materials occur in drift that predates the last major glaciation, the Kluane, and, on the basis of geomorphological and stratigraphic evidence, indicate three separate non-glacial intervals (Denton and Stuiver, 1967). Organic debris in outwash related to the oldest recognized glaciation, the Shakwak, and organic debris from a silt bed beneath ice-contact stratified drift of the younger Icefield Glaciation, and peat from sinuous stringers contained in the Icefield till itself, proved too old for radiocarbon analyses (Y-1355, 1481, 1486). Organic debris in outwash overlying the Icefield till, however, yielded three finite dates ranging from 30,100 to 37,700 years (Y-1356, 1385, 1488). These deposits are overlain by the Kluane till.

The finite datings appear to represent a non-glacial interval, the Boutellier, of about 10,000 years duration which may be the equivalent of the mid-Wisconsin interval of the mid-continent. The Silver non-glacial interval between the Icefield and Shakwak Glaciations, from which the older age-datings were obtained, was much longer than the Boutellier interval judging by differences in the depth of oxidation of the underlying materials, but neither the duration nor the climate of the interval is known. It may be either early Wisconsin or Sangamon; if the latter, then the Shakwak Glaciation and its outwash deposits may be Illinoian.

Near Fort Selkirk on Yukon River a Pleistocene lava with a striated surface, overlies gravel, sand, and silt that in turn overlies glacial till. Charred wood found 4 feet from the top of the bedded unit was dated at >38,000 years B.P. (I-GSC-27). Another record of forest cover predating the last glaciation was found near Mayo on Stewart River. An abraded log was found in the base of a till lens occurring near the top of a 100-foot section of sand and gravel, the whole overlain by bedded silt. The till lens at this site is considered to be the same as the upper till at numerous sites along the river nearby, and to represent the glacier which apparently terminated some 8 miles below Mayo Landing. The wood was dated at >35,000 years (I-GSC-180). Nearby, wood from a silt, sand, and minor gravel unit overlain by about 10 feet of till and 30 to 50 feet of thinly bedded silt and sand, was dated at >46,580 years (GSC-331). Also on Stewart River, below McQuesten, old wood was found in an ash lens beneath 10 to 15 feet of organic silt (GSC-342).

Further interesting evidence of former forest cover is the report of wood from a 200-foot cut-bank on Liard River. A thin soil zone, including a volcanic ash horizon, and about 100 feet of brown till overlies about 100 feet of crossbedded sand. Wood from 30 feet above river level was dated >40,100 years (GSC-412).

CLASSICAL WISCONSIN AND POSTGLACIAL EVENTS

The last continental ice sheet had three main component parts, the Laurentide Ice Sheet, the Cordilleran Glacier Complex, and the Queen Elizabeth Islands Glacier Complex.[1] The Laurentide Ice Sheet is commonly regarded, on the basis of ice-flow patterns, as comprising the Labrador and Keewatin sectors which, although confluent at the Wisconsin glacial maximum, became discrete areas of outflow during deglaciation. The former retreated to one or more centres or ice divides in northern Quebec and Labrador, and the latter to an ice divide west of

[1] The term 'glacier complex' is used in the broad sense to include ice sheet, ice cap, piedmont glacier, and valley glacier.

Hudson Bay. These major areas of late ice flow themselves split up into smaller short-lived units of outward-flowing ice prior to the final dissipation of the main remnants. Deglaciation also resulted in other major components of Laurentide Ice Sheet. An ice sheet remained and was nurtured locally in Foxe Basin and, later, on Baffin Island. This ice sheet, the Foxe–Baffin Glacier Complex, was independent of the Keewatin and Labrador sectors at least in late Wisconsin time, and parts remain today. Also, early in the deglacial process, major independent ice caps formed in the Appalachian regions, for instance, in Newfoundland, and in part may have been independent of the Labrador sector throughout most of the Wisconsin. The above ice masses and stages in their deglaciation are shown in Figure 15.

The last continental ice sheet is described according to its major component parts regardless of the degree of interdependence experienced during their build-up, at their maximum, or during decay. These are: Appalachian Glacier Complex, Labrador sector of Laurentide Ice Sheet, Keewatin sector of Laurentide Ice Sheet, Foxe–Baffin Glacier Complex, Cordilleran Glacier Complex, and Queen Elizabeth Islands Glacier Complex. Due to the size of the areas within the zone of influence of the major glacier units chosen, it may be necessary to discuss some matters according to smaller areas or to particular glacial or deglacial events. As the ice sheets and glaciers waned and the land surface was uncovered, the forms implanted by the ice were left exposed or were covered by glacial debris washed out from the receding ice margins on land, in lakes, or in the sea. The complex record of glacial and postglacial features and deposits remaining today afford a means of tracing the paths of the receding ice margins and interpreting the deglacial and postglacial events.

Appalachian Glacier Complex

The Appalachian Glacier Complex occupied most if not all of the Appalachian Region in Canada. This includes the island of Newfoundland, the Maritime Provinces, and the Appalachian Mountains of southeastern Quebec. The build-up of glaciers in this region prior to the Wisconsin maximum is little known. The physiographic province is, in large part, a highland region of abundant snowfall, which under the Wisconsin climatic conditions may have developed a number of independent ice caps early in the glacierization process. Under the influence of prevailing northeasterly moving moisture-laden winds from the continental interior and from surrounding bodies of water, these ice caps may have spread radially by differential accretion and preferentially towards the southwest. There are many records of glacial striae and grooves, and of rock dispersal, that trend at right angles to what is commonly considered the ice-flow trend at the Wisconsin maximum and during later retreat. The Appalachian Region may well have been largely ice-covered before confluence with Labrador ice was established along the St. Lawrence River and Strait of Belle Isle. Early ice movements are thus postulated as local and independent of the outward growth of the Labrador sector of Laurentide Ice Sheet. It is probable that the Appalachian Glacier Complex was comprised of two main parts, one over Newfoundland and the other over the Maritime Provinces; both were confluent with the Labrador sector of Laurentide Ice Sheet for some time during the Wisconsin but confluence with each other along Laurentian Channel, through Cabot Strait, was short-lived at the Wisconsin glacial maximum.

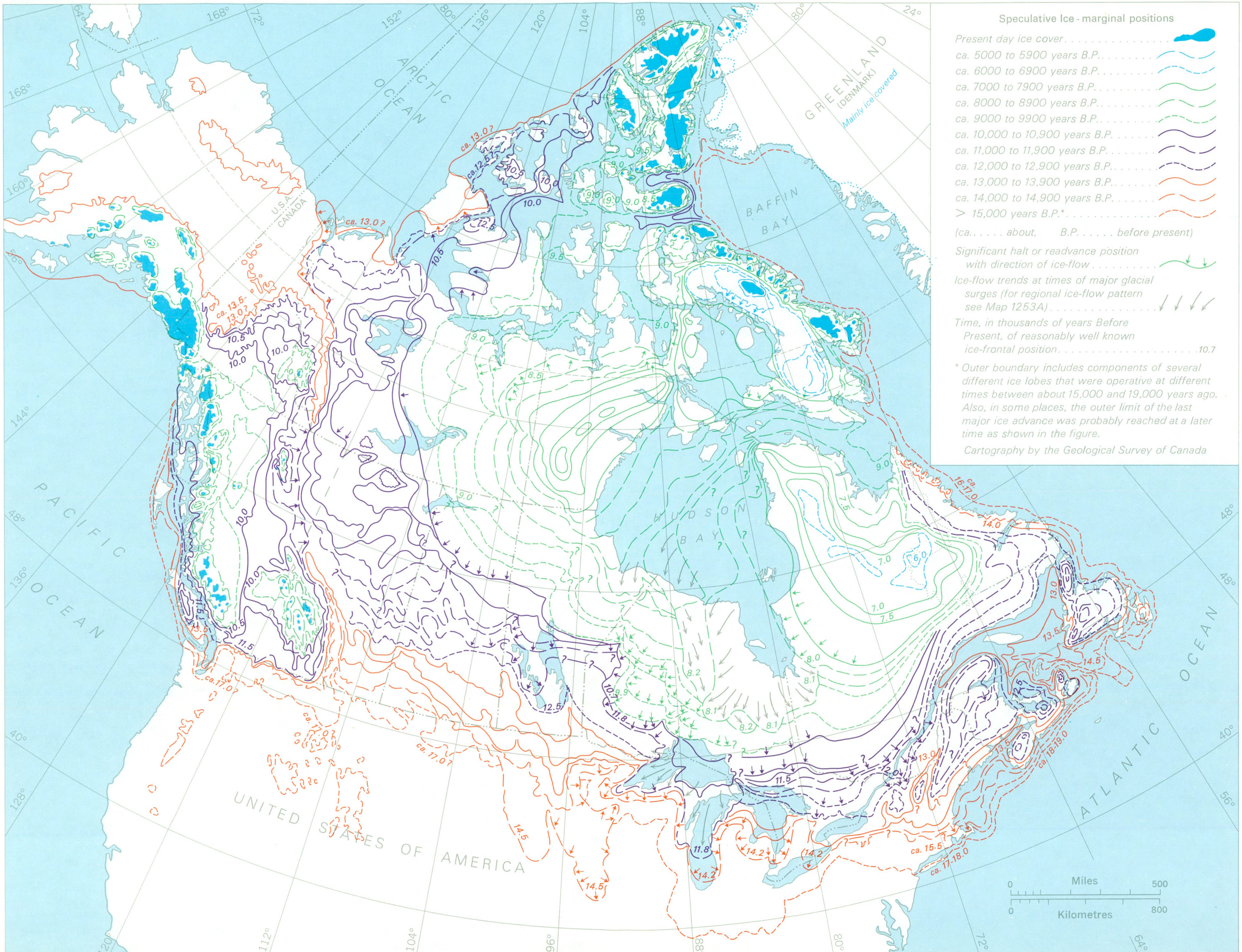
The pattern of deglaciation as indicated by ice-flow features of all types (Map 1253A) gives little evidence of general ice-frontal retreat towards Quebec and Labrador. The influence of the maritime environment is obvious. Only on the mainland of Nova Scotia is there a regional southeasterly trend to suggest glacier movements from, or retreat towards, a Labrador centre, but this trend could equally well indicate a centre of outflow from New Brunswick or Maine. Though Labrador erratics have been reported on the Gaspé Highlands, a significant overriding of the Gaspé and New Brunswick Highlands by Labrador ice of Wisconsin age has not been proven.

Newfoundland

Glacial events. On Avalon Peninsula, Henderson (1960) found no evidence of a general ice invasion from the west and concluded that the peninsula had its own ice cap from which active glaciers flowed northward and southward down the channels now occupied by Trinity, Conception, St. Mary's, and Placentia Bays, and thence onto the Grand Banks. A few cobbles from the main part of the island occur along the east side of Trinity Bay at about 20-to-25-foot elevation and hence close to the marine limit; this suggests that the main island ice remained active after considerable recession of the Avalon ice cap and as the sea encroached on the present shores of Avalon Peninsula.

Henderson believes that the last active glacier on Avalon Peninsula lay in St. Mary's Bay and moved debris into the central part of the peninsula as it fanned out towards Trinity and Conception Bays; free flow southward along St. Mary's Bay was impeded by a 'baymouth' threshold. Transverse lineaments in the morainal materials of central Avalon are evident on airphotos, although such patterns are not invariably convex in the direction of flow. Henderson believes that the glacier down-wasted in St. Mary's Bay with marginal retreat taking place on all sides.

MacClintock and Twenhofel (1940, p. 1731) concluded that the whole island was glaciated during the Wisconsin and that " . . . ice spread as a complete cap from the Long Range Plateau, the Central Plateau and the Avalon Peninsula outward in all directions, to beyond the present shore lines of the island." Their studies gave no indication of pre-Wisconsin events as earlier postulated by Coleman, but rather served to establish a gross chronology for the Wisconsin, based mainly on observations in St. George's Bay area in southwestern Newfoundland.



GSC 1967

FIGURE XII-15. Stages in the deglaciation of Wisconsin ice.

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The oldest drift—"St. George's River Drift"—includes till, ice-contact gravel, and marine silt, all of which display a topography suggestive of ice blocks surrounded or buried by marine beds. Following a significant ice-frontal retreat from the shores the "Bay St. George Delta" was built, complete with marine fossils like those of the present day. With re-advance the ice overrode the delta, deposited an upper till, and built an end moraine system near the coast; these deposits comprise the "Robinsons Head Drift." A major retreat of the ice towards the central plateau areas followed, but a still-stand produced the "Kittys Brook" moraine system in the valleys. Further recession resulted in the higher knobs of the plateau protruding above the ice sheet and becoming over-steepened by the outward-moving ice. Still later the ice occurred only as local valley glaciers on the steep sides of the plateaux. Some glaciers deposited small moraines and formed cirques, many of which still contain perennial snow and ice. The work of K. Widmer in the Hermitage Bay area, on south coast of Newfoundland, shows the same close association between 'late' ice and marine overlap.

On the main part of the island there was strong outward ice flow (Jenness, 1960). Fiords are particularly well developed along the south and northeast coasts. There are large areas of fluted and drumlinized terrain, and of ribbed moraine (Pl. XII-7). The lineations, together with recorded striae, grooves, and boulder train data, suggest a very complex ice-flow pattern with erratic, shifting centres of active flow during the waning of the island ice cap. The last active ice caps appear to have been on Newfoundland Highland.

The matter of Labrador ice occupying much of Newfoundland at the Wisconsin maximum, as postulated by Flint (1940) on the basis of increasing height of raised strandlines northward along the west coast, is open to question. The writer believes that Newfoundland maintained its own active ice cap throughout Classical Wisconsin time though it was confluent with Labrador ice in Gulf of St. Lawrence during most of this period. MacClintock and Twenhofel (1940) reported that overriding Labrador ice was possible but there was no real evidence, and concluded that the problem would remain unresolved until the transportation of drift boulders was studied in terms of regional geology of mainland and island alike. Cooper (1937), however, actually records mainland erratics dispersed over the highest part of the northern end of Great Northern Peninsula and associated with southeast-trending striae. Such glaciation by Labrador ice is to be expected in such close proximity to the mainland. On aerial photographs a system of De Geer moraines along Strait of Belle Isle is evident, which indicates late southwestward flow along the strait into Gulf of St. Lawrence,

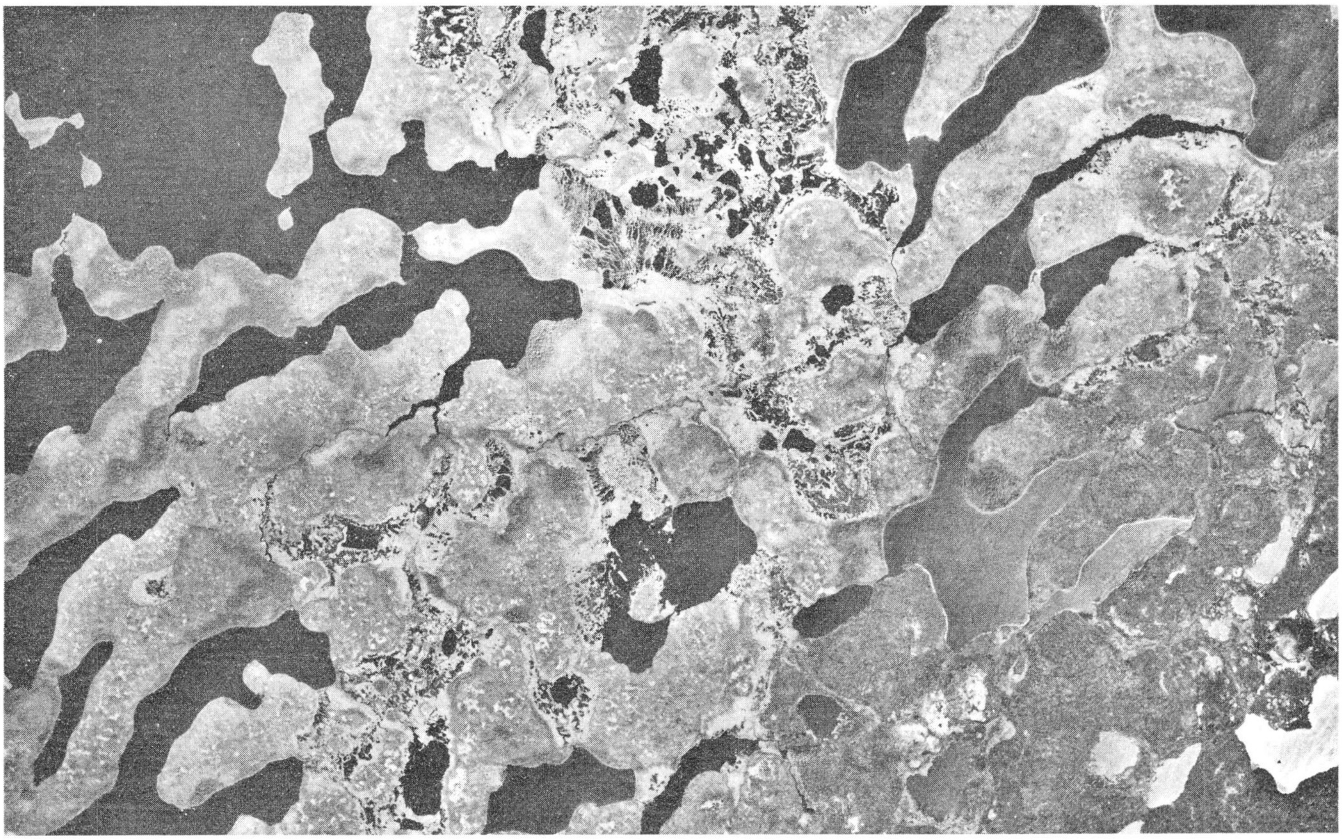


PLATE XII-7. Ribbed moraine near Meca Pond in eastern Newfoundland. Vertical airphoto. The rib ridges show no evidence of ice flow. Scale 1 inch to 1,320 feet.

probably of confluent Newfoundland and Labrador ice. As marine waters extended northeastward up the deep channel between Quebec and Newfoundland, and thence into the shallow Strait of Belle Isle, Newfoundland ice would be separated from Labrador ice. At about this stage of deglaciation, about 12,000 years ago, Mecatina Plateau northwest of Strait of Belle Isle was effectively diverting Labrador ice towards the northeast, away from the Strait of Belle Isle.

On northern Newfoundland variations in the trend of De Geer moraines suggest a short-lived active centre of outflow from the area of former Newfoundland and Labrador ice confluence at the extreme northern end of the peninsula. Late ice flow from the Newfoundland Highland changed from southwestward to northwestward as the centre of outflow shifted back to the Long Range highland and a system of lobate end moraines was formed in the vicinity of Ten Mile Lake.

The writer's observations in the central-west coast region support the concept of an active Newfoundland ice cap in late Wisconsin time. There was strong ice flow from the interior both along major valleys parallel the bedrock structures and also across rugged terrain towards the west coast.

Marine events. The interpretation of data on maximum marine overlap is complicated by the fact that 'late' ice has prevented the registration of the highest shorelines in some areas, as in southern, southwestern, and probably also northeastern coastal Newfoundland. This phenomenon implies that local as well as regional postglacial rebound is reflected in the resultant picture of marine overlap.

On Avalon Peninsula, Henderson (1960) found evidence of late-glacial marine overlap only along the east coast of Trinity Bay and on Avalon Isthmus. The isobase of zero uplift may trend northeastward along the east side of Placentia Bay, cross the isthmus at Norman's Cove, and thence pass just inland from the east coast of Trinity Bay. Henderson found that the elevation of raised marine features increased northwestward along the isthmus and also northward along the east shore of the island proper to Bonavista Bay and beyond. The 100-foot isobase passes along the southeast shore of Burin Peninsula and along the axis of Bonavista Bay. The isobase data may reflect an early period of marine overlap broadly concentric about the eastern side of Newfoundland.

The highest shoreline features decrease in elevation westward along the south coast. K. Widmer records wave-cut benches on islands at the mouth of Fortune Bay and on nearby mainland points at an elevation of 100 to 110 feet. The marine limit is about 70 feet at head of Fortune Bay and 30 feet near Burgeo. Between Burgeo and Port aux Basques raised shore features are absent. Similarly, along the southeastern side of St. George's Bay, on the west coast of Newfoundland, there is evidence of late ice advances into the sea that may have prevented the registration of high shoreline features (Flint, 1940; MacClintock and Twenhofel, 1940). Flint has reported, nevertheless, that the level of maximum marine overlap increases from zero at the southwest end of Newfoundland to 100 feet in head end of St. George's Bay, and to 200 feet in Bonne Bay, with isobases trending N70°E. At a later time, during a pause in the uplift, the sea carved benches in bedrock along the west coast and on the north coast, termed the "Bay of Islands surface." The trend of the isobases on this surface is N80°E, the zero isobase lying along the north shore of St. George's Bay, the 100-foot isobase passing through Bonne Bay, and the 250-foot isobase passing through Hare Bay. Study of aerial photographs of this area reveals another prominent bench near sea level, possibly that recorded by Cooper (1937) as a wave-cut bench at 30 feet elevation.

On the north coast raised marine features exhibit three still-stands, two of which may correlate with those recorded on the west coast. The third set may reflect the early period of marine overlap recorded along the eastern shores of the island. Marine shells from Baie Verte, considered to be related to a former sea level at 180 feet elevation, gave dates ranging from 11,520 to 11,950 years B.P. (GSC-55, 75, 87). As there is evidence of marine features and deposits well over 200 feet and perhaps up to 250 feet, it is evident that deglaciation of the north coast had begun well before 12,000 years ago. Marine shells from the west coast of Newfoundland have also been radiocarbon dated; these indicate marine invasion into St. George's Bay prior to 13,420 ± 190 (GSC-598), and into Bay of Islands before 12,600 ± 170 years B.P. (GSC-868).

Cape Breton Island, Nova Scotia

It has been presumed that the Laurentide ice invaded the Gulf of St. Lawrence and passed southeastward across Cape Breton (Goldthwait, J. W., 1924). Glacier ice certainly flowed southeastward through Cabot Strait, scouring St. Paul Island and the northern tip of Cape Breton and also passed through the Strait of Canso, but elsewhere the regional ice-flow trends are poorly preserved and little understood. The writer believes that Cape Breton Highlands were not glaciated by Labrador ice but rather by a local ice cap, and that outward ice flow was not extensive.

There is evidence, however, of a glacier in Northumberland Strait on the west side of Cape Breton Island. Valleys between 200 and 250 feet elevation appear to have been filled, as if by glacial imponding, and later terraced by stream action. E. H. Muller records a river section a mile north of Southwest Margaree where 8 to 10 feet of gravel overlies about 20 feet of red lake clays, 10 feet reddish gravel, and 5 feet of blue cohesive till. It may well be that the glacier that imponded lakes in valleys along the west coast of Cape Breton was the one in Northumberland Strait, which may represent confluent ice from the highlands of Nova Scotia and New Brunswick from the south and west.

Goldthwait (1924) reported northeast-trending striae over eastern Cape Breton and refuted an older hypothesis that this ice flow was from Newfoundland. He noted ice-flow features and drift 'tails' that unmistakably proved a northeast to northward ice flow but he was concerned as to the source area. Though the more northward-trending striae on the east coast require an ice flow from the continental shelf towards the present island he suggested a possible source area south of Strait of Canso. Lacking information on such a centre, however, and with knowledge of east and east-northeast striae along Northumberland Strait he favoured a flow from New Brunswick with, presumably, a radial expansion as the ice emerged from Strait of Canso and Chedabucto Bay. This event preceded his Acadian Bay ice lobe which he considered the main Wisconsin Glaciation. The late H. L. Cameron, however, indicated south-southwest-trending striae east of Bras d'Or Lake and an end moraine system thought to be formed by this glaciation. Some indications of former southwestward ice flow are also evident on aerial photographs (Map 1253A). The anomalous southwest ice-flow trend of eastern Cape Breton appears to be developed on a broad morainal tract. D. R. Grant (GSC) considers that this trend represents a younger ice flow that oriented the drift and lightly striated the bedrock. On the southeast side of many shore outcrops he found evidence of intense scouring action by glaciers directed towards the northeast. The older striae, grooves, and *roches moutonnées* and the trough-like topography of Chedabucto Bay suggest a funnelling of ice down the trough with northward splaying or lobing along eastern Cape Breton. The younger glacier movements appear to have been southeastward through Strait of Canso and northeastward along Bras d'Or Lake basin, and also southward and southeastward out of the lake basin towards the sea. Neither the ice-flow trends in eastern Cape Breton, their indicated source areas, nor the indicated order of glacial events are in accord with the concept of the Labrador ice overriding the island and extending onto the continental shelf.

Information on marine overlap in Cape Breton is limited. A prominent subhorizontal feature at about 200 to 250 feet is present over long stretches of both eastern and western shores. H. L. Cameron reported limited marine overlap in Aspy Bay on northern Cape Breton, and along the southeast coast, but also proglacial lakes in several valleys in-filled to about 250 feet. In view of the semi-coincidence in levels some of the raised coastal features may be ice-contact phenomena, or else both valley-fills and coastal terraces are graded to a common sea-level stand, possibly of pre-Wisconsin age. The writer has not seen any evidence of marine overlap on Cape Breton Island.

Mainland Nova Scotia

Glacial events. The pattern of ice-flow features in Nova Scotia south of the Cobequid Mountains is orderly; at the Wisconsin maximum the ice moved southeasterly across the region. This ice-flow direction is exhibited in many places by drumlins composed of fine-textured till. In areas underlain by granite or quartzite the till is silty to sandy and light coloured. In a zone along LaHave River the till is olive-grey and derived from local slates. Elsewhere to the northeast the drumlins are composed of an anomalous red till that overlies locally derived tills, and is believed by Grant to have been derived largely from Pleistocene sediments scoured from Bay of Fundy, Minas Basin, and Northumberland Strait. L. H. King (The Bedford Institute) located off-shore moraines that may represent the Wisconsin maximum or a somewhat later stage of retreat. He believes that the off-shore moraines probably formed at a time of low sea level when relatively thin ice calved into marine embayments.

As the Wisconsin glacier waned the ice front along the Atlantic coast retreated inland to the northwest. Absence of marine overlap along this coast suggests a relatively thin ice sheet near its terminus. As deglaciation proceeded marine waters invaded the Bay of Fundy giving rise to a concentric pattern of retreat towards the Nova Scotia Uplands. Late stage upland centres of outflow were active. Drumlins have been re-oriented in some areas and eskers appear at variance with earlier ice-flow features. Late ice in the central upland region remained active or was reactivated so that it moved granitic drift northward across Cornwallis–Annapolis Valley and deposited it on the Triassic trap rocks of North Mountain, and along the Fundy shore (MacNeill, 1953; Hickox, 1962). That the central Nova Scotia Uplands should retain an ice cap late in the period of deglaciation is not surprising in view of the maritime setting; it is at present an area of heavy snowfall. The precise age of this upland ice cap is not known and there are currently no dates on relevant materials, but it is known that the sea invaded the western side of Bay of Fundy prior to $13,325 \pm 500$ years B.P. (I-GSC-7).

The pattern of ice-flow features in northern Nova Scotia is highly irregular. The Cobequid Mountains and Antigonish Highlands appear to restrict the general southeast flow so typical of the region to the south and east. There was some glacier flow northward from these highlands towards Northumberland Strait. Glacial striae and boulder trains indicate northward movement in the vicinity of Pictou and New Glasgow. Cobequid-type igneous stones occur sporadically to the north of the mountains in the red sandy tills of the lowlands; they are especially noticeable near Pictou and on Pictou Island. Goldthwait (1924) believed the former were derived from the north flank of the mountains by eastward-moving glaciers from New Brunswick in advance of the Acadian lobe incursion from the north. Others contend that the last ice to invade the lowlands was from New Brunswick rather than from the north. The occurrence of Cobequid igneous stones in the red till of the Carboniferous Lowlands required some northward transport; this might have resulted from stream deposition prior to emplacement of the last red till sheet but later northward glacier movements are more probable as such stream transport is not effective with the present relief. Late active glaciers on the eastern Cobequid Mountains and Antigonish Highlands are indicated also by the occurrences of eskers and associated outwash, with north-dipping beds, as along River John near Scotsburn and Maryvale. In Barney's River area, at the east end of Merigomish Island, the outwash is graded to, or below, present sea level. Furthermore along the coast near Malignant Cove kame terraces, with northeast-dipping crossbeds, occur up to elevations of about 100 feet; these formed while the same late glacier occupied Northumberland Strait.

The western end of Cobequid Mountains does not appear to have harboured a late ice cap. In this region glacier retreat was northward and meltwater poured through gaps in the western Cobequid Mountains to deposit extensive valley trains with kettled deltas in Minas Basin. Small elongate moraine ridges on the north-central part of the Cobequids may be end moraines. The northeasterly trend of Joggins moraine (Wickenden, 1941), south and east of Amherst, has been assumed to indicate glacial retreat towards the northwest. However, both northwest and southeast of this moraine, fluting and striae trend south-southwest and the ice flow was undoubtedly southward; this ice-flow direction is the reverse of that farther east and was most likely formed at a somewhat earlier date as a calving bay developed in Bay of Fundy.

Marine events. Postglacial changes of level in Nova Scotia are complicated. The Fundy embayment is the only area affording evidence of former emergence but the whole province is now involved in submergence. In southwest Nova Scotia the isobases of differential uplift trend northeastward, roughly parallel to the Fundy shore, with the zero isobase at Yarmouth; uplift is 120 feet at Digby and 150 feet on Long Island, southwest of Digby. The isobase trend changes abruptly in Minas Basin. This may result from changes in amount and rate of uplift due to early opening of Bay of Fundy as a calving bay in the ice front, to presence of late ice north of Minas Basin, and possibly also to the structural influence of Cobequid fault. For these reasons, projection of the Fundy isobase data northward from Minas Basin may not be valid. Borns and Swift (1966) report glaciofluvial deposits overlying glaciomarine deltaic deposits; the top surface of the former slopes from approximate 140 feet elevation at Advocate Harbour to 60 feet near Truro whereas the surface of the lower unit slopes from elevation 130 feet (110 feet above high tide level) in the west to mean high tide at Five Islands, about halfway between these places. The writer, however, believes the upper unit is basically glaciomarine and that the relict marine surface slopes from 120 feet west of Parrsboro to 20 feet at east end of Cobequid Bay. He believes that 'late' ice prevented marine waters from entering the Truro lowland, and also prevented development of higher-level strandlines near Advocate Harbour. The tidal range in Minas Basin is today about 40 to 55 feet but it was undoubtedly different in the past; for this reason the marine limits are referred here to high tide.

There is no evidence of marine overlap along most of the Northumberland shore. A sea bench on the northeast-trending coast east of Arisaig records a former sea-level stand some 5 to 10 feet above the present, and comparable sea-level stands have been recorded elsewhere along this and the George Bay coast; all these are believed to be pre-Wisconsin. In the extreme northwestern part of the province, however, the zero isobase of postglacial uplift is believed to pass through Northport, with maximum marine overlap increasing westward to perhaps 50 feet at the provincial boundary. The evidence is, however, rather inconclusive.

Prince Edward Island

Evidence of glaciation in the western part of Prince Edward Island has long been recognized (Chalmers, 1895) by an abundance of igneous and other stones in the drift. These were derived from westerly or northwesterly sources, whereas the bedrock is red sandstone and shale, mainly of Permian age. The general absence of foreign stones in the drift of the central and eastern parts of the island led early workers to regard the drift mantle as a regolith; but undoubted glacial till is present in all parts of the island (Prest, 1962; Frankel, 1966). It has also been assumed that the last major glacier to override the whole island was from the north—the Acadian Bay lobe of Goldthwait (1924). The distribution of erratics along the north shore lends some support to this concept, but south-trending striae are generally lacking whereas east-trending striae occur along or near the coast. A few north-south trending striae have been recorded from central parts of the south coast but the sense of movement is unknown. These striae predate the last major glacier movements along Northumberland Strait. The general southeasterly trend of ice-flow features across most of the island, coupled with the prevalence of foreign stones in the western end, suggests glaciation from New Brunswick. The island was deglaciated, however, over a period of time during which significant lobing of the ice fronts occurred. On the basis of miniature crag-and-tail features on the south shore west of Borden it is known that there was late westward flow along Northumberland Strait. But numerous glacial striae along the south shore between Borden and Hillsborough Bay, however, trend northeast-southwest rather than along shore, and similar trends occur in the interior on the southwest side of the central higher parts of the island. This suggests that late ice flowed southwestward from the interior towards the Straits and thence westward beyond Borden. Also eastern and north–central parts of the island are characterized by an abundance of glaciofluvial deposits and an anastomosing system of eskers, whereas elsewhere these features are uncommon; this situation is thought to reflect the final decay of remnant ice on central and eastern Prince Edward Island.

Marine overlap in western Prince Edward Island reaches a maximum of 75 or 80 feet along the west coast. The marine limit in the northwestern part of Malpeque Bay is about 30 feet and in the southeastern corner about 10 feet. The zero isobase appears to be near Borden and probably trends southward across Northumberland Strait to Cape Tormentine, New Brunswick. There is no evidence of marine overlap in central or eastern Prince Edward Island. Shells from northwestern Prince Edward Island, believed related to a sea-level stand about 50 feet above the present, were dated at 12,410±170 years and 12,670±340 years (GSC-101, 160).

Magdalen Islands, Quebec

The Magdalen Islands in the central part of the Gulf of St. Lawrence are commonly regarded as lying squarely in the path of Laurentide Ice Sheet as it spread southward or southeastward towards the Atlantic. There is, however, no evidence of Wisconsin Glaciation above the limit of postglacial marine overlap, and below this limit there is only a till-like material that was deposited under water.

At Amherst Harbour, Goldthwait (1915) observed a 15-foot sandy deposit, resembling glacial till and containing only local rock materials overlying deeply decayed buff weathering, grey sandstone. As some of the stones had striations parallel to their long axes he concluded that a glacial origin was more valid than a marine drift origin. Coleman (1920) also noted the complete absence of glacial features on Amherst Island aside from the sandy drift, but he concluded that the thin margin of the continental ice sheet was afloat at a time of higher sea level.

