CANADA
DEPARTMENT OF MINES
HON. W. A. GORDON, MINISTER; CHARLES CAMSELL, DEPUTY MINISTER

GEOLOGICAL SURVEY
W. H. COLLINS, DIRECTOR

MEMOIR 168

Late-Glacial Correlations and Ice Recession in Manitoba

BY
Ernst Antevs

OTTAWA
F. A. ACLAND
PRINTER TO THE KING'S MOST EXCELLENT MAJESTY
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Preface

The disappearance of glacial Lake Ojibway in northern Ontario and Quebec, when the ice front stood north of Cochrane, sets practically a limit to the extension of the varved clay chronology in the region south of James bay. A study of the varved clays from the last stages of glacial Lake Agassiz in northern Manitoba was, therefore, carried out during the summer of 1929.

In order to correlate the Late-Glacial ice borders and events in northern Manitoba with those in western Quebec and eastern Ontario, the existing data concerning the direction of the ice flow, the moraines, etc., were compiled. This led to an attempt at a general correlation of the Late-Glacial of the East and the Middle West. The results of these various investigations are herewith presented. Furthermore, the probable correlation of the Late-Glacial of North America and of Europe, as well as the basis of attempted correlations between clay varves from America and Europe, are discussed in some detail. A suggestion for a division of the late Quaternary of North America is also presented.

Late-Glacial Correlations and Ice Recession in Manitoba

CHAPTER I

DIVISION OF THE LATE QUATERNARY OF NORTH AMERICA

BASIS OF DIVISION

A well-defined, detailed division of the late Quaternary of North America is desirable both for a clearer conception of the sequence and relationships of the events and conditions in the continent and for purposes of comparison with other parts of the globe. Evidently, the more universal the general basis of such a division, the better will the division be. The widest spreading, changing condition that can come under consideration seems to be temperature. Variations in temperature are the main causes producing glacial and interglacial epochs, and mainly control the rate of growth and waning of ice-sheets, the shifting of climatic belts, the migrations of plants and animals, and the fluctuations of sea-level. In a broad sense temperature changes are simultaneous over the entire northern or southern hemisphere, and, perhaps, the entire world, though the effects of such temperature changes may not be consummated at the same rate everywhere.

On this basis, the Quaternary period, characterized by animals and plants of modern types, may be taken to comprise the time since the beginning of the last Ice Age (in North America),[1] and the Pleistocene may be taken to comprise the time from the beginning of the last Ice Age until the end of the last glacial epoch. This is the sense of these terms as adopted by the U.S. Geological Survey (Willmarth, 1925, pages 45, 49).

The Pleistocene, as is usually done, is best divided into glacial and interglacial epochs. The end of the Pleistocene, that is, the end of the last (Wisconsin) glaciation in North America, may be set at the time when the temperature in the southern and greater part of the once glaciated area had risen to equal the present temperature. At that time fairly large ice remnants may have still existed in Labrador and Keewatin.

[1]While the English-American term "Ice Age" comprises all the Quaternary glaciations, the Scandinavian "istid" and the French "période glaciaire" (Haug, 1911, pp. 1761, 1768, 1897) mean both the whole Quaternary and a glacial epoch, and the German "Eiszeit" means a glacial epoch. "Eiszeit" means glacial epoch even in Albrecht Penck's title "Die Ursachen der Eiszeit" (Sitzungsber. Preuss. Akad. Wiss., Phys.-math. Kl., 1928, VI, pp. 76-85), as the reader is told on the last page of the paper. The German equivalent for "Ice Age" is "Eiszeitalter".

In Denmark and southern Sweden (K. Jessen, 1920, page 269; Munthe, 1925, Plate 7, page 76) the rise of the temperature to that of present times occurred in Boreal or Ancylus age, i.e. about 8,700 years ago, the date most commonly taken as marking the limits between the Late-Glacial and the Post-Glacial epochs. In North America the end of one epoch and the beginning of the other may have occurred at about the same date.

The transition from growth to waning of the last ice-sheets, marking the greatest change of temperature and of other climatic factors since the corresponding event of the next preceding glaciation, seems to be the most suitable occurrence from which to count the late Quaternary epoch, as pointed out by De Geer (1911, page 466). The late Quaternary epoch extends to the present. The time from the climax of the last glaciation until the end of this epoch may be called, also in accordance with common usage in Sweden, the Late-Glacial. Marked changes in the rate of waning of the last ice-sheets are logical events for a division of the Late-Glacial epoch, as also pointed out by De Geer. And the rise of the temperature to equal that of the present, then to above it, and finally the fall to the present temperature appear to be the most universal points of division of the Post-Glacial epoch.

TABLE I

Divisions of the Quaternary in North America. The Iowan drift may represent a separate glacial epoch or a stage of either the Illinoian or the Wisconsin glaciations. In Canada only two or three glaciations are on record.

<table>
<tr><td rowspan="14">Quaternary or Last Ice Age</td><td colspan="2" rowspan="4">Post-Glacial</td><td rowspan="7">Late Quaternary</td><td rowspan="4">Post-Glacial</td><td rowspan="2">Late</td><td>Modern</td><td></td></tr>
<tr><td>Recent</td><td></td></tr>
<tr><td>Middle</td><td colspan="2">............(Temperature higher than today)</td></tr>
<tr><td>Early</td><td colspan="2">............(Temperature as today)</td></tr>
<tr><td rowspan="10">Pleistocene</td><td rowspan="4">Wisconsin</td><td rowspan="3">Late-Glacial</td><td>Younger</td><td colspan="2"></td></tr>
<tr><td>Middle</td><td colspan="2"></td></tr>
<tr><td>Early</td><td colspan="2">............(Climax of last glaciation)</td></tr>
<tr><td colspan="5">5 Wisconsin</td></tr>
<tr><td></td><td colspan="5">IV Peorian</td></tr>
<tr><td></td><td colspan="5">4 Iowan</td></tr>
<tr><td></td><td colspan="5">III Sangamon</td></tr>
<tr><td></td><td colspan="5">3 Illinoian</td></tr>
<tr><td></td><td colspan="5">II Yarmouth
2 Kansan</td></tr>
<tr><td></td><td colspan="5">I Aftonian
1 Nebraskan Jerseyan</td></tr>
</table>

TERMINOLOGY

Names of geological divisions should, if possible, be general, self-explanatory, and fully neutral. They should not tend to convey a specific idea, since this may prove to be incorrect; they should not refer to only a special region, for this limits their usefulness and necessitates a number of parallel sets of terms; they should not require a detailed, unimportant knowledge in order to be understood; they should not apply to a climatic, floristic, or faunistic condition, as do "Boreal Age," "Dryas Age," and "Yoldia Age," since, during the waning of the ice-sheets, these conditions at any point were in a constant state of change, and formed migratory belts, so that, for instance, a zone with arctic climate and biota bordered the retreating ice edge and this itself was followed by a succession of belts. A uniform division and terminolgy is as necessary for the Quaternary as for the older periods.

Perhaps the most neutral way to indicate different stages of a period or an epoch is the one most generally employed, namely by addition of the adjectives "lower," "middle," and "upper"; or the partly equivalent words "older", "middle", and "younger" ("newer"); or "early", "middle", and "late". The first-mentioned terms, so suitable for pre-Quaternary periods—Lower, Middle, and Upper Jurassic, for instance—are not so good for the late Quaternary, because the deposits of the older part of this epoch are normally not overlain by the deposits of the younger part. Still, the terms have been adopted in archæology—Lower, Middle, and Upper Palæolithic. On the other hand the French terms" Quarternaire inférieure", "Q. moyen", and "Q. supérieure" (Haug, 1911, pages 1769, 1776) are used in a natural sense. Better for the late Quaternary are the terms "older", "middle", and "newer" ("younger") employed in Great Britain to differentiate the three Pleistocene drift sheets distinguished there, and the terms "early", "middle", and "late" used in the United States. Besides, these terms have perfect correspondence in generally accepted Scandinavian and German terms, in the Swedish "äldre" or "tidig"-, "mellan"-, and "yngre" or "sen"-; and the German "Alt"- or "Früh"-, "Mittel"-, and "Jung"- or "Spät"-. Generally the terms "early", "middle", and "late" seem preferable because of uniformity; for the terms late Quaternary and Late-Glacial are already established. However, the expression "Late Late-Glacial" is incongruous and for it may be substituted the term "Younger Late-Glacial".

LATE-GLACIAL EPOCH

The waning of the last North American ice-sheet was characterized by three stages: first, the relatively rapid retreat in the peripheral zone; then, the slow melting in northernmost New England, southern Quebec, and Ontario, and the northern parts of the Great Lakes region; and, finally, the rapid release of the central area. These stages, largely determined by changes in temperature, our general norm of division, may be chosen as defining the subdivisions of the Late-Glacial epoch. The change from one stage to the next was gradual and, therefore, it is, at least at present, best to select events in the history of the Great Lakes as dating division points. It so happens that the second stage practically coincides in time with the most prominent Late-Glacial stage of the Great Lakes, Lake Algonquin. The inauguration and the termination of this lake being well-marked events, they may be taken as divisional points.

At the inauguration of Lake Algonquin the ice border stood at St. Johnsbury, the Adirondacks, Rome, Albion, Hamilton, Toronto, Port Elgin, Alpena, Munising (and ? Winnipeg). And at the termination of this lake the ice front was at Maniwaki, Mattawa, Franz, and Long lake, and probably at Neteianga lake, Trout lake, and Wekusko lake (*See* Figures 1, 2). These ice borders, therefore, also define the Early, Middle, and Younger Late-Glacial epochs. The Early Late-Glacial, then comprises the time of the retreat of the ice from the terminal moraines to St. Johnsbury and Toronto; the Middle Late-Glacial, or age of Lake Algonquin, that of the recession from Toronto to Mattawa; and the Younger Late-Glacial that of the withdrawal from Mattawa to somewhere in Labrador—the time until the temperature in southern Canada and the United States had risen to the same level as now.

These epochs may be subdivided in accordance with the principle adopted for the major divisions, namely, changes in temperature, and also with respect to important events in the history of the large glacial lakes. A general subdivision of the very earliest stages of the ice retreat cannot be made until a more detailed correlation between the Late-Glacial of New England and of the Middle West has been obtained. However, regional substages have been distinguished by Leverett (1929, page 19) in the Middle West on the more pronounced oscillations of the ice border and on successive culminations of the different ice centres. The moraines, marking the limits of the oscillations, are given in Table II, page 8. Substage 1 comprises the formation of the Shelbyville moraine and the recession to the Bloomington morainic system. Substage 2 comprises the Bloomington morainic system and the retreat to the Kalamazoo-Mississinawa morainic system; and so forth. Substages 1 and 2 correspond to the culmination of the Labrador ice; substage 3 to that of the Patricia ice; and stage 4 to that of the Keewatin ice. Beside these oscillations, other oscillations, especially that marked by the Niagara Falls moraine and recorded at Claremont, N.H., may serve as limits of divisions. The length of time represented by the Early Late-Glacial seems, according to varve chronologies and estimates, to be some 11,500 years (Antevs, 1928, pages 151, 152, 164).

The Middle Late-Glacial can, at present, be subdivided only according to the lake and sea stages distinguishable in the basins and lowlands, the fluctuating rate of recession of the ice being little known. The stages, their relationship, and the corresponding ice borders are given in Table III, page 10. The correlations are discussed on pages 9-17.

Although lack of varved clay has so far thwarted the attempts to establish an exact varve chronology of the Middle Late-Glacial epoch, the length of time of some of its stages can be roughly estimated in various ways, viz., from the changes of level, physiographic alterations, distance of ice recession, and so forth.

Some idea of the time involved in the last stages of the epoch may be had from the changes of level in Ottawa valley. The rates of the vertical movements at the disappearance of the ice load are, perhaps, not yet determined anywhere in America, but the rates obtained in Sweden may indicate their approximate magnitudes. In Ångermanrland, in the centre of the Scandinavian ice-sheet, the uplift of the land during the first 700 years after the uncovering by the ice averaged 0·12 to 0·13 metre a year (Lidén, 1913, page 28). In the region of Filipstead, north of lake Vänern, it seems during the first 220 years after the release from the ice to have amounted to about 0·09 m. annually (Granlund, 1928, page 25). And at Karlsborg, on lake Vättern, the earliest recorded uplift appears to have been about 0·11 m. a year (Westergård, 1926, page 67). Let us, therefore, assume that the vertical movements in Ottawa valley during Middle Late-Glacial time averaged 0·1 m. (0·328 feet) annually. The uplift of some 450 to 500 feet (137 to 152 m.) during the First Champlain Sea (Antevs, 1925, page 70), would then represent about 1,370 to 1,520 years. And the sinking of some 250 feet (76 m.) in the early part of the Second Champlain Sea and the later uplift of more than 200 feet (61 m.) may represent some 760 and 610 years respectively. Considerable time being consumed by stand-still at the turning points, the Champlain Sea and the intervening Ottawa Land stage may be estimated at 4,000 years.

The First Port Huron stage of Lake Algonquin may have been short in comparison with the following stage, still it must represent a fairly long time, certainly more than 1,000 years, for during it the ice front in the region south and east of lake Ontario probably retreated from the Albion moraine to the Philadelphia moraine, in the meantime forming the Oswego and the Watertown moraines (*See* Taylor, 1924).

Lake Iroquois and Lake Frontenac present contradictory aspects, suggesting that important conditions in their history are unknown. The north shore of Lake Iroquois has not been located northeast of Trent river, 21 miles north-northwest of Trenton, and probably did not extend much farther (Coleman, 1904, page 357; Johnston, 1916). The ice border at the time of drainage of the lake may have stood just north of Huntsville, at Stony lake, Tweed, and Lansdowne and in New York at the Dekalb moraine running through Malone to Covey Hill. During the existence of Lake Iroquois the ice front may, consequently, have retired but little farther than during the contemporary First Port Huron stage. However, the Iroquois shore is strongly developed, being about as mature as the shore of lake Ontario, and suggests that Lake Iroquois may have lasted for a few thousand years. On the other hand, no definite shore-line has been identified with Lake Frontenac, except on the slopes of the Adirondacks (Taylor, 1915, page 445; Coleman, 1922, page 49), though the ice border during its existence appears to have receded from the northern part of Stony lake and Tweed to the south of Renfrew, to Ottawa and Hawkesbury, a distance of 60 to 75 miles, and the lakes thus must have existed for a considerable time, probably at least 1,000 years. In all, the Middle Late-Glacial epoch perhaps comprises some 10,000 years.

The first half of the Younger Late-Glacial is at present best subdivided in accordance with the history of the Glacial Lakes (Table IV, page 18).

Lake Ojibway originated on the north side of the divide north of lake Huron just before or at the beginning of the epoch. Lake Barlow started forming when the ice border had retired beyond the drift barrier that holds the modern lake Timiskaming, that is probably about 500 years later. When the ice front stood between the height of land and Ramore and at the south shore of lake Abitibi, that is 1,100 years after the uncovering of the mouth of Montreal river and lake Timiskaming, the two lakes merged (Antevs, 1925, pages 58, 74; 1928, page 142). The part of the resulting huge lake that lay north of the divide was during the years 2022 (or 2015) to 2027 drained to Hudson bay, while the southern part was lowered (Antevs, 1928, page 103). At the time of the drainage the ice front probably stood some 50 miles north of Cochrane and at the junction of Kenogami and Albany rivers.

The latter part of the Younger Late-Glacial is little known. However, subsequent to the drainage of Lake Barlow-Ojibway, the ice front readvanced to Iroquois Falls and Nellie lake and deposited a moraine. It then withdrew to beyond Cochrane, but advanced for a second time forming the moraine on which Cochrane is now situated. It again retreated, but pushed slightly forward for a third time (Antevs, 1928, page 150). The length of this period of oscillations, after the drainage of Lake Barlow-Ojibway, represents probably more than 1,000 years.

The later retreat is still less known, but it seems as if the epoch may have lasted for nearly 2,000 years after the ultimate uncovering of Cochrane, and the entire Younger Late-Glacial may comprise some 6,000 years.

POST-GLACIAL EPOCH

As already stated, the Post-Glacial epoch is considered as having commenced when the temperature in the southern parts of the previously glaciated area had risen to equal that of the present time. In eastern North America the ice-sheet was then probably confined to Labrador peninsula. This was probably, in round figures, 9,000 years ago.

The Post-Glacial epoch may be best subdivided on the basis of the temperature changes, i.e. the rise of the temperature to above the present stand, and its fall to the present stand serving as division points. In southern Sweden the summer temperature was higher than today, from about 7,000 to about 3,000 years ago. The three subepochs thus distinguished may be called the Early, Middle, and Late Post-Glacial. The Middle Post-Glacial, then, is the age of the Post-Glacial temperature maximum. During the latter and greater part of this age, furthermore, the sea-level stood, possibly, higher than at present.

This proposed division can, perhaps, be applied to the whole globe and may, therefore, be useful for correlations. The several divisions made of the Post-Glacial in Europe are regional or local.

CHAPTER II

CORRELATION OF THE LATE-GLACIAL OF THE EAST AND THE MIDDLE WEST IN NORTH AMERICA

INTRODUCTION

According to the proposals made in the preceding chapter, the Late-Glacial is divided into the subepochs Early, Middle, and Younger. The Early Late-Glacial comprises the time of uncovering of the belt between the terminal moraines and St. Johnsbury, the Adirondacks, Rome, Albion, Hamilton, Toronto, and Port Elgin. The Middle Late-Glacial coincides with the life of Lake Algonquin and comprises the time of ice retreat from the line mentioned to Maniwaki, Mattawa, and Franz. The Younger Late-Glacial, finally, comprises the time from the withdrawal of the ice front from Mattawa valley until the temperature rose to equal that of today in the northern United States, that is until the eastern ice-sheet had become limited to the Labrador peninsula.

EARLY LATE-GLACIAL

The striking separation by a marked re-entrant south of Buffalo, of the Pleistocene ice-sheets into a large lobe over New England and New York and a still larger lobe over the Great Lakes region, was not so much due to any topographic obstacle in the Buffalo region as to the location of the ice centre due north of New England in western Labrador and to exceptionally easy flow in the basins of the Great Lakes.

The peripheral drift of the last glaciation in the Middle West appears to have been deposited by successively culminating ice lobes coming from the northeast, Labrador; from the north, Patricia; and from the northwest, Keewatin (Leverett, 1929, pages 18-20). The Labrador ice laid down the extreme Wisconsin drift in Ohio, Indiana, and Illinois; the Patricia ice deposited the peripheral drift (the young red drift) in Wisconsin; and the Keewatin ice deposited the outer Wisconsin drift (the young grey drift) in Minnesota, Iowa, and the Dakotas, partly on top of the young red drift. These facts are, by Leverett, interpreted as due to culminations of the snowfall successively in Labrador, Patricia, and Keewatin.

The wide separation of the two terminal moraines in southeastern Massachusetts and in eastern Long Island, their westward convergence, the overriding of the inner moraine over the outer in western Long Island, and the occurrence of only one morainic line west of this point, suggest a westward shift of the main flow of ice also in the New England-New York lobe. It also seems probable that the New England lobe reached its climax earlier than did the eastern part of the Great Lakes lobe. The correlatives of the Shelbyville and Bloomington morainic systems of the

Middle West (Leverett, 1915, pages 30, 62) are, therefore, probably to be expected inside the terminal moraines of Long Island. Which moraines they are, cannot be said.