Alcock (1941) reported boulder clay on Amherst and Entry Islands, ground moraine on Grindstone Island, an end moraine on Coffin Island, and large erratics at elevations over 200 feet on Grosse Isle. He saw no evidence of marine uplift, thus disagreeing with Chalmer's interpretation of beaches and terraces to about 115 feet and Coleman's marine overlap to about 200 feet. Prest (1957) considered the Coffin Island end moraine to be a kettled deposit of ice-contact stratified drift, or kame moraine, with the beds dipping inland to the west, and that its bouldery surface was the result of ice rafting during a period of marine overlap. The boulders are mainly foreign to the islands and are believed derived from a northern source. He confirmed the presence of marine gravels on Grindstone Island to a maximum elevation of 120 feet. Alcock considered the 'till' on Amherst Island to be a deposit let down gently from floating ice, an opinion shared by the writer in view of the contact relations between drift and bedrock.

The exact reason for the presence of striated local stones in the marine drift mantle of Amherst Island remains obscure; these stones were probably derived by glacier action some distance beyond the present shores and later ice rafted into their present position. Marine or lacustrine submergence of at least 120 feet elevation is indicative of ice near at hand; this is also indicated by the ice-contact stratified drift on Coffin Island. Glacier ice evidently reached the islands from the north, but only shelf ice reached the southern shores.

New Brunswick

Glacial events. The pattern of ice-flow features in New Brunswick clearly reveals a south to southeast trend in the western and southern parts, and an east to northeast (and/or southwest) trend in eastern parts. Granitic rocks from the Precambrian Shield occur in the Saint John River valley, on the adjoining Chaleur Uplands (Lee, 1955), and in an end moraine along the northern flank of the Chaleur Uplands. It is thus clear that Laurentide ice did reach the upland as well as flow strongly down the Saint John River valley, and probably also down the Matapedia Valley in Gaspé into Chaleur Bay, but elsewhere glaciation was relatively light. Alcock (1948) found evidence of strong glacial flow from the northern end of New Brunswick Highland northeast towards Chaleur Bay and east towards Gulf of St. Lawrence, and concluded that such movements must have preceded as well as followed the arrival of Laurentide ice in New Brunswick. He assumed that Laurentide ice over-topped the highlands on the basis of south-trending striae in southern New Brunswick and particularly in the Moncton area, but striae trends in this area are now known to be diverse. It is more likely, judging by ice-flow patterns and an end moraine (Map 1253A), that ice flowed mainly around the highlands rather than over them. Flint (1951) suggested that the ice responsible for the northeast- and east-trending striae was of local derivation and subsequent to the main Laurentide glaciation. South of Bathurst, the ice-flow features visible on airphotos, suggest that some late ice flow was out of Chaleur Bay and towards the south-southwest but this has not been confirmed. Farther south, between Newcastle and Moncton, ice-flow features and an east-west end moraine also suggest a southerly ice movement as if from Miramichi Bay. On the other hand, but a short distance southeast of the ice-flow features and closer to Moncton, miniature crag-and-tail features observed on a pebbly sandstone indicate ice flow to the northeast parallel to the regional drainage.

On both the mainland and islands in the extreme southwest corner of New Brunswick, Alcock noted *roches moutonnées* and striations trending southeast to east, in contrast to the general south-southeast trend typical of the western part of the province. He concluded that there had been an ice movement from the mountains of Maine, presumably prior to the main Laurentide glaciation.

Marine events. Chalmers (1890) noted marine terraces along the Fundy shore both southwest and northeast of Saint John. In the former area the deposits had a maximum altitude of 225 feet and contained shells; in the latter area the highest terrace deposits were only 125 feet and no shells were seen. In Saint John River valley at Fredericton, Lee (1959) found estuarine deposits at 125 feet. It is possible that the lower part of Saint John River valley and the New Brunswick side of Bay of Fundy were covered by late ice while the high-level terraces were being formed southwest of Saint John. Shells from the seacliff 5 miles west of Saint John are dated at 13,325 ± 500 years (I-GSC-7).

Along the east coast of New Brunswick large tracts of low-lying ground formerly covered by the sea show little or no evidence of marine action; there were apparently no significant halts as the land rose from the sea and, in general, marine sediments appear to have been removed by erosion or incorporated into the soil profile. Local areas of marine sand, gravel, and poorly washed sediments, and the disposition of ice-rafted boulders do provide some evidence of former marine action. The marine limit varies from zero at the east end of Cape Tormentine to 100 feet at Moncton. It is 150 feet about 15 miles west of Richibucto and 225 feet at Newcastle and Bathurst. Farther west along Chaleur Bay the marine limit is not evident, probably as a result of late ice in Chaleur Bay during the period of maximum overlap along the east coast. The isobases of differential uplift trend northeast along the Bay of Fundy shore, northward through Moncton area, and northeastward again along the east shore of the province; this gives them an open S-shape. The effects of late ice and the possible influence of Cobequid fault in the amount and rate of uplift on either side of Bay of Fundy, together with the lack of age determinations, preclude correlation across the bay. The uplift in western Prince Edward Island, however, appears in accord with data from eastern New Brunswick.

Gaspé Peninsula, Quebec

Glacial events. Much has been written concerning glaciation of Gaspé Peninsula; this is succinctly summarized by McGerrigle (1952). Cirque and local ice-cap development preceded Laurentide glaciation of Gaspé Peninsula. Shield erratics have been found on the highest parts of the peninsula. There is no proof, however, that the erratics were emplaced during the last major glaciation. McGerrigle notes the general sparseness of erratics in eastern Gaspé, and the seeming absence of anorthosite in a 50-mile-wide belt along the central part of the peninsula. It is noteworthy that neither McGerrigle nor Brummer (1958) reports Canadian Shield stones in the Béland–Upper York River highland area though they occur in lower areas farther east. It is perhaps noteworthy that granite gneiss erratics in southeastern Gaspé are only recorded along or close to the shore highway and their mode of emplacement is uncertain. The Matapedia Valley was the main ice-flow route into Chaleur Bay.

A highland ice-cap phase followed the maximum phase of Laurentide glaciation and, as the ice receded from the north side of Gaspé, this ice cap remained active and transported boulders northward as far as St. Lawrence River (McGerrigle, 1952). On the eastern end of the Shickshock Mountains, erratics derived from lower areas some miles south of the mountains indicate that the highland ice cap once had a centre of outflow lying south of the highlands proper. As the ice cap waned it broke up into minor ice caps that gave rise to radial flow as from the Tabletops and Béland–Upper York River highlands. The local ice-cap phase of glaciation finally gave way to cirque and valley glaciation.

Marine events. The extent of marine overlap along the southern and eastern coasts of Gaspé is not precisely known. Marine sediments are reported at 180 feet on the north side of Chaleur Bay and at 224 feet on the eastern end of the peninsula. The writer observed gravel deposits up to a maximum of 180 feet near Grand Cascapedia, but regarded these as outwash deposits graded to a wave-base perhaps as low as 120 feet above the present sea level. The steep outcrops of the eastern end of the peninsula do not lend themselves to the preservation of marine strandlines. A marine terrace at Prével has an elevation of only 90 feet and at Cap-des-Rosiers Est the upper limit of near-shore sediments is at 75 feet. At present there is evidence of coastal 'drowning' at the head end of Gaspé Bay. Along the western part of the north shore marine deposits have been recorded up to 300 feet, but 4 miles east of Mont Louis the writer did not see any evidence of marine overlap above the surface of a small delta at an altitude of 95 feet.

Appalachian Mountains of Quebec

Flint (1951) has summarized the voluminous literature pertaining to an Appalachian centre of glacial outflow (that either preceded or followed glaciation) from the highlands of Maine, New Hampshire, Vermont, and adjoining Quebec. In the last area the problem concerns the evidence of northward flow towards the St. Lawrence from Notre Dame and Green Mountains. Flint concluded that ice has flowed northward, at least locally. Recent studies by Gadd (1966), however, do not support the concept of glacial transport to the north. It may be that glaciation of the western part of the Appalachian Highlands, south of Quebec City, was unlike that in the east where late northward flow is evident. Lee (1962) reported striae and crag-and-tail indicative of northward ice flow in the Rivière du Loup–Trois Pistoles area. These were formed when a calving basin developed in St. Lawrence Valley while ice remained over the drainage divide in Notre Dame Mountains.

Labrador Sector, Laurentide Ice Sheet

Little is known about the initial development of the Labrador sector of the Laurentide Ice Sheet or the course of glacier movements that led to the ice cover of the entire region between the Labrador coast and western

Manitoba, and possibly beyond. It is probable that in response to a change in climate, the Lake Plateau of central Quebec–Labrador developed an ice cap by "instantaneous glacierization" (Ives, 1957), followed by radial outflow. The early influence of an open Hudson Bay in the development and differential expansion of the ice cap towards the west was favoured by Hare (1951) and Derbyshire (1962) but not, on climatological grounds, by Barry (1960). The prevailing southwesterly winds from the mid-continent no doubt contributed to southwestward expansion of the ice sheet towards Ontario and the Great Lakes region. There were probably local centres of excessive snow accumulation in the marginal zone of the expanding ice sheet; these may account for glacier lobations and resulting striae which in many areas diverge markedly from the general, deglacial, trend.

Early in Classical Wisconsin time the glaciers flowed off the Laurentian Highlands into St. Lawrence Valley and, upon filling the valley, sought escape northeastward towards the Gulf of St. Lawrence and southwestward towards the Ontario basin. Gadd (1966) reports striae indicative of southwesterly ice movements from the north flank of the western end of Notre Dame Mountains south of Quebec City, which he believes were formed during an overriding phase of glaciation. Strong eastward ice flow was long-continued down Saguenay River and thence northeast down St. Lawrence River as indicated by ice-flow features on the land and a very deep channel in St. Lawrence River beginning at the mouth of the Saguenay. The channel extends east-southeastward across Gulf of St. Lawrence through Cabot Strait and thence to the edge of the continental shelf, and is known as Laurentian Channel. Ice-flow features give some indication that the ice-flow divide in St. Lawrence Valley may have migrated westward from well east of Saguenay River to somewhat west of the Saguenay during Wisconsin time. Gadd suggests also that during the Wisconsin there was a shift of the centre of outflow in the Laurentian Highlands towards the west and that by the time the glacier was waning in St. Lawrence Valley the ice-flow direction was southeastward across the valley. He reports evidence of strong ice flow south–southeastward up the Chaudière River valley at this later stage.

The Laurentide ice upon filling St. Lawrence Valley and flowing along it sought escape southward through cols in the eastern part of Notre Dame Mountains and hence combined with elements of the Appalachian Glacier Complex; ultimately, as earlier mentioned, it may have overtopped the highest mountains in Gaspé, but its southward extent was perhaps more limited than has been generally recognized.

Study of the mineralogy of tills of southwestern Ontario led Dreimanis, *et al.* (1957) to conclude that glacier movements into Ontario and Erie basins were first from the north-northeast and only later from the east along the axis of the lake basin. The seeming reversal of the order of major periods of ice flow between Erie basin and Trois-Rivières region may be a result of their wide geographic separation and also to the different parts of Wisconsin time being considered. The occurrence of cobbles and boulders of Gowganda conglomerate in southwestern Ohio also indicates strong southerly movements from the Huron basin prior to the incursion of the Erie ice lobe. Also, in northwestern Ontario the writer found pebbles and cobbles of oölitic jasper, derived from the Sutton Mountains–Nowashe Lake belt west of James Bay, or from the Belcher Islands on the east side of Hudson Bay, and of a characteristic greywacke reportedly derived from Cape Jones at northeast end of James Bay, perhaps as much as 600 miles from their outcrop area. These indicate an early southwesterly ice-flow direction that is somewhat divergent from the deglacial flow pattern.

Information on the retreat of the Labrador sector of Laurentide Ice Sheet is naturally far more abundant than on the advance or build-up stages. The present surface bears a record of the last glacial events, modified only in part by postglacial changes, whether erosional or depositional. After the Wisconsin glacial maximum it is probable that there was thinning of the ice sheet over a broad marginal zone as the ice front retreated. The following discussion attempts to follow the major changes and events as the ice front receded from various parts of the country.

St. Lawrence Lowlands

In southern Quebec the margin of active ice receded northward from Notre Dame Mountains into St. Lawrence Valley, but flow probably continued both up and down the valley from near the mouth of Saguenay River for some time before giving place to a general southward flow across the valley. During this latter stage the ice advance was restricted by the north side of Notre Dame Mountains and a system of end moraines was deposited, the lowest of which is known as the Highland Front moraine (Gadd, 1964, 1966). The whole system represents a lowering of the surface of about 1,000 feet. The St. Antonin Moraine (Lee, 1962) is the easternmost end moraine of the Highland Front system, and near Rivière du Loup the glacier that produced it was moving down valley and calving into the sea which occupied the lower part of the valley.

Farther west, in the wide part of St. Lawrence Valley, meltwater was ponded between the Appalachian Highlands, including Adirondack Mountains, and the ice front to form glacial Lake Vermont, with discharge south down Hudson Valley. The Drummondville Moraine (Gadd, 1960) was probably built during the Fort Ann phase of this lake. As the ice front receded northward the lake expanded northeastward along the south side of St. Lawrence Valley until it was able to discharge into the sea near Quebec City. The last short-lived lake is presumed herein to represent the Trenton phase of the post-Iroquois lakes as recognized in the Ontario basin by E. Mirynech, and is considered to be the last phase of con-

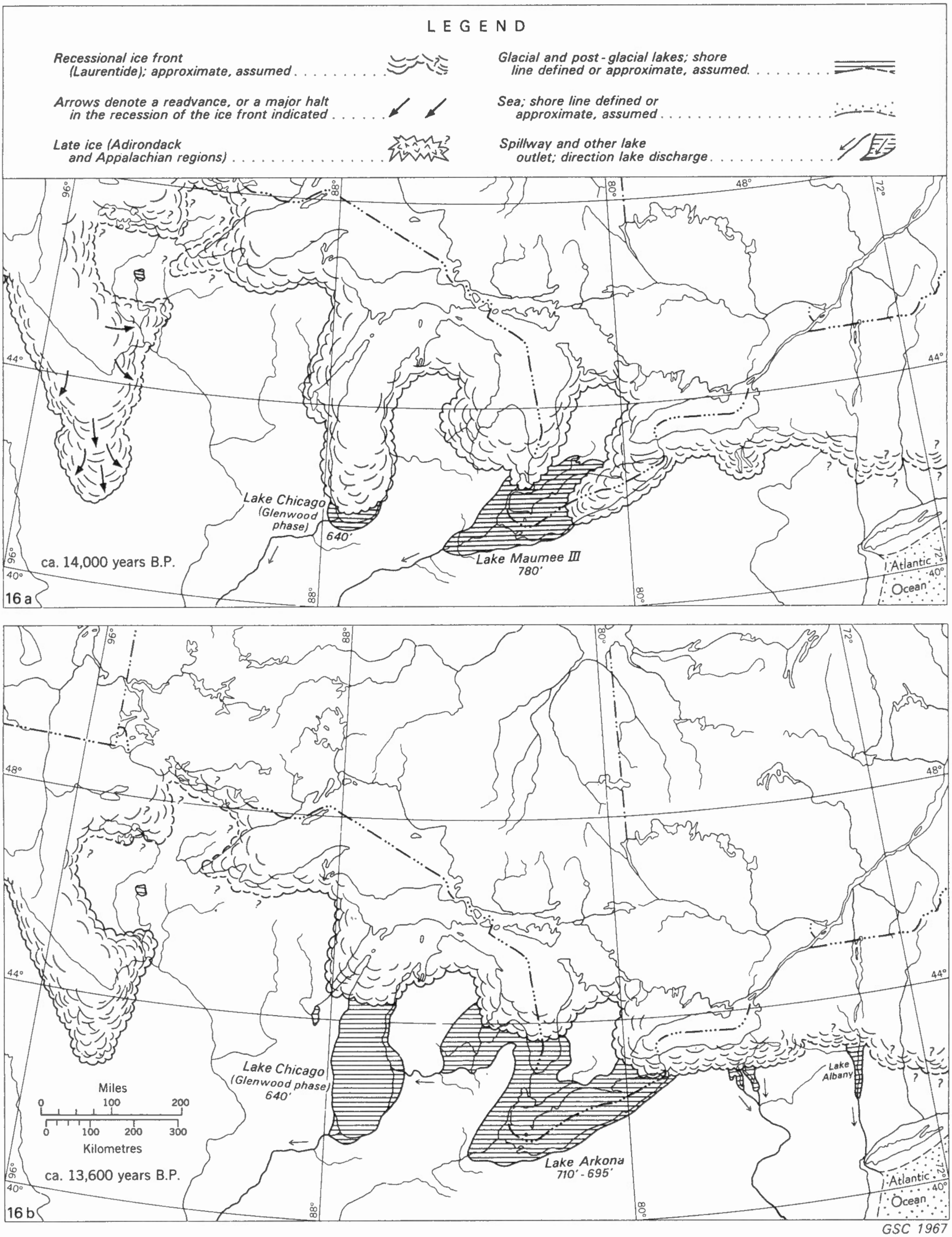


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada.