TABLE II

Correlation of Ice Borders and Lakes and Chronology of Early Late-Glacial

Morainic systems (ice borders)		Lapsed time in years: From year of deposition of first measured varve at Hartford	Lapsed time in years: Total	Glacial lake stages
Niagara Falls-Claremont—Lake Winnepesaukee..		3,500 3,200	11,500 (?)	Lake Warren
		−2,671	....	Lake Wayne drains to Hudson river
(4) Des Moines-Port Huron-Alden-Finger Lakes-Catskill-Northampton		1,800	....	Lake Whittlesey
		1,450	....	Lake Arkona
		−0	7,500 (?)	Lake Maumee
	Southern New England			
(3) Kalamazoo-Mississinawa				
	No correlation: Long Island			
(2) Bloomington				
(1) Shelbyville				

A means of correlating stages in the uncovering of New England and the Great Lakes region is to be found in the varved clays (Antevs, 1922, page 99, Plate VI). During the years 2671-2676, after the formation of the oldest measured varve at Hartford (same as years 5671-5676 *See* Antevs, 1922, pages 55, 99), there occurred, as recorded by an enormous increase in the annual sedimentation at Albany and points to the south, "the first escape eastward, through the Mohawk, of the waters of the Great Lakes". The ice front at this time stood north of Brattleboro, Vt., and Albany, N.Y., and west of the Hudson trended somewhat north of west judging from the striæ. The drainage must have been of great magnitude, and as just stated, most probably marks the first opening of Mohawk valley for the waters of the Late-Glacial Great Lakes. This probably took place during the stage called Lake Wayne, the successor of Lake Whittlesey and predecessor of Lake Warren (Taylor, 1915, page 386).

During the existence of Lake Whittlesey the ice front rested on the Port Huron-Alden morainic system, one of the most prominent morainic lines in the Great Lakes region. The halt and readvance of the ice border in New England next previous to the opening of the drainage system by way of Albany, occurred at Northampton in Massachusetts during the years 1,450-1,800 (4,450-4,800) after the deposition of the oldest measured varve at Hartford (Antevs, 1922, Plate VI). The maze of moraines south of lake Ontario is not surely untangled, but the Port Huron-Alden moranic system probably connects with the oscillations at Northampton by means of the Finger Lakes moraines and the moraines on the west side of Catskill mountains. Each readvance is so marked as to make a correlation in the other region highly probable. The Port Huron morainic system runs in wide curves from Buffalo across the regions west of lake Ontario and south of Georgian bay, lake Huron, and lake Superior (Leverett, 1915, page 62; 1929, page 19). Westward its correlatives are the moraines of the Iowa and the Dakota ice lobes, i.e. the outermost moraines of the Wisconsin glaciation.

Some time after the already mentioned first drainage of the Great Lakes through Mohawk valley, the ice border pushed forward probably to form the Niagara Falls moraine (Taylor, 1915, pages 392-398). About years 3,200-3,500 (6,200-6,500), dating from the year of formation of the oldest measured varve at Hartford, there was also a marked advance at Claremont and lake Winnepesaukee, N.H. The two oscillations may be the same. The trend of the ice border from the Niagara region to New England is not known, but probably was nearly due east. The westward continuation of the Niagara Falls moraine may run slightly north of the Port Huron moraine.

MIDDLE LATE-GLACIAL

Lake Algonquin was inaugurated in the southern part of Huron basin, when the waters of the Huron-Erie-Ontario basins fell from the Lundy beach. Immediately after, lake Erie was separated from the glacial waters of the Ontario basin and ceased to be a glacial lake (Taylor, 1915, pages 408, 442). The lake thus initiated in the southwestern part of the Ontario basin underwent changes, as the ice border oscillated, but before long Lake Iroquois was established, when a low outlet was opened past Rome through the Mohawk and the Hudson (Taylor, 1915, page 444). Practically speaking the beginning of the middle stage of uncovering from the ice, therefore, coincides with the beginning of Lakes Algonquin, Erie, and pre-Iroquois. Lake Algonquin terminated with the withdrawal of the ice from Mattawa valley. At this latter event the uplift of North Bay region had practically ceased (*See* Antevs, 1925, page 60); and after it no appreciable uplift seems to have taken place for a very long time. This

TABLE III

Correlations of Ice Borders and Lakes of Middle Late-Glacial

Morainic systems (ice borders)	Glacial lake stages	
	Lake Algonquin	
(?Wekusko-Trout-Neteianga lakes ?) Franz-Mattawa-10 miles north of Maniwaki	Third Port Huron	Second Champlain Sea
	Second Fenelon Falls	
North of Pembroke-North of Gracefield	Port Huron-Chicago	Ottawa land
	First Fenelon Falls (Kirkfield) Stage	First Champlain Sea
Hull-Hawkesbury-Delson		
Stony lake-Tweed-Malone-Covey hill		Lake Frontenac
		Lake Iroquois
Northeast corner of Georgian bay-Stony lake-Napanee-Philadelphia	First Port Huron	
		Pre-Iroquois Lakes
Munising-Alpena-Port Elgin-Hamilton-Albion-Rome-Adirondacks-St. Johnsbury		

makes it highly probable that the uplift that largely emptied Lake Algonquin just before its final drainage past Mattawa, was the same as that which caused the Second Champlain Sea to recede to the lower tracts of Ottawa valley. Lake Algonquin may, consequently, be the correlative of all the stages in the Ontario basin and in Ottawa valley from the pre-Iroquois lakes until the second Champlain Sea, inclusive.

Another starting point for correlation is to be found in the continuance of the deep and wide channel of the Algonquin River to and perhaps below the present level of lake Ontario at Trenton (Johnston, 1916, page 14; Coleman, 1922, pages 31, 51). This channel could have been excavated during the First Champlain Sea or later, for the Iroquois shore stands near Trenton 386 feet above lake Ontario (Johnston, 1916), and the final drainage level of Lake Iroquois amounted to 290 feet and that of Lake Frontenac to more than 217 feet, a total of more than 507 feet. The uplift which shifted the outlet of Lake Algonquin from Fenelon Falls (Kirkfield) to Port Huron-Chicago may consequently have been the same as that which caused the First Champlain Sea to recede. This uplift had, perhaps, already started during the last stage of the life of Lake Iroquois, for the shore of this lake is slightly split in the northeastern corner (cf. Coleman, 1904, page 362; 1922, page 48). At Healy Falls, the northernmost point at which the Iroquois shore is observed, 22 miles north-northwest of Trenton, the difference between the highest strong gravel beach ridge and a lower strong beach is 10 feet, the altitudes being 689 and 679 feet respectively (Johnston, 1916, page 13).

The correlation between the termination of the First Fenelon Falls stage and the withdrawal of the First Champlain Sea is supported and specified by observations by Johnston (1916, page 15) in the region southwest of Fenelon Falls. Sections exposed on the shore of Lake Simcoe 11 miles southwest of Beaverton show a distinct break in the sands of Lake Algonquin, and the upper series, at a height of 730 to 735 feet contains fossil freshwater shells of several species. Fossil shells of the same kinds were found near Roches Point and at Wilfred in the highest beach of Lake Algonquin at altitudes of 775 and 800 feet, respectively. All the species live today in Georgian bay.

These conditions record three stages of Lake Algonquin, viz.: (1) a high stand; (2) a low stand; and (3) the highest stand. During the second stage the water-level stood at least 50 feet below the highest beach and considerably below the threshold at Fenelon Falls. Lake Algonquin may then have drained past Port Huron and Chicago, whereas the lake during the first and third stages discharged past Fenelon Falls. The highest beach may derive from a late date of Lake Algonquin, judging from the fossil shells. The conditions under discussion seem to suggest oscillations of the land. They recall the events in Ottawa valley during the Champlain Sea (Antevs, 1925, page 72; 1928, page 101). And from the above correlation of the uplifts that ended the First Fenelon Falls stage and the First Champlain Sea, it may be concluded that the three stages of Lake Algonquin here considered corresponded to the three stages of the Champlain Sea, viz., the First Champlain Sea, the Ottawa Land stage, and the Second Champlain Sea.

From the correlations made it follows that the First Port Huron and the First Fenelon Falls stages may correspond to the pre-Iroquois lakes, Lake Iroquois, Lake Frontenac, and the First Champlain Sea stages.

At the inauguration of Lake Algonquin and the other lakes mentioned, the front of the waning ice-sheet probably was located on the south shore of lake Superior east of Munising 86° 40′ west (46° 25′ north), (Leverett, 1929), near the north end of Michigan peninsula, just south of Georgian bay, south of lake Simcoe, and just northwest, west, south, and east of lake Ontario (Taylor, 1913; 1915, pages 410, 469, Plate 21; 1924; 1925). It is probably marked by the moraines running through Toronto, Hamilton, Albion, Rochester, Syracuse, and Oneida to Rome, thence northwestward to north of the western end of Oneida lake, and so in a general northeastward direction into the Adirondacks, where it has not yet been traced (Taylor, 1924). Judging from the course of somewhat younger morainic lines, the ice border under consideration may have curved around the Adirondack mountains to the region southwest of Plattsburg. From there it may have continued about due east to the vicinity of St. Johnsbury. The exceptionally rapid shift of the continuous ice front from the eastern Mohawk valley to the northwestern slopes of the Adirondacks was no doubt due to insufficient supply of ice. An extensive field of ice on the southern slopes of the mountains may have lost motion (Cook, 1924).

At the introduction of the First Fenelon Falls (Kirkfield) stage, or the opening of a lower outlet of Lake Algonquin through Sturgeon, Stony, and Rice lakes, and Indian and Trent rivers, to the Ontario basin, the ice front, judging from striæ, may have stood a little inside of the northeastern corner of Georgian bay, several miles south of Huntsville, on Stony lake—the key point, at Napanee, and at or near the Philadelphia moraine in New York state. In New York the ice border, in other words, may have parallelled St. Lawrence river at a distance of 15 to 25 miles from the northern slopes of the Adirondacks.