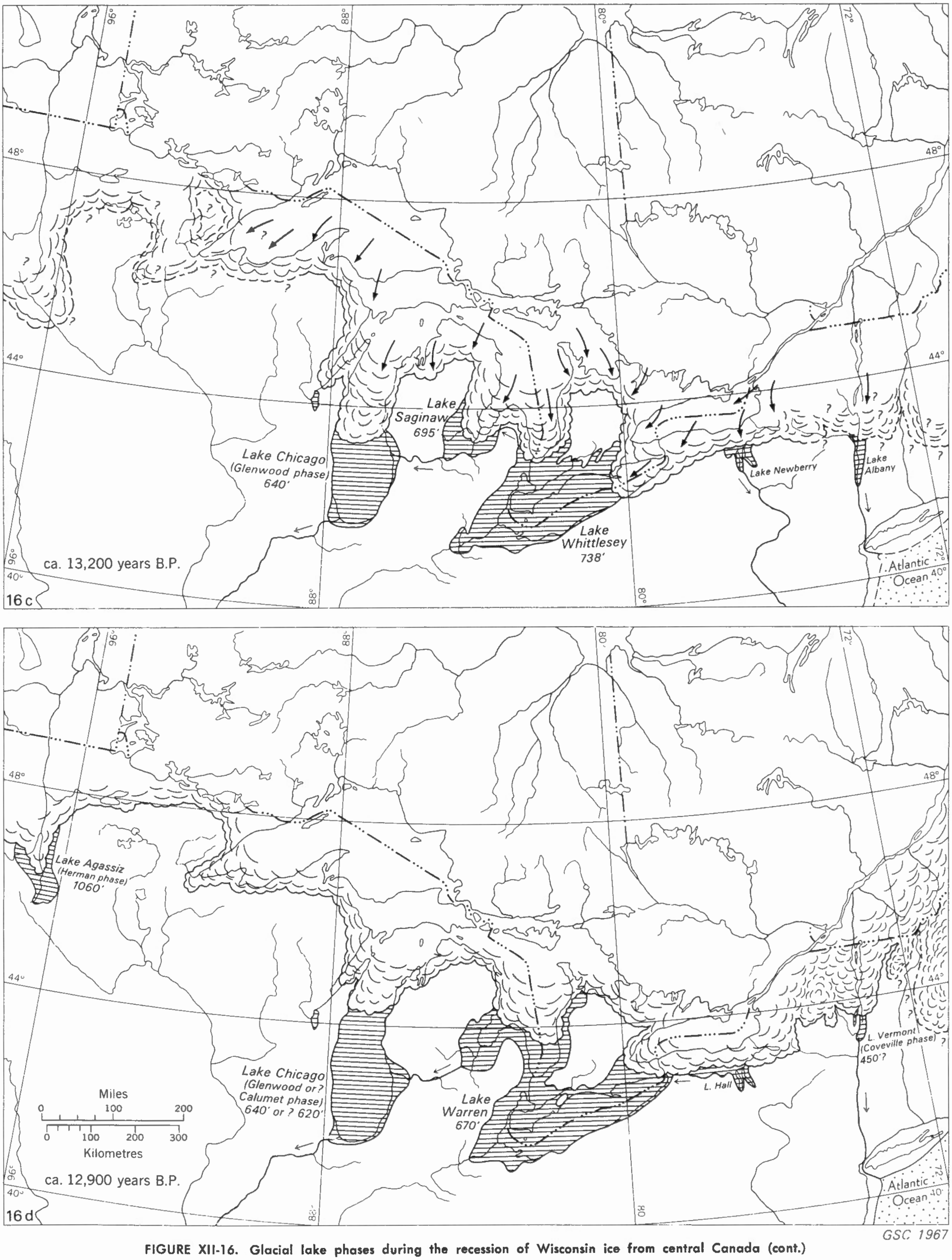


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

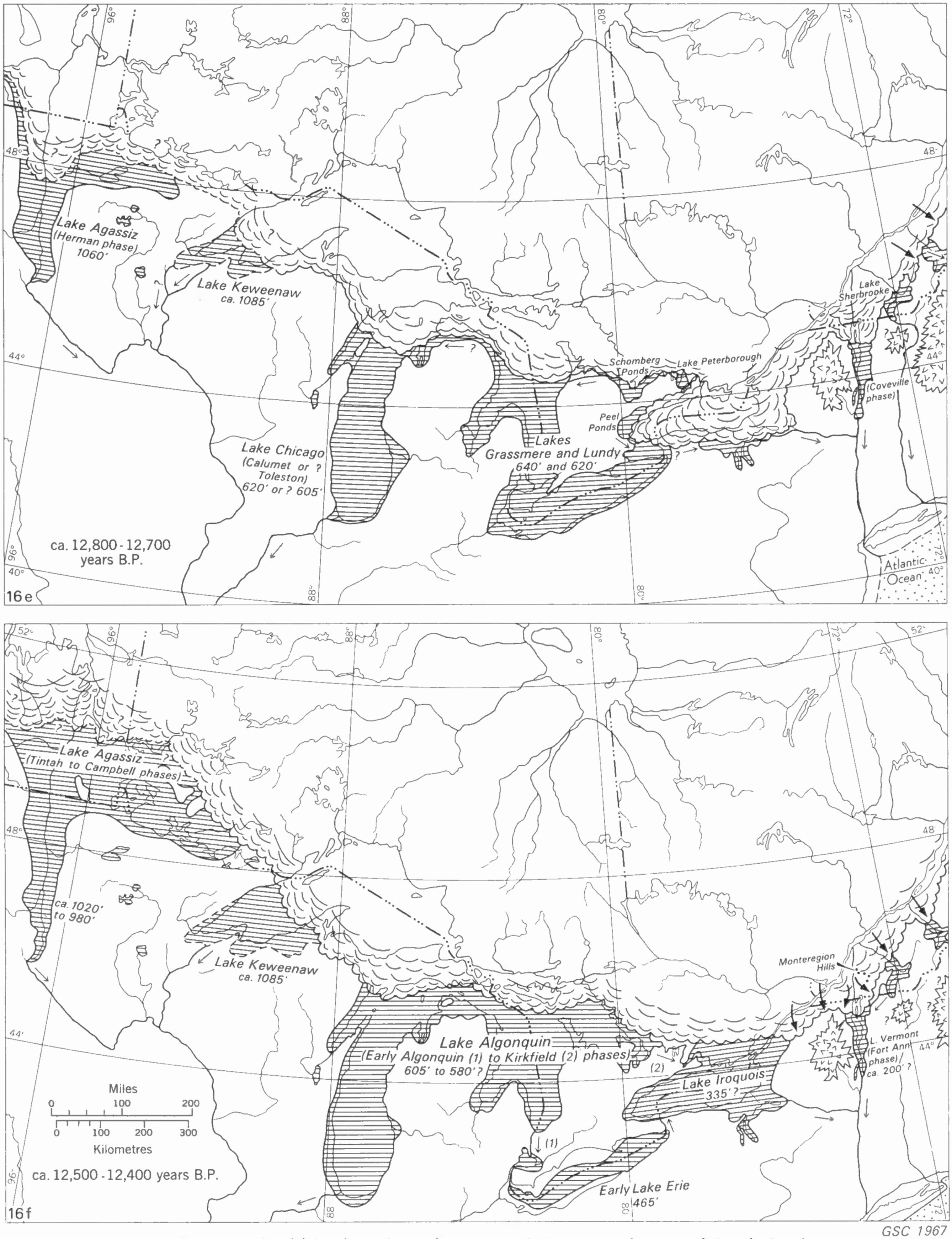


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

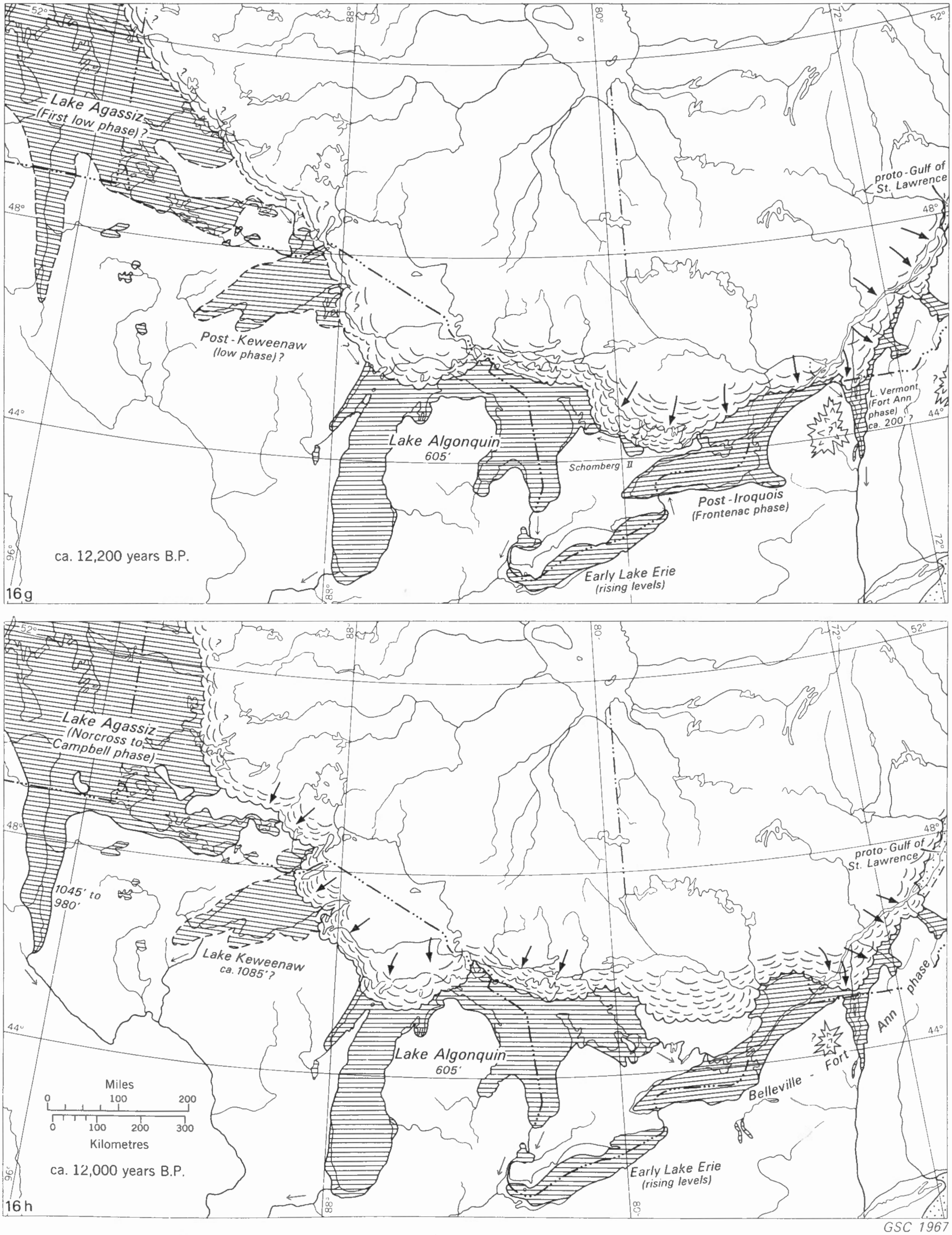


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

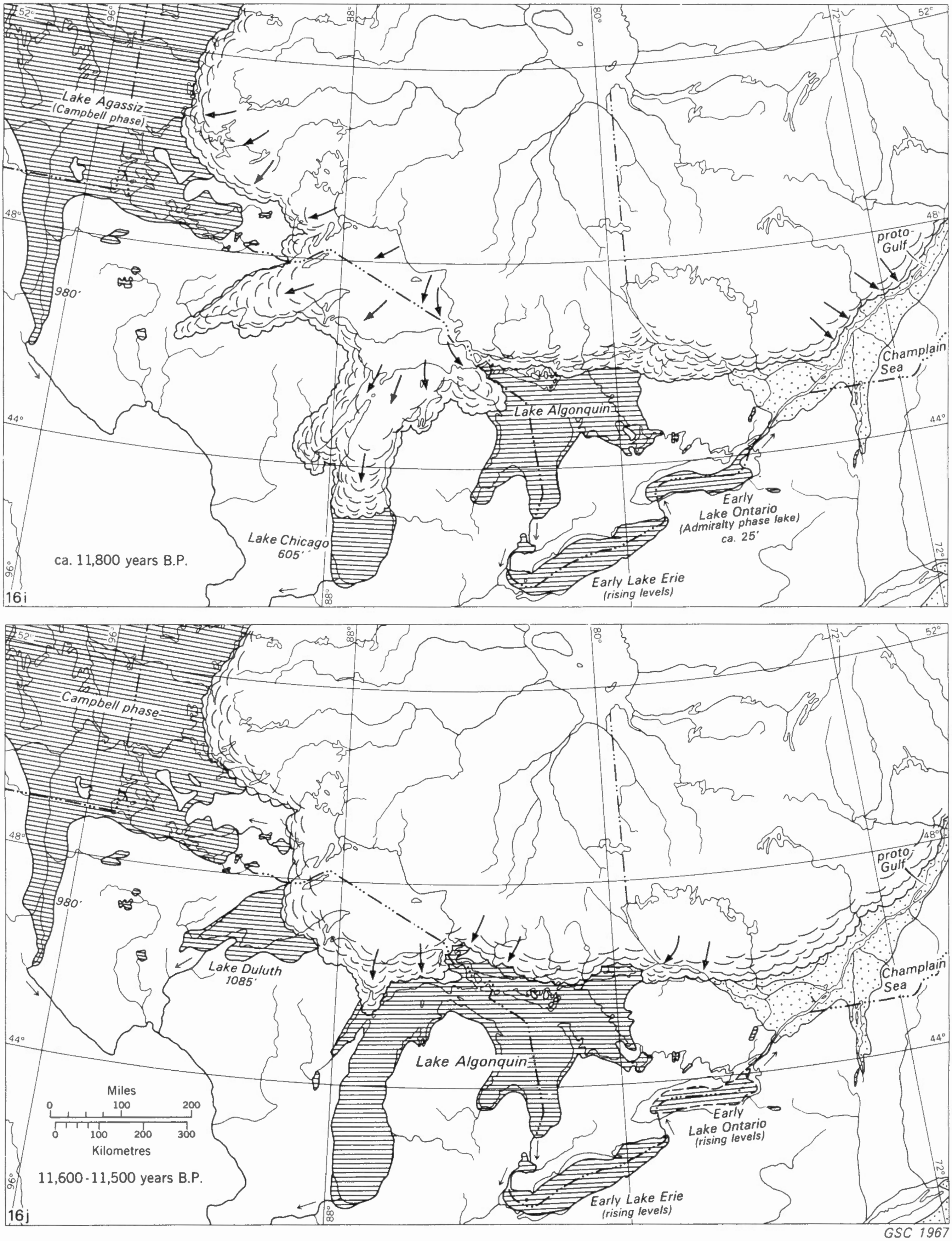


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

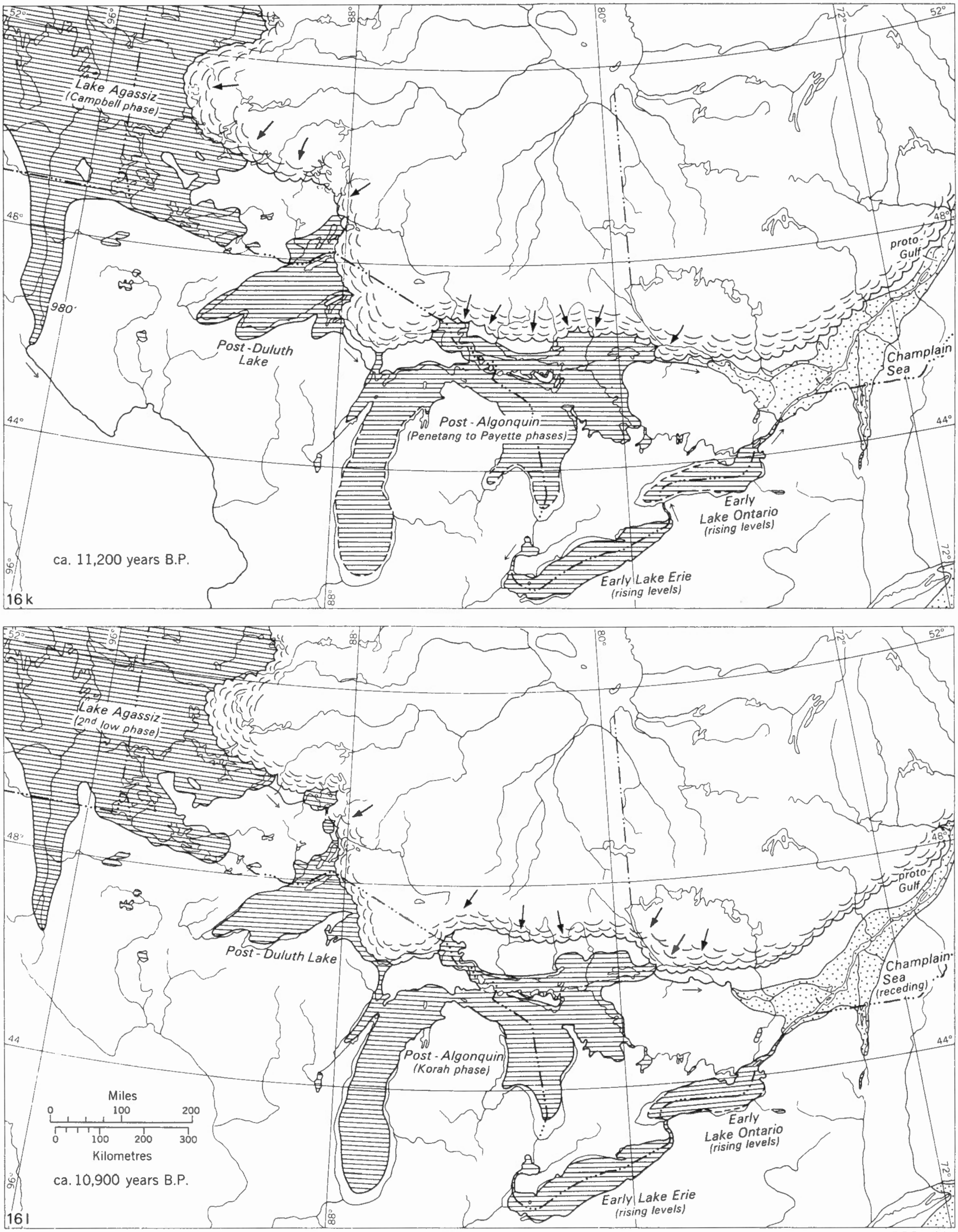


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

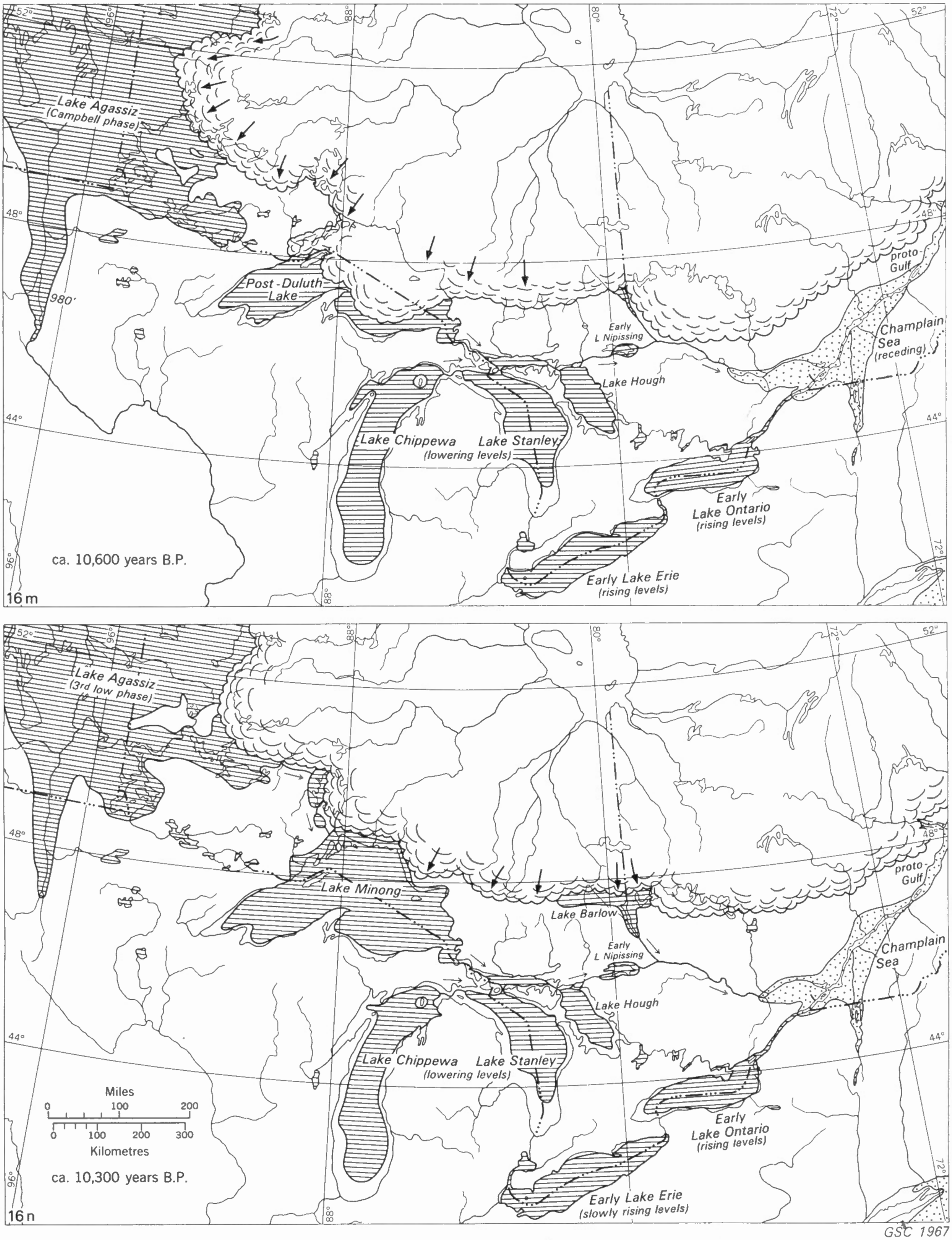


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

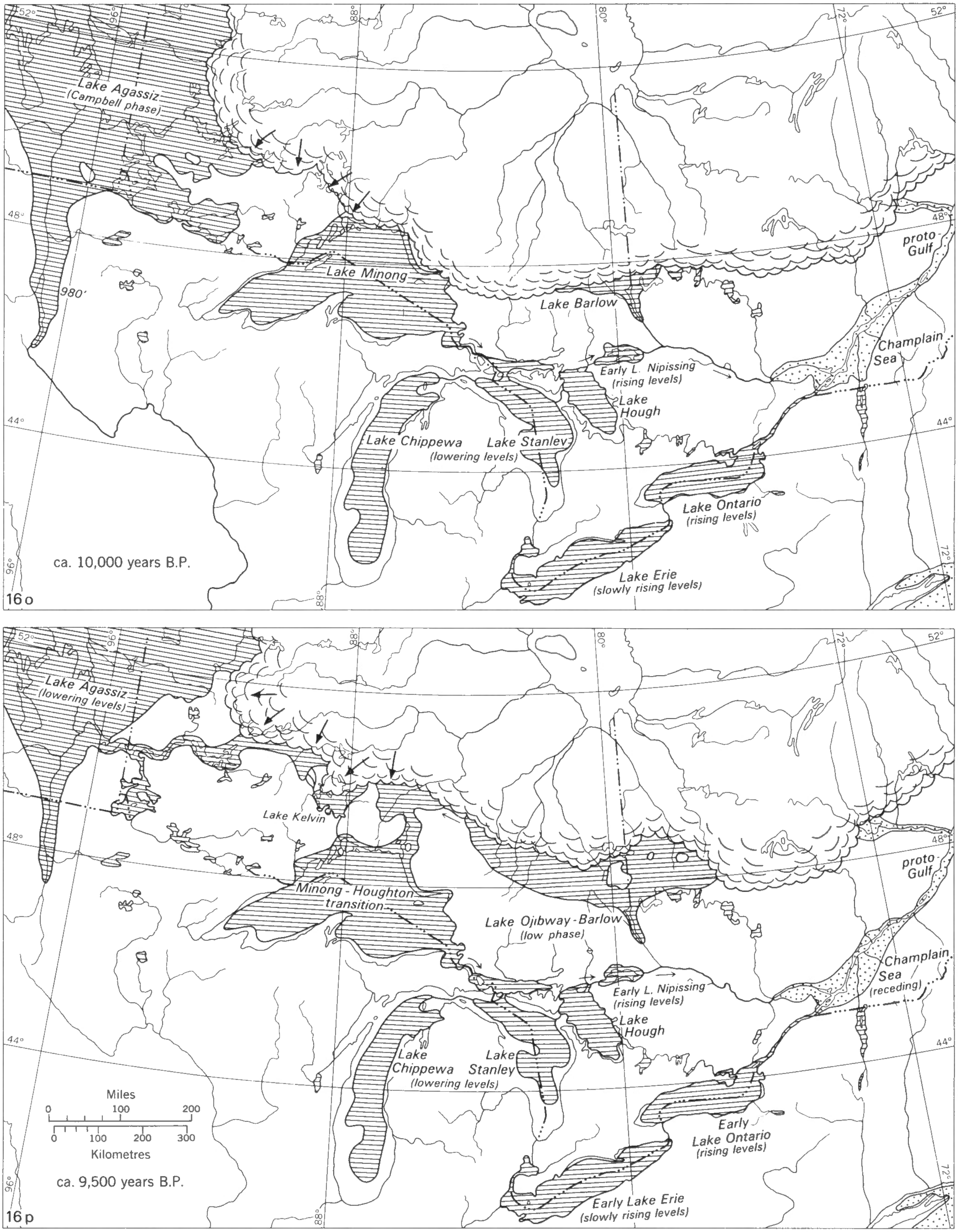


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)

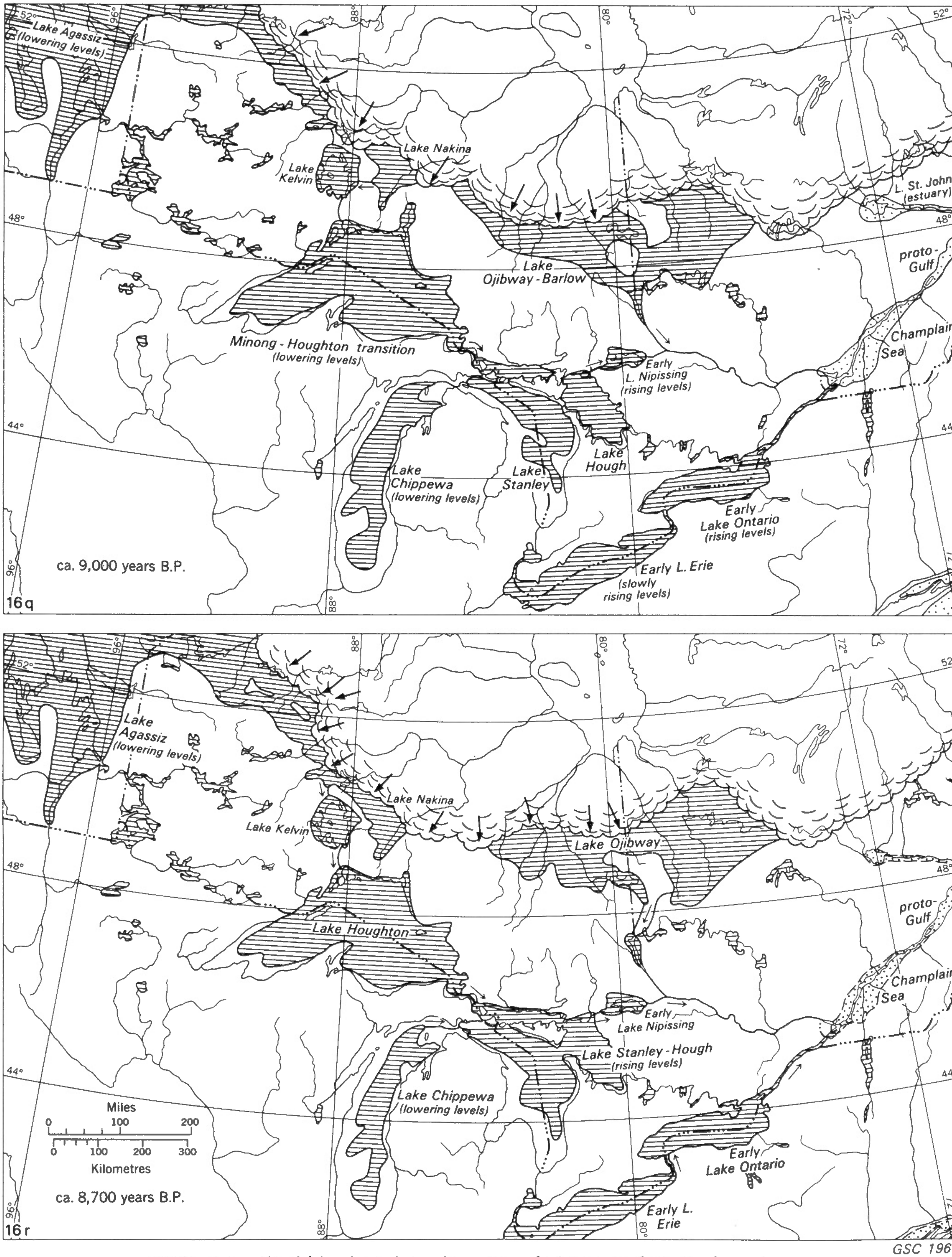
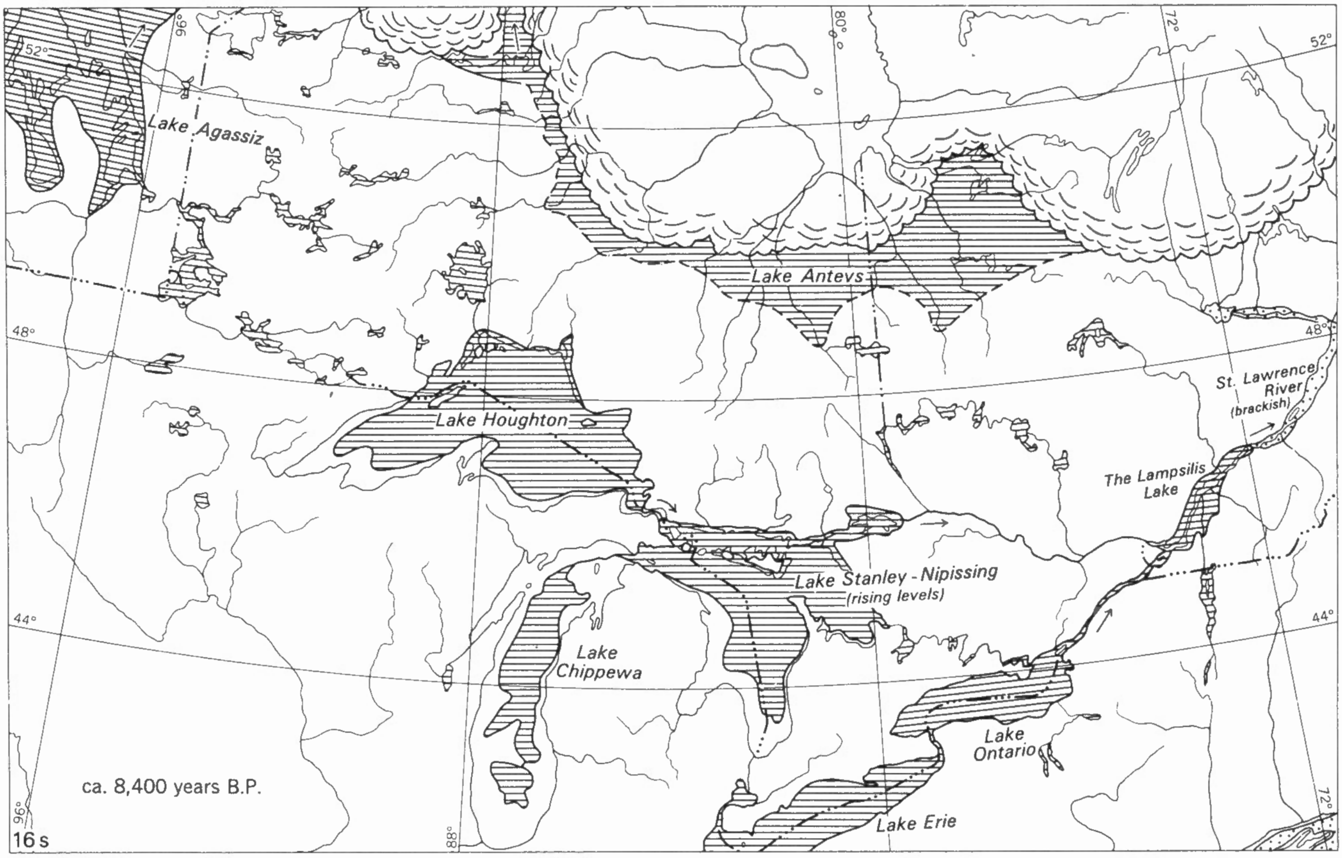


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)



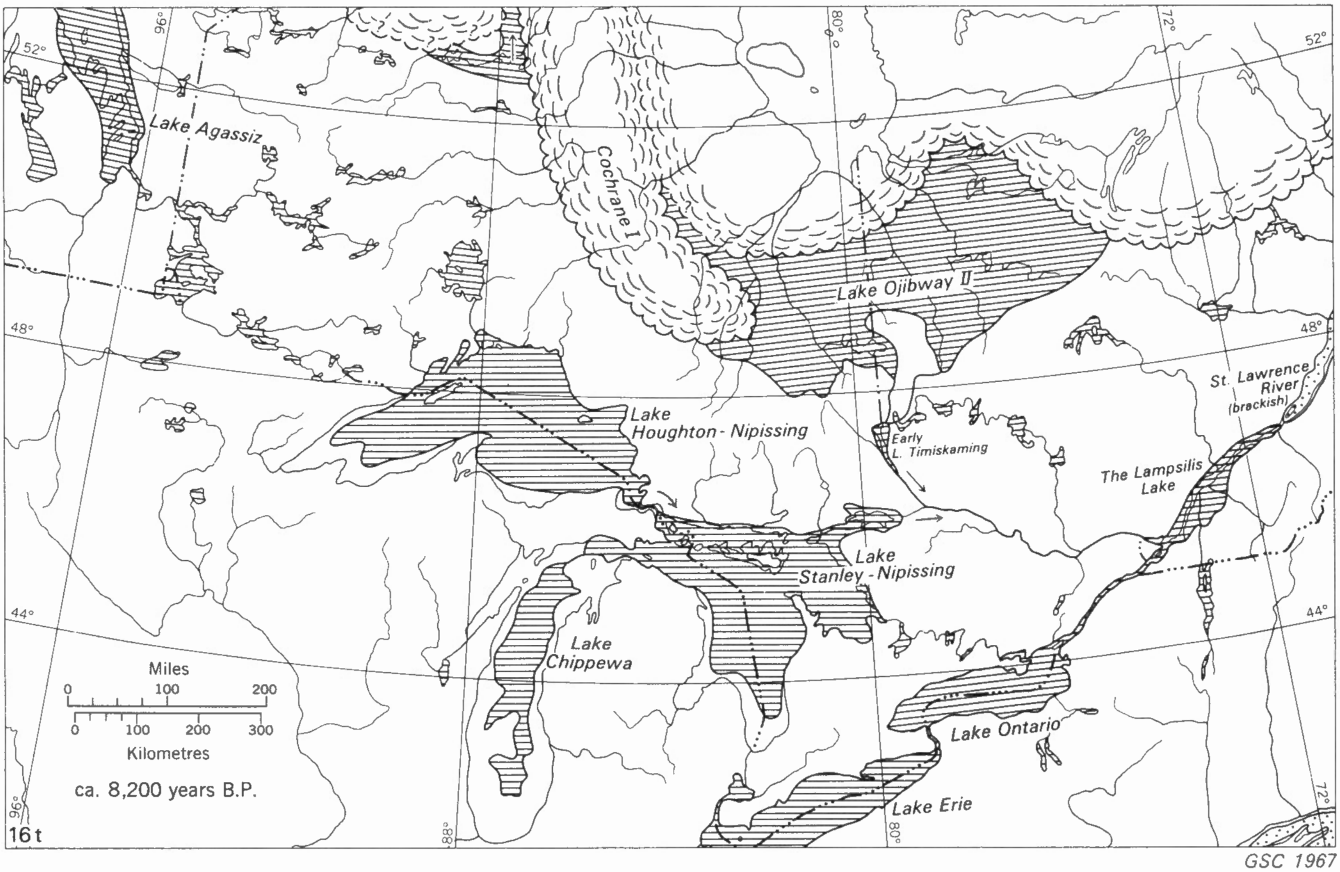


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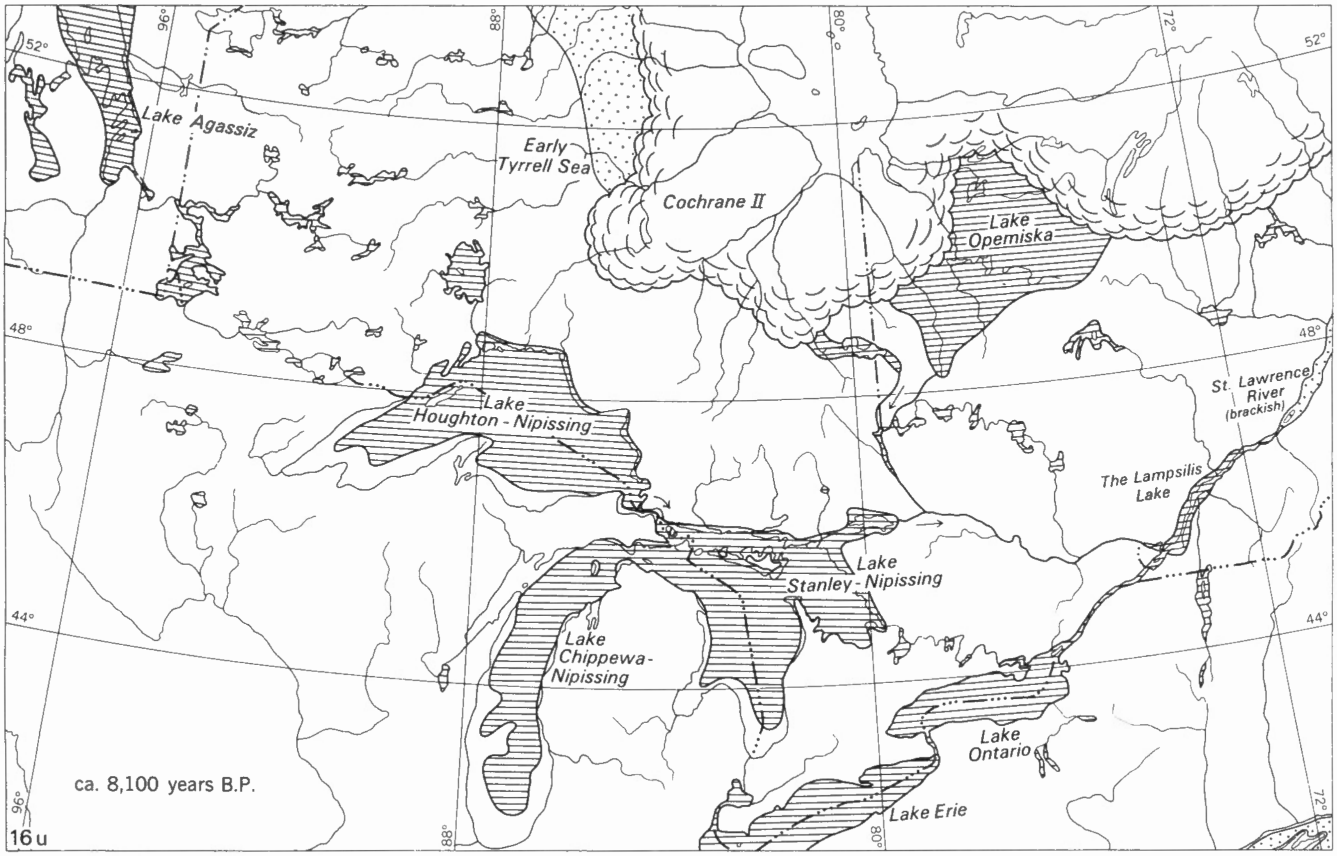
FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)



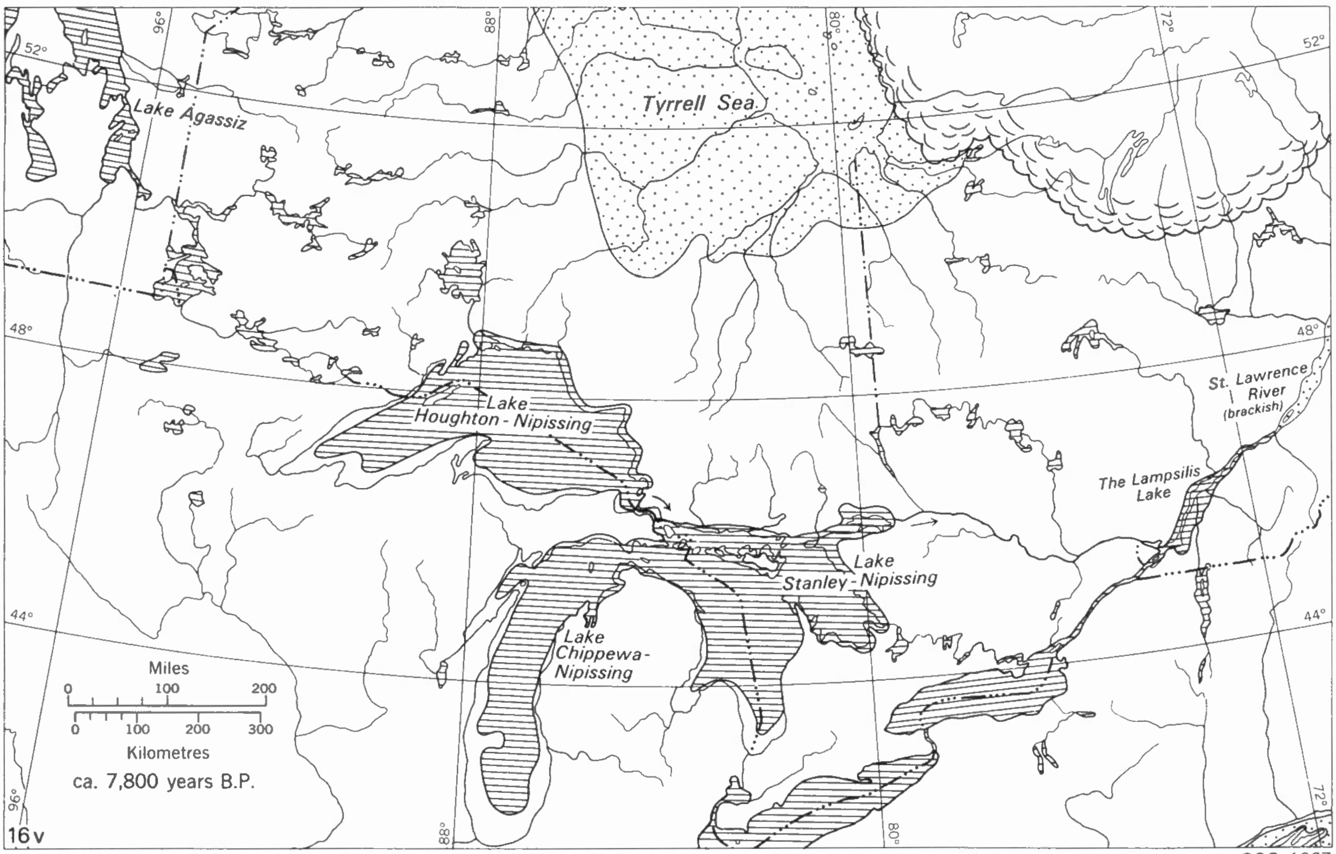


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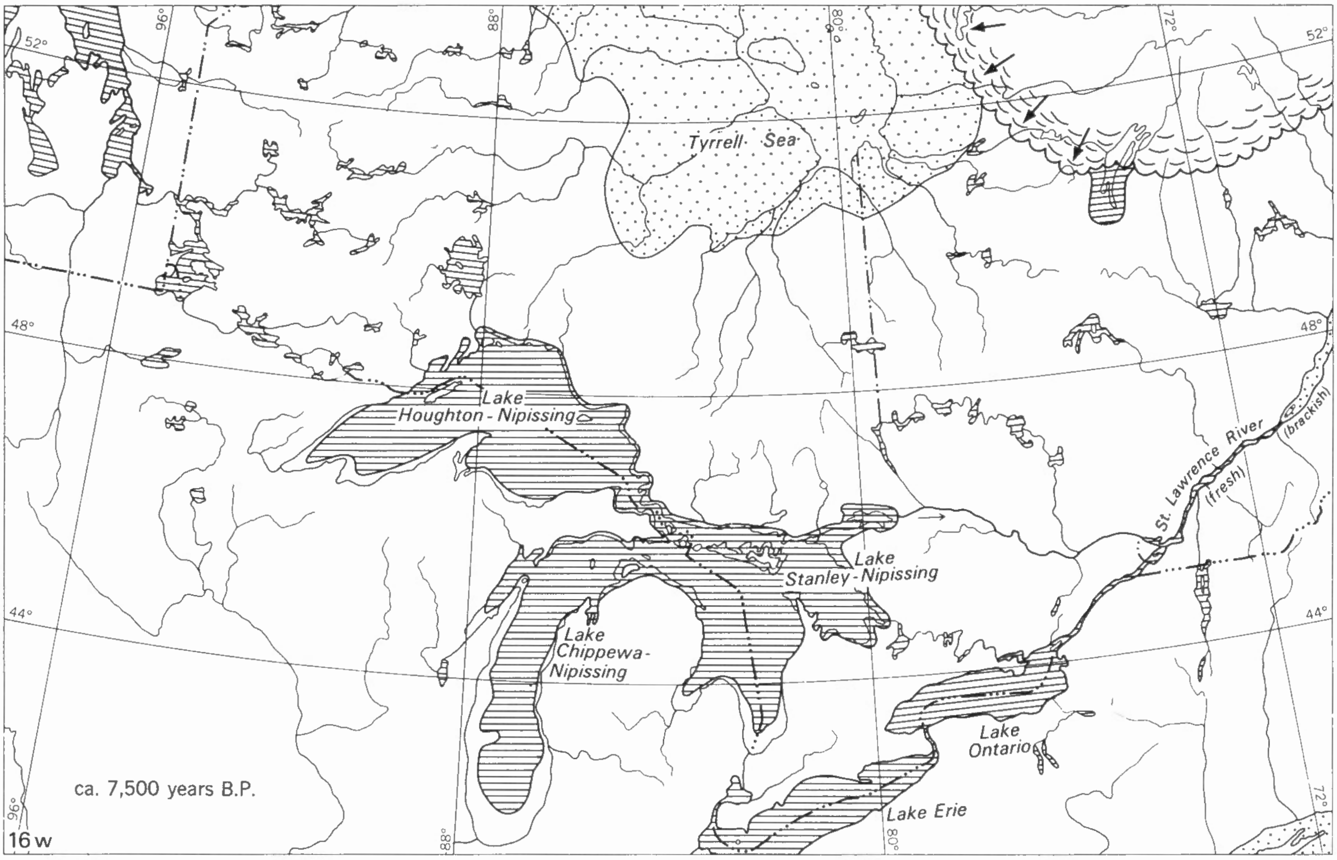
FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (cont.)



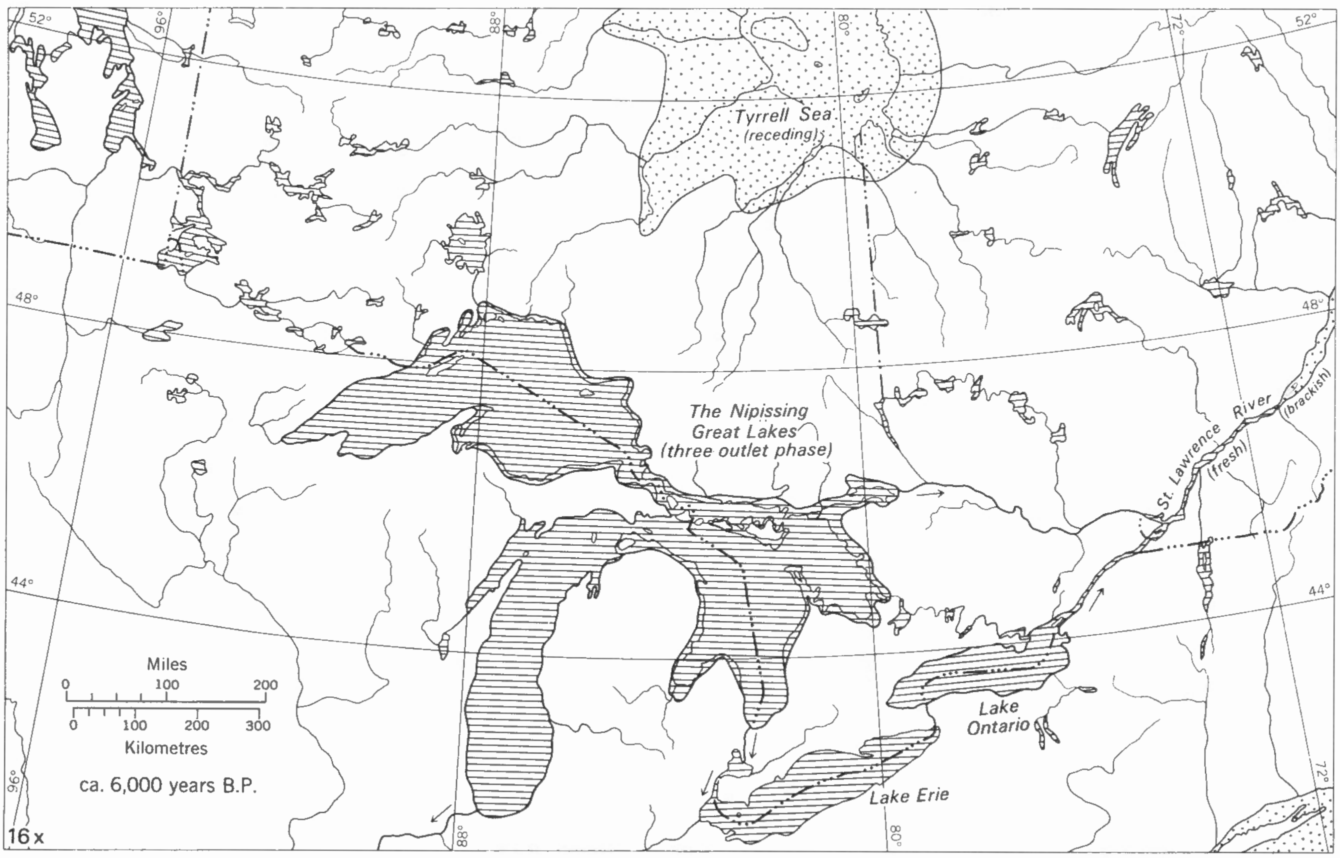


FIGURE XII-16. Glacial lake phases during the recession of Wisconsin ice from central Canada (conc.)

fluent Iroquois–Vermont waters. According to N. R. Gadd an ice lobe occupied Chaudière River valley and probably formed the last barrier between the glacial lake and the sea (Fig. XII-16h). Recession of the ice front from Chaudière Valley allowed the lake to drain to sea level whereupon the sea encroached the upper St. Lawrence and Ottawa River valleys. This inland sea in St. Lawrence Valley above Quebec City constitutes the Champlain Sea (Gadd, 1964). Labrador ice remained active and built an end moraine, presumably in the sea, parallel to the valley in the vicinity of St. Narcisse (Karrow, 1959). This moraine may be traced westward some 100 miles to near Lachute, Quebec (Parry and MacPherson, 1964). As the ice front receded from St. Lawrence Valley the Champlain Sea attained its maximum northward extent. Strandlines occur up to 750 feet above sea level north of Trois-Rivières and Montreal. Thereafter differential uplift brought about regression of Champlain Sea and development of the modern river system (Figs. XII-16k-w).