At the termination of Lake Iroquois the ice border seems to have stood but little north of the line mentioned, viz., at Huntsville, Stony lake, Tweed, and at or near the Dekalb moraine which trends through Canton, Potsdam, and Malone to Covey Hill.

At the drainage of Lake Frontenac and breaking in of the sea over the St. Lawrence lowland, the ice border stood south of Delson, situated about 8 miles south of Montreal, about at Hawkesbury, north of Ottawa, and south of Quyon and Renfrew (Antevs, 1928, pages 137, 164). During the Ottawa Land stage it was located several miles north of Pembroke, for there are to be found clays from the first as well as from the Second Champlain Sea (Antevs, 1928, page 220). It probably stood north of Gracefield in Gatineau valley, for the wash plain at Kazabazua is probably of the same age as those a few miles northeast of Fort Coulonge, at Pembroke, and on the south bank of the Ottawa to 7 miles northwest of Petawawa (*See also* Antevs, 1928, page 138).

When Lake Algonquin, after a great differential upheaval and an emptying out by a moderate drainage at Mattawa, changed to Nipissing Great Lakes, the ice front stood at Mattawa, for it must have been ice that up to this time had prevented the water of Lake Algonquin from taking

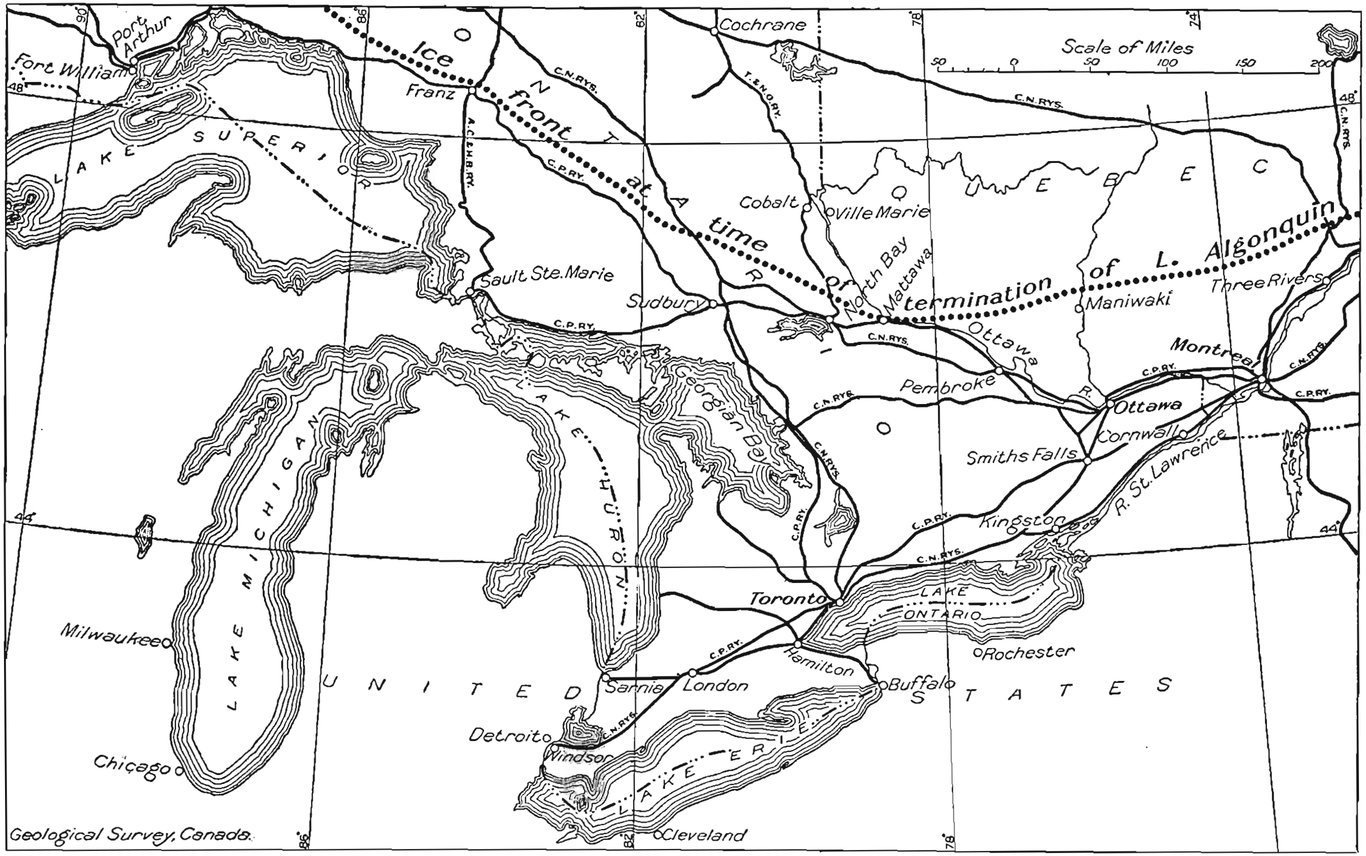

Figure 1. Index map of southeastern Ontario, showing position of ice front at time of termination of Lake Algonquin; the position of ice front has been in part determined by plotting all recorded glacial striæ.

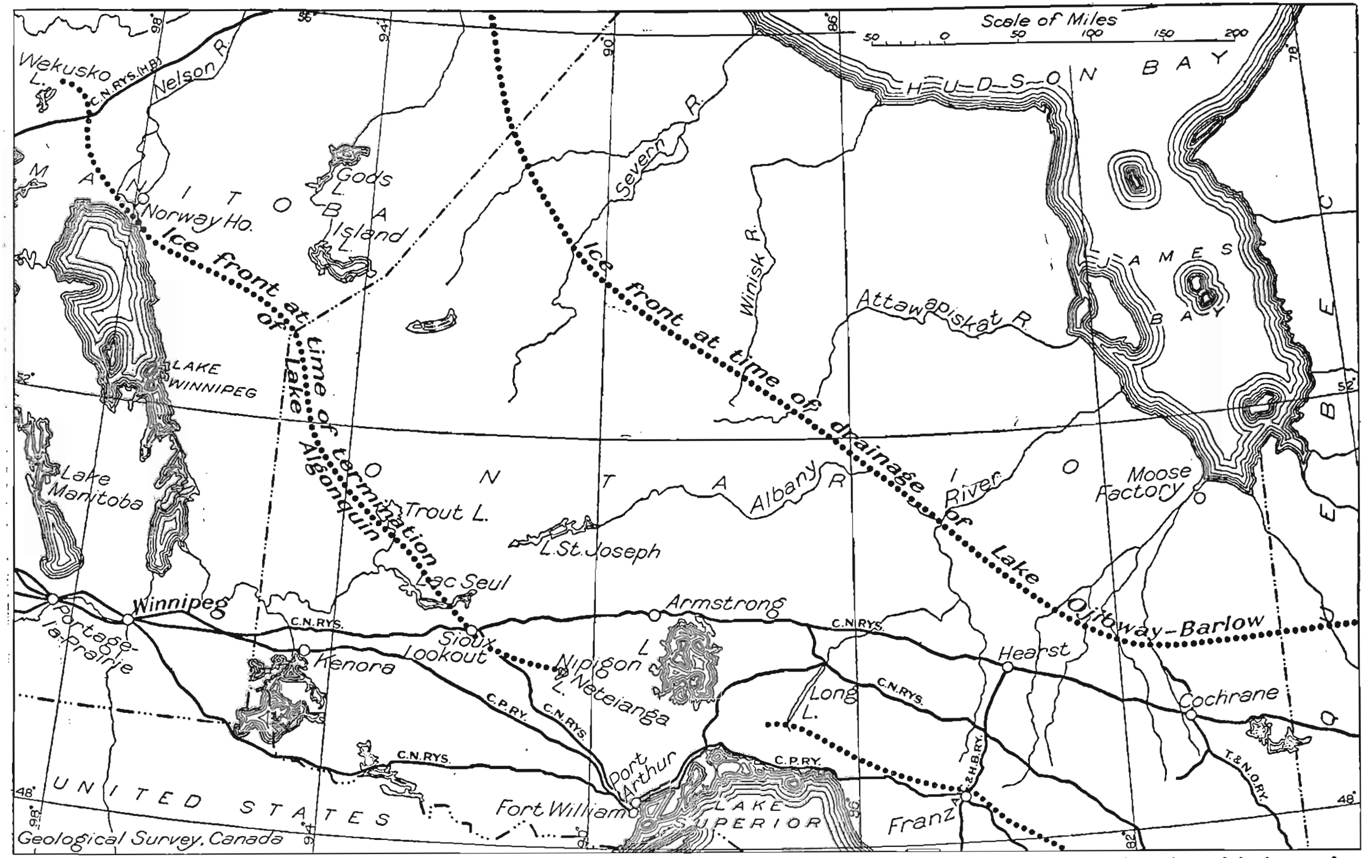

FIGURE 2. Index map of western Ontario, showing position of ice front at time of termination of Lake Algonquin and at time of drainage of Lake Ojibway-Barlow; the position of ice front has been in part determined by plotting all recorded glacial striæ.

this lower outlet. From the junction of Mattawa and Ottawa rivers the ice border, judging from the striæ, may have trended first due east and then slightly north of east to a point some 10 miles north of Maniwaki, situated 70 miles north of Ottawa.