In the Great Lakes region, thinning of the Labrador sector of Laurentide Ice Sheet and retreat of the ice margin from the farthest point of advance, some 150 miles south of Lake Michigan, began about 17,000 years ago. The ice front retreated into the Lake Erie basin about 14,500 years B.P. Meltwaters were ponded between the basin rim and the receding ice front to form the first of the glacial Great Lakes, Lake Maumee, with discharge southward into Mississippi River (Hough, 1958, 1963, 1966). Thereafter, fluctuations of the ice front and differential uplift of the land, consequent upon removal of the ice load, combined to produce a complex system of lakes and spillways throughout the Great Lakes region and extending into the upper St. Lawrence Valley.

The present indicated history of the Great Lakes is the result of studies by geologists, geographers, and others that date back well before the turn of the century. The early work established the fundamental concepts pertaining to glacial lakes and the framework upon which later work was based. Studies made during the last decade, supported by radiocarbon datings, have made necessary some major changes in the basic chronology and sequence of events. The most recent attempts to correlate events between the various lake basins and discharge routes are by Hough (1963, 1966), Wayne and Zumberge (1965), and Chapman (1966). Two contrasting correlation charts are, in fact, introduced by Wayne and Zumberge to indicate differing concepts held by various authors. On the Canadian side of the Great Lakes available data and radiocarbon dates of lake deposits remain in conflict with the generally accepted correlations.

The writer has attempted to harmonize viewpoints, or otherwise point out discrepancies, concerning deglacial events in the well-documented history of the central Great Lakes and upper St. Lawrence regions with events in southern Manitoba and in James Bay region. The account invokes non-uniform fluctuations of several major ice lobations—as a result of variably delayed responses to climatic changes—to overcome some problems raised by radiocarbon dating of lake and sea deposits in widely separated regions (Figs. XII-16 a-x). It is based also in part on regional ice-flow trendlines and morainal positions developed during deglaciation (Map 1253A).

The southern glacial Great Lakes. The first of the glacial Great Lakes formed in the western end of Erie basin and is known as glacial Lake Maumee (Fig. XII-16a). It expanded into southern Huron basin as the Huron ice lobe receded. Successive water levels, resulting from use of two different outlets as the ice margin fluctuated, were at 800, 760, and 780 feet a.s.l.[1] During the high phases discharge was to the south via Wabash River, and at the low phase to the west, via Grand River, into glacial Lake Chicago in the Michigan basin and thence south, in both phases, to the Mississippi River system. Further ice retreat resulted in final abandonment of the Wabash River spillway and reopening and modification of the Grand River spillway to form glacial Lake Arkona, with its unwarped strandlines developed at 710, 700, and 695 feet a.s.l. (Fig. XII-16b). Continuing retreat of the ice front resulted in low-level lakes that are recorded by deposits in the United States and locally in Canada. A major glacial advance that built the Port Huron moraine system and a correlative moraine that crosses the base of Long Point in Erie basin gave rise to glacial Lake Whittlesey in western Erie basin, with prominent beach at 738 feet a.s.l. (Fig. XII-16c). Discharge was to the northwest via Ubly River channel into Saginaw basin and thence to Lake Chicago and Mississippi River.

Fluctuations of the ice front resulted in expansion of the lake both in the north and east, although at lower levels because of down-cutting of the outlet across Michigan Peninsula. These lake phases are known as Warren and Wayne and, in the main, they occupied the whole of Erie basin and southern part of Huron basin (Fig. XII-16d). Hough related them to fluctuations of the Port Huron (Mankato) ice. The successive lake levels are reported as 680, 670, 660 (Wayne), and 685 feet a.s.l. Continued recession of both Huron and Ontario ice lobes resulted in still lower lake levels known as Grassmere at 640 feet and Lundy at 620 feet a.s.l. (Fig. XII-16e). Discharge was believed to have been eastward, for the first time, along south side of Ontario basin into the Mohawk and Hudson River systems. Hough has favoured an outlet around the northern end of Michigan Peninsula and thence via glacial Lake Chicago and Illinois River to the Mississippi, pointing to the similarity of elevations of major lake phases in the lake basins concerned. Leverett and Taylor (1915), however, thought that by Lundy time the Huron basin waters drained southward to Erie basin. Further retreat of the Ontario basin ice lobe facilitated discharge eastward; early Lake Erie was established in

[1] a.s.l.—above Atlantic mean sea level. The given elevation of a glacial lake is the present elevation of the unwarped part of the inferred lake basin.

Erie basin and early Lake Algonquin in Huron basin with the controlling sill at Port Huron (Sarnia) at 605 feet a.s.l. (Fig. XII-16f). Early Lake Erie occupied only a small part of the present lake basin due to depression of the outlet at Buffalo (Fort Erie), but differential uplift later raised the outlet area with consequent filling of the Erie basin (Lewis, 1966).

When the ice sheet receded from the southern part of Lake Ontario basin, a lower outlet was uncovered at Rome, New York, and the Lundy Lake was lowered rapidly. Glacial Lake Iroquois was established in the Ontario basin at about 335 feet a.s.l. (Fig. XII-16f). The spillway led eastward through Mohawk River valley to Hudson River and thence south to the Atlantic Ocean. Due to prolonged use of Rome outlet, many excellent shoreline features were developed. The writer believes, however, that the shoreline features at 1,100 feet southwest of Covey Hill and ascribed to main Lake Iroquois are the work of a younger and lower lake phase.

The main Iroquois shoreline in the Ontario basin is now warped upward along a line trending N20°E with the hingeline situated south of the lake basin. The shoreline is about 360 feet at Hamilton, 460 feet at Rome, and 700 feet at Watertown, New York (Coleman, 1937). The indicated uplift gradient between Rome and Watertown is about 6 feet per mile. If this gradient and trend of uplift is projected beyond Watertown, the isobase of differential uplift through Cherabusco, New York, would be about 1,270 feet. An increase in the rate of tilt northward, as probably occurred, would give a higher figure but this would be offset by any swing towards the north in the line of maximum uplift, as does occur in the Adirondack region. Short spillways, at elevations of 1,305 and 1,290 feet, have been reported south of Cherabusco (MacClintock and Terasmae, 1960; MacClintock and Stewart, 1965) and were believed related to local lakes or pondings in front of Fort Covington ice with discharge westward into Lake Iroquois. Though the Iroquois shoreline is reported at only 1,100 feet southwest of Covey Hill there is evidence of shoreline features to about 1,250 feet, 1¼ miles south of Cherabusco. This is considered to be an Iroquois shore rather than that of a local lake. A small embayment of Lake Iroquois extended eastward from Cherabusco as a re-entrant between the receding northern and eastern ice fronts around Adirondack Mountains. This ice, probably Fort Covington, built an end moraine and ribbed moraine complex south and southwest of Covey Hill. C. S. Denny (U.S.G.S.) believes that an outlet opened southeast of Ellenburg, New York, which allowed main Lake Iroquois to breach the end moraine and discharge into glacial Lake Vermont farther south in Lake Champlain valley. In the northeast, strandlines indicate a lowering of about 150 feet during operation of this outlet; the real lake level, however, was lowered perhaps only 75 feet. The new lake level, formerly considered main Lake Iroquois, is herein referred to as the Ellenburg phase of the post-Iroquois lakes.

As glacier recession was resumed, Covey Hill outlet (sill elevation 1,010 feet) was uncovered and the post-Iroquois lake level was lowered a further 75 feet. This lake phase was named glacial Lake Frontenac and is herein considered the Frontenac phase of the post-Iroquois lakes (Fig. XII-16g). The Covey Hill outlet today consists of a 60-foot dry waterfall or cliff, a 75-foot deep lake with an unknown thickness of sediment at its base, and a mile-long gorge about 130 feet deep and 300 to 600 feet wide. When the ice withdrew from the northern and eastern flanks of Covey Hill there was a major drop in lake levels of some 125 feet. The short stand at this level, probably occasioned by an ice-marginal fluctuation on the northern sides of Covey Hill, is termed the Sydney phase lake by E. Mirynech.

In the Trenton embayment in eastern Ontario the lake was lowered a further 30 to 75 feet according to Mirynech, prior to a significant halt responsible for development of the Belleville beach. Isobases drawn on the Belleville beach would place the strandline on Covey Hill at about 750 feet. This is the same as that of the Fort Ann phase of glacial Lake Vermont which expanded northward as the eastern side of Covey Hill was uncovered by the ice (Chapman, 1937). The now confluent water bodies discharged southward along the earlier established route to Hudson River and the sea (Fig. XII-16h).

Rapid ice-marginal retreat in St. Lawrence Valley allowed the Belleville–Fort Ann phase lake to expand northward and also to extend a long arm northeastward between the ice front and the eastern Quebec Uplands which opened an escape route eastward to the sea at Quebec City. Water levels first dropped about 40 or 50 feet and, according to Mirynech, the Trenton shoreline was formed in the eastern part of Lake Ontario basin. This temporary halt in lowering lake levels was probably occasioned by an ice-marginal fluctuation in conjunction with a topographic barrier near Quebec City. As Labrador ice receded from Chaudière River valley the Trenton phase lake drained to sea level; marine waters then interchanged with the remaining lake waters in upper St. Lawrence Valley to form Champlain Sea (Fig. XII-16i). The sea extended up St. Lawrence River to Brockville and, as the ice front receded, up the Ottawa River to Petawawa. The Lake Ontario basin waters were drained to the Admiralty low water phase, probably less than 25 feet above Champlain Sea. These last events occurred shortly after 12,000 years B.P. as Champlain Sea shells have been dated 11,880 ± 180 years (GSC-505).

The indicated maximum drop in lake levels from main Iroquois (presumably 335 feet a.s.l.) to the level of Champlain Sea is more than 335 feet. This is probably due to differential uplift during the later phases of lake history and to the difference in real sea level during early Iroquois and Champlain Sea time as compared to the present. Also, the main Iroquois data may well be in some error.

The Algonquin Lakes. As ice recession continued in the Huron basin, early Lake Algonquin occupied the southern part of the basin and, also, Georgian Bay and Lake Simcoe lowland. Numerous strandlines have been named in various parts of the Huron and Michigan basins applicable to a long 'Algonquin' interval of ice retreat and differential uplift, but it is evident that the lake phases and their implications are not yet adequately known.

Early Lake Algonquin at 605 feet a.s.l. was drained southward, possibly first via Chicago and later via Port Huron, into Lake Erie until such time as Lake Simcoe lowland became free of ice and discharge was possible via the Kirkfield-Fenelon Falls and Trent Valley river system into Lake Iroquois (Fig. XII-16f). Continued use of this outlet system, by the Kirkfield phase lake (at about 580 feet a.s.l.), was blocked by a readvance of the Simcoe ice lobe (Deane, 1950) and discharge returned to the Port Huron outlet (Fig. XII-16g). Retreat of the ice front gave access to the Kirkfield outlet system again, but in the meantime differential uplift is considered to have raised the drift-filled outlet; the water plane may thus have remained at about 605 feet a.s.l. for a lengthy period (Fig. XII-16 h,i,j).

The main Algonquin strandline in the Huron basin is a prominent feature, in many places marked by a strong shore bluff. According to L. J. Chapman, it is not found south of Point Clark, midway up the eastern shore, having been undercut by the present lake. It extends northward into Georgian Bay and eastward around Lake Simcoe to the Kirkfield outlet where, due to differential uplift, it has a present elevation of 870 feet (Deane, 1950). The strong Algonquin strandline may be a result of operation of both the Kirkfield–Fenelon Falls and Port Huron outlets during a halt in isostatic uplift of the region. The writer considers it likely that down-cutting at Fenelon Falls greatly reduced the discharge at Port Huron over a lengthy period. Deane believed the Fenelon Falls channel was about 30 feet deep. Extensive gorges were cut in limestone along Trent River valley above the Sydney strandline of the post-Iroquois lakes; thereafter discharge was appreciably lessened.

The gradient of the warped main Algonquin strandline in Lake Simcoe and Lake Couchiching areas increases northward from 2.8 to 4.0 feet per mile along a line N21°E. Farther north, the highest strandline recorded by Chapman (1954, 1966) of 1,070 feet elevation at Huntsville and 1,245 feet elevation at Sundridge shows a further increase in gradient to 5.0 and 6.3 feet per mile. This high rate of tilt militates against rapid differential uplift during the time represented by the highest shorelines and, accordingly, main Lake Algonquin must have extended northward as far as Sundridge without any appreciable change of level while the ice front retreated in the Huron–Georgian Bay region. Between Sundridge and Trout Creek, however, Chapman (1966) records that the highest strandlines were lowered northward by about 50 feet, and he relates this lowering to differential uplift. If the main Algonquin water plane is projected northward to Trout Creek, however, the amount of lowering would be about 125 feet. Such sudden uplift during the existence of the lake in this area seems unlikely. It is more likely that a period of ice-marginal and subglacial discharge occurred at Fossmill, the waters flowing eastward through Petawawa and Barron Rivers into the Champlain Sea in Ottawa River valley. If the main Algonquin strandline was projected northward to the Fossmill outlet sill at Kilrush (elevation 1,145 feet) it would be at about 1,415 feet elevation. Thus flow at Fossmill outlet over the Kilrush sill might have lowered Lake Algonquin by as much as 270 feet had this not been offset to some degree by differential uplift over a fairly lengthy period and provided the area remained above sea level.

This two-fold process, ice-controlled discharge and differential uplift, may account in part for the long-standing controversy as regards parallelism or convergence of some of the post-Algonquin beaches, namely Ardtrea, Upper Orillia, Lower Orillia, Wyebridge, Penetang, Cedar Point, and Payette. According to Deane (1950) the Fenelon Falls outlet remained in use until after the Ardtrea phase. There had been about 50 feet of differential uplift by Upper Orillia time, and Lake Simcoe was separated from 'Algonquin' waters following the Penetang phase (Fig. XII-16k.)

Prior to this event, during the life of Lake Algonquin, there was a major re-advance of the ice sheet in the Superior and Michigan basins, but apparently it was of little consequence in the Huron basin. This ice advance, the Valders, is interpreted as a glacier surge from northeast of Lake Superior (Fig. XII-16i). The highland east of Lake Superior may well have restricted the flow of ice into the Huron basin while it lobed far southward into Michigan basin. This lobation persisted as the Valders ice receded (Fig. XII-16j) and hence glacial Lake Algonquin was able to invade the Sault Ste. Marie area while being restricted at the north end of Michigan basin and virtually excluded from Superior basin (Hough, 1958).

Glacial Lake Algonquin did not enter the Sudbury basin; strandlines occur at various elevations but all are well below a projected main Algonquin water plane. Farther west, however, higher strandlines are evident and 3 miles northwest of Sault Ste. Marie they reach an elevation of 1,025 feet. This strandline is clearly related, on the basis of isobase trend and rate of uplift, to the highest strandline recorded on Manitoulin Island. Three miles south of Little Current this is given as 1,013 feet, and is reported to be main Algonquin. Projection of the isobase eastward, however, shows it to be about 25 feet lower than the main Algonquin shoreline east of Huntsville. This suggests that Manitoulin Island was ice covered until after some 25 feet of uplift had occurred, probably during transition from the Ardtrea to Upper Orillia phase.

In the northern part of Lake Simcoe area the difference in elevation between Upper Orillia and Payette strandlines is about 140 feet (Deane, 1950, Fig. 7). Thus operation of Fossmill outlet system over an extended period, with a maximum draw-down at Fossmill of about 220 feet, readily accommodates all the strandlines recognized by Deane. Had there been no uplift of the Fossmill area during the life-span of these post-Algonquin lakes, the Sheguiandah and Korah lake phases, which represent a lowering of 70 feet below the Payette strandline at Sault Ste. Marie (Hough, 1958), might also correlate with discharge at Fossmill over the Kilrush sill. It is probable, however, that as differential uplift was already in progress an additional 50 to 100 feet of uplift may have occurred in the Fossmill area over a period of several hundred years. Accordingly the Sheguiandah and Korah lake phases may be correlated with use of subsidiary routes into the Petawawa River valley. One of these subsidiary routes is through Sobie and Guilmette Lakes, the controlling sill position being east of the latter lake at about 1,125 feet elevation. This site, some 3 miles north of the Kilrush sill in the direction of maximum uplift, permits a further draw-down of 40 feet, and is considered by the writer to correspond with the Sheguiandah phase lake, 35 feet below the Payette level at Sault Ste. Marie. A second outlet route became available when the ice front receded into the Mattawa River valley and exposed part of the Amable du Fond River valley; this led southward to Mink Lake and the Petawawa River system. The sill at Mink Lake is at 1,075 feet elevation and lies on the same isobase as the Guilmette sill (elev. 1,125 feet). Thus a maximum draw-down of 50 feet is possible but there may well have been some 15 feet of uplift as the ice withdrew into the Mattawa valley. The writer therefore correlates the Korah lake phase, about 35 feet below Sheguiandah stage at Sault Ste. Marie, with use of the Amable du Fond outlet over the Mink Lake sill (Fig. XII-16l).

Discharge of the post-Algonquin lakes eastward at Fossmill, and by subsidiary channels, to Petawawa and Barron Rivers resulted in deposition of a large delta, composed largely of sand and some gravel, in the Champlain Sea. Judging by the size of the delta, the spillway must have lain along an active ice front. Gadd (1963), in fact, located a small end moraine in Ottawa River valley near the mouth of Petawawa River. Boulders in the Petawawa–Barron Rivers spillway system are unusually large, in places averaging 2 to 3 feet in diameter. They form ridges and hummocks up to 20 feet high that appear to be bars built by torrential streams.

While the discharge of the post-Algonquin lakes was controlled by Mink Lake sill, immediately prior to eastward discharge down Mattawa River valley, the water depth over North Bay sill was about 400 feet—a calculation based on an uplift rate of 6.5 feet per mile in a direction N20°E and substantiated by strandline positions north of North Bay. As Hough (1958) considers the Payette phase lake had a surface elevation of 465 feet, the Korah phase was about 400 feet. Hough has given the elevation of Lake Stanley, the lowest level lake in Huron basin, as about 190 feet though it may have been appreciably higher according to C. F. M. Lewis. It is thus evident that following perhaps 200 feet of draw-down, as the post-Algonquin Lake discharged eastward along Mattawa River valley, the Huron basin part must have become separated from the lake remaining in the Nipissing Lake basin, which was then lowered a further 200 feet to expose the North Bay sill (Fig. XII-16m). There was also a separate low-level lake in Georgian Bay basin; W. M. Tovell proposes to name it Lake Hough in recognition of J. L. Hough's work on the documentation of the low-level lakes in Michigan and Huron basins.

During the period of ice retreat eastward along Mattawa River valley, the lowering of lake levels must have been accomplished mainly by subglacial discharge. The writer has found no evidence of a surface outlet in the valley between Amable du Fond and a point 5 miles southeast of Mattawa, Ontario. By the time surface (ice-marginal) discharge to the Ottawa River valley began, an over-all lowering from the Korah lake level in Nipissing–Mattawa Valley of about 370 feet had already taken place. Surface discharge then lowered the lake a further 30 feet at which time North Bay sill emerged. Withdrawal of the ice from Ottawa Valley between points 7 and 20 miles east of Mattawa only served to steepen the spillway gradient east of North Bay.

Superior Lake Basin

The Duluth–Minong–Houghton Lakes. Glacier retreat in Superior basin was in a northeasterly direction, resulting in ponding of meltwater at the western end. The history of Superior basin lakes and ice retreat is dealt with by Farrand (1961) and Zoltai (1965). The early phases of Superior basin lakes, as recognized by Farrand, did not affect Canadian shores. Glacial Lake Duluth may have formed a beach on Isle Royale just east of the International Boundary at 1,060 or 1,075 feet elevation, and Zoltai shows Lake Duluth extending along the boundary west of Lake Superior. Lake Duluth discharged southwestward to the Mississippi Valley (Fig. XII-16j). Retreat of the glacier subsequently opened a lower route into Michigan basin that resulted in a series of short-lived lakes. These were formerly considered equivalent to the Algonquin lake phases of Huron basin, but are now thought to be independent and are referred to by Farrand as the post-Duluth lakes (Figs. XII-16j–m). Beaches referable to some of these lakes occur on Isle Royale and west of Fort William. The latter, however, partly represent a separate and earlier lake in the Kaministikwia River basin (Zoltai, 1965).

When glacier retreat from the eastern end of Superior basin allowed the post-Duluth lakes free access to Whitefish Bay they may have become briefly confluent with post-Algonquin lakes in Huron basin. Somewhat later the glacier receded entirely from Superior basin. Farrand refers to the resulting lake as Lake Minong; it was responsible for the highest shoreline features in the northeastern end of the lake basin. The upwarped beaches occur between elevation 950 and 1,000 feet on the north shore. Farrand relates Lake Minong to Sheguiandah lake phase in Huron basin but the writer considers it to be somewhat younger and with its outlet at Sault Ste. Marie (Figs. XII-16n, o).

Below the Minong strandline on the north shore of Lake Superior there are a succession of beach ridges or small wave-cut bluffs, depending on the character of the shore, that Farrand relates to post-Minong lake phases. They have a vertical range of over 50 feet. The lowest and best formed shoreline at 750 to 765 feet near Dorion is known as the Dorion beach and appears to have resulted from outlet adjustments rather than differential uplift (Figs. XII-16p, q). Another adjustment was responsible for a further lowering of several tens of feet, and development of a shoreline that Farrand considers correlative with the low-level lake phases in Huron basin. He named the low-level lake in Superior basin as Lake Houghton and reports that it discharged to the Huron basin by the proto-St. Marys River at Sault Ste. Marie. He considers the elevation of the lake as about 360 feet a.s.l. Its shorelines have been destroyed along the east side of Lake Superior by a younger lake but are preserved on the north shore as a result of differential uplift (Figs. XII-16r, s).

Upper Great Lakes Region

The low-level, Nipissing, and post-Nipissing lakes. As mentioned earlier, glacier retreat from the eastern end of Mattawa River valley served to drain the lakes in the upper Great Lakes basins to very low levels. The indicated amount of draw-down at North Bay also necessitates establishment of separate lakes in the Lake Nipissing, Georgian Bay, and main Huron basins (Fig. XII-16m). Erosion of outlets from each of these basins allowed for continuing draw-down of the upper lakes over a very long interval during which differential uplift was raising the northeastern end of the drainage system. Thus the lowest levels of Stanley, Chippewa, and Houghton were not established simultaneously. These lakes remained until their outlets were drowned as water encroached from the east consequent upon the continuing uplift of the North Bay region. A speculative configuration of the lakes is shown in (Figs. XII-16m–w). Dependant upon the assumptions made as regards both the ages of the several post-Algonquin lake phases and the amount and rate of differential uplift during their life-span and during the subsequent low-level lake phases, various ages may be deduced for these low-level lakes. Lake Stanley may have merged with Lake Chippewa as early as 10,400 years B.P. rather than as late as 8,100 years B.P. as shown. Perhaps an age of 9,500 years B.P. is more likely than either extreme but it is necessary to allow sufficient time for the re-excavation of Mackinac channel between the Michigan and Huron lake basins.

When North Bay was raised to 605 feet a.s.l. discharge began again at Port Huron and at Chicago. Continuing differential uplift finally raised North Bay outlet above the lake surface and drainage was entirely by the southern outlets. Long use of the southern outlets resulted in a rather stable lake level and a consequent well-marked shoreline, that of the Nipissing phase of the Great Lakes (Fig. XII-16x). The confluent water bodies in Huron, Michigan, and Superior basins with discharge at Port Huron and Chicago are referred to as the Nipissing Great Lakes. The Nipissing terrace is well displayed a short distance west of North Bay at 700 feet elevation whereas the present shore of Lake Nipissing is at 648 feet. The highway to Sault Ste. Marie crosses the Nipissing terrace many times.

The Nipissing phase of the Great Lakes is usually dated at about 4,200 years B.P. Many organic deposits from positions below the Nipissing strandline, in various parts of the Great Lakes area, have been given ages ranging back to more than 7,000 or 8,000 years B.P.; these are commonly referred to the interval of pre-Nipissing rising water levels. However, recent dates from Little Pic River, north shore Lake Superior, indicate the age of a Nipissing phase lake to be about 6,000 years B.P. Wood from silty clay beds beneath a 15-foot capping of sand that forms a terrace at the highway bridge over this river was dated at 5,920 ± 120 and 5,960 ± 120 (GSC-83, 103). The terrace has an elevation of about 700 feet. Farrand (1961) considered the deposit Nipissing/Algoma transition as he determined the elevation at 692 feet and found a cobble ridge at 718 feet which he considered to be the Nipissing beach. The writer and S. C. Zoltai noted a transition from plant-bearing silty clay, with shells in the upper few feet, upward into shell-bearing (mainly *Sphaerium sulcatum*) sand. Tiny plant tissues from the basal few inches of the sand were dated at 6,100 ± 160 (GSC-285). The writer therefore considers the plant-bearing silty clay, the shell-bearing sand, and the boulder beach to represent the same lake phase, probably the Nipissing. On the eastern side of Lake Huron, 4 miles northeast of Owen Sound, wood has been found in the basal part of an extensive gravel ridge that L. J. Chapman considers an undoubted Nipissing beach. The gravels are 18 feet thick, pass down-slope into sand, and rest on a silty clay. The wood from the sand-clay contact was dated at 5,770 ± 130 (GSC-347). Fourteen miles to the east at Meaford, wood was found in clay beneath 8 feet of sand with surface elevation of 605 feet that is also considered related to a Nipissing phase lake. The wood was dated 6,300 ± 150 years (L-312).

Radiocarbon dates on basal organic materials from small lakes at the Nipissing level on Manitoulin Island (Lewis, 1968) prove that a Nipissing phase was near maximum about 5,500 years ago. Similar evidence from North Bay reported by Lewis indicates that the spillway ceased to function before 5,000 years B.P. It is thus evident that the Nipissing Great Lakes (605 feet a.s.l.) formerly considered to be in the order of 4,200 years B.P. are in fact a combination of lake phases reflecting first the use of three outlets and later the use of only the two southern outlets, the whole ranging from about 6,000 to 4,200 years B.P. The three-outlet phase may be termed Nipissing Great Lakes I (Fig. XII-16x) and the two-outlet phase Nipissing Great Lakes II.

Erosion of the southern outlets of Nipissing Great Lakes II lowered the water level until a segment of Port Huron spillway system became stabilized by a combination of bedrock and concentration of boulders at the same time that Chicago outlet was rock-controlled; the resulting halt in the lowering of the outlets initiated the Algoma phase lake at about 596 feet elevation (Hough, 1958). This event may have occurred about 4,000 years ago (GSC-301). As the northern parts of the lake basins were slowly uplifted, a lateral shift in the Port Huron channel into more easily eroded materials brought about a resumption of down-cutting. Uplift of the sill at Sault Ste. Marie gave rise to a separate lake in Superior basin known as the Sault stage (Farrand, 1961) possibly about 2,000 to 3,000 years ago. This was followed by a sub-Sault stage and finally Lake Superior at 602 feet elevation. In the meantime down-cutting at Port Huron lowered the Huron and Michigan basin waters to their present 580 feet elevation.

Northwestern Ontario

Retreat of the ice in northern Ontario beyond the Superior and Huron basins is dealt with by Zoltai (1961, 1963, 1965) and Boissonneau (1966, 1968). Prior to its retreat from Superior basin, the Laurentide Ice Sheet west of Superior had advanced southwest into Minnesota. As it receded northward into western Ontario it remained in contact with glacial Lake Agassiz over a wide front; the history of this lake is discussed separately. Fluctuations in the rate of ice flow resulted in construction of a number of noteworthy end moraines. The most southerly end moraine system in western Ontario trends southeastward from Lake of the Woods to Rainy Lake (Map 1253A). Following an ice-frontal retreat of about 75 miles to the northeast the Steeprock Moraine was constructed; this extends from near Steeprock Lake southeastward to International Boundary about 25 miles west of Lake Superior. Shortly thereafter the much more extensive Eagle-Finlayson Moraine was constructed; this may be traced for about 175 miles southeastward from 30 miles northwest of Kenora to within 30 miles of Thunder Bay, where it is truncated by an end moraine left by a Superior basin ice lobe. Following a major retreat in western Ontario, a re-advance of the ice front was responsible for the Hartman Moraine which Zoltai (1965) considers to be of Valders age. At its northwestern end it curves sharply to the north and then northeast until truncated by the younger Lac Seul Moraine. When the ice front was at the Hartman Moraine, major ice lobes were also active in Nipigon and Superior basins and built the correlative and adjoining Dog Lake and Marks Moraines respectively. The latter has been traced to a high bedrock area southwest of Thunder Bay and hence the 'Valders' ice border may cross the International Boundary only a few miles west of the Superior shore.

Some 10 to 20 miles northeast of Hartman Moraine, the ice sheet built the extensive Lac Seul Moraine, which curves around the western side of Lac Seul. To the southwest this moraine merges with the Kaiashk Interlobate Moraine (Zoltai, 1965), which was built in part between the ice lobes that produced the Hartman and Dog Lake Moraines, but Zoltai has suggested that the northeastern end of the Kaiashk Moraine is a reconstructed end moraine of the Lac Seul lobe. A retreat followed by a re-advance of the ice front, of at least 20 miles, is indicated by overridden varved clays in Lac Seul area (Zoltai, 1965). Lac Seul Moraine extends northwestward into Red Lake–Lansdowne House area (Prest, 1963). On the southwestern side of Trout Lake it rises sharply some 270 feet but on the distal side it slopes more gently to the southwest and displays many excellent beaches and terraces of glacial Lake Agassiz. The moraine was produced beneath this lake and only three small parts remained above lake level. To the northwest the moraine is lower and less distinct but a series of large, wave-modified ridges of sand and gravel, unlike adjacent reworked eskers, may be morainal and indicate the former position of the ice lobe. The end moraine appears to loop northward and then eastward towards Windigo Lakes where it is truncated by end moraines of the minor Windigo ice lobe. Ice-flow features and eskers also indicate the convex form of the Lac Seul ice front. The westward lobation of the ice sheet was probably a result of active calving into the deep water of a Lake Agassiz embayment in the ice front (Fig. XII-16m).

Following a short period of recession in western Ontario, a halt in retreat or minor re-advance of the ice sheet produced the Sioux Lookout Moraine on relatively high ground west of Lake Nipigon. To the northwest are sporadic De Geer moraines which delineate the ice-frontal positions during calving into Lake Agassiz. Following a retreat of at least 80 miles the northern part of the ice sheet halted to form the Whitewater Moraine, and an eastern lobe of the ice re-advanced to form the Nipigon Moraine along west side of Nipigon basin (Zoltai, 1965). The writer believes that only the northern part of this end moraine system was built at this time (Fig. XII-16p), the southern part being older.

Greater activity of a northern component of the receding ice sheet relative to an eastern component gave rise to a series of recessional end moraines in Windigo Lakes area in northwestern Ontario and to an interlobate moraine south of these lakes.

The next major moraine system left by the fluctuating though generally receding ice sheet in northwestern Ontario was the extensive and generally southeast-trending Agutua Moraine (Tyrrell, 1913; Prest, 1963). A 25-mile re-advance is indicated by subtill lake clays along Otoskwin River. The highest part of the morainal system is at a major bend in Albany River west of Miminiska Lakes where it rises 500 feet above the river. The Crescent Moraine system (Zoltai, 1965) north of Lake Nipigon is probably a correlative of Agutua Moraine.

After construction of Agutua Moraine the northern ice sheet receded some 20 miles to the northeast before halting long enough to leave several short segments of end moraine on-line with belts of De Geer moraines. The eastern ice sheet at this time appears to have thrust actively forward—the Miminiska ice advance—and overrode parts of the Agutua Moraine in the vicinity of Albany River, and left a complex of end moraine segments, interlobate moraine, eskers, and ice-flow features that serve to delineate its form. To the southwest, in the vicinity of Ogoki River, this same ice advance appears to have constructed the Nakina Moraines (Zoltai, 1965).

The northern lobe, following a further retreat of about 30 miles, halted to construct a moraine near Big Beaverhouse on Winisk River. This moraine trends northwestward from south of Wunnummin Lake to Little Sachigo Lake. The eastern lobe left little evidence of its ice-frontal position; it may have been responsible for short segments of moraine west of Lansdowne House, and may have extended south-southeastward through Ogoki Lake. A short distance eastward from this lake the ice-flow features are masked or destroyed by the younger Cochrane ice advance from the north (*see under* Northern Ontario and Western Quebec).

Glacial Lake Agassiz. Glacial Lake Agassiz occupied large areas in Minnesota, North Dakota, Saskatchewan, Manitoba, and Ontario as Laurentide Ice Sheet receded north of the Mississippi drainage divide. As the lake expanded northward its southern limit contracted and, although the total area covered by Lake Agassiz was more than 200,000 square miles, it was probably no more than about 80,000 miles in extent at any time (Elson, 1967).

The history of glacial Lake Agassiz is, in part, intricately related to the period of ice-sheet recession from northwestern Ontario previously described. During this recession the outlets were controlled by position of the ice front and by topography in northwestern Ontario. Many of the strandlines were formed as a result of the operation of eastern outlets that discharged into Nipigon basin and thence southward into Superior basin.

Though the existence of a former huge lake in the Red River basin in the United States was recognized as early as 1822 and named in honour of Louis Agassiz in 1879, the first comprehensive work was that of Upham (1895) and later that of Leverett and Sardeson (Leverett, 1932). Many have also contributed new information on the Canadian part of the lake basin, particularly Johnston (1946) and Elson (1957, 1967). Information on eastern outlets discharging into glacial Lake Kelvin in Nipigon basin and the extent of Lake Agassiz in northwestern Ontario has resulted from field studies by Zoltai (1961, 1963, 1965) and Prest (1963). Elson has established a multifold history of Lake Agassiz. He recognizes a series of progressively lower lake levels related to successive operation of southern, northwestern, eastern, and northern outlets, with some rises in lake level due to ice-marginal fluctuations in the spillway areas. He has related strandlines in the Red River basin to these various outlets. The writer has also attempted to restore the configuration of some Lake Agassiz phases relative to the eastern outlets using the rates of tilt as determined by Johnston, the spillway positions and elevations given by Zoltai, his own data on lake levels and ice-margin positions in northwestern Ontario, and recently available contoured topographic maps.

Lake Agassiz began to form as Laurentide ice receded northward into the headwaters of Red River basin along the Minnesota–Dakota boundary, and it expanded northward into Manitoba and northwestern Ontario bordering on both Keewatin and Labrador sectors of the ice sheet. Discharge was through Lake Traverse at the south end of the Red River system into Minnesota River of the Mississippi River system. Lake Agassiz was lowered some 80 to 90 feet as the outlet was eroded to a bedrock sill. The lowering was irregular, perhaps due to formation of boulder armaments along the spillway that were periodically removed by increase in rate of discharge. Four lake phases, the Herman, Norcross, Tintah, and Campbell, and numerous minor intermediate strandlines are recognized in southern Manitoba and northwestern Ontario (Figs. XII-16d–f).

Elson (1966, p. 92) has suggested that a lowering from and return to the Norcross level was related to a short period of eastward discharge into Lake Superior via a Dog River spillway route in northwestern Ontario (Figs. XII-16g, h). The writer believes that this early phase of discharge may account for the first low water phase of Lake Agassiz perhaps about 12,000 years ago (Elson, 1966, Fig. 6). Sandy peat from alluvial fill at about the Campbell level in Assiniboine River valley and associated with a deepening lake was dated at 12,400 ± 420 years (Y-165). When an ice advance plugged the Dog River outlet, discharge was returned to the south. Lake levels rose to the Norcross level due to in-filling of the Lake Traverse outlet route in the interim. As this sediment was removed by erosion Lake Agassiz was lowered again to the Campbell level, controlled by the bedrock sill (Figs. XII-16i, j). The ice front at this time was at the Hartman and Dog Lake Moraines which are considered to be of Valders age.

When the ice-sheet margin again receded, the Dog Lake spillway was not re-opened; eastward discharge was not possible until the ice front receded northeastward beyond Sioux Lookout, whereupon the lake began to discharge into Nipigon basin (Fig. XII-16k). This event occurred about 11,000 years ago. A re-advance of the ice front closed the eastern outlets, returned the lake level to the Campbell strandline, and built the Lac Seul Moraine (Figs. XII-16l, m). Retreat of the ice again uncovered outlets into Nipigon basin with resulting lowered lake levels (Fig. XII-16n). A re-advance of the ice to build the Sioux Lookout Moraine about 10,000 years ago again closed the eastern outlets and raised Lake Agassiz to the Campbell level for the last time (Fig. XII-16o). Thereafter glacier retreat resulted in operation of ever-lower eastern outlets; a succession of prominent beaches was built in parts of the Red River basin. As the lake surface was lowered, more and more ground was exposed west of Nipigon basin and spillways lengthened westward (Fig. XII-16p).

During operation of the lower eastern outlets the ice receded from the region northwest of Nipigon basin (Prest, 1963) and exposed a lower outlet route that led southeastward along the ice margin to the northern end of Nipigon basin. As there was a major retreat of the northern ice front prior to the Agutua advance, low ground may have allowed discharge from Lake Agassiz southeastward into glacial Lake Barlow–Ojibway; if so, the Agutua re-advance has obliterated all trace of these outlets. Some readjustments in the drainage routes no doubt took place as the ice front receded from Agutua Moraine, and the routes were deranged again as the ice advanced to the position of the Nakina Moraines. At this time Lake Agassiz was confluent with glacial Lake Nakina (Zoltai, 1965) and the outlet was into the northern end of the Nipigon basin (Fig. XII-16r).

The retreating ice sheet remained in contact with Lake Agassiz until it receded from the position of the Big Beaverhouse Moraine. Discharge may have remained to the southeast, possibly into Barlow–Ojibway, as the moraine was built, but then changed to northward, probably down Echoing River valley into Hudson Bay Lowland. The drop in lake level occasioned by this event severed Lake Agassiz from the ice-marginal lakes remaining along the ice front on the Ontario part of the Hudson Bay watershed (Fig. XII-16s). At a still later date Lake Agassiz discharge was diverted to the Nelson River system and as the last confining ice melted the glacial lake gave place to early phases of Lakes Winnipeg, Winnipegosis, and Manitoba.

Northern Ontario and Western Quebec

Glacial Lake Barlow–Ojibway. In northern Ontario, east and northeast of Lake Superior, glacier retreat was northward towards James Bay, and in the adjacent parts of western Quebec it was mainly northeastward towards central Quebec. Lake waters were ponded between the ice front and the height-of-land (Figs. XII-16m–q). The manner by which this great lake remained at high levels astride the Hudson Bay–St. Lawrence drainage divide area and discharged southward along the valley of Lake Timiskaming has long been a subject of controversy. The outlet was plugged by a moraine some 10 miles north of Temiscaming, Quebec, and it is likely that lake levels were stabilized as a result of boulder concentrations in the spillways. New gradients and lake levels probably resulted from periodic isostatic adjustment. Progressive shallowing of the lake is indicated by sand horizons in the deposits of varved clay and by a greater rate of tilt on the highest strandlines in the southern part of the basin relative to those in the northern part. The writer determined a rate of tilt from the outlet to Larder Lake of about 4 feet per mile, and from Larder Lake to Plamondin Hill of 2.1 feet per mile. Hughes (1965) states that the maximum tilt of short segments of lower strandlines varies from 2.1 to 3.8 feet per mile and that the maximum rate of tilt must be in excess of 3.8 feet. The location of the hinge-line has not been established but it is probably to the south of the outlet. It is possible that differential uplift of the northern part of the basin took place along a second hinge-line in the Larder Lake–Noranda area.

A plot of the maximum waterplane from Plamondin Hill to Larder Lake indicates a minimum depth of water, over the present drainage divide near Noranda, of about 300 feet. It is thus evident that a combination of uplift and erosion of the outlet near Temiscaming must have lowered the lake by this amount before glacial Lake Ojibway was confined to the north; it then discharged from a point 15 miles west of Noranda (Fig. XII-16r) southward into ancestral Lake Timiskaming. The drift plug at Temiscaming has been channelled to a depth of over 350 feet; a small part of this no doubt occurred in post-Barlow–Ojibway times.

Limited information on maximum strandlines east of Malartic, Quebec, indicates that lake levels had lowered appreciably before the Quebec highlands glacier receded from that part of the Lake Barlow–Ojibway basin. A puzzling matter concerning Lake Barlow–Ojibway is the sudden deepening of the lake, indicated by increase in varve thickness and decrease in secondary sand, in varve year 1528 when the ice front was in the vicinity of Cochrane (Hughes, 1965). Perhaps discharge had been westward into Superior basin for some time prior to this varve year, whereupon a re-advance of the ice front closed this outlet, deepened the lake, and returned discharged to Timiskaming channel. This re-advance may correlate with formation of either the Crescent or the Nakina Moraines.

During retreat of the ice front northward from the Hearst–Cochrane region a prolonged halt occurred in the vicinity of Fraserdale, 60 miles north of Cochrane. A major east-west kame moraine composed of silt, sand, and some gravel was deposited in contact with the ponded waters on its south side (Boissonneau, 1966). This moraine may correlate with interlobate moraine west of Hearst and north of Hornepayne. These deposits are mainly sand and gravel with an appreciable local content of limestone. All these moraines may correlate with the Nakina Moraines (Zoltai, 1965) farther northwest; indeed there are no other comparable pre-Cochrane moraines in northwestern Ontario.

The lake in front of the moraine near Fraserdale is believed to have discharged southward, via an outlet near

Noranda, into Lake Timiskaming basin. This drainage route is in accord with the Barlow–Ojibway isobase trends and an indicated lowering of lake levels by about 300 feet in Lake Abitibi region. When the ice retreated north of the kame moraine, however, discharge and complete drainage must have been towards Hudson Bay, either laterally or subglacially. Antevs (1925, p. 77), with regard to the immediate pre-Cochrane stage of Barlow–Ojibway, reported that discharge into Hudson Bay took place across thin remnant ice in James Bay. Later (1931) he thought that drainage around the ice front to the sea occurred in the vicinity of the mouths of Hayes and Nelson Rivers, Manitoba. The writer considers that the lake, here named glacial Lake Antevs (Fig. XII-16s), may have drained northwestward along a line of weakness between a large mass of semi-stagnant ice in northwestern Ontario and a major glacier in James Bay and southern Hudson Bay, and discharged into the sea in the easternmost part of Manitoba. It is assumed that the sea had entered Hudson Bay via Hudson Strait and then along the line of a southwest-trending bottom channel in Hudson Bay, and effectively severed the Keewatin and Labrador sectors of the Laurentide Ice Sheet at this time.

Cochrane phase of recession. The next major event in the recessional history of the Laurentide Ice Sheet is the emplacement of the Cochrane till in Hudson Bay Lowland. Ice from Hudson Bay evidently moved southward more or less along the flank of the Shield (Cochrane I); somewhat later there was a southwestward and southward thrust from James Bay region (Cochrane II). The Cochrane ice front is seldom marked by end moraine. As pointed out by Hughes (1956) "The term 'Cochrane moraine' . . . is misleading, and should be dropped in favour of 'Cochrane till,' for the Cochrane till is a stratigraphic unit recognizable over a considerable area." The Cochrane limit is merely a line beyond which Cochrane till has not been found. Beyond the till limit the Cochrane event is represented by a series of varved sediments (Connaught sequence) containing pebbles lithologically similar to those of the till, but contrasting sharply with pebbles in underlying varved sediments (Frederickhouse sequence) that comprise a large part of Barlow–Ojibway deposits. Hughes reports counting about 60 varves of Connaught sequence in individual exposures. Maximum varve thickness is attained at varve 25, and this is considered to correlate with emplacement of the Cochrane till farther north. The Connaught sequence also includes varved sediments deposited in a shallow remnant of glacial Lake Antevs during recession of the Cochrane ice.

The typical Cochrane till is a stone-poor, blue-grey clay till that rests on varved sediments. It takes on a very pale pinkish tint when weathered, and is a yellow-brown when oxidized. Being calcareous, it contrasts markedly with the older, sandy, and non-calcareous tills of the Timmins–Cochrane region (Hughes, 1965). The clayey character of the till is believed to be due mainly to the incorporation of lake clays as the Cochrane ice advanced southward.

Boissonneau (1966) refers to an ice-front retreat of several miles to the north of the kame moraine near Fraserdale on Abitibi River, prior to the advance that deposited a thin mantle of clayey till over parts of it. The writer has observed the Cochrane clay till overlying varved sediments near Coral Rapids (364 feet elevation). The Cochrane till has been recognized as far north as the mouth of Little Abitibi River at 200 feet elevation but in this area its clayey character is produced from older, dark grey clay and light buff limestones. The ice front, prior to the Cochrane advance, probably receded to the vicinity of Coral Rapids where it impinged on a land surface at a present elevation of under 400 feet a.s.l.

Recession of the ice sheet down the Hudson Bay watershed from Timmins at 1,029 feet elevation to Coral Rapids at 364 feet elevation, a distance of 110 miles, would seem to indicate a lowering of the ice-sheet surface by about 600 feet. As the most southerly point where Cochrane till has been recognized is 12 miles north of Timmins at an elevation of about 960 feet, this would seem to require a thickening of the ice sheet of some 500 to 600 feet during the Cochrane advance. Such a thickening of the ice sheet is improbable at this late stage of deglaciation especially when marine water of the Tyrrell Sea (Lee, 1960) flooded much of Hudson Bay Lowland shortly thereafter (I-GSC-14, 7875 ± 200 years). It is more probable that the emplacement of Cochrane till is somehow related to the incursion of the sea into Hudson Bay. The two main surges of ice, envisaged by the writer as responsible for the Cochrane ice-flow features and emplacement of the clayey till evident on Map 1253A, are shown in Figures XII-16t, u.

Northern Quebec and Labrador

The retreat of the last ice sheet in this region was mainly towards the higher terrain of the interior, and the pattern of ice-flow features and eskers is generally radial about an interior U-shaped area south and southwest of Ungava Bay within which the last remnants of ice melted. Glacial lakes were not common but a few were extensive and long-lived.

Recession from Labrador and Ungava coasts. At the last glacial maximum the ice sheet in northern Labrador extended major tongues through cols in the Torngat Mountains, spread laterally in the coastal areas, and calved into Labrador Sea a short distance off the present coast. The upper limit reached by these valley glaciers has been referred to as the Saglek level by J. T. Andrews and the event itself to the Saglek Glaciation (Andrews, 1963; Løken, 1962). The elevation of the trimline, which becomes lower to both the north and east, indicates the surface slope of the ice sheet which lay west of the Torngat Mountains. Farther south, in the coastal Kaumajet and Kiglapait Mountains which at one time were believed to have remained unglaciated, erratics on the higher peaks indicate ice flow from the west at some earlier glacial maximum. This was considered Wisconsin (Tomlinson, 1959) but, in the light of J. T. Andrews work, is not likely Classical Wisconsin. Various levels of ice movements have been noted in coastal Labrador by many workers but it is not always clear which observations are related to the last regional glaciation (Daly, 1902; Wheeler, 1958; Ives, 1958a, b, 1960a; Tomlinson, 1959; Andrews, 1963). All agree that cirque glaciation in the coastal mountains was minor, thus supporting the concepts of Tanner (1944). Cirque development followed the retreat of the ice sheet and probably also took place in the Little Ice Age. According to Tomlinson, many nunataks and a large strip of continental shelf appeared as the ice sheet retreated from the coastal area, prior to transgression by the sea.

Shells from Eclipse Channel near the northern end of the Torngat Mountains indicate that the sea encroached on the northern Labrador coast prior to 9,000 ± 200 years ago (L-642). The shells were at an elevation of 95 feet in a region where Løken has placed the marine limit at 185 feet. On the basis of pollen analysis, Wenner (1947) placed deglaciation of the central Labrador coast prior to 10,000 years B.P.

Torngat Mountains were mainly nunataks at the maximum of the last major glaciation (Ives, 1958a; 1960a, Fig. 3), though overridden during an earlier glaciation formerly considered maximum Wisconsin (Ives, 1957). As the Labrador ice thinned and the supply of inland ice was cut off at the cols, the trunk glaciers in the throughgoing mountain valleys stagnated. The mountains themselves supplied little ice; some cirque glaciers formed, but seldom cut through the lateral moraines of the trunk glaciers. Løken (1960) has reported some late, mountain ice caps in the Abloviak River drainage basin east of Ungava Bay and, on the nearby Labrador coast, has noted features around Ryans Bay indicative of a re-advance of valley glaciers late in the deglacial process. In Okak Bay area between the coastal Kaumajet and Kiglapait Mountains, Tomlinson (1959) also found evidence of a late stage re-advance that formed push moraines prior to the final waning of the ice sheet. Both there and farther south Wheeler (1958) and J. T. Andrews noted high-level lateral moraines formed when the ice front lay east of the present coast, and also prominent younger moraines believed to represent a late glacial re-advance. These eastward re-advances across the low, southern Torngat Mountains may indicate that in the region south of the Torngat the Labrador ice was less restricted than to the north and flowed readily towards the coast. During retreat it maintained an active ice front but, as a result of a rapid thinning, did not form moraines.

Glacial lakes. As the ice sheet thinned west of Torngat Mountains, the eastern outlet glaciers retreated and left numerous lateral and end moraines along the Labrador coast. Later the ice margin receded west of the drainage divide and meltwaters were ponded in two of the major stream valleys east of Ungava Bay (Ives, 1958b, 1960a; Løken, 1962). At the same time, farther south the ice sheet receded across the plateau between Torngat Mountains and George River basin. Discordance between the drumlinoid and esker trends may indicate that recession of the ice sheet was not along the same path as the flow during the earlier advances (Matthew, 1961).

In the George River drainage basin the meltwaters were ponded to form numerous high-level, ephemeral lakes, followed by the long-continued glacial Lake Naskaupi (Ives, 1960b). Several major stands of this lake are recognized. The remarkably well developed shorelines of glacial Lake Naskaupi II are related to discharge down Koraluk (formerly Kogaluk) River to Labrador Sea. Other discharge routes are via the Fraser, Harp, and Kanairiktok Rivers and via the George River–Lake Michikamau col but their respective lake phases have not been determined. At its maximum development, glacial Lake Naskaupi was about 200 miles long and 30 miles wide, occupying the entire George River basin except for the ice-dammed northern end. The correlative glacial Lake McLean (Ives, 1960b) formed in the upper part of Whale River basin immediately west of George River, and discharged at its southeastern end into Naskaupi II. It was about 70 miles long and 30 miles wide, and was confined on both northwestern and southwestern sides by the ice sheet. The strandlines of these lakes are tilted up towards the southwest and the Labradorean centre (Barnett and Peterson, 1964).

The development of the Naskaupi and McLean strandlines, which are cut in bedrock in places, required a long stand of each of the several lake phases. This requires rock-controlled sills at the outlets, and an active confining ice barrier in Ungava Bay (Ives, 1960b). The esker and ice-flow pattern south of Ungava Bay appear to be in accord with the concept of an Ungava Bay ice dome, but most evidence as to the sense of glacier movement south of the bay indicates northward rather than southward flow with recession of the ice sheet towards the interior (Matthew, 1961). Northward flow would appear to be a natural consequence of incursion of the sea into Ungava Bay. Within George River basin itself Matthew noted that certain ice-flow features indicate westward flow where only eastward flow might be expected because of the ice-dammed lakes. Thus there has been considerable controversy over the position of the ice divide. The writer considers that three periods of ice-divide or ice-dome migration are needed to satisfy the field data and the following glacial history is suggested: (i) an early north-trending ice divide must have been present within George River basin (Map 1253A) to account for westward ice-flow features within the basin; (ii) as Ungava Bay and the lowland to the south filled with ice the ice divide shifted westward; (iii) eastward ice flow was maintained in George River basin during the early stages of deglaciation but did not remove all evidence of the earlier westward movement; (iv) with the incursion of the sea into Ungava Bay a major rearrangement of the ice flowage resulted; (v) a short-lived ice divide probably then lay a short distance west of glacial Lakes Naskaupi and McLean, trending slightly east of north; (vi) ice flow was everywhere towards Ungava Bay, rapid southward recession of the ice divide followed, and the glacial lakes drained northward; (vii) the ice divide then lay south of Ungava Bay and no longer had a north-trending extension; subsequent changes in position took place in the interior as recession continued.

As the ice sheet thinned due to climatic amelioration, glacier retreat from northern New Quebec, west of Ungava Bay, followed as a natural consequence of the opening of Ungava Bay and Hudson Strait. Ice dispersion was outward towards the coast. As the ice sheet thinned and receded over the height-of-land near the coast, meltwater was ponded along the ice front southwest of Hudson Strait. The highest lake at about 1,800 feet elevation was ponded in the headwaters of Povungnituk River and drained eastward into Joy Bay (Ives, 1960a). As the ice margin receded southwestward down Povungnituk River valley it became concave to the northeast, and successive lower lake phases discharged northward to Hudson Strait and eastward to Ungava Bay (Map 1253A). Maximum lake levels were at about 1,500, 1,200, and 1,000 feet a.s.l. and several systems of deep spillway channels occur. Data available do not warrant conclusions regarding the direction or amount of postglacial uplift. Matthews (1962) outlined small lakes along the ice front south of Cape Wolstenholme. These presumably drained westward or southwestward and were somewhat younger than at least part of the abovementioned larger lake system.

Recession from Hudson Bay Coast. As Labrador ice receded inland from northern Hudson Bay it calved actively into the sea, and between Cape Wolstenholme and Port Harrison left a magnificent series of De Geer moraines; these preserve the lobate form of the ice front in most valleys along the coast. These coastal moraines may correlate with a similar belt of De Geer moraines well inland from east coast of James Bay. The lack of De Geer moraines in the coastal belt east of James Bay may indicate that the ice was too thick for their formation at the time the sea encroached on the region; east of Richmond Gulf, the near absence of De Geer ridges is more likely due to thin ice relative to the depth of the sea.

Recession to the interior. Inland from the area of marine overlap, ice recession left the usual pattern of eskers and

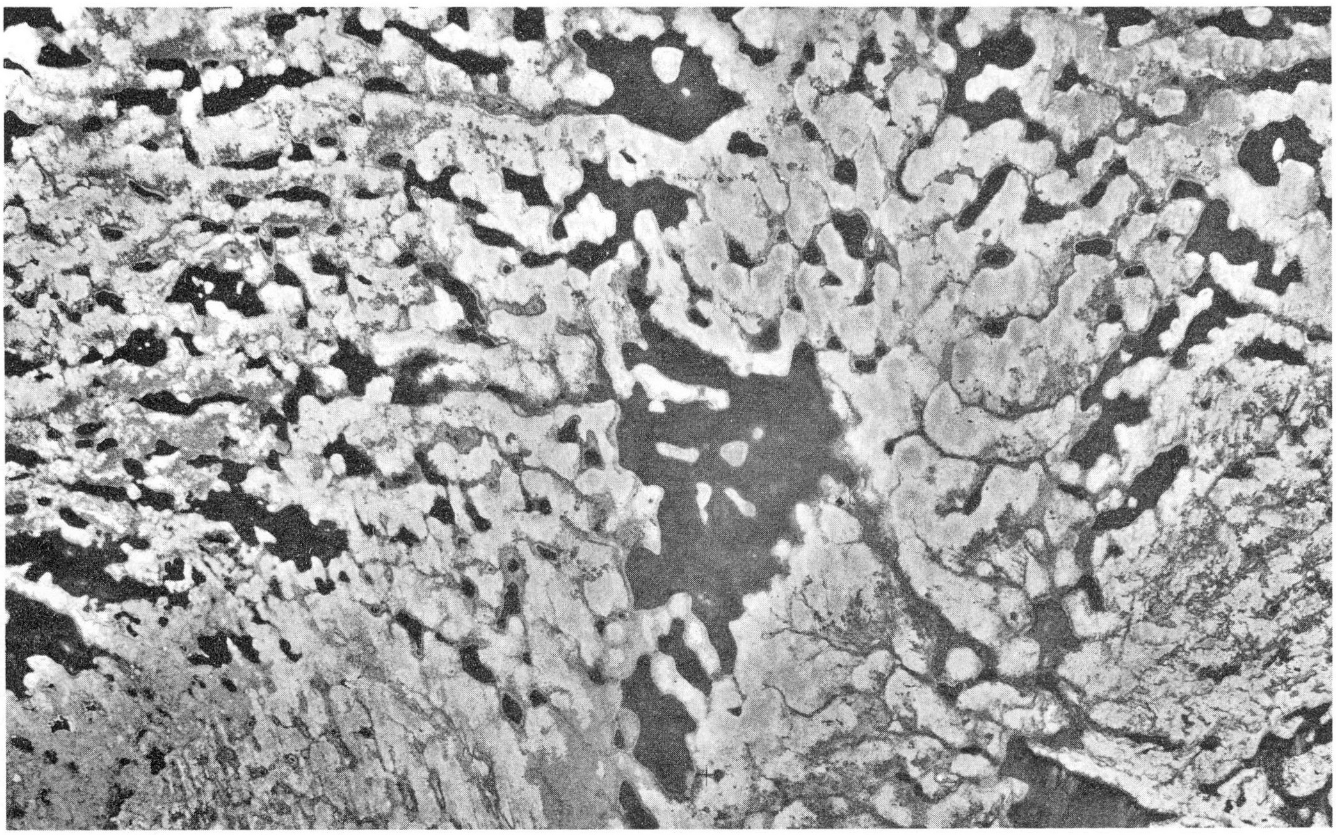


PLATE XII-8. Fluted ribbed moraine south of Ungava Bay, Quebec. Ice flowed northward. Photo suggests close temporal association between ribbing and fluting. Vertical airphoto; scale 1 inch to 1 mile.

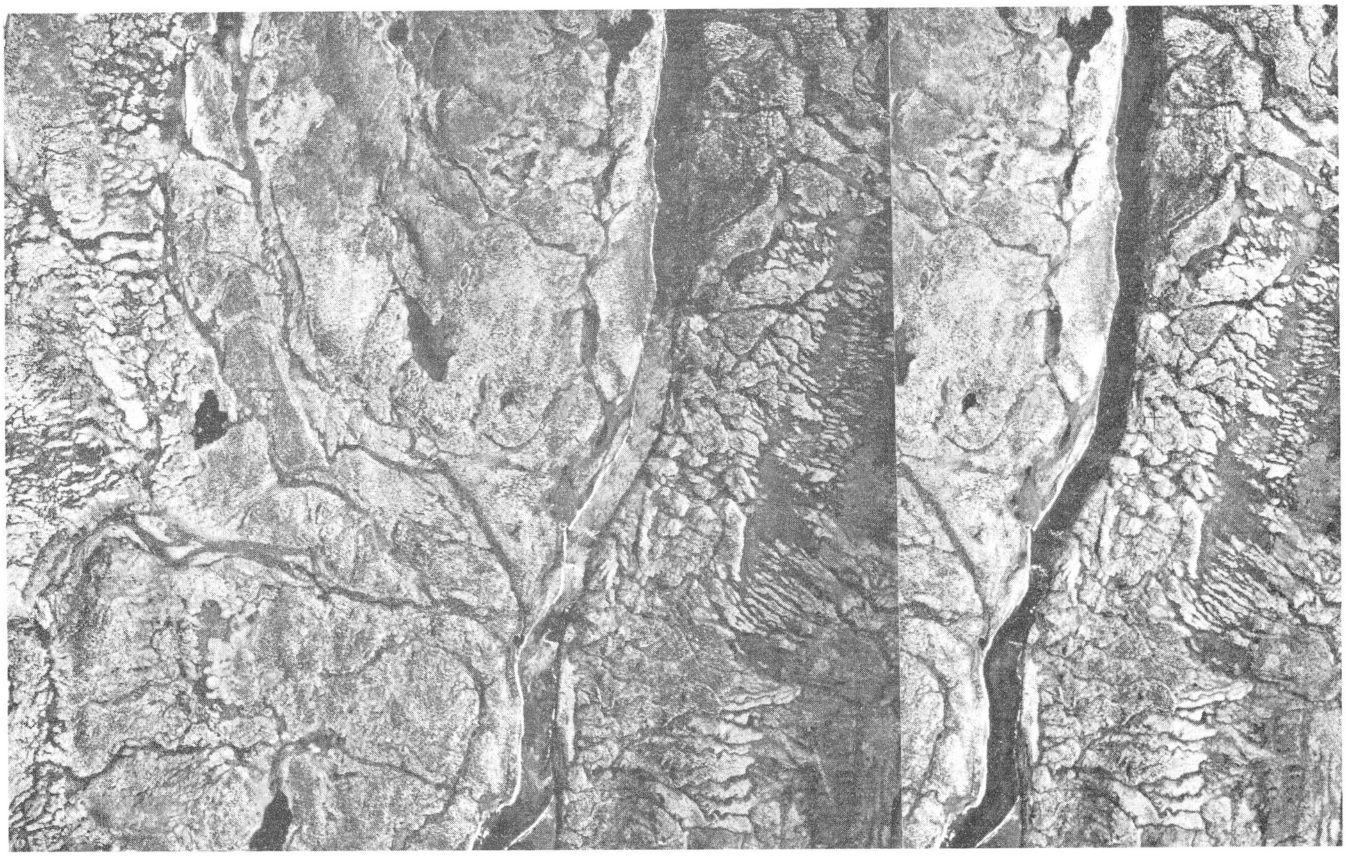


PLATE XII-9. Ice-marginal and subglacial meltwater channels, Mushalagan River, Central Quebec. The ice receded northward. Series of small meltwater channels formed subglacially or in much crevassed ice; the larger channels are considered ice-marginal channels. Stereoscopic pair; scale 1 inch to 3,330 feet.

ice-flow features characteristic of the Canadian Shield. Ribbed moraine is prevalent in many places associated with other ice-flow features and with eskers (Pl. XII-8). Final recession was probably to a number of scattered centres or ice divides in the interior of northern Quebec peripheral to Ungava Bay (Henderson, 1959b; Ives, 1960c; Derbyshire, 1962). One of the last remnant bodies of ice was in Howells valley west of Schefferville where there was much shifting of the late centres of outflow, with topography in control of the final ice movements. Late ice remnants were also present in Knob Lake depression and in the Lac le Fer–Swampy Lake depression. Subglacial and submarginal meltwater channels (Pl. XII-9) were the most effective forms of drainage in the final dispersal areas (Ives, 1959). Derbyshire believed that subglacial channels were first controlled by ice thickness and differential hydrostatic pressure, but as the remnant glaciers thinned they became much crevassed and waterlogged and the topography directly controlled the subglacial drainage.

From the pattern of ice-flow features and eskers (Map 1253A) it is clear that there was no major final dispersal centre, but rather a complex shifting of centres in response to changing climate. The Schefferville region, however, was a major centre of dispersal. This is indicated by the presence of various sets of striae in the region and by the directions of differential isostatic uplift determined from glacial lakes to the south, east, and northeast of this area (Barnett and Peterson, 1964).

A radiocarbon date on gyttja beneath peat in Chibougamau district, Quebec, indicates that pond deposits were formed as early as 6,960 ± 90 years B.P. Pollen extends through the uppermost 3 feet of silty clay of the glacial lake deposit. It is therefore likely that the ice front had receded from this area prior to about 7,500 years ago. Bog bottom dates from Grand Falls, Hamilton River (Morrison, 1963) and from near north end of Ashuanipi Lake, 100 miles to the west-southwest (Grayson, 1956) range from 5,250 ± 800 to 5,575 ± 250 years B.P. In all these sequences palynological evidence indicates a time gap between the base of the peat and the underlying glacial deposits. The dates indicate the initial development of widespread peat deposits and not the time of deglaciation of the area. This region was probably deglaciated prior to 7,000 years ago. A radiocarbon dating on a composite peat sample from near the base of a peat bog near Marymac Lake, 170 miles north-northwest of Schefferville, indicates that lake deposits had given place to peat prior to 6,400 ± 900 years ago. Grayson estimated the start of bog formation as closer to 8,000 than 7,000 years ago. Deglaciation of this area between 8,000 and 7,500 years ago is in keeping with incursion of the sea into Ungava Bay about 9,000 years B.P. and into Hudson Bay by 8,000 years B.P.

Keewatin Sector, Laurentide Ice Sheet

The Keewatin sector of the Laurentide Ice Sheet, or simply Keewatin ice sheet, is that part of the continental ice sheet which left a pattern of transverse morainal elements and longitudinal ice-flow features and eskers as it receded towards the Keewatin Ice Divide, an elongate area northwest of Hudson Bay. It covered the whole of western Canada east of the Cordillera, and an adjoining part of the north-central United States, except for a few small areas near its terminus near the International Boundary where the ice sheet was relatively thin. It also covered some of the arctic coastal islands and the southern side of Melville Island. At its maximum it was in contact with Cordilleran ice on the west, and Arctic, Baffin, and Labrador ice on the north and east. Tyrrell (1897) states that the name Keewatin is derived from Cree Indian meaning north or north wind and he used the term as applied to the ice sheet which he believed had its "gathering ground" northwest of Hudson Bay, in District of Keewatin, Northwest Territories. He was aware of the difficulty of establishing a glacier in this inland region of low altitude but nevertheless envisaged an active and shifting centre of outflow operating over a long period of time. The term Keewatin Ice Divide is now applied to a linear zone, almost 500 miles long, lying northwest of Hudson Bay, around which eskers and ice-flow features are arranged in a roughly radial pattern. It represents "the zone occupied by the last glacial remnants rather than a centre of ice dispersal" (Lee, *et al.*, 1957).