Inside the northeastern bays of lake Superior there are series of abandoned beaches and erosional terraces referable to Lake Algonquin (Coleman, 1922, pages 25-27). Especially important are the beaches at Goudreau lake (1,330 feet), situated 30 miles northeast of Michipicoten Harbour, and at Obatonga lake (1,380 feet) and Pokei lake (1,445 feet), 27 and 35 miles northwest of Michipicoten Harbour, respectively. On Cache lake, 13 miles east of Heron bay, beaches have been traced to an altitude of about 1,025 feet (Taylor, 1896, page 255). On the northernmost part of lake Superior beautiful terraces and beaches occur, especially at Jackfish bay, Terrace bay, Schreiber, and Mazokama (88 degrees west) (Lawson, 1891, pages 275, 272, 269; Taylor, 1896, page 255). The highest record of water-level on the coast is a delta of 1,020 feet altitude at Schreiber. Farther inland, on the south side of the divide between the Great Lakes and Hudson bay, at the south end of Long lake (87° west, 49° 7′ north), there are, according to information kindly supplied by Mr. T. L. Tanton, handsome beaches at high levels (perhaps 1,200 feet). From this it follows: that the ice border at the time of the termination of Lake Algonquin ran from 10 to 25 miles north of lake Superior; that, although in the region north of lake Huron varved clays and silts seem to extend to a maximum altitude of only 820 feet, and generally only to about 730 feet (Collins, 1925, page 97), yet the gravel and silt plains crossed by the Canadian Pacific railway between Sudbury and Franz (Coleman, 1922, pages 23-30) were probably deposited in Lake Algonquin, and that the ice front at the termination of this lake may have stood north of the Canadian Pacific Railway line and north of Wanapitei lake (*See* Figures 1, 2).

Glacial Lake Agassiz, in Manitoba, first emptied to the gulf of Mexico through lake Traverse, Big Stone lake, Minnesota river, and the Mississippi; then eastward to the lakes occupying the basins of the Great Lakes; and lastly northward to Hudson bay. Since beaches of the Lower Campbell stage, according to information by W. A. Johnston, extend to the region of The Pas, and since the lake began to drain eastward during the second next stage, the Blanchard (Upham, 1896, pages xxiii, 443), the ice front at the shifting of the outlet may have stood a good distance north of The Pas.

When the new eastward outlet was employed, Lake Agassiz near Arden, 30 miles west of the south end of lake Manitoba, stood about 1,000 feet above the present sea-level (Upham, 1896, page 441). Judging from the course of the isobases of Lake Agassiz in Dakota and Minnesota, and of Lake Algonquin, Arden may have undergone about the same amount of uplift as Winnipeg and the region just north of Rainy lake on the International Boundary. The waterplain of Lake Algonquin, if extended to the divide running 25 to 35 miles west of lake Superior, would here, a little south of the boundary, stand at 1,000 feet altitude. The amount of tilt of the region just west of lake Superior since the opening of the eastward outlet of Lake Agassiz is not known from observations, but is perhaps

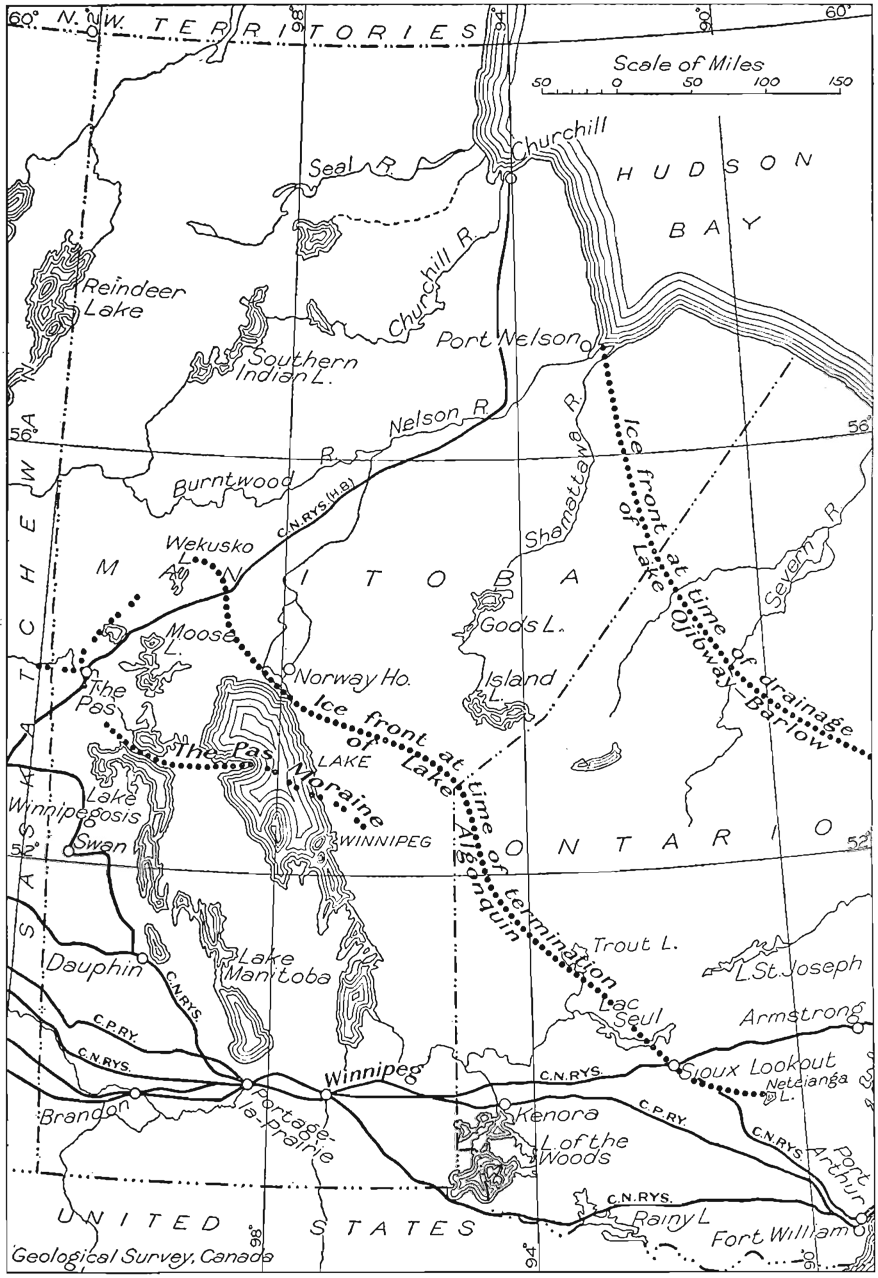

Figure 3. Index map of Manitoba, showing position of ice front at time of termination of Lake Algonquin and at time of drainage of Lake Ojibway-Barlow and position of The Pas moraine; the position of the ice front has been in part determined by plotting all recorded glacial striæ.

of the same order as that in the northern part of Lake Algonqion. In this lake, from a line between Sault Ste. Marie and the base of the Keweenaw peninsula (Leverett and Taylor, 1915, page 439) to Pokei lake, north-northwest of Michipicoten Harbour, it amounted to 425 feet in 120 miles, i.e. to 3·5 feet a mile. With a tilt of this amount Lake Agassiz did not overflow the divide in the distance from northern Minnesota to latitude 49½ degrees north, 85 miles northwest of Port Arthur, where the watershed has an altitude of between 1,500 and 1,600 feet. But perhaps the lake drained across at about 1,400 feet altitude at Neteianga lake (90° 30′ west, 49° 45′ north), 110 miles northwest of Port Arthur.

From Neteianga lake (*See* Figure 2) the ice border, judging from the striæ, may have run west-northwestward past the south end of Sturgeon lake to Sioux Lookout, thence northwestward across lac Seul to the large moraine known to extend, on the southwest side of Trout lake, for some 35 miles (Dowling, 1896, page 51; Bruce and Hawley, 1927, page 48). In its continuation the border probably led over the angle in the boundary line between Ontario and Manitoba to the north end of lake Winnipeg and thence across Grass river west of Pakwa lake to the big loop of this river north of Wekusko lake, where the ice border receded very slowly and perhaps halted.

Few observations of striæ have been made in the region west of lake Nipigon and the probable course of the ice front east of Neteianga lake (*See* Figure 2) is debatable. On the one hand westerly to almost westerly striæ on the western shore of Nipigon lake seem to indicate that the ice disappeared west of the lake earlier than over its basin, and that the Neteianga ice front may have trended first eastward and then southeastward to south of the lake. There is here a broad, morainic belt (A. W. G. Wilson, 1910, page 104) that might be a correlative of the Trout Lake moraine. On the other hand, it seems probable that there was formed a deep re-entrant in the ice front over the lowland between lake Superior and lake Nipigon and over Nipigon basin which, if ice-free, stood under deep water of Lake Algonquin. Terraces observed by Coleman (1922, page 29) in the region east of the lake indicate this. The divide between the Great Lakes and Hudson bay forms two-thirds of a circle around the lake and in the northeast runs at a distance of 20 to 30 miles. Although, consequently, the supply of ice equalled that in the region to the east, the depletion probably was greater because of calving. Therefore, it is possible that the ice border from Neteianga lake ran northeastward and eastward around the north end of lake Nipigon.

This uncertainty about the course of the Trout Lake-Neteianga Lake ice border as well as about that of the Long Lake-Mattawa (*See* Figure 1) ice border east of lake Nipigon makes it difficult to form an opinion about their relationship. However, the fact that they point towards each other suggests at least approximate contemporaneity. And the fact that the Mattawa-Long Lake ice front marked the very beginning of a long age of relatively rapid retreat after a long halt, and the fact that the Trout Lake-Neteianga Lake ice border also marks the end of a pronounced halt or retardation while, somewhat nearer the ice centre, in a belt from Setting lake to the first railway crossing of Nelson river, the ice recession is known to have been rapid, make it probable that the two ice borders approximately correspond to each other.

YOUNGER LATE-GLACIAL

If the ice border withdrew at the same time from Mattawa valley and from the loop of Grass river north of Wekusko lake, northern Manitoba, and if the rapid uncovering of the belt about Setting and Landing lakes was contemporaneous with that of Timiskaming region, the final drainage and disappearance of Lake Agassiz occurred during the middge age of Lake Barlow, and Lake Agassiz may have corresponded to the period of ice recession from northern New England to north of lake Timiskaming and may have persisted for between 10,000 and 15,000 years.