Keewatin Ice Divide is an area of few eskers and flow features as compared to adjacent areas (Lee, 1959). Those present indicate that the area of late ice dispersal migrated westward during the last stages of activity, the result of marine transgression into Hudson Bay. Thus, in many places, ice-flow features that were formed by ice moving in opposing directions merge against or into one another. The origin and initial development of Keewatin ice, however, remains conjectural.

Southern Interior Plains

The thickness of glacial drift on the Prairies varies greatly, probably averaging around 200 feet and locally is known to be more than 1,000 feet. Its general character is different from that of glacial drift in eastern Canada. The Prairies include great tracts of hummocky drift (Pls. XII-10, 11) that in places show transverse ridging (Prest, 1968). These features appear to reflect the clayey nature of the Prairie tills which have been derived from dominantly shale strata, and some of which contain much montmorillonite. Major end moraines per se are relatively uncommon and largely a matter of interpretation. Many of the great drift ridges formerly regarded as end moraines of southwestward-moving glaciers are now considered to be drift-mantled bedrock ridges, and for the most part mark marginal rather than end positions of lobate ice fronts. Furthermore, there has been much thrusting and folding of bedrock by glacier pressures, including induced pore-pressure (Slater, 1927; Byers, 1960; Kupsch, 1962) which have made it generally difficult and commonly impossible to differentiate between recessional drift ridges and ice-thrust ridges mantled with drift (Prest, 1968). Both types are included under the transverse ridge symbol on the glacial map (Map 1253A). Great blocks of bedrock, some perhaps measurable in miles, are underlain by till along Oldman River in southern Alberta (Stalker, 1963). L. A. Bayrock reports similar masses east of Edmonton. Sheets of bedrock resting on and capped by till have been encountered in drilling in Saskatchewan, according to E. A. Christiansen. Whitaker (1965) reports similar occurrences in southern Saskatchewan and also refers to very large masses of stratified drift that occur as

PLATE XII-10
Hummocky disintegration moraine near Cypress Hills, Saskatchewan.

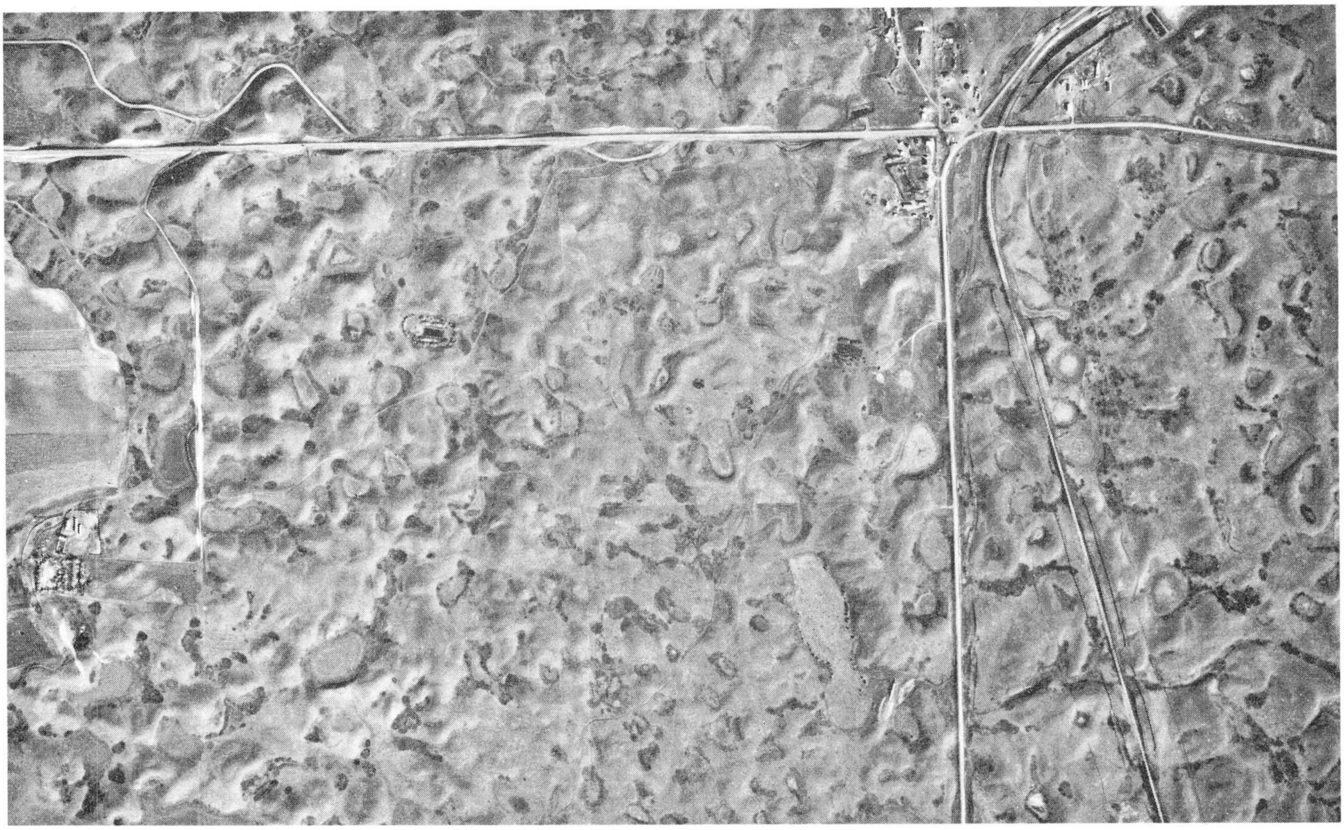


PLATE XII-11. Hummocky disintegration moraine east of Lake Johnstone, southern Saskatchewan. Occurs on western side of ice-thrust moraine of the Coteau moraine complex; local relief is about 80 feet. Town of Crestwynde in upper right corner. Vertical airphoto; scale 1 inch to 1,320 feet.

erratics and presumably were frozen when moved. The obvious thrusting action involved in the emplacement of ice-thrust ridges and discrete sheet-like erratics between tills, coupled with the sliding of debris in hummocky disintegration moraine and consequent inversions of topography, militate against easy development of a late glacial chronology for the Interior Plains. At present it is difficult to document an orderly general pattern of ice retreat despite the large amount of work done.

Ice-flow directions. On the southern Interior Plains the ice-flow and recessional moraine patterns indicate former glacial movements towards the southwest, south, and southeast. Local topographic features were instrumental in directing the late phases of glacier flow. It is also evident that the regional northwest trend of bedrock and the regional slope to the northeast have played important roles in deflecting the southwest flow from the Shield, towards the south and southeast. Southwest-trending ice-flow features in some relatively high areas on the Plains may be related to movements of a southwestward expanding ice sheet, but it is more likely that these features are related to south and southeastward moving glaciers which locally overtopped the valleys containing them. Shield stones indicative of a distant source are present along the length of the Rocky Mountain Foothills, but their provenance has not been determined. At the Wisconsin glacial maximum Keewatin ice was in contact with Cordilleran ice, or overlapped ground vacated by this ice, along the entire length of the Foothills in Alberta, except for a small area in the Porcupine Hills of southwestern Alberta (Douglas, 1950).

In Manitoba the ice-flow pattern indicates that regional glacier movements were to the southeast. Cretaceous rocks from the west side of Red River valley, near the International Boundary, were transported more than 400 miles across Paleozoic and Precambrian terrain and incorporated in till of the Grantsburg sublobe east of Minneapolis, according to H. E. Wright, Jr. (Univ. Minnesota). This indicates long-continued ice flow parallel to the presumed retreat flow pattern. Glacier retreat was mainly towards the northwest, but near Lake Winnipeg basin the influence of the Labrador ice has complicated the ice flow and the age relationships between various patterns have not been established. Along the Manitoba escarpment, the prominent hills and valleys brought about lobations of the late ice fronts and hence locally the ice receded towards the east and northeast.

In southern and central Saskatchewan the ice-flow patterns are exceptionally intricate and the order of late-glacial movements varies greatly from one area to another

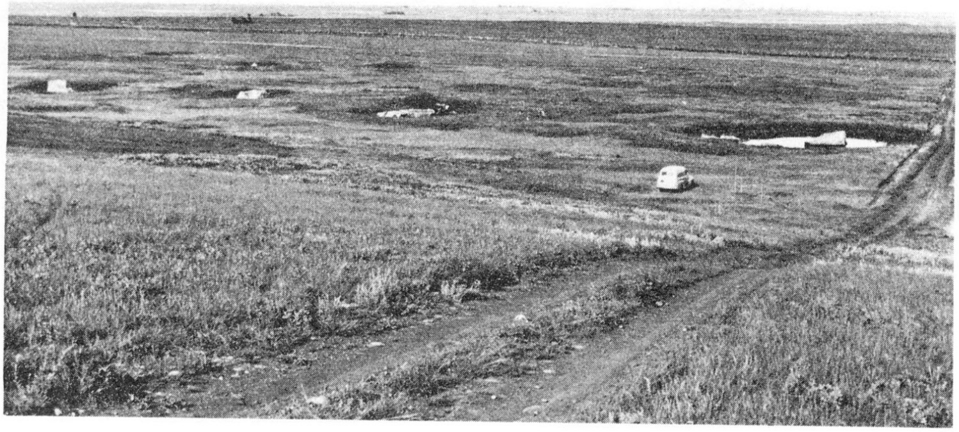


PLATE XII-12

Blocks of pebbly quartzite of erratics train from Athabasca Valley glacier, Claresholm, Alberta. The depressions around the boulders are mainly due to buffalo wallowing. The largest boulder, now broken in three pieces, measured about 50x13x4 feet.

(Parizek, 1964; Christiansen, 1965). There appears to be a regional recession of the ice sheet towards the north-northeast. The major ice-frontal positions, however, may only reflect thinning of the ice sheet and a lowering down the regional northeast slope while the ice flow remained essentially south and southeast along the slope. The longitudinal and transverse glacial lineaments appear to indicate this general trend but with many variations due to local topographic effects.

In southern Alberta the glacial features clearly indicate a predominant south to southeast trend. Keewatin ice must have flowed southward for an extended period for Cordilleran ice from Athabasca River valley merged with it and emplanted immense blocks of pebbly quartzite upon it; these were transported southward as far as Montana (Stalker, 1956; Mountjoy, 1958). These erratics were distributed along a path generally only a few miles wide and more than 400 miles long. They comprise the Foothills erratics train (Pl. XII-12). A. MacS. Stalker believes that this train of erratics represents the limit or near-limit of the Classical Wisconsin ice sheet. The group of quartzite blocks at Okotoks, Alberta, are considered to be parts of a former single block that weighed about 18,000 tons.

Regional ice recession was to the northwest parallel to the regional structure, with minor recessional shifts to the north and northeast occasioned by local high ground and by northeast-trending valleys. In northern Alberta the direction of ice recession gradually shifted to the northeast towards the Shield and hence the ice-marginal positions and moraines trend northwest. There were, however, many lobations and variations in glacier movements during the period of recession that are now revealed by a complex pattern of ice-flow features.

Glacial lakes. Meltwater was commonly ponded along the receding ice margins, but other than glacial Lake Agassiz, already discussed, the lakes were generally ephemeral with poorly marked shorelines. The bottom deposits are generally thin but in some areas are thick enough to completely mask hummocky terrain. Where thick lake sediments were deposited prior to the melting of buried ice blocks a hummocky terrain formed which closely resembles that of disintegration or dead-ice moraine.

In southern Saskatchewan, the ice front along the upper part of Qu'Appelle River ponded meltwater on its south side to form glacial Lake Regina (Johnston and Wickenden, 1930; Christiansen, 1961). This lake at the 1,900 to 1,950 feet level discharged southeastward via the Souris River system into Mississippi River. As ice recession continued into north-central Saskatchewan, the Qu'Appelle River valley served as a spillway for a series of lakes ponded in the Saskatchewan River valleys, discharging into glacial Lake Agassiz. These lakes received meltwater from the ice front and from glacial lakes as far northwest as the Rocky Mountains of east-central Alberta. As ice recession continued, the upper reaches of Assiniboine River in eastern Saskatchewan then served as a spillway into the Agassiz basin. The precise relationship of the various lake stages in Saskatchewan River valley with the glacial lakes in Alberta or with glacial Lake Agassiz have not been established.

In southern Alberta glacial lakes were numerous but small; meltwaters readily escaped to the southeast. Glacial lakes were more extensive in northern Alberta where the ice front plugged the major river valleys that trend northeast, and lateral discharge was difficult. Glacial Lake Edmonton occupied part of the North Saskatchewan River valley near Edmonton and at one stage extended into the Athabasca River valley (Gravenor and Bayrock, 1956; and Taylor, 1958). Deposits in the western end of the basin occur at altitudes up to 3,400 feet, but southeast of Edmonton lie below 2,500 feet. Discharge was to the southeast, at first into Red River and later into Battle River and the Qu'Appelle Valley to Lake Agassiz. A succession of lakes in Smoky River and Little Smoky River valleys west of Athabasca River are described by Henderson (1959a) and St. Onge (1966). The highest lake phase at 2,800 feet elevation is known as glacial Lake Rycroft. It extended west and northwest along the convex ice front into Peace River valley. Discharge was via the Pass Creek spillway from the southeastern end of the lake across a low divide at the head of Iosegun River into the Athabasca and presumably into a low stage of Lake Edmonton. St. Onge (1966) reports a radiocarbon date on shells pertaining to glacial Lake Rycroft of 12,190 $\pm$ 350 years (GSC-508).

By the time the ice front had receded so that discharge could take place into Lesser Slave Lake basin, the ice-marginal lake in the Smoky River valleys had lowered some 800 feet but remained in the Peace River valley and occupied a somewhat larger area than did Lake Rycroft. This phase is known as glacial Lake Fahler. It discharged eastward over a sill 10 miles west of High Prairie at an altitude well below 2,000 feet (Henderson, 1959a) into a lake in the Lesser Slave basin that was confluent with water ponded in the Athabasca River valley. These waters discharged across a sill east of Lac La Biche, at an altitude of 1,850 feet, and thence flowed via Saskatchewan River to glacial Lake Agassiz. The succession of lakes in Peace River valley, including Rycroft, Fahler, and lower levels were termed glacial Lake Peace by Taylor (1958) and Mathews (1963).

The Lake Fahler phase came to an end when the ice front had retreated about 150 miles down Peace River valley and thereby opened lower discharge outlets that led to the Arctic Ocean. According to Henderson (1959a), the Lake Fahler basin above the town of Peace River was drained prior to a re-advance of the ice sheet that returned the lake to its former level with discharge eastward to Lesser Slave Lake basin. Retreat of the ice sheet again resulted in a drainage reversal. Numerous spillways led westward from Peace River valley to Fort Nelson River in British Columbia, permitting northward discharge via Liard and Mackenzie Rivers. Glacial Lake Peace was lowered by a succession of outlets at about 1,800, 1,500, and 1,200 feet. Further ice retreat allowed discharge northward along Hay River into the Mackenzie. This lake phase would appear to correlate with a high level phase of glacial Lake Tyrrell (Taylor, 1958). As the ice sheet receded northeastward and eastward in northern Alberta, this lake expanded into the Lake Athabasca basin and the lower reaches of Athabasca River. When the Slave River valley became ice-free, glacial Lake Tyrrell merged with glacial Lake McConnell which discharged to Mackenzie River from the western end of Great Slave Lake (Craig, 1965a).

Northern Interior Plains and Mainland Arctic Plain

The character and thickness of drift in the northern Interior Plains vary greatly. Craig reports that the till, in the region west of Great Slave Lake, is silty to clayey and stony. Drumlinoid forms are locally very abundant; some of those west of Great Slave Lake are up to 100 feet high and display surface furrowing with a relief of 10 to 15 feet. Hummocky and kame moraines are limited but north of Great Bear Lake hummocky moraine is widespread and there are many end moraines. Mackay (1958) states that the till in Darnley Bay area is rich in silt and clay due to prevalence of carbonate rocks. Eskers, though not abundant, are dispersed widely over the northern plains.

Craig reports drift depths up to 380 feet in the area west of Great Slave Lake and up to 150 feet in river bank sections along Mackenzie River, parallel to the mountain front, but the bedrock is not exposed. Drill-holes east of Inuvik on lower Mackenzie River indicate a maximum drift thickness of 230 feet. The drift is probably very thick in the moraines north of Great Bear Lake. Ice-thrust features are common along the Arctic coast (Mackay, 1963).

At the climax of the last or Classical Wisconsin Glaciation, the Keewatin ice west of Great Slave Lake does not appear to have made contact with the Cordilleran Glacier Complex although it reached the mountains west of lower Liard River. There was, rather, a mainly unglaciated zone separating them along the mountain front that was 10 to 30 miles wide and almost 200 miles long extending north from the British Columbia boundary; this zone did, however, contain some valley glaciers. Another zone, slightly larger, occurs west of Great Bear Lake where the Keewatin ice flowed northwestward along the mountain front. These mainly unglaciated areas have been little studied but there is geomorphological evidence along the eastern mountain front that the last ice sheet did not reach as high as an earlier ice sheet. Keewatin ice extended as elongate lobes into some of the valleys along the eastern sides of the unglaciated areas and reached altitudes of above 4,500 feet. These two relatively unglaciated areas may be joined by a similar narrow strip or corridor to the large unglaciated region in northwestern Yukon and Alaska.

Ice-flow directions. West of Great Slave Lake the last ice sheet appears to have met the mountains at almost right angles, but farther north it was deflected parallel to the mountain front and Mackenzie River valley. It also moved parallel to the mountain front over most of the Plains east of Mackenzie River. As the ice sheet thinned and the ice front receded into the Great Slave Lake basin active flow there remained westward with the result that late ice-flow features cut across the trend of some earlier formed features parallel to the regional trend. This late ice was forced to flow around Horn Plateau west of Great Slave Lake and the ice front was, therefore, deeply indented. Elsewhere between Great Slave and Great Bear Lakes a complex pattern of ice-flow features discernible on airphotos suggests some differences in regional flow coupled with local lobations due to topography as deglaciation took place. With continued eastward recession towards the Shield the ice front straightened along the length of the Plains escarpment, some 20 to 30 miles west of the

Canadian Shield edge. Segments of end moraine and some minor moraine occur near Lac la Martre (Craig, 1965a). Craig suggests that this ice front extended southward to Great Slave Lake and beyond. Morainal features, discernible on airphotos of the region north of Lac la Martre, are suggestive of a continuation of the same 'straightened' ice front.

Keewatin ice did not extend beyond the present Mackenzie Delta except west of the river mouth. West of Peel River, the last ice sheet was not so extensive as an earlier ice sheet (Hughes, 1965), but it did extend west along Arctic Coastal Plain as far as Kay Point. Much information on the Pleistocene deposits and history of Mackenzie Delta area is given in a comprehensive report by Mackay (1963).

Between Mackenzie and Anderson Rivers ice-flow features trend north along the strike of the Paleozoic rocks but south of Eskimo Lakes trend northeastward; the reason for this deflection is not known. Mackay (1963) considers the morainic topography on the north side of Eskimo Lakes, in a belt 5 to 10 miles wide, as the terminal zone of a major ice sheet. He also noted smaller end moraines in the Campbell–Sitidgi Lakes area east of Inuvik, and also ice-flow features that trend north-northeast and northwest. He also suggested that the northeastern end of the Tuktoyaktuk Peninsula, which separates the Beaufort Sea from Eskimo Lakes and Liverpool Bay, might be unglaciated and that the limit of the last major ice sheet lay more or less along the middle of the peninsula. Fyles (1967a), however, suggests that the limit of the last ice sheet may have lain along the length of the south side of Tuktoyaktuk Peninsula, and that it may not have reached the northern part of Cape Bathurst.

In general, ice retreat north of Great Bear Lake was towards the southwest, parallel the strike of both Proterozoic and Paleozoic rocks. The complex pattern of ice-flow features, however, indicates that during the late stages of deglaciation the ice margin was highly lobate. Near the Arctic coast the late ice-frontal lobations were further complicated by Amundsen Gulf lobe (Mackay, 1958). The interplay of this ice lobe with that north of Great Bear Lake, and of local draw-down effects in vicinity of Great Bear Lake basin are described by Craig (1960).

Glacial lakes. As Keewatin ice receded from the northern Interior Plains meltwaters occupied the major depressions of Great Bear and Great Slave Lakes, and flowed northward down Mackenzie River. When the ice front lay along the edge of the Canadian Shield a vast lake extended from Great Bear Lake basin southward through Great Slave Lake basin to the lower Athabasca and Peace River valleys. Craig (1965a) has named this confluent lake phase glacial Lake McConnell. As isostatic uplift drained the lake to lower levels it separated into smaller lakes ancestral to the present Great Bear and Great Slave Lakes. In the latter lake basin strandlines are warped upward to the east. Craig suggests a minimum uplift rate of about 2 feet per mile in Great Slave Lake basin based on the presence of a delta at 500 feet elevation at Fort Simpson and of strandlines up to about 925 feet near the Canadian Shield edge. Great Bear Lake became separated from Great Slave Lake when the water levels fell below about 750 feet, the elevation of the present drainage divide between them.

Pingos and ground ice. Postglacial features of great interest in the lower part of Mackenzie Delta and in Eskimo Lakes–Liverpool Bay region are the pingos or conical ice-cored hills (Porsild, 1938). Mackay (1962, 1963) states that the pingos are elliptical to oval, 100 to 2,000 feet across, 10 to 150 feet high and generally asymmetrical. They occur in former lake basins or drainage channel depressions. The ice cores are believed to roughly conform to the shape of the pingo but perhaps have steeper slopes, and to bottom close to the surrounding ground level. The ice is overlain by a mantle of sand capped by vegetal matter, and is underlain by sand similar to that of the surrounding depression areas. The tops of the pingos are frequently breached or broken. Pingos with diameters of more than about 600 feet do not stand as high as the smaller ones, and those with the greatest base are merely bulge-like swellings in the lake basin sediments. Pingos result from a squeezing process as permafrost encroaches upon unfrozen saturated sands beneath what were formerly relatively large and deep lakes. They may exist for hundreds to thousands of years.

Great lenses of ground ice also constitute important elements of the landforms of this northern region according to Mackay. Fyles (1966) stresses the difficulties of stratigraphic analyses where such ice wedges are prevalent.

Arctic Lowlands

This part of the Arctic Archipelago lies between the mainland and the Queen Elizabeth Islands and includes the northern limit of Laurentide Ice Sheet. The drift generally reflects the variable character of the underlying Proterozoic and Paleozoic bedrocks.

Large morainal belts with fresh glacial landforms characterize an eastern part of Banks Island and are related to the Classical Wisconsin Glaciation (Fyles, 1962). Drift is varied and generally heaped into steep-sided hills but in places occurs as broad, smooth, hills and ridges, with intervening gentle depressions. Relief is several hundred feet. Marine beaches are locally well developed.

The last continental ice sheet reached the south coast of Melville Island perhaps as recently as about 11,000 years ago, and was responsible for the Winter Harbour Moraine (Fyles, 1967b). The moraine is composed of bouldery drift and stratified deposits and comprises a belt of hilly and ridged topography 50 miles long and averaging 2 miles wide. It is less extensive than a former glaciation.

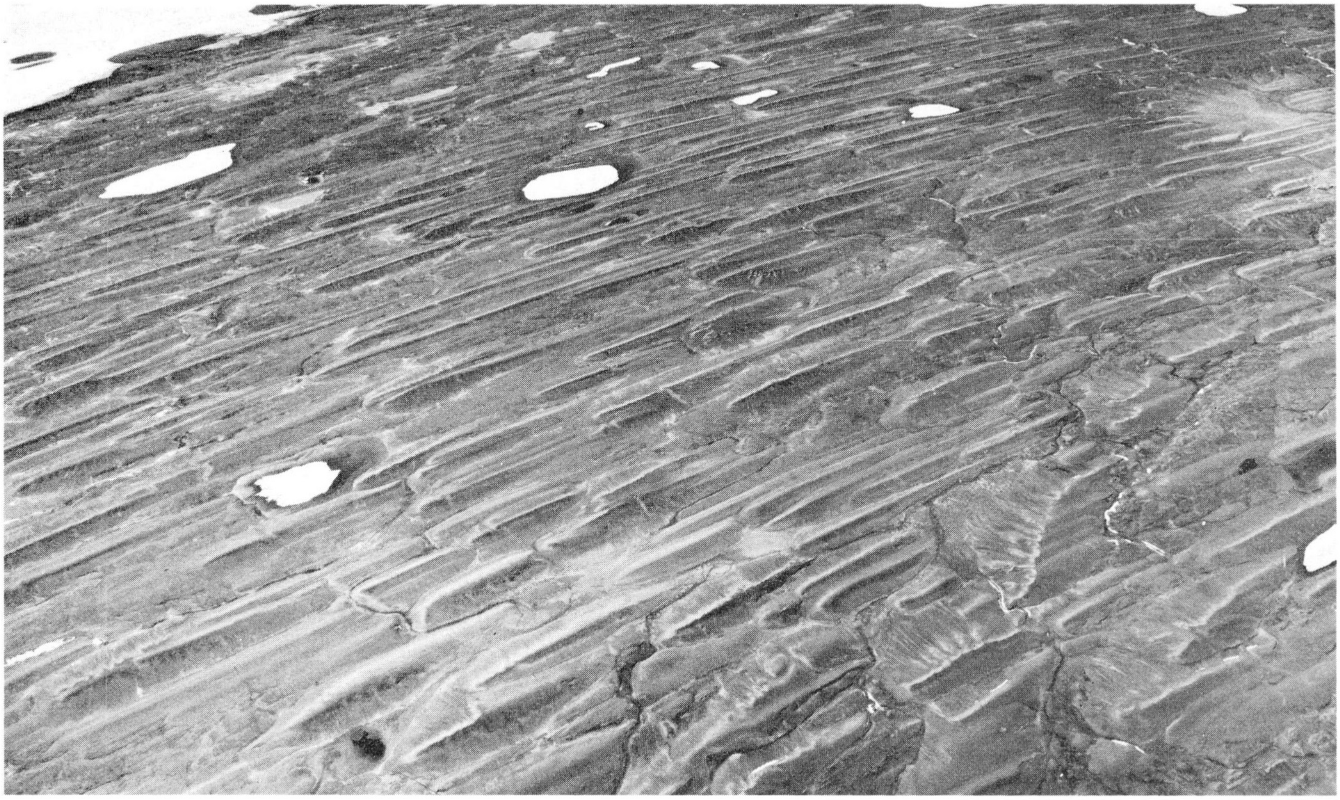


PLATE XII-13. Drumlin belt, Stefansson Island, Northwest Territories. View to the west. Ice flowed northward into Viscount Melville Sound.

Fyles (1963a) reports that the drift on Victoria Island characteristically has a light coloured matrix of slightly sticky, dense, loam or sandy loam. Numerous and exceedingly variable linear landforms are built of glacial debris or glacially deformed strata. These belts include landforms that are generally ascribed to end, kame, and hummocky moraine. The moraine ridges range from major features several hundred feet high, traceable more or less continuously for tens of miles, to miniature forms only a few feet high and a mile or so long. The complexities and interrelations of these great areas of hummocky and ridged moraine make it difficult to delineate end moraines per se. The drift is generally a few tens of feet to more than 100 feet thick and locally exceeds 500 feet. Drumlinized forms are characteristic of most lowland areas (Pl. XII-13) and there the drift is only a few tens of feet thick with bedrock protruding in some places. Eskers are numerous in the low-lying eastern and southern parts of Victoria Island, but they are few and short elsewhere; the longest is 120 miles. The eskers are usually less than 150 feet high but some knots at the junction of two or more esker-tributaries are several hundred feet high. Most consist of sand and gravel but some have material similar to that of the moraines.

An end moraine ridge and kame hills occur on northwestern Prince of Wales Island forming a linear zone of thick drift (Craig, 1964a). Eskers are rare on Prince of Wales and Somerset Islands and on Boothia Peninsula but are numerous on King William Island. They generally parallel the ice-flow direction, but on Prince of Wales Island some are transverse. Kames are striking features of central Boothia Peninsula (a Canadian Shield promontory within the Arctic Lowlands) and western Prince of Wales Island. They occur as broad zones that appear to be ice frontal and also as individual hills and groups of hills and ridges. They range from a few tens to about 200 feet high, and are rarely as much as 400 feet above the general ground level. Glacial landforms are uncommon on the northern parts of Prince of Wales Island and especially so on Somerset Island. These Paleozoic uplands are monotonously featureless but a thin layer of silty, rubbly glacial till occurs in many places, and some ground moraine topography, isolated kames, and meltwater channels are present.

Ice-flow directions. Ice-flow features and transverse morainal elements in the Arctic Lowlands are very complex, but nevertheless the maximum extent of the Wisconsin ice sheet can be delineated in a general way. The Amundsen Gulf ice lobe crossed Darnley Peninsula and reached the eastern side of Franklin Bay but does not appear to have crossed Cape Bathurst. It glaciated at least part of southern Banks Island and the east coast of Banks Island (Craig and Fyles, 1960). J. G. Fyles believes it may have extended westward along M'Clure Strait to the western end of Banks Island. Keewatin ice reached some distance inland on southern Melville Island, and farther east crossed Somerset Island and Boothia Peninsula (Fraser, 1957; Craig, 1964a). The ice sheet, at the glacial maximum, probably flowed northwesterly and northerly across Prince of Wales and Somerset Islands but later as it thinned and receded the deep channels between the islands influenced the ice flow. On southern Prince of Wales Island the last ice flow was eastward across the north-trending drumlin fields left by the earlier regional flow. A late ice lobe in Peel Sound may account for easterly trending striae on the west side of Somerset Island. Northwest- and northeast-trending drumlin fields occur on parts of King William Island. Craig (1964a) believes that the last ice movements over King William Island and Boothia Peninsula were northeastward.

On Victoria Island the regional flow was westward and northwestward but successive changes of late flow occurred as the ice sheet thinned (Fyles, 1963a). In places diverse drumlinoid trends occur side by side or are superimposed one upon the other. Flow directions are also controlled by relatively minor topographical features, suggesting that the ice sheet was lobate and thin. As the ice thinned, it withdrew from the higher parts of western Victoria Island, retreating to the southeast. Active ice tongues, fed by the main ice sheet to the south and east, remained in the depressions now occupied by the various arms of the sea. The moraines of western Victoria Island are lateral to these ice tongues. In the east-central lowland the ice probably flowed northwesterly in the early stages of wastage but with continued wastage, which took place progressively from west to east and north to south, the ice covering the lowland became thinner, and the direction of ice flow was deflected to an ever increasing degree by minor topographic irregularities. Glacier lobes in the low areas produced fan-shaped areas of ice-flow features trending northward and southwestward.

In the southeastern part of the island the morainal ridges are smaller, lower, and shorter, compared to the great morainal complexes of the western parts. This part of the island was the last to be deglaciated. Strongly flowing ice from the mainland constituting the receding Amundsen Gulf lobe passed down Bathurst Inlet and crossed Coronation Gulf, building a fan-shaped drumlin field across the southwestern end of the island. Apparently the last ice on the island covered the low ground bordering the east and southeast coasts while ice tongues still occupied M'Clintock Channel and Queen Maud Gulf (Fyles, 1963a).

Western Canadian Shield

In Canadian Shield areas west and north of Hudson Bay the character of the drift and the glacial features themselves are rather unlike those of the plains and lowlands. The till matrix is generally silty to sandy and lacks

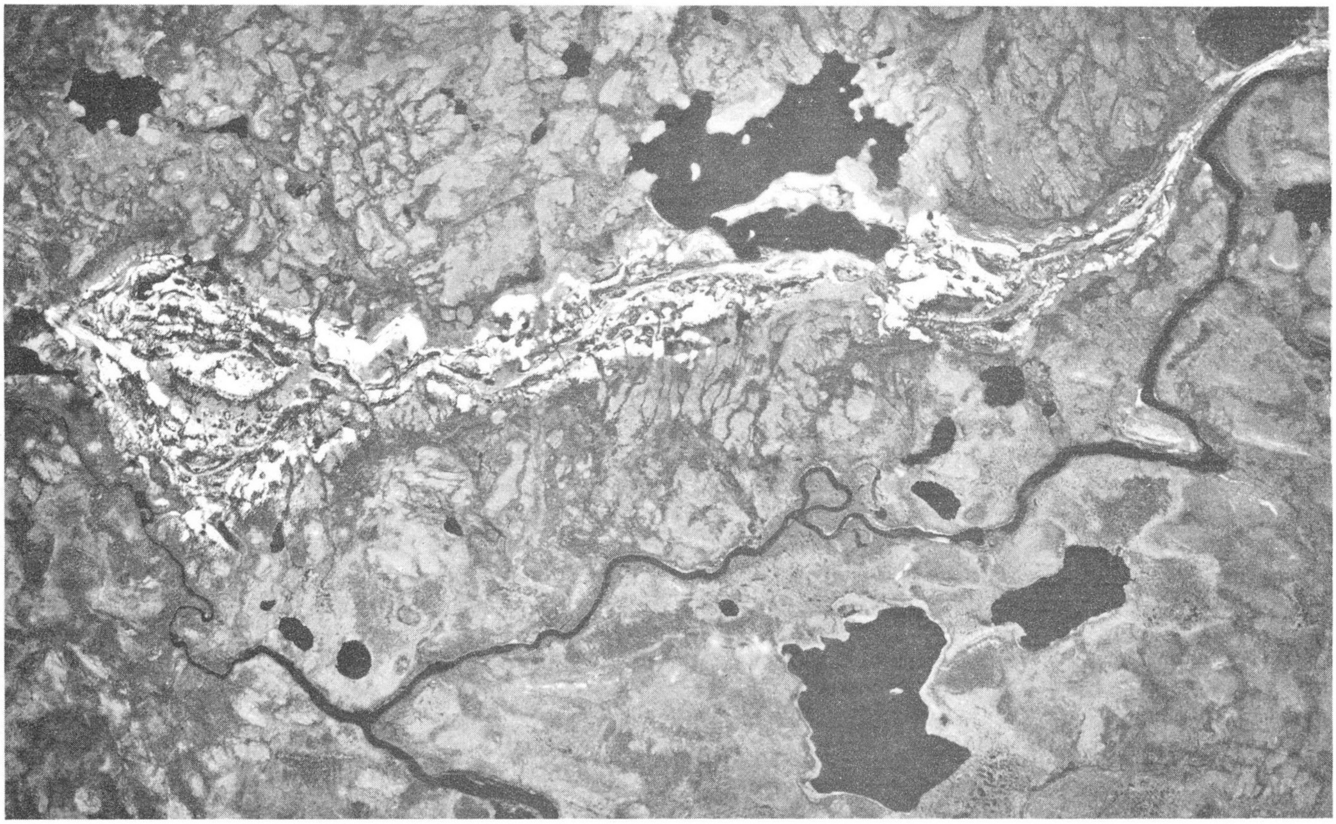


PLATE XII-14. Esker complex west of Ennadai Lake, Northwest Territories. Meltwater flowed southward. The large esker 'knot' is the south end of a system that is more or less continuous northeastward for about 60 miles to Kazan River. Vertical airphoto; scale 1 inch to 3,300 feet.

cohesive properties. The western part of the Shield is commonly mantled with drift in the form of ground moraine and ribbed moraine, which has a somewhat hummocky appearance but is unlike the hummocky terrain of the disintegration or dead-ice moraine of the plains. Eskers are widespread and numerous; they include the simple, sinuous ridges and the great complex multiple systems with widths of a mile or more (Pl. XII-14). The eskers and the remarkably continuous belts of drumlins, drumlinoid features, and glacial flutings delineate ice-flow directions during recession of Keewatin ice. End moraines are not common, but where present are clear linear features transverse to the ice-flow direction. Relief is commonly several tens of feet to about a hundred feet but locally kames, esker knots, and moraine knobs stand several hundred feet above the surrounding terrain.