TABLE IV

Ice Borders, Lakes, and Part Chronology of Younger Late-Glacial

	Lapsed time in years			
	From year of deposition of bottom varve at mouth of Montreal river	Total		
Retreat and oscillations of ice border				
Readvance of ice border to Cochrane and Iroquois Falls				
Ice border at (?) mouth of Nelson river-Severn lake-south of Winisk lake-junction of Albany and Kenogami rivers—50 miles north of Cochrane	−2,025	−3,025 ?	Disappearance of Lake Barlow-Ojibway	
			Barlow-Ojibway	
Ice border at Ramore-south shore of lake Abitibi	−1,100	−2,010 (?)		
Drainage (?) of lake Agassiz	−0	−1,000 (?)	Ojibway	Barlow
Ice border at (? Wekusko lake-Trout lake-Neteianga lake?)-Franz-Mattawa-10 miles north of Maniwaki		0		

Among later ice borders in the east, that at the time of the sudden disappearance of Lake Ojibway is especially important. Curiously, this ice border is not known by direct observations, for subsequent to the drainage of Lake Ojibway, a very considerable advance of the ice took place, obliterating most records of the former recession. From the clays in the region south of Cochrane we can, however, form some idea about its position. During its original recession the ice front passed Cochrane about 1,500 years after the uncovering of the mouth of Montreal river in lake Timiskaming (Antevs, 1928, pages 106, 142, 148). Thick varves suggest rapid retreat during the next 200 years. During this time the ice border may have retired to at least 30 miles north of Cochrane. The subsequent events are less surely known, but most probably the ice withdrew at a diminished rate for another 300 years, until it stood some 50 miles north of Cochrane, when Lake Ojibway was completely drained. Wherever the ice front stood, the disappearance of the lake occurred during the years 2022 (or 2015) to 2027. The fact that the ice border subsequently readvanced to Iroquois Falls, Nellie lake, and a point 3 miles north of Frederick House lake, i.e. probably for a total of some 70 miles, and that, judging from the striæ, the ice moved due south from James bay, indicates that the drainage took place from the northwestern end of the lake to Hudson bay. Since Lake Ojibway at the time of its disappearance had deep water over Cochrane region it there surely had an elevation of over 1,000 feet. Since, furthermore, the land then sloped northward relatively to its modern stand, the drainage may have taken place when land somewhat over 1,000 feet high was uncovered. The land rising very gently from the west shore of James bay and the southwest shore of Hudson bay, a general elevation of over 1,000 feet, is met first between Attawapiskat and Winisk lakes, about $87\frac{1}{2}$ degrees west and $52\frac{1}{2}$ degrees north. It may have been here or somewhat farther west that Lake Ojibway was drained. Since the district of Patricia, and Manitoba slope from the regional divide on the 52nd parallel gently towards Hudson bay, and the land north of latitude 53 is below 1,000 feet elevation, the drainage probably took place around the ice front to the sea in the vicinity of the mouths of Hayes or Nelson rivers in eastern Manitoba. The ice front probably formed a gently convex curve from a point some 50 miles north of Cochrane to the junction of Albany and Kenogami rivers, trended thence along a nearly straight line to Severn lake, and finally curved to the vicinity of the mouth of the Nelson (*See* Figure 3).

CHAPTER III

CORRELATIONS OF NIAGARA GORGE AND THE GREAT LAKES

Since it was found by Johnston (1928, page 26) that the entire Whirlpool Rapids gorge and the eddy basin of Niagara canyon were excavated previous to the last glaciation, and thus form parts of the same pre-Wisconsin gorge as do the Whirlpool and the St. David buried gorge, Taylor's (1913, page 29; 1913a, page 25) original and generally accepted correlation of the different gorge sections with the stages of the Great Lakes is, in part, no longer valid. A revision has recently been made by Taylor (1929, pages 258-261). Taylor's old and new correlations are here set forth in brief:

	1913	1929	
1	First gorge..		Port Huron stage of early Lake Algonquin
2	Old narrow gorge...		Fenelon Falls (Kirkfield) stage of Lake Algonquin
3	Lower Great gorge, including Eddy basin, but not the Whirlpool	Lower Great gorge to the Whirlpool; and clearing of the Whirlpool, Eddy basin, and part of the Whirlpool Rapids gorge	Port Huron-Chicago stage of Lake Algonquin
4	Whirlpool Rapids gorge (and Cantilever gorge—a 900-foot long section at the southern railway bridge)	Clearing of upper part of Whirlpool Rapids gorge and of Cantilever gorge; and formation of a 600–800 foot long, narrow, and shallow gorge, subsequently obliterated	Nipissing Great Lakes
5	Upper Great gorge...		Present Great Lakes

Taylor's new correlations and time estimates give 2,000 to 2,500 years as the duration of the Nipissing Great Lakes and about 3,000 years as that of the modern Great Lakes. They give in other words 5,000 to 5,500 years as the length of time since the ice border left Mattawa valley. This is clearly too short a time. The lower end of the Upper Great gorge may not be the only location at which gorge cutting could take place during the Nipissing Great Lakes.

The existing Great Lakes were preceded by the Nipissing Great Lakes, a long stage during which the three upper lakes emptied through Mattawa and Ottawa rivers, and only lake Erie drained through Niagara river. During this stage a shallow gorge section must have been formed. The first shallow canyon below the falls—Whirlpool Rapids gorge not considered—is

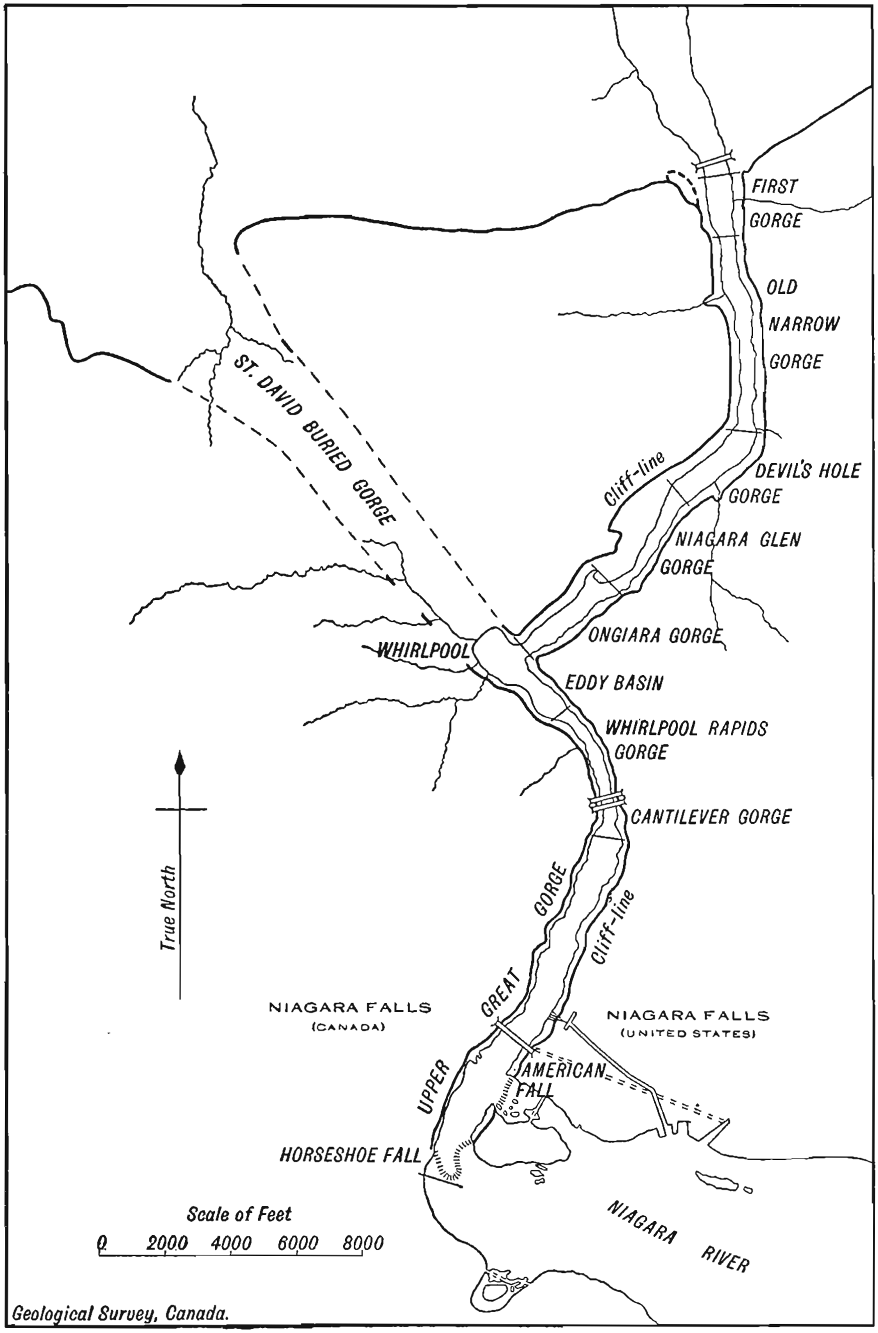

Figure 4. Niagara gorge.

the Niagara Glen gorge (*See* Figure 4). This gorge, which forms the middle section of what Taylor calls the Lower Great gorge, extends for the whole length of the Glen from the north end of Foster flats to Crips eddy, or from a place 2,800 feet southwest of the bend at the south end of the Old Narrow gorge to a place 3,300 feet northeast of the Whirlpool. It is 3,400 feet (1,050 m.) long. The top width of the gorge, except at Wintergreen terrace, averages between 1,600 and 1,700 feet, and the width of the river ranges from 280 to 600 feet (Spencer, 1907, page 102), the Niagara Glen pressing the river against the cliff on the American side. The water depth in the rapids is probably smaller than in the wider gorge above it and the channel is reoccupied by big boulders obstructing the flow, so that the water is piled up 20 to 21 feet higher than below the Glen (Spencer, 1907, page 103). This section may be the correlative of the Nipissing Great Lakes. Since these lakes persisted for several thousand years, the mean annual rate of recession of the falls would have been very small, probably only about 5 inches (13 cm.). This is a reasonable figure considering the fact that the water mass probably was about three times the volume of the modern American fall (Taylor, 1913a, page 24), which with a width of 1,000 feet does not recede measurably beyond what it retreats by weathering (Taylor, 1929, page 259).