Ice-flow directions. As Keewatin ice receded from the southern Interior Plains onto the Shield the ice front appears to have formed subparallel to the Shield edge and ice flow was roughly radial to this boundary. A halt in northern Saskatchewan is marked by the Cree Lake Moraine (Sproule, 1939; Tremblay, 1961a, b). It has been traced by means of airphotos northwestward to the western end of Lake Athabasca and eastward into Manitoba for about 500 miles. It varies from an end moraine-outwash complex to a single discrete ridge or a series of minor ridges. Thereafter, retreat towards Keewatin Ice Divide appears to have been uninterrupted by other major halts or re-advances.

In the Great Slave and Great Bear Lakes region recession was also continuous and no major end moraines were formed. The most significant result of this long period of retreat was the development of a marked discordance in the pattern of eskers and ice-flow features between two major ice lobes, a discontinuity that persisted as the ice sheet receded towards Keewatin Ice Divide (Fyles, 1955; Craig, 1957; Craig and Fyles, 1960; Blake, 1963; and Craig, 1964b). The discontinuity in the general radial pattern of flow features suggests that two distinct major glacier lobes formed during recession of the ice sheet. The break in flow pattern may readily be traced from the west side of Dubawnt Lake west-northwest to a point north of Aylmer Lake, and then curving northward past Contwoyto Lake to Coronation Gulf and beyond. To the north and east the flow features lie at a sharp angle to this 'break', whereas to the south and west they tend to be parallel or diverge only slightly from it. The line of demarcation is also revealed by a complex esker system between Dubawnt and Aylmer Lakes which lies a few miles south of the break. This esker continues for about 225 miles and except near the western end all tributary eskers lie only on the south side. Beyond Aylmer Lake a series of closely spaced eskers mark the continuation of the esker system towards Great Bear Lake (Craig, 1964b).

Craig and Fyles (1960) suggest that topographic control was responsible for developing the discontinuity. South and west of the break the ice was covering higher ground, moving slowly and stagnating locally, whereas the ice on the lower ground north of it was actively flowing. Blake (1963) also invokes a difference in activity between two lobes to account for the discontinuity northeast of Contwoyto Lake. Craig indicates that some northward-trending eskers north of it were overrun nearly at right angles by the last-formed lobe of the northern glacier. Southwest- and south-trending drumlinoid ridges west of Dubawnt Lake on the north side of the discontinuity are at right angles to the trend of ice-flow features on either side. These features are in part parallel to older striae recorded by Tyrrell (1897), but Craig considers them related to lobation within the northern part of the ice sheet as it receded towards Keewatin Ice Divide.

Ice retreat from the region north and east of the discontinuity was marked by the formation of end moraines. Two of the moraines, each about 25 miles long, lie east-northeast of Contwoyto Lake and trend slightly west of north. The westernmost moraine, with a maximum relief of nearly 400 feet, lies close to the discontinuity and apparently represents the terminal position of ice moving southwest (Blake, 1963). The eastern moraine is a recessional moraine. Several segments of east-west trending end moraine are evident southwest of Queen Maud Gulf. A prominent moraine extends northeastward from Back River, at the 106th meridian, to MacAlpine Lake for a total, including breaks, of more than 200 miles. This moraine varies from a few tens of feet to 4 miles wide, and from 10 to 250 feet high. The ice front apparently curved eastward to Chantrey Inlet (Blake, 1963; Craig, 1961) and was responsible for two more segments of end moraine some 60 and 25 miles long. Many segments of end moraine that may correlate with that west of Chantrey Inlet extend eastward to Committee Bay for another 225 miles. In this latter area, however, there are several discordant segments and correlation of ice-frontal positions is uncertain. Correlation with end moraines farther east, beyond the sphere of influence of Keewatin ice, is even more tenuous. A younger end moraine of the northern lobe of Keewatin ice also extends northeastward from Back River at the 106th meridian for a distance of 100 miles.

The position of the extension of the ice fronts south of the discontinuity are uncertain, but according to Craig (1964b, Fig. 7) there may be a significant jog eastward. Direct correlation of moraines across the discontinuity has yet to be established.

Glacial lakes. As Keewatin ice receded eastward across the Shield the east and northeast drainage systems became ice free and meltwaters were ponded to form several extensive glacial lakes (Map 1253A). A large lake occupied the Lake Athabasca basin and spilled northward over the rock sill in Slave River at an altitude of 700 feet (Craig, 1965a).

Somewhat earlier a glacial lake formed behind the Cree Lake Moraine in the Cree Lake and Pipestone (Cree)

River valley, the strandlines lying at about 1,800 feet. It may have drained southward through a break in the moraine and then discharged westward down the deep Clearwater River channel to Athabasca River valley. There is, however, an outlet channel from the southwestern end of the glacial lake basin, through which discharge took place into a lake in Richardson River valley at an elevation of about 1,400 feet; this was probably an eastern part of glacial Lake Tyrrell. As the ice front retreated northeast down Pipestone River the glacial lake in that valley discharged laterally to Lake Athabasca basin and successively lower levels were established. When the receding ice fronts cleared the eastern end of Lake Athabasca basin the two water systems merged at about 1,000 feet elevation. During this interval of retreat, a glacial lake formed in the Wollaston Lake basin at a maximum altitude of about 1,600 feet and discharged southeastward into Reindeer Lake basin. Later, lower outlets to Pipestone River valley were established. Reindeer Lake basin was occupied by meltwater when the ice front receded from the moraine at its south end. The outlet through the moraine was at about 1,200 feet a.s.l. and led directly into glacial Lake Agassiz lying south of the moraine.

Extensive glacial lakes also formed in Northwest Territories. Craig (1964b) found evidence of short-lived glacial lakes in the Artillery Lake–Lockhart River basin, east of Great Slave Lake. Beaches occur up to about 1,300 feet around Clinton and Holden Lakes. As the major southern lobe of Keewatin ice receded eastward from this area water was ponded in the headwaters of the northeast-sloping Thelon and Dubawnt Rivers drainage systems. The highest lake phases were about 1,250 feet and discharge was probably southward. Lower lake levels resulted as the ice receded eastward. Retreat of the major northern lobe of Keewatin ice allowed the water in the Thelon basin to discharge northward to the Back River system and lake levels were established at about 800 and 700 feet. At about this time Hyper–Dubawnt was at an altitude of 900 feet. The above lake phases are figured and described by Craig.

South of Dubawnt Lake, glacial Lake Kazan (Lee, 1959) occupied part of Kazan River and Ennadai, Kasba, and other lake basins. It drained eastward towards South Henik Lake and thence presumably into the sea in Hudson Bay. The strandlines around Ennadai Lake (1,070 feet elevation) are at an altitude of 1,260 feet (Pl. XII-15). Later, as the ice melted from Keewatin Ice Divide the lake discharged via Kazan River.

Marine overlap. At the same time as some of the above-mentioned lakes were present the sea was encroaching along the central-north coast (as early as 10,200 years

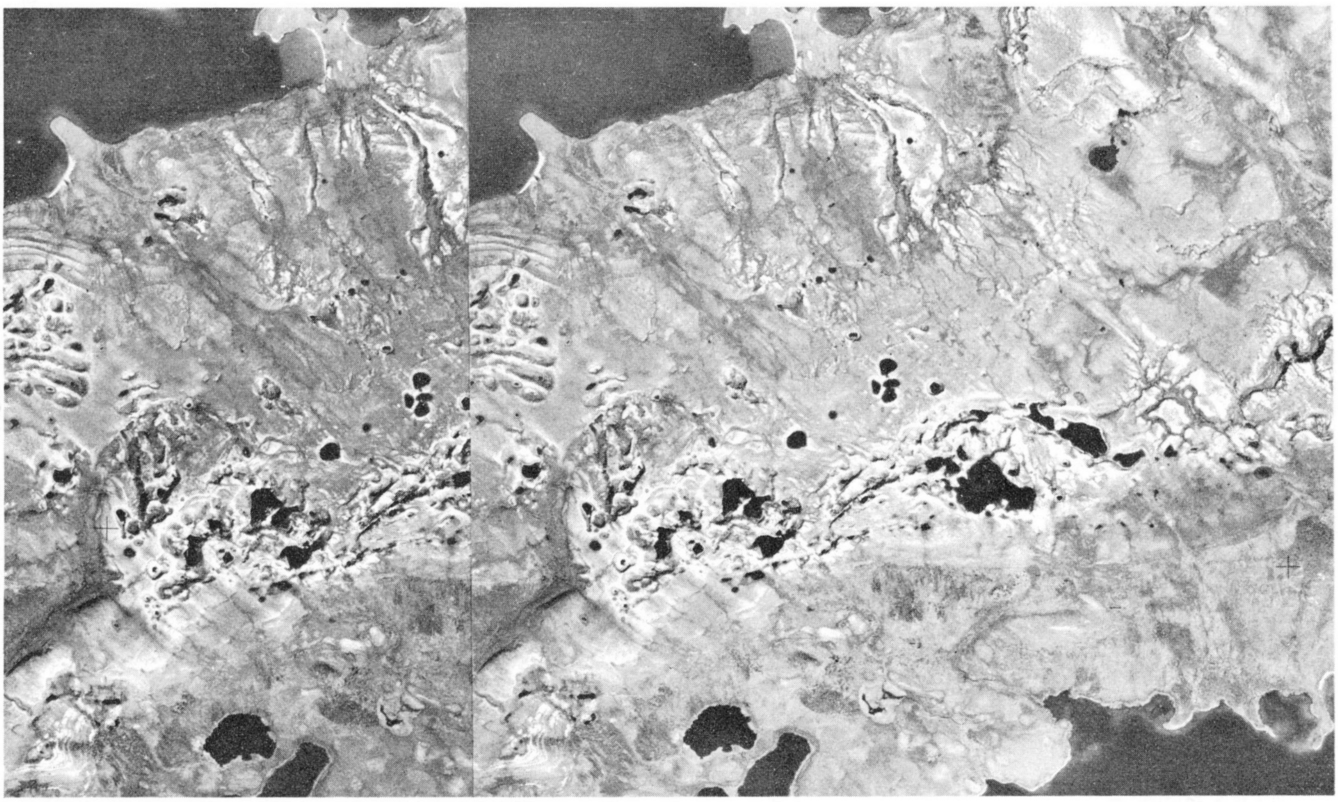


PLATE XII-15. Fluted and kettled ice-contact deposits with raised beaches and frost polygons, Ennadai Lake, Northwest Territories. Ice-contact deposits, 50 miles west of south end Keewatin Ice Divide, were fluted by an ice advance and later terraced and channelled by meltwater. Glacial Lake Kazan washed the lower slopes of the deposits. Some of the buried ice melted during and after the glacial lake phase. Stereoscopic pair; scale 1 inch to 3,100 feet.

B.P., Craig and Fyles, 1960). The sea was also in contact with a part of the ice front as the end moraine was formed at MacAlpine Lake only 8,200 years ago. By means of radiocarbon dating of shells from close to marine limit, Blake (1963) has estimated that the ice retreated about 175 miles from Kent Peninsula to MacAlpine Lake in about 1,000 years. Retreat towards Keewatin Ice Divide was also rapid, for the sea had encroached from Hudson Bay to near its maximum limit in the divide area by 7,200 years B.P. (Lee, 1960). It is clear also that isostatic rebound was rapid during this period of retreat for there is a general decrease in elevation of the marine limit with proximity to the divide. The rate of land emergence has decreased exponentially to the present (Lee, 1960; Craig, 1961). For the Bathurst Inlet area, Blake (1963) suggested an average rate of 1.5 feet per century over the past 2,000 years and probably less at present.

The marine limit west of Hudson Bay appears to decrease from south to north; it is about 650 feet in northern Manitoba, only 480 feet at Kaminuriak Lake, and 400 feet at Wager Bay. Along the Arctic coast the marine limit is greater, being over 650 feet west of Committee Bay and decreasing westward to 500 feet at Chantrey Inlet. It then increases westward to 750 feet at the southern end of Bathurst Inlet and drops off rapidly northwestward to 200 feet in Darnley Bay. The marine limit is not synchronous everywhere and it is probable that more than one centre of postglacial uplift has been active in this vast region.

Foxe–Baffin Glacier Complex

The glacial and deglacial history of Baffin Island is rather complex as a consequence of its physiography, latitude, and climate; it involved a major sector of Laurentide Ice Sheet, several local ice caps, and cirque and valley glaciers. The events in some parts of the present land areas were unlike those in others even where contiguous. On Meta Incognita (formerly Kingait) Peninsula, along the south coast of Frobisher Bay, Mercer (1956) found that unmodified cirques were common near the end of the peninsula. Along the central part of the bay the cirques showed evidence of later modification by ice from the interior, but towards the head of the bay they were absent entirely because that region was covered by interior ice prior to the interval of cirque glaciation. He also showed that cirque glaciers did not form anywhere along Frobisher Bay shore in late Wisconsin time as the interior ice receded. He considered that the interval of cirque formation was long and that it occurred during the Wisconsin. Bird (1963), however, favours development during an interglacial period. Some cirque glaciation, however, did take place in the mountainous terrain along the coast north of Cumberland Sound in late Wisconsin time. Thus despite sound field observation in several areas, any attempt to generalize for the entire island must be made with caution.

It is generally agreed that there was an ice dome over Foxe Basin at some stage during the last major glaciation but there are two views concerning its age: (1) that it existed at about the time of the last glacial maximum; and (2) that it only existed during a later stage of deglaciation. In either, the ice dome was concomitant with many local ice caps as well as cirque and valley glaciers along the eastern and northern highland areas (Ives and Andrews, 1963; Bird, 1963; Andrews and Sim, 1964; Falconer, *et al.*, 1965; Andrews, 1966). The relationships between the Foxe Basin ice dome and other sectors of Laurentide Ice Sheet are not well known. The writer considers that the ice dome was a dispersal centre for a substantial part of Classical Wisconsin time. It was probably confluent with Keewatin and Labrador sectors of Laurentide ice on the west, southwest, and south during a large part of the Wisconsin but maintained its own sphere of influence throughout. Ice from the local ice caps on Brodeur and Borden Peninsulas, together with huge valley glaciers flowing northward along Admiralty and Navy Board Inlets, probably merged with ice from Devon Ice Cap in Lancaster Sound at the last glacial maximum.

Glaciation

It was formerly held that initiation of glaciers on Baffin Island took place in the eastern mountain rim and that migration of the ice towards the interior resulted in piedmont glaciers that later expanded into ice caps and then merged to form an ice sheet. As the mountain rim is the eroded eastern edge of a westerly tilted upland it is unlikely that the mountain glaciers could contribute materially to the formation of the interior ice sheet; rather the ice probably accumulated on the larger central tract of the upland surface. As the westward slope of the upland is gentle and its valleys shallow, the ice must have flowed westward over a broad front, presumably as a plateau ice cap which gradually expanded and its margin moved progressively westward (Mercer, 1956; Ives and Andrews, 1963). Glacierization resulted from depression of the snowline below the level of the uplands as the Wisconsin climatic regimen changed. Rapid, large-scale development of the ice cap on the uplands starved the smaller glaciers in the coastal mountains so that the mountain ice caps and glaciers were never much more extensive than they are today. Perhaps the Penny Ice Cap is an exception. Penny Highlands are a remnant of a dissected older upland higher than the main Baffin surface (Bird, 1959). J. D. Ives believes that, because of its proximity to Labrador Sea and the moisture-laden winds moving up Davis Strait, Penny Highlands may have developed an extensive ice cap earlier than other parts of Baffin Island.

The early-formed interior Baffin Island ice caps presumably merged into an ice sheet which expanded differentially westward and fed ice into Foxe Basin. Glacierization of Melville Peninsula may have contributed eastward-flowing ice to Foxe Basin, and Laurentide ice may have flowed northward into the basin, though there is no evidence of these events. Laurentide ice, however, must have prevented or hindered early southward ice flow from Foxe Basin which therefore filled and later became a dispersal area (Ives and Andrews, 1963, Fig. 19). An ice dome over Foxe Basin is indicated by the tilting of marine strandlines in central Baffin Island towards the basin (Andrews, 1966). Analysis of carbonate content of drift in many places around Foxe Basin indicates outward flow from the basin (Andrews and Sim, 1964) though sporadic distribution of Precambrian crystalline limestone in central Baffin Island is a possible source of some carbonate. Blake (1966) did not find Paleozoic erratics or any increase in carbonate content of drift east of Amadjuak Lake in southeastern Baffin Island. There is, however, good evidence of early radial flow from Foxe Basin across the narrow central part of Baffin Island (Ives, 1962, 1964; Ives and Andrews, 1963; Andrews, 1963). There the ice sheet reached elevations of over 3,000 feet and escaped eastward through major valleys in the mountain rim and onto the continental foreland. Paleozoic erratics found north of Barnes Ice Cap were probably transported northeastward from Foxe Basin and survived through the later short period of outflow from proto-Barnes Ice Cap (Ives and Andrews, 1963, Fig. 22).

Foxe Basin ice also flowed eastward across Baffin Island through low ground to Cumberland Sound and probably southeastward to Frobisher Bay. It flowed southward across Foxe Peninsula and possibly also across part of Southampton Island (Bird, 1953). Strong and probably long-continued westward flow took place across Melville Peninsula where the bedrock is strongly scoured and *roches moutonnées* are common (Blackadar, 1958; Sim, 1960a; Craig, 1965b). There is no direct evidence of northward flow out of Foxe Basin but it may perhaps be inferred from evidence of northeast ice movements east of Steensby Inlet. It is of course possible that substantial ice caps on northern Baffin Island inhibited or prevented northward flow during the Wisconsin Glaciation.

At the glacial maximum some local ice caps occupied parts of the eastern and northern 'mountain' belt and also Brodeur and Borden Peninsulas, and may have prevented the incursion of interior ice to adjacent parts of the higher ground, and protected parts of the coastal foreland from the ravages of glaciers issuing along the fiords. Thus small parts of the eastern coastal lowland as well as some nunataks remained unglaciated at the last glacial maximum (Løken, 1966).

Deglaciation

Following the last glacial maximum, the Clyde phase of Ives and Andrews (1963), there was a period of glacier thinning and ice-marginal recession. This was followed by still-stands and short re-advances comprising the Cockburn phase of glaciation. The Cockburn Moraine System of end and lateral moraines marks the eastern ice-frontal positions of the interior ice sheet. In northeastern Baffin Island the outermost end moraines of the system lie on the high interfiord mountain flanks and descend into the fiords as lateral moraines; the inner moraines lie mainly on the upland plateau (Falconer, *et al.*, 1965). The ice was 1,500 to 2,000 feet thick in the heads of the fiords at this stage. Radiocarbon datings on shells from close to the marine limit established upon withdrawal of the ice from the coastal areas suggest that some of the eastern moraines formed about 8,200 years ago. The marine limit varies from about 220 feet elevation at some fiord heads to about 50 feet at the outer coast. The inner moraines on the plateau must be somewhat younger. Blake (1966) reported that a major end moraine in southeastern Baffin Island, believed a part of the Cockburn Moraine System, was forming for several hundred years after 8,200 years B.P. and suggested that a major end moraine on Foxe Peninsula was another correlative moraine. The Cockburn Moraines extend across northern Baffin Island and along the west side of Melville Peninsula (Falconer, *et al.*, 1965). Parts of this moraine in northern and northwestern Baffin Island may have formed somewhat earlier, perhaps as much as 9,000 years ago. The Cockburn Moraine System therefore appears to occupy a range in time from about 9,000 years to less than 8,000 years ago. This must have been a period of waning ice and consequent rapid change in ice-dome configuration. According to J. T. Andrews the east coast mountain glaciers were less extensive during the Cockburn phase of glaciation than they are at present. The Penny Ice Cap and several other major bodies were separated from the interior ice during this period.

Following the Cockburn phase of glaciation, the Foxe Basin ice dome underwent rapid disintegration and the sea invaded Foxe Basin about 7,500 to 7,000 years ago. The centre of ice dispersal was thus shifted northeastward onto Baffin Island causing ice flow towards Foxe Basin, whereas it had earlier been away from the basin. On the northeastern side of Baffin Island the ice margin receded west of the drainage divide and glacial lakes formed. Along the western side of central Baffin Island glaciers now occupied the fiords and inner bays. Marine overlap left strandlines ranging from about 350 feet near the coast to 100 feet farther inland, while glaciers in the valleys remained in contact with the sea. An end moraine in lower Isortoq River valley is associated with a sea level coincident with the marine limit (Andrews, 1966). Andrews equated the marine limit 'still-stand' with an increase in ice load on Baffin Island and a concomitant reduction in isostatic recovery. He named this period of thickening ice and moraine formation the Isortoq phase of glaciation. It would appear to correspond with the proto-Barnes Ice Cap phase of glaciation (Ives and Andrews, 1963, Fig. 22). Radiocarbon dates indicate that it occurred less than 7,000 but more than 5,500 years B.P. Andrews found that the marine limit was about 230 feet elevation some 15 miles from the mouth of Isortoq River and about 300 feet at the mouth, indicating a differential upwarp in direction S35°W towards the centre of Foxe Basin, at a rate of about 5.3 feet per mile.

The ice receded from Foxe Basin coast west of Nettilling and Amadjuak Lakes by 6,700 years B.P. and the marine limit was established west of these lakes at about 350 feet. Ice apparently remained in Amadjuak Lake basin during the Isortoq phase of glaciation; it was gone before 4,500 years B.P., and possibly as early as 5,500 B.P.

As the proto-Barnes Ice Cap shrank, glacial lakes remained along its northeastern side and continued to discharge into Baffin Bay. Glacial lakes were ponded in the middle Isortoq River valley, north and northwest of Barnes Ice Cap, during the period from about 5,000 to 2,000 years ago. As recently as 1,000 years ago a minor expansion of the ice cap blocked the river again and gave rise to another glacial lake, which drained in relatively recent times as the ice cap assumed its present position, size, and shape. The southwestern edge of the present ice cap with recent end and marginal moraines is shown in Plate XII-16. The east coast mountain glaciers are believed to have expanded in the last 2,000 years, reaching their maxima in the last few centuries (Harrison, 1964). Within the last 200 to 400 years a large part of the interior uplands was covered with thin ice and snowfields, but this has successively melted off until today only 2 per cent is covered, exclusive of Barnes Ice Cap (Ives, 1962). Falconer (1966) showed that in northern Baffin Island a small ice cap of about 3 square miles and never more than about 200 feet thick, developed as recently as 300 years ago in an area of patterned ground first deglaciated about 1,000 years ago. This small ice cap was inactive and served to protect the underlying polygonal structures and their vegetal cover—now emerging as the ice recedes. Deglaciation is still in progress and hence the term postglacial is inappropriate on Baffin Island.

In northernmost Baffin Island, Craig (1965b) reports that the glacial landforms and nearly all the surficial deposits are due to the last or Classical Wisconsin Glaciation. There is, however, a dearth of glacial landforms particularly on the Paleozoic rocks of Jones–Lancaster Plateau. Erratics are present throughout the region but are scarce on the northern part of central Brodeur Peninsula. Linear morainal zones with a hummocky surface and poorly defined higher ridges occur in a few places and although some mark terminal positions others afford little evidence of the position and movement of the ice that formed them. The most extensive linear zone is along south side of Bernier–Berlinguet Inlet, where it has been traced for 90 miles and in places rises 200 feet above adjacent land. The glacier appears to have moved westward and expanded radially. On the north side of the inlet there is also evidence of later ice flow from an ice cap to the north-northeast. Glaciers may have flowed eastward into Admiralty Inlet and northward along the inlet according to Craig. Also glaciers may have flowed westward off Bylot

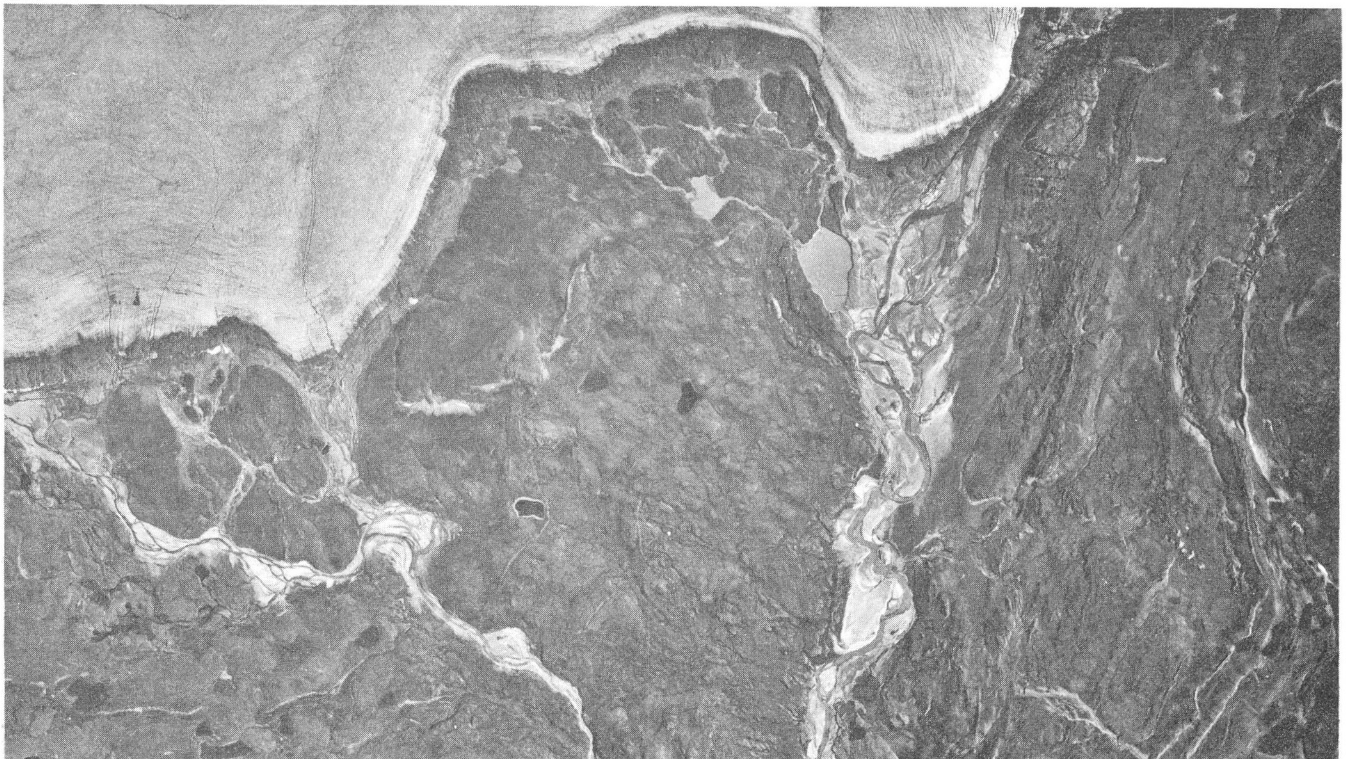


PLATE XII-16. Barnes Ice Cap with recent marginal and end moraines, Baffin Island, Northwest Territories. Vertical airphoto of upper reaches MacDonald River, southwest side Barnes Ice Cap. Scale 1 inch to 4,700 feet.

Island. Large areas of hummocky moraine are found in northern Baffin. These seldom afford any evidence of former ice positions or movement but along the west side of Brodeur Peninsula were probably formed by glaciers from the plateau. Radiocarbon dates on marine shells from raised beaches in northernmost Baffin Island indicate that maximum marine overlap occurred between about 9,000 and 7,000 years B.P.

As the above events were taking place on Baffin Island a somewhat comparable sequence of events occurred on Melville Peninsula (Sim, 1960; Ives and Andrews, 1963; Falconer, *et al.*, 1965; Craig, 1965b). Following the period of regional northwest ice movements over the peninsula, probably during both the Clyde and Cockburn phases, there was extensive thinning of the glacier complex and later incursion of the sea into western Foxe Basin. Sim reports flooding of the east coast of Melville Peninsula to about 500 feet a.s.l., and Craig reported marine overlap of the southeast coast to 485 feet. An ice cap probably occupied most of interior Melville Peninsula before breaking up into two or three local ice caps. Radial outflow was probably maintained by the larger southern body. In the centre of the peninsula a myriad of meltwater channels indicate a former small ice cap, and farther northeast eskers suggest a southeasterly flow of meltwater from another ice remnant (Sim, 1960; Craig, 1965b).

Deglacial events in northern Southampton Island, which may have been covered in part by Foxe–Baffin Glacier Complex, are little known. Bird, however, reports a late ice cap in the central part of northeast coast from which radial flow took place.

Queen Elizabeth Islands Glacier Complex

At present ice fields and valley glaciers are common features of the higher eastern part of the Queen Elizabeth Islands and small ice caps occur on Meighen and Melville Islands. Not all are remnants of the former more extensive ice complex; some have developed in recent times. Relatively fresh glacial features are generally lacking, possibly due to the inactive character of the cold arctic glaciers as compared to temperate-climate glaciers, but mainly to the character of the bedrock and to intense solifluction processes (Fyles and Craig, 1965). At its maximum development the glacier complex probably occupied the entire archipelago except for the coastal part of southern Melville Island which was covered by the Laurentide Ice Sheet. Active ice flow in the western part of the archipelago was largely restricted to the channels between the islands. Present data on ice-flow directions and on marine overlap suggest that the ice was thicker over the eastern and southern parts of the archipelago and that recession was towards these regions; perhaps the glacier complex preferentially developed adjacent to the Greenland and Laurentide Ice Sheets. Final recession and wastage may have taken place on the individual islands. Blake (1964) considers that some parts of Bathurst Island were deglaciated by 10,000 years B.P. and that extensive areas were ice free by 9,000 years B.P.

Prince Patrick Island. Glacial striations and other features at the head end of Mould Bay were formed by ice moving southward from the interior of the island (Tozer and Thorsteinsson, 1964). These features are attributed to the action of a small local ice cap, presumably of Wisconsin age, that developed subsequent to the establishment of the main drainage system. Mould Bay and the adjacent inlets were deepened and scoured. J. G. Fyles has noted boulders at the southwestern end of the island that were transported from the east and probably indicate a westward channelling of ice flow from Prince Patrick and Melville Islands. He also has noted that marine overlap is greater on the eastern side of the island, suggestive of thicker ice eastward.

Melville Island. Tozer and Thorsteinsson (1964) have concluded that the many fiord-like inlets in western Melville Island were probably originally stream-cut valleys that have been scoured by ice from a local ice cap. They also noted many glaciated valleys in interior western Melville Island. J. G. Fyles found erratics on both northern arms of the island that were far north of their outcrop area and presumably were deposited during the last regional glaciation. The present glaciers on western Melville Island are not remnants of the Wisconsin ice cap, for they occur in areas with youthful, little modified, postglacial ravines. The ice caps are probably less than 200 feet thick.

The maximum marine overlap is on the eastern side of the island, the highest raised shore features lying at 235 feet elevation (Henoch, 1964). Shells from the highest deposits were dated at 9,075 ± 275 years (I-730) and more from 175 feet of elevation were dated at 8,275 ± 320 years (I-GSC-21). The marine limit appears to decrease westward as no evidence of marine overlap was noted in western Melville Island. Henoch suggests that the depression of the east coast was produced by an ice cap centred near Sabine Peninsula, whereas depression of the south coast was caused by Laurentide ice.

Bathurst Island. The Queen Elizabeth Islands Glacier Complex occupied the whole of Bathurst Island and apparently restricted the northward expansion of Laurentide ice. Blake (1964, 1965) found abundant meltwater channels, patches of drift with ice-flow features, some dead-ice topography, rare striations, wave-modified esker and end moraine remnants, and other data indicative of former glacier cover. The ice may have been thicker over the northern end of the island than the southern as marine limit is about 450 feet in the north, only about 350 feet in the western and southwestern parts, and is rarely more than 300 feet along the east-central and southeast coasts. Recession appears to have been towards a remnant glacier in the interior of the island. Ice flowed northward off the island and may have reached Lougheed Island for

Fyles (1965) has noted erratics of rock types found on Bathurst Island. Marine overlap on Lougheed Island reaches an altitude of 300 feet. Shells from 25 to 100 feet above sea level were dated at 8,200 ± 180 years (I-GSC-24).