The unusual width of the Niagara Glen gorge in relation to the small river that is assumed to have carved it may be explained by the great width below, by the very flat floor of Lockport dolomite which caused the water sheet to spread, and by the separation of the water into two falls by a narrow island, now vanished, extending from the northeastern end of Foster Flats to a point just northeast of Wintergreen terrace. The conditions here were much the same as they are at the present falls except that the greater part of the water passed on the American side of the island, so that the American fall retreated upstream the more rapidly and tapped the Canadian stream (Gilbert, 1896, page 222). The Wintergreen flats and their slopes are parts that had not yet been carved away by the Canadian stream when it was captured.

Increase in width of the bottom part of the gorge at the southwest end of the Glen to three times that in Foster rapids, or to 900 feet, and above here an increase to an average width of 600 feet indicates a large augmentation in the volume of the water, as pointed out by Gilbert (1896, pages 222, 235) and Spencer (1907, pages 188, 191, 361). Increase in the water depth—the water 1,200 feet below the Whirlpool is 100 feet deep—supports this conclusion. The width between the cliffs is 1,600 feet just above the Glen and averages 1,200 feet. Since the transition at the southern end of the Glen is the latest record in the canyon of a marked increase in the volume of the water, it may mark the introduction of the modern Great Lakes. This view was already held by Spencer (1907, pages 149, 368), though he, naturally, thought that the Whirpool Rapids gorge was post-glacial. The increase of the water mass probably took place when the falls were at the southwestern end of Wintergreen terrace, or 200 feet northeast of Crips eddy. The narrow width of the uppermost 500 feet of the gorge, at its junction with the Whirlpool, together with the reefs there suggested by the turbulent rapids, seems to show that the east wall broke away before the vertical fall had receded to the Whirlpool.

The length of this section, which may be called the Ongiara gorge after the park on its American side, is thus 3,500 feet, and the part formed by slow recession of the falls is about 3,000 feet (900 m.).

The Ongiara gorge thus may have been formed, together with the Upper Great gorge, after the uplift of the northern part of the Great Lakes had diverted the outflow from Mattawa river to St. Clair and Niagara rivers. Assuming that the rate of recession of the falls in this section was of the same order as the modern, the Ongiara gorge represents about 1,000 years. The failure of the wall below the whirlpool before the vertical falls had reached it, and the rapid growth in width and depth of the Upper Great gorge at its lower end, indicates that the Whirlpool, the Eddy basin, and the Whirlpool rapids and Cantilever gorges were quickly cleared of the bulk of the drift that clogged them, and that the fall very soon shifted to the upper end of the old pre-Wisconsin gorge.

According to the correlation here presented the entire canyon below Niagara Glen corresponds to Lake Algonquin. This lake appears to have had a somewhat more complicated history than held by Taylor (1915), the following stages being discernible: (1) First Port Huron stage; (2) First Fenelon Falls (Kirkfield) stage; (3) Port Huron-Chicago stage; (4) Second Fenelon Falls stage; and (5) Third Port Huron stage.

During the Third Port Huron stage, corresponding to the last part of the Second Champlain Sea, Niagara river may have carried the entire discharge of lakes Algonquin and Erie. The stage was of short duration, perhaps some 600 to 700 years. It was preceded by the Second Fenelon Falls stage, during which only the waters of lake Erie drained through Niagara river. The Third Port Huron stage may, consequently, correspond to the wide section between the Glen and the river bend, the oldest part of what Taylor called the Lower Great gorge. This section may be called Devils Hole gorge after the ravine on the American side carrying this name. The Devils Hole gorge is 2,700 feet long. The river is 700 to 800 feet wide, and the cliffs 1,500 to 1,600 feet apart. No soundings have been made in the gorge, but, judging from the behaviour of the water, the depth may be moderate, that is some 50 to 70 feet.

The straight lower part of Niagara canyon can, from its much greater water depth in the oldest part, be divided into two sections called by Taylor the First gorge and the Old Narrow gorge. The First gorge extends for 2,000 feet upstream from the Niagara escarpment. The average top width is 1,400 feet. The mean water depth is 140 to 150 feet. As suggested by Taylor, it may be a correlative of the First Port Huron stage of Lake Algonquin.

The remaining section, the Old Narrow gorge, extends from a point 2,000 feet south of the escarpment to the bend below Niagara University. It is $1\frac{1}{8}$ miles (1·8 km.) long. It has a nearly uniform width between the cliffs of, on an average, 1,200 feet. It was originally narrower, for the talus slopes are fully twice as wide as those in the upper sections of the canyon, due to greater age and to thicker shale and thinner capping dolomite in the walls. The width of the river is 500 feet and its depth is probably 60 to 70 feet. This gorge may correspond to the three remaining intermediate stages of Lake Algonquin. Of these, the First Fenelon

Falls stage may represent several thousand years, the others probably about 1,000 years each. The short duration of the Port Huron-Chicago stage, the moderate increase in water because the discharge of the upper lakes was divided between Port Huron and Chicago, and the later crumbling of the canyon walls, may explain the absence of a wider gorge from this stage.

The correlations made between the Niagara gorge and the Great Lakes are recapitulated in the following table:

1	First gorge	First Port Huron stage of Lake Algonquin
2	Old Narrow gorge	First Fenelon Falls stage Port Huron-Chicago stage Second Fenelon Falls stage
3	Devils Hole Gorge	Third Port Huron stage of Lake Algonquin
4	Niagara Glen gorge	Nipissing Great Lakes
5	Ongiara gorge Clearing of the Whirlpool, Eddy basin, Whirlpool Rapids gorge, and Cantilever gorge	Modern Great Lakes
6	Upper Great gorge	

CHAPTER IV

CORRELATION OF THE LATE-GLACIAL OF NORTH AMERICA AND OF EUROPE

FACTORS CONTROLLING ICE WANING AND THEIR CLIMATIC SIGNIFICANCE

The only basis of correlation between the Late-Glacial of North America and of Europe is the climatic changes. The best records of the climatic variations during this epoch are the rate of waving of the ice-sheets and the amount of ice melting. The changes of flora and fauna are not known in North America, and, furthermore, are not suitable for the present purpose, because the biota were in migration, and because they do not present any basis for an absolute chronology.

The conditions governing the rate of waning of the ice-sheets were nourishment and depletion. Normally the ice-sheets during their disappearance were nourished, and the retreat of the ice front indicated excess of depletion over alimentation. Dissipation took place through agencies working upon the surface of the ice-sheet, upon the ice front, and beneath the ice-sheet. The agencies working upon the surface of the sheet were heat, insolation, wind, rain, and melt-water. Those working on the front, where it ended in water, were the melting activity of the water and its buoyancy tending to induce calving. The agency attacking from below was circulating water.

From studies of modern glaciers and ice-sheets it is known that among the factors enumerated, heat insolation and calving were by far the most important (*See* Antevs, 1925, pages 46-56; Sandford, 1926). Heat causes both melting and evaporation, but on an ice surface warmed to the melting point, thawing will exceed evaporation, because of the very great heat of vaporization.

The relative effectiveness of heat and insolation has been little studied; but Maurer (1914) found that horizontal ice blocks in shade and in bright sunshine at Zürich disappear at rates of as 5 to 6. From observations made on modern glaciers it may be concluded that rains were of relatively small importance as causes of the waning of the ice-sheets, because the rains may have normally been rather cold. For the same reason melt-water may have had an insignificant part in the depletion. Wind may have been the most important climatic factor next to heat and insolation, for evaporation increases about in proportion to the square of the velocity of the wind; the winds descending from the ice-sheets were dry and sometimes dynamically heated; and evaporation takes place at any temperature.

Because of its low temperature, water in which the ice front terminated may have caused little melting, but very much greater destruction by detaching icebergs. Its effects evidently would be greatest in wide and deep lakes with powerful waves. Dissipation accordingly took place by melting, evaporation, and calving.

The ice melting, then, was essentially determined by heat and insolation. Since the sun is the principal control of the earth's climates, the solar radiation evidently influenced the ice melting, yet, as will be shown, did not determine its annual variations. The sun's radiation appears to increase with growing numbers of sun-spots up to a maximum and then to decline very slowly (Abbot, 1929, page 48). At times of sun-spot maxima, of high solar constant, the earth's temperature is slightly lower than during sun-spot minima (Mielke, 1913; Köppen, 1914; Abbot, 1923; *See* Antevs, 1921). As this apparent paradox may have its cause in increased cloudiness during sun-spot maxima (Abbot, 1923), high solar radiation may be attended by both low temperature and low insolation at the earth's surface. In other words, the ice melting may have been small during times of strong solar radiation and numerous sun-spots. It is also significant that evidences of the most important solar cycle, the 11·5-year sun-spot cycle, are very scarce in the records of the varved glacial clays (Brooks, 1928; Antevs, 1929). From this it follows that factors and conditions without connexion with the fluctuations in the solar radiation may have largely determined the variations in the rate of ice melting. As, therefore, the varves do not to any marked degree record the changes in the solar radiation from year to year, De Geer's (1926, pages 254, 275) application of the term "solar curve" to the normal varve is inappropriate, as also pointed out by Brooks (1930). But perhaps the choice of name was largely due to circumstances, for De Geer states:

> "Such a general curve, intended to show the normal variation in the heat arrived from the sun to the earlier glaciated, now temperate regions of the earth, I originally used to call the normal curve. But this name having afterwards been applied to the average curve from a single valley, or that of the Connecticut River, it may better be substituted by the quite unambiguous word, the solar curve."