Cornwallis Island. According to Thorsteinsson (1958) the island was glaciated at least once. Striae at Resolute Bay on the south coast and Read Bay on the east coast indicated ice flow from an island ice cap. U-shaped valleys and other evidence of glacial scour in eastern coastal areas also indicated ice flow from the island. A series of rock basin lakes associated with *roches moutonnées* some 12 miles inland from Read Bay indicates northward ice flow. As glacial erratics of both igneous and metamorphic rocks occur in many places at elevations above the marine limit it is probable that Laurentide ice reached the island prior to the ice-cap stage but the age of this event is not known. Thorsteinsson places the postglacial marine limit at 425 feet elevation and notes that the well-defined beaches occur below 275 feet.

Devon Island. Ice caps on the highlands and plateaux of Devon Island cover about a quarter of its surface. Though the main ice cap has a maximum altitude of about 6,100 feet it is presumed to have a thickness of only 1,000 to 2,000 feet. It is passive except for a few distributaries in the eastern highland which flow in deep channels. Elsewhere the role of the ice caps appears to have been that of a protective cover and the smaller ice caps are known to be frozen at their base. Meltwater is carried by supraglacial streams. The physiography of the present unglaciated parts led Roots (1963) to conclude that the former, larger ice cap was also largely protective rather than an agent of erosion. He noted, however, that Griffin Inlet on the west coast was glacially sculptured by west-flowing ice and that north of there the ice margin was locally active, accounting for a series of lakes. Lakes in the west-central part of the island were also formed by west-flowing ice from the larger ice cap as it diverged southward to Maxwell Bay and north to the north shore of the island. Near the northwestern edge of the large ice cap a succession of stream channels and numerous ice-marginal meltwater channels indicate former melting but the ice near the present edge does not display these features. In the northwestern part of the island a more extensive ice cap than the present one was responsible for some 'hummocky' moraine south of Norwegian Bay and along the west coast south of Grinnell Peninsula, but the peninsula itself, like many of the islands, shows no evidence of having been glaciated. Glacial striae and crystalline erratics on Beechey Island, at the southwestern end of Devon Island, are believed due to northwestward ice flow. A granitic erratic found west of Jones Sound was also attributed to westward-flowing ice.

Brock, Borden, and Mackenzie King Islands. On Brock Island, Fyles (1965) observed an arcuate belt of ridges that he regarded as a thrust moraine formed by a glacier flowing westward through Wilkins Strait. No other glacial landforms were recognized on Brock or Borden Islands although a few erratics are present. The marine limit on Mackenzie King Island lies between 50 and 100 feet elevation and on Brock Island at about 50 feet elevation. It is unknown on Borden Island, but a beach crest at the west end is 10 feet above sea level.

Ellef Ringnes Island. Evidence of glaciation on Ellef Ringnes Island is rare, in large part due to solifluction and also to the nature of the bedrock. Crystalline erratics occur northwest or west of their outcrop and well above all evidence of marine overlap. Two gravel ridges trending west and one trending southwest are interpreted as eskers (St. Onge, 1965; Fyles, 1965). The marine limit is 150, 110, and 75 feet on the east-central, south, and west-central coasts of the island respectively. There is no information on the Pleistocene of Amund Ringnes Island.

Meighen Island. Fyles and Craig (1965) report the occurrence of numerous, large striated boulders on Meighen Island and the occurrence of striae, ascribed to westward-moving ice, on nearby Fay Islands. They believe that late Pleistocene glacial ice emanating from Axel Heiberg Island extended westward to cover these islands. The present ice cap has a summit elevation of just over 900 feet and a maximum thickness of about 400 feet (K. Arnold, 1965). Arnold has determined over the past ten years about 15 feet of thinning at the summit of the ice cap and at the northern edge, but notes that a slight deterioration of climate could cause an increase in the ice cap. He observed postglacial marine features to about 50 feet above sea level but places the marine limit at about 80 feet.

Axel Heiberg Island. On west-central Axel Heiberg Island meltwater appears to have been greater in early postglacial times than during the present arid climate (Müller, 1963). Because of the frozen ground and lack of vegetation, when run-off occurs now it is rapid, eroding channels and depositing valley alluvium and fans. The short valleys that occur between the glaciers and the sea are U-shaped only near the ice front and become V-shaped with valley trains and small fans in their lower ends (Rudberg, 1963). Some partly dissected trough-shaped valleys also occur in parts of southern Axel Heiberg not now occupied by glaciers. Abundant striae in the west-central part of the island indicate two periods of ice movement, the first from the north and east and the second from the southeast and east-southeast. Rudberg did not find undisputed evidence of far-travelled erratics in this area but red granite, presumably from Ellesmere Island, occurs on Schei Peninsula on the northeastern side of Axel Heiberg. There also, inland ice-flow features indicative of late ice movements trend northward. South of Schei Peninsula, J. G. Fyles observed red granite erratics that had been transported northwestward, roughly parallel to Eureka Sound; such movements represent former major trunk glaciers seeking escape from between Ellesmere and Axel Heiberg Islands. Elsewhere striae appear to record local, late, ice flow down small valleys and fiords as the glaciers receded inland towards the higher areas.

Two main stages of glaciation, the younger definitely Wisconsin (Boesch, 1963) and the older unknown, are recognized in west-central Axel Heiberg. As the Wisconsin glacier waned and receded onto the island the sea encroached on the uncovered ground, reaching altitudes now some 250 feet above sea level. Shells from the highest features have been dated at 9,000 ± 200 years (L-647F). Isostatic rebound was fairly rapid as the basal layer of a peat deposit lying at 80 feet elevation was dated at 4,210 ± 100 years (Bern). The Thompson glacier has not advanced more than 2 kilometres in the last 4,000 years according to Müller, and also the climate of the last 9,000 years never favoured a glacial advance of more than a few hundred feet beyond present positions. It is currently forming a push moraine of huge blocks of stratified material up to 60 feet thick. Organic matter in the lower part of some blocks has been dated at 6,200 ± 100 years (L-647A) and a piece of driftwood from a high part of the push moraine was dated at 5,325 ± 270 years (Gx-0144). This seems to indicate that formation of the push-moraine commenced after the hypsithermal period and that 5,300 years ago the sea occupied the area now covered by the snout of Thompson glacier. This implies that sea level was at least 50 feet higher than at present. Marine shells from a terrace 120 feet above sea level have been dated at 5,330 ± 195 years (Gx-0143).

Ellesmere Island. Ellesmere Island is by far the largest of the Queen Elizabeth Islands and supports the largest ice caps in Canada. In its mountains and plateaux the former ice cover probably formed a complex of coalescent valley glaciers and mountain ice caps, which conformed in a general way to the topography and were similar to the major glacier complexes on the island today (Fyles and Craig, 1965). In some places the distribution of erratics indicates a former ice cover that was thick enough to maintain a regional flow across the grain of the topography. Christie (1967) reports that boulders of granite and gneiss, probably from Greenland occur over a 10-to-25-mile-wide coastal belt of Hazen Plateau. The maximum westward extent of the Greenland ice is not known, as late Ellesmere ice has since advanced southeast and east across Hazen Plateau to the coast. At the glacial maximum the whole of northern Ellesmere Island was under ice cover and shelf ice probably extended well beyond the present north shore (Hattersley-Smith, 1961). The coastal ice caps on northern Ellesmere Island west of the northernmost tip, and disposition and shape of stream valleys and fiords which indicate former trunk glacier flow in western and northern parts, may be a result of westerly and northwesterly snow-bearing winds and high altitude. Taylor (1956) and Smith (1961) however, have suggested that the fewer fiords on Hazen Plateau may be partly a result of confluence of Greenland and Ellesmere ice.

The valleys of northern Ellesmere are regarded as normal preglacial valleys that have been straightened and deepened by strong glacier flow. Disraeli Fiord has a water depth in excess of 965 feet, and seismic data indicate a submarine canyon, trending northwest from the fiord mouth, about 2,800 feet deep (Crary, 1956). Despite this great depth the glacier in Disraeli Fiord appears to have overridden Ward Hunt Island, giving it a *roches moutonnées* shape and emplanting gabbro erratics from Ellesmere Island to an elevation of 1,100 feet (Hattersley-Smith, 1961). The present streams are deep and mostly V-shaped, but near their head, at the glacier margin, they are U-shaped, and tributary valleys are hanging relative to the trunk valleys.

As the ice receded from the north coast the sea inundated the land to a maximum altitude of 400 feet, as on Ward Hunt Island, but the marine limit varies greatly from place to place presumably due to the presence of late ice. Marine shells from 125 feet above sea level have been dated at 7,200 years. Cyclically-bedded silt, sand, and organic materials, the last as much as 4 inches thick, occur at elevations up to 70 feet at the head of Clements–Markham Inlet. These have been interpreted as deposits formed during the climatic optimum but J. G. Fyles reports that they also occur at various altitudes in many valleys and may have a considerable age range.

The character of the northern valleys has led all workers to conclude that the present ice cap is little changed since the climatic optimum. At present there are again extensive areas of shelf ice up to 50 feet thick and 12 miles wide, and also land ice near the coast. Hattersley-Smith (1961) considers these to be expressions of the colder climate following the climatic optimum (about 5,000 B.P.) whereas the interior ice fields are largely remnants (maximum of 2,700 feet thick) of the main Wisconsin glacier complex. He has recorded the advance of valley glaciers over raised beaches and into V-shaped valleys, and the growth of low ice caps on areas of raised beaches. He concludes that the climate has ameliorated in the past few decades and, although the main, high-level ice caps and outlet glaciers in northern Ellesmere Island show little or no change in areal extent from year to year, thinning is taking place in the lower reaches of the outlet glaciers and both thinning and recession of the small low-level ice masses have occurred. On the north coast of the island the surface wastage and occasional calving by the ice shelf reflects the warmer trend of recent decades.

In the Lake Hazen and Hazen Plateau regions Ellesmere ice sheet receded towards the United States Range (Christie, 1967). Directional ice-flow features and distribution of erratics indicate east and southeast ice flow across the plateau. Even though ice-flow features are present, the stream valleys show little other evidence of major trunk glaciers; truncated spurs, hanging valleys, and over-deepening are rare. A piedmont glacier must have formed on the plateau by coalescence of glaciers stemming from the United States Range. Lake Hazen basin was excavated along one contact of a wedge-shaped body of weak rocks overlying stronger rocks, and the lake is at least 864 feet deep.

In the western part of northern Ellesmere Island fiord-like valleys, glacial striae, and the distribution of erratics also show outward movement of trunk glaciers. Erratics at high elevations indicate regional flow across a region of ridges and valleys with a relief of more than 3,000 feet (Fyles and Craig, 1965). They report that during deglaciation the complex of glaciers and ice caps in western and central Ellesmere Island retreated progressively away from the outer coasts, up fiords and valleys, and into various high plateau and mountain areas. They also note that end moraines are not abundant and occur mostly in valleys within a few miles of an existing ice cap or glacier. The inferred period of shrinkage, on the basis of a few radiocarbon dates, is thought to be late Classical Wisconsin. An end moraine in central Ellesmere Island about 3 miles from the western margin of the ice cap was built more than 6,400 years ago, and another less than a mile from a modern glacier, more than 4,200 years ago; this indicates that the rate of recession over the past 6,000 years has been rather slow. Müller (1963) reports shells from an altitude of 350 feet at Eureka that dated 9,550 ± 250 years (L-647b).

Cordilleran Glacier Complex

The Cordilleran Glacier Complex is the system of intermontane, piedmont, and valley glaciers that developed in the Cordilleran physiographic province. Growth of the glacier complex stemmed from the passage of moisture-laden Pacific air masses over the various mountain systems.

PLATE XII-17. Klutlan Glacier, St. Elias Mountains, Yukon Territory. Note the marginal and medial moraines, drift-mantled terminal zone with small ponds, meandering meltwater channels on glacier surface, and termini of valley-side streams at glacier margin. Glacier flows east and north at head-end Generc River valley. Vertical airphoto; 1 inch to 5,200 feet; stereoscopic pair.

It is probable that extensive alpine glaciation preceded development of the major ice fields. Snowfall was, without doubt, heavier on the Coast Mountains than on mountains in their lee and it was also heavier on the western sides of all mountain ranges than on their respective eastern flanks. As glacial conditions continued, however, glaciers formed on the eastern flanks and later contributed to in-filling of interior valleys, plains, and plateaux.

Glaciation

Coast and St. Elias Mountains. The westward-moving glaciers of the Coast and St. Elias Mountains were largely dissipated in the Pacific Ocean. Farther south a glacier probably occupied much of Hecate Strait and extended to Queen Charlotte Islands. A major glacier from the southern part of the Coast Mountains occupied the Strait of Georgia, probably nourished in part by glaciers from the high central part of Vancouver Island. It moved southwestward across the southern part of the island where it scoured mountains to at least 5,100 feet (Fyles, 1963b), and southward to beyond Seattle, Washington (Armstrong, *et al.*, 1965). In Fraser Valley the last major glacial period is termed the Fraser Glaciation (Armstrong, *et al.*, 1965) and it probably represents the same geologic-climatic episode as the Classical Wisconsin Glaciation of the mid-continent region. The build-up and advance include a period of alpine glaciation, the Evans Creek Stade and an ice sheet glaciation, the Vashon Stade, when ice occupied the lowlands of southwestern British Columbia and adjoining parts of Washington. On Vancouver Island all the deposits relatable to the last regional glaciation are referred to as Vashon drift (Fyles, 1963b). The last regional ice sheet is believed to have entered the northern end of Strait of Georgia after 25,000 years B.P. but it did not reach southern Vancouver Island until after 19,000 years B.P.

Central interior. Except for those along the Pacific Coast, the glaciers from all parts of the mountain systems flowed into the intermontane areas and sought escape along the river valleys. Because of topographic confinement, the interior glaciers thickened until they covered all but the highest peaks of the Interior Plateau, a belt of relatively less precipitation. In central British Columbia the ice attained a thickness in excess of 6,000 feet according to H. W. Tipper, and may have thickened until its surface became a major area of accretion standing somewhat higher than the confining mountains. Ice flow was mainly eastward from the Coast Mountains across Nechako Plain to where it was deflected north and south along Rocky Mountain Trench. Ice flow at the Wisconsin maximum paralleled the trend of southern Cassiar Mountains, but locally the ice flowed across the ranges towards Liard Plain.

Southern interior. The first ice to enter Okanagan Valley probably stemmed from gathering grounds in Monashee Mountains to the north and east (Nasmith, 1962). The southern tributary valleys show little evidence of glaciation; apparently there was little tributary ice from the adjacent plateaux and highlands. The valley glacier was then joined by the major ice sheet from Coast and Cascade Mountains which spread south and southeast over Columbia Plateau and reached a terminus some 90 miles south of the International Boundary. At the boundary the ice sheet reached an altitude of 7,200 feet, and it was accordingly some 6,300 feet thick over Okanagan Valley. Meltwater flowed south down Columbia River to the Pacific. A great thickness of lake sediments resting on outwash has been drilled in Okanagan Valley and in one place the valley bottom lies below sea level. The valleys, therefore, probably attained their present form prior to the last glaciation.

Eastern region. The glaciers in the eastern part of southern Rocky Mountains were formed by local precipitation and were not in contact with the ice sheet west of the continental divide. However, north of Athabasca Valley eastward movement of ice across the divide did occur and some glaciers from the upper part of the Fraser River system flowed into Athabasca River valley. The southern limit of strong ice flow across the Rocky Mountains lies about 100 miles south of Peace River.

Piedmont glaciers undoubtedly were present locally south of Athabasca Valley, but most ice was contained in the valley glaciers that flowed eastward through the Foothills onto the Plains to where the ice either dissipated or merged with Keewatin ice. The time relations between the two glacier systems have not been established everywhere. South of Crowsnest Pass the valley glaciers only reached the Foothills and had receded into the mountain valleys before the Keewatin ice had spread that far west. Keewatin drift extends some miles up the vacated valleys overlapping the Cordilleran drift (Stalker, 1960, pp. 72-73). Farther north, the Cordilleran glaciers and Keewatin ice did not merge during the Wisconsin but were separated by Porcupine Hills, a narrow, elongate belt of relatively high hills (Douglas, 1950). In the Foothills west of Calgary, Cordilleran glaciers had also attained their maximum development and receded before Keewatin ice reached the area and overlapped some 5 to 20 miles onto the recently vacated ground. Farther north Cordilleran ice in Athabasca Valley merged with Keewatin ice and was deflected southward east of the Foothills as earlier mentioned.

In Peace River valley the Cordilleran ice also appears to have reached a limit some distance east of the later western limit of Keewatin ice. There, however, a re-advance of Cordilleran ice is recognized; a relatively late Cordilleran glacier from Rocky Mountains, reached to within 15 miles of Fort St. John after Keewatin ice had begun to recede (Mathews, 1963). North of Peace River the relationships of the Cordilleran and Keewatin glaciers are little known.

In the Liard Plateau and the adjacent plains Cordilleran and Keewatin ice made contact in only a few places. The Cordilleran glaciers were not very extensive in the Mackenzie Mountains nor along the eastern side of Selwyn Mountains as this was apparently a region of low precipitation. Glaciers from the eastern Selwyn Mountains and to a lesser extent from the western side of Mackenzie Mountains partly filled the intermontane area. Only elongate valley glaciers flowed eastward and northward towards the unglaciated or older glaciated terrain. Valley glaciers that developed on the dry, eastern side of Mackenzie Mountains were short and seldom reached low altitudes. In places they extended somewhat beyond the western limit of Keewatin ice.

Northern interior. The limit of the Wisconsin Cordilleran ice in western Yukon is shown on Map 1253A. The limit probably corresponds with the McConnell Moraine and glacial advance of Bostock (1966) in central Yukon Territory. This limit is marked by fresh ice-marginal landforms, particularly in Stewart River valley a few miles below Mayo. Glaciers moved northward from areas of heavy snowfall in the St. Elias, Coast, and Cassiar Mountains, westward from the Selwyn and Pelly Mountains towards the dry Yukon Plateau, reaching somewhat short of the limit reached by pre-Wisconsin ice sheets. Near the ice margin innumerable monadnocks protruded above the glacier surface. The Ogilvie Mountains, which separate the Yukon and Porcupine Plateaux, also nourished glaciers; these moved outward from several of the higher areas within the mountain complex but did not reach the plateaux. Meltwaters on the northern side of Ogilvie Mountains flowed northeastward via upper Peel River until they encountered Keewatin ice in the lower part of the river basin; then, joined by meltwater streams from the Keewatin ice, escaped northward along Eagle River to Porcupine River, Yukon River, and finally Bering Sea. As deglaciation proceeded and Keewatin ice vacated lower Peel River valley, meltwater from Ogilvie, Selwyn, and Mackenzie Mountains joined that from the Keewatin ice and followed Peel River to Mackenzie Delta and Beaufort Sea.

Deglaciation

West coast. The last regional glaciation (Vashon) on southern Vancouver Island is represented by a till and associated deposits (Fyles, 1963b). Both are generally sandy, especially in the lowlands where southwest-moving ice from Strait of Georgia overrode the sand deposits of the Quadra non-glacial interval. As the ice thinned, southwest flow was maintained for a time through mountain passes and valleys and received additions from glaciers stemming from high parts of the island. As recession continued island-derived ice became dominant. According to J. G. Fyles, ice retreat from Strait of Georgia was probably underway by 14,000 years B.P., and by 12,800 years B.P. the western side of the strait was open so that much of southern Vancouver Island was ice free. Along the mainland coast, thinning and marginal recession was prolonged by active accumulation in the high parts of Coast and Cascade Mountains. The last major ice lobe in Fraser Lowland was in contact with the sea until after 11,000 years B.P., when a major re-advance into the sea—the Sumas Stade—deposited a widespread layer of stony and locally shell-bearing marine clay varying from a few feet to over 500 feet thick. Deposition was probably from icebergs except near the ice margin where shelf ice was present (Armstrong, *et al.*, 1965). North along the Pacific Coast the ice progressively lingered longer in contact with the sea, probably to about 10,000 years B.P. Along the Alaskan coast the glaciers had receded to or near their present positions by 8,000 to 7,500 years B.P.

Southern and central interior. Deglaciation of the southern part of Cordilleran Glacier Complex was accomplished largely by downmelting and stagnation of the ice mass as a whole, with no clearly defined halts or re-advances (Nasmith, 1962). Lowering of the ice surface ultimately left the plateaux and highland areas bare, and when the remaining ice was confined to the valleys its surface area was greatly reduced. It seems likely therefore that withdrawal of the ice proceeded much more rapidly in the early stages of melting than in the later stages because the sharp reduction of the surface area may have been sufficient to bring ablation into equilibrium with ice accumulation. In Okanagan Valley lateral meltwater channels were cut at altitudes of 3,700 and 2,800 feet near the International Boundary. Later melting may have produced a nearly flat ice surface with meltwater flowing on top of the ice. The valley glacier was still thick enough to flow, however, and a northward ice-frontal retreat was maintained. Farther north up the valley the Okanagan ice lobe was more active and lateral meltwater channels were cut. A minor climatic change resulted in development of a few morainal ridges. Several glacial lakes formed in Okanagan Valley region. Glacial Lake Penticton was the largest lake and had the longest history. A beach of this lake is tilted upward to the north at a rate of 3.5 feet per mile. Volcanic ash in talus, alluvium, lake beds, and peat deposits covers a broad region and is correlated by Nasmith with the Glacier Peak eruption about 6,700 years B.P.

The last glaciation northwest of the Okanagan Valley is represented by a single till (Fulton, 1965). In many places, this rests on proglacial sediments that pass down into non-glacial sediments about 20,000 years old, suggesting correlation with the Fraser Glaciation of the Coast Mountains and Vashon Glaciation of Vancouver Island. According to Fulton, regional downwasting of the ice sheet followed, without oscillatory marginal fluctuations. This was the result of confinement of the ice to intermontane positions. In general, the southern margin first retreated northward and later west of north. As the ice sheet thinned it broke up into a number of stagnant or semi-stagnant lobes occupying the major valleys and lowlands. During the early phases of retreat meltwater was ponded in the lowlands and discharge was to the south. Upon further recession the meltwater flowed eastward into the Okanagan River system and thence via Columbia River to the Pacific Ocean. This route was used during deglaciation of most of the southern part of the Interior Plateau. Finally, however, recession of the ice sheet into Coast and Cascade Mountains opened the Fraser Valley and discharge took this more direct route to the Pacific (Mathews, 1944; Fulton, 1965).

Elsewhere in southern British Columbia less is known of the general pattern of deglaciation. The Cordilleran glaciers retreated to a number of mountain ranges where snowfall was heavy, maintaining for a time active glaciers in the major valleys well beyond their area of nourishment. As recession progressed, the valley glaciers retreated to within the confines of the surrounding mountainous belt and finally left the valleys entirely.

Relatively late ice is thought to have remained in northeastern Columbia Mountains, plugging the Rocky Mountain Trench in the vicinity of Big Bend, and in adjacent parts of Rocky Mountains. These mountains have glaciers at present, but whether they are remnants of the former sheet or were newly formed is unknown. A region of late ice is also postulated for a large part of southern Coast Mountains. H. W. Tipper (GSC) believes that as recession progressed the eastern flow from Coast Mountains lost contact with westward flow from Columbia Mountains. The latter, then directed largely by topography, turned north-northwest and south-southeast. The south-southwest trending ice divide between longitude 120 and 121 degrees (*see* Map 1253A) marks this line of separation. Most of the flow from Coast Mountains was northeast towards the Interior Plateau and upper Fraser Valley. Nechako Plain was covered by a broad, east to northeast flowing ice sheet from central Coast Mountains (Armstrong and Tipper, 1948). This part of the mountains at present has fewer glaciers than parts farther north and south. Glacial lakes were ponded in the valleys of Fraser, Nechako, and Stuart Rivers by active glaciers in the lower part of Fraser Valley; discharge was north along Parsnip River in Rocky Mountain Trench and down Peace River into lakes dammed by Keewatin ice in Alberta. It is thus possible that meltwater from glaciers originating in the Coast Mountains may at one time have flowed eastward via Peace River to the Interior Plains and thence via the Great Lakes system into the Atlantic Ocean.

Northern interior. Centres of late glacial activity persisted in the northern parts of Coast Mountains, and in the Cassiar and Skeena Mountains. In the Cassiar Mountains, reversals of ice-flow direction occurred during deglaciation, evident from the distribution of erratics and from ice-flow trends and ice-marginal meltwater channels (Watson and Mathews, 1944). Glaciers from the western side of Cassiar Mountains reached to the Pacific, but the relationships of these ice movements with those of ice in the Coast Mountains astride the outlet valleys have been little studied. During glacier recession, the Cassiar Mountains had an ice cap above 7,000 feet elevation from which glaciers flowed southwest and northeast. In the southern part, ice from the higher Coast and Skeena Mountains to the west passed through the Cassiar Mountains, along Dease River valley, flowing in a northeasterly direction and reaching to about 7,000 feet elevation. Alpine glaciation both preceded and followed the last major glaciation (Gabrielse, 1963).

In parts of the southwestern Yukon the last major glaciation, the Kluane (Denton and Stuiver, 1967), probably correlates with the McConnell advance in central Yukon Territory. Radiocarbon dates on organic materials from beneath Kluane drift indicate that the glaciation began about 30,000 years B.P. and persisted to at least 12,500 years and perhaps to less than 10,000 years ago. The Kluane glaciers, stemming from ice fields at altitudes of 12,000 to 15,000 feet on the eastern flank of St. Elias Mountains, moved northeastward through Kluane Ranges. It then flowed northwestward along Kluane River valley to Donjek and White Rivers, and northward to beyond Snag (Bostock, 1952; Krinsley, 1965). In the valley southeast of Kluane Lake the elevation of the surface of the glacier was 6,100 feet, near Donjek River it was about 5,000 feet and at its northern limit about 2,500 feet. The Kluane glaciers deposited three distinct tills which are associated with outwash and lake deposits. The glacier that deposited the uppermost till sheet in Kluane Valley had a minimum elevation of 2,500 to 3,200 feet. The glacier retreated along the paths of its earlier advance without any significant pauses or halts although large masses of ice stagnated in Kluane Lake area. Loess, derived from Kluane outwash, was deposited as a thin blanket over all older deposits up to an altitude of 4,500 feet. The Kluane Glaciation was followed by a period of soil formation (Denton, 1967). The soil was developed prior to the neoglaciation of St. Elias Mountains, about 2,800 to 2,600 years ago, which Denton correlates with the Little Ice Age. Following a relatively warm interval the glaciers were again active from about 600 years ago to the present century but are currently undergoing a fluctuating recession. The Donjek and Kaskawulsh Glaciers attained their neoglacial maxima about 300 years ago.

ECONOMIC CONSIDERATIONS

The surficial deposits or overburden that mantle the bedrock in most parts of Canada are mainly the result of glaciation. The direct glacier deposits or till, together with varied mixtures of ill-sorted or well-sorted materials, comprise the moraine that characterizes much of Canada's surface. Some large areas of gently rolling ground moraine in southern Canada have proved suitable for agricultural purposes but, in general, the irregular surface and variably stony materials limit or hinder large-scale farm operations. The more hilly areas of ground moraine, along with disintegration, interlobate, end, and kame moraines, therefore serve as ranchlands or forest sites, and are important also for water storage. Some of these morainal tracts with their myriad lakes provide game preserves, park lands, and recreational areas important to our tourist industry. Areas of glaciofluvial deposits are similarly important, particularly in regard to water supply.

Vast regions in southern Canada were formerly covered by glacial lakes and the resulting flat areas of fine-grained soils comprise the farm lands that supply most of the grains and vegetables essential to support the ever-expanding urban and industrial areas. In many parts of Canada glaciation was instrumental in providing better soils than we might otherwise have had, for glacial processes determined the mixing and size-sorting of materials that characterize some of our best farm lands. Elsewhere the glaciers probably removed excellent soils and transported them beyond our borders, leaving only bouldery or rocky terrain in their place.

Similarly, glaciation has played a vital role in our mineral and mining development. In parts of the Canadian Shield, Cordilleran, and Appalachian Regions the glaciers have stripped-off the weathered mantle of many orebodies, removing and dispersing the valuable secondary ones. Glaciation thus exposed the primary mineral deposits that foreshadowed the age of the prospector in Canada with consequent development of our mining industry. Elsewhere, orebodies perhaps once exposed by glacier action have been buried by younger glacial deposits and remain to be found by drilling and geophysical surveys. Others may be found through study of the surficial geology. Boulder tracing along the line of the last ice movements, as deduced from striae and other ice-flow features, is a simple technique that has been practised and proven successful on many occasions. More refined systems such as the glacio-focus method of localizing the source areas of minor constituents in drift (Lee, 1964, 1968) should prove successful also in the search for orebodies. A comprehensive knowledge of glacial events and glacier transportation will be even more necessary in future years as Pleistocene deposits are thoroughly examined for clues as to the loci of valuable minerals.

Glaciation has played a key role in the development of our towns, cities, and industrial areas. These are generally located in drift-covered parts of plains and lowlands, or in valleys in mountainous or rolling areas, rather than on adjacent more rocky and irregular-surfaced terrain. Because of the relative ease of excavating in unconsolidated materials as compared to bedrock there is still a tendency, despite modern heavy equipment, for urban expansion to avoid the areas of rock outcrop. This practice is engulfing the better arable lands at an alarming rate. Though some of our cities are in areas of little or no drift, most include areas of heavy drift. In some cases the bedrock is deeply buried and only the largest buildings are founded on bedrock, all others being anchored in the drift mantle. Where possible, spread-footings are located on till or gravel layers, and experience has shown that great care must be exercised to evaluate the physical properties of the drift mantle. The character and thickness of the drift mantle also have a direct bearing on the costs of water and sewer installations, and the construction of roads and sidewalks. Drift mantle may yet prove to be the deciding factor with regard to the type of rapid transit systems to be planned for many urban areas.

Sand and gravel are in great demand in all urban areas, and as a result supplies are being rapidly depleted within many miles and generally some tens of miles of these centres; whereupon, due to excessive transportation costs, crushed rock is used as a substitute. Most supplies of sand and gravel are from glaciofluvial deposits but locally river, lake, and seashore materials are employed. Despite competition from crushed rock, the use of sand and gravel for construction purposes alone remains at a high level. The 1965 production and value of sand and gravel used in road construction, railway ballast and in concrete, asphalt, and mortar mixes is given in Table XII-3. These materials are in large part of Pleistocene age, but some preglacial gravels are included.

A Pleistocene gravel deposit of unusual interest, which was not used for construction but rather as an iron ore, was that at the Canadian Charleston mine south of Steep Rock Lake, Ontario. There a gravelly end moraine included sufficient 'float' to be mined on a commercial basis as iron ore. Pebbles, cobbles, and boulders of high grade hematite and goethite were the main components of the gravel deposit. These were deposited from an ice lobe that had gouged them from beneath Steep Rock Lake. A high grade iron pellet concentrate was produced on the property. Although operations were confined to the summer months, and there was no production in 1961, some 642,957 tons of high grade iron concentrate was produced from about 4.5 million tons of gravel between 1959 and 1964 when operations ceased.

Clay deposits of both lacustrine and marine origin were used extensively in former years in the manufacture

TABLE XII-3 | *Production and value of sand and gravel (by F. E. Hanes, Mines Branch)*

Province	1965 short tons	1965 $
Newfoundland	4,063,734	3,684,891
Prince Edward Island	412,064	374,081
Nova Scotia	6,574,387	4,498,803
New Brunswick	4,491,514	2,594,846
Quebec	40,507,369	19,583,351
Ontario	75,082,026	55,297,474
Manitoba	9,757,104	6,767,068
Saskatchewan	8,570,008	5,615,794
Alberta	13,163,941	10,661,383
British Columbia	20,484,706	12,662,016
Totals	183,106,853	121,739,707

of brick, tile, and related products. Their use for brick and tile has been largely discontinued in areas where shale is available, and the amount still being used for this purpose is a small part of the over-all production of clay products valued at somewhat more than $31 million annually. Bricks made from Pleistocene clays are found in most brick buildings in the older parts of our cities and in the older farm houses.

Major construction projects such as water storage and power dams, superhighways, new railways, causeways, and port facilities must take surficial deposits into account, as some need to be removed and alternate materials hauled to the sites. Landslides involving these materials may occur where highways cut through hills or valley walls or where stream erosion undercuts valley sides. Clays or clay tills may be sought for impermeable blankets in water reservoirs, or permeable fill may have to be found for some earthfill dams. The glacial history of such regions should be known, for it may dictate the amount of testing necessary to evaluate the stratigraphic sequence at the construction site proper.

The effects of differential uplift need to be evaluated where major construction projects with a long-term life-span involve large bodies of water. Differential uplift between opposing ends of the storage basins of only a small fraction of an inch per year may be important, and its postglacial record is requisite. The rate of differential uplift in Hudson Bay, based on short-term tide gauge records, may be about 2 feet per century (Barnett, 1966). Such information, together with the rate of postglacial uplift, would provide a useful curve of land–sea relations important in planning major harbour facilities in the region.

Waste disposal is becoming an ever-greater problem in industrial and mining areas. Areas of relatively impermeable clay are at times sought for waste disposal pits for corrosive or noxious materials, but the mineralogy of the clay should be determined with regard to possible chemical reactions within the deposit. Waste products that are less objectionable and that will break down in a relatively short time may be pumped into some drift-covered waste lands, but great care has to be exercised to prevent contamination of the groundwater.

Surficial deposits have a direct bearing on Canada's water resources for they determine the rate of run-off and amount of water storage. They hold much water that would otherwise be lost by rapid run-off, and they grudgingly but positively discharge this to the rivers and streams. Without this constant to semi-constant supply, many of our river systems would vacillate between raging torrents and dry valleys.

Thus our heritage of glacial deposits is fundamental to the very character of this land and affects the progress of our agricultural, forestry, mining, and fishing industries, urban and industrial developments, and recreational facilities, all of which influence our daily way of life and our future plans.

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