Rapid rate of ice disintegration on land indicates high air temperature, windiness, and strong insolation. Waning of the ice being the result of dissipation exceeding nourishment, the rate of retreat on land approximately records the total summer heat and insolation, provided the nourishment and the thickness were uniform. The variations in ice thickness may have had relatively small importance, but the fluctations in the ice supply by changes in snowfall and in ice flowage were of greater consequence. Thus, in a deep lake in which the ice moved rapidly, the ice front may have retreated but slowly, may have been stationary, or even may have moved forward in spite of strong and rapid disintegration by calving and ablation. Although in cases like this the rate of the retreat does not record the ice melting and the temperature, the thickness of the clay varves does since it registers the ice melting. Therefore, slow recession or halt combined with thick varves signifies considerable ice supply and ice melting, high summer temperature. Normally, however, the rate of the ice retreat on land records the total summer temperature and hence furnishes a means of correlation on the basis of variations in temperature.

MODE OF ICE WANING

An ice-sheet ending on land dissipates by melting and evaporation. Evaporation occurs both at low and high temperatures, and from the entire surface of the ice-sheet. Melting begins at 32 degrees F., and is consequently limited to those parts of the ice surface that are warmed to the temperature stated (internal melting and bottom melting not considered—*See* Antevs, 1925, page 54). Both evaporation and melting being greatly dependent on temperature, the ablation, or surface depletion, decreases from a maximum at the ice edge to a minimum in the cold and calm central parts.

In western Greenland, at latitude 69 degrees north, the snow-line—the upper limit at which the annual snow is melted and melt-water lakes occur in summer—seems to lie at, roughly, an altitude of 1,500 metres (4,900 feet) and 80 km. (50 miles) inside the ice edge. In eastern Greenland it lies lower and farther out (de Quervain, 1920, page 150). In northern Greenland, at about 80 degrees north, the snow-line lies at 750 to 1,000 m. altitude (2,460 to 3,280 feet) at the edge of the ice-sheet, and increases in elevation inward. It occurs, on an average, only some 25 km. (15 miles) inside the ice border (Koch, 1928, page 373). According to other observers the snow-line occurs at essentially lower levels than those given above, all over Greenland, and in northeastern Greenland locally even at sea-level (*See* Kayser, 1928, page 264). During the crossing of the Greenlandic ice from latitude 70 degrees north on the west coast, to latitude 66 degrees north on the east coast, June 20 to July 22, 1912, de Quervain (1920, page 116) encountered at from 180 km. (112 miles) inside the western edge to 170 km. (106 miles) from the eastern edge, a sharply marked central area in which the temperature averaged -10 degrees C. (14 degrees F.) and seldom reached as high as -5 degrees C. (23 degrees F.), whereas in the bordering belts the temperature by day normally rose above the melting point. In this marginal zone of melting much or all of the melt-water that is formed by day may freeze again by night, when, because of the cooling effect of the ice, the temperature is low.

The width of the zone of melting of the Pleistocene ice-sheets evidently must have varied with temperature and insolation. It would be greatest at low latitudes far away from the sea. In favourable regions it may have reached 100 or more miles. It is important to note that melting, the principal mode of ablation, increased in amount outwardly from nothing at the inner limit of this zone; it was not equal over a broad belt, for instance New England, as assumed by Flint (1929, page 269; 1930a, page 67).

Ablation taking place in a marginal zone sloping gently outward but approximately level will first expose the summits of mountains and hills. Thereafter the ice will melt faster over the highlands than over the lowlands because of absorption of heat by the exposed ground and its reflection and transmission to the ice (cf., however, Kayser, 1928, pages 376-378). In regions with moderate relief the ice over the lowlands may, under these conditions, have disappeared shortly after the ice on the bordering hills. A debris cover could not very well be washed down onto the low land ice to retard its melting, as believed by Flint (1929, page 259), since the surface of this ice all the time would lie above the lowermost, ice-free part on the hillsides. The deep valleys in which the ice might be expected to remain for a considerable time after it had disappeared from the adjacent highlands, almost all stood under water which caused destruction by its buoyancy, so that such valleys also became ice-free approximately as early as the highlands (*See* Antevs, 1925, page 52). Accordingly, the Pleistocene ice-sheets underwent appreciable depletion over relatively narrow margins, and normally as they waned preserved a marked front comparable with that of modern ice-sheets.

In at least two instances in North America large areas of ice seem to have become detached from the ice-sheet, viz., on the southern slopes of the Adirondacks (Cook, 1924) and southeast of the lower part of the St. Lawrence (Antevs, 1925, page 66). Since ice tends to form a level surface these ice bodies may have retained slow motion for some time. The remnant of the last European ice-sheet even moved up hill, or westward from its centre east of the mountain range on the Swedish-Norwegian boundary until it had practically disappeared (Frödin, 1925, Plate IV).

When these detached ice fields became thin, they may have broken up in smaller fields and blocks of ice, because of holes and moats developed over and around hill summits and ridges, and because of crevasses widened by melting.

The solidity of the ice within the marginal zone may have differed greatly. Where the marginal ice moved slowly over level ground, crevasses may have been few. Where a tongue of ice rapidly pushed forward in a large basin, the great difference between the rate of flow of the thick central ice and of the thin marginal ice more or less anchored on the higher land bordering the valley may normally have produced a broad belt with numerous and large fissures extending obliquely inwards and backwards into the ice lobe. The marginal areas of the repeatedly advancing south-Baltic lobes of the last ice-sheet, viz., northern Germany, the Danish Islands, and southern Skåne, also present widespread features generally attributed to dead ice (von Bülow, 1926, 1927; Wolff, 1927, page 359; Röpke, 1927-28; Andersen, 1929, page 366; Madsen, 1929; Antevs, 1925, page 93). The dead ice here, then, was originally slow-moving or anchored marginal ice, now receiving little or no supply, that became fissured by tension caused by flow of the ice in the central parts of the lobe. Since the numerous cracks offered vulnerable spots for attack by the ablating agencies, the ice field gradually fell apart.

Where the marginal belt of the ice-sheets moved over a strongly broken country with ridges, terraces, and steep slopes it may, especially if the topographic features were transversal to the direction of the ice flow, have become considerably cracked and faulted. Most of the crevasses may have been transversal, and the number may have decreased rapidly from the ice edge inward, judging from the conditions in the marginal belt in the Greenlandic ice-sheet. On the east side of Greenland fissures were found by Nansen to be restricted to the outer 8 miles and on the west to the outer 25 to 30 miles, although, according to various observers, crevasses locally occur much farther inland (*See* Hobbs, 1911, page 129; Kayser, 1928, page 385). The entire ice in North East Land, Spitsbergen, seems to be much broken up (Sanford, 1926, pages 201, 217). The fissure belt of the Pleistocene ice-sheet, since it was subjected to greater motion, was probably much wider than that of the Greenland ice. Since, however, ice near the melting point quickly welds together, so that permanent fissures in a glacier in temperate regions are only surface features (Gilbert, 1904, page 198; Hess, 1904, page 157), and since snow and freezing melt-water may have filled the cracks, it follows that open crevasses extending to the ground, in other words dissected ice, may have formed only a narrow zone, occupying a fraction of the belt of melting.

PROBABLE CORRELATION

The climatic changes, as recorded by variations in the rate of waning of the ice-sheets, which seem to be the only basis for a correlation between the Late-Glacial of North America and of Europe, have also been taken as a foundation for a division of the Late-Quaternary of North America (*See* page 2). The Late-Glacial has been divided into the three sub-epochs, Early, Middle, and Younger.

The Early Late-Glacial comprises the time of uncovering of the belt between the terminal moraines and a string of moraines running through the northern part of Michigan peninsula, south of Georgian bay, south of lake Ontario, and north of the Adirondacks to St. Johnsbury.

The average width of this zone is 250 miles in round figures. The time involved has been estimated at 11,500 years. The rate of depletion of the ice in this zone varied considerably, but was on the whole moderate. Moraines and other features mark several halts and oscillations.

The Middle Late-Glacial is taken to comprise the time of uncovering of the zone extending from the above-mentioned line running through the Great Lakes area, to a line running from a little north of lake Superior east-southeastward to Mattawa and thence eastward to north of Maniwaki. The width of the zone varied considerably due to ice lobes in the large lake basins. The mean width, where retreat took place largely on land, probably is 150 miles or somewhat less. The length of time represented is not well known; but 10,000 years may be a fair estimate. The rate of depletion of the ice in this intermediate belt was very slow, judging from the relative narrowness of the zone and the long time suggested by the contemporaneous lake stages. Thin varves and absence of agreement between rather closely located clay profiles also indicate a long time.

The Younger Late-Glacial comprises that time from the uncovering of Mattawa valley until the temperature in southern Canada had risen to equal that of modern time, that is probably until the ice had become confined to Labrador peninsula and to a small area in Keewatin. The length of the subepoch was probably some 6,000 years. The decay of the ice was rapid, though interrupted by a very marked reversal in Cochrane region.

The Late-Glacial epoch of Europe has by De Geer (1910, page 1146; 1911; 1912, Plate 1; 1926, Plate 3) been divided into the Dani-glacial, Gothi-glacial, and fini-glacial subepochs. In view of recent mappings and correlations of ice borders in Germany and Denmark, this division needs revision. Firstly, a new subepoch may properly be distinguished for the time of release of the peripheral zone. A suitable demarking line to the subsequent stage is the Pommerania (Inner Baltic) moraine and its correlative in Denmark, the East Jylland moraine (*See* Woldstedt, 1929, pages 166, 209; Madsen, 1928, page 114, Plate 2). This subepoch may appropriately be called the *"Germani-glacial"*. The "Germani-glacial" then comprises the time of the last ice retreat from the terminal moraines, i.e. the Brandenburg moraines (Woldstedt, 1929, page 208), to beyond the site of the Pommerania-East Jylland moraines. The probably large advance of the ice to form the moraines just mentioned is perhaps also best referred to the Germani-glacial, but the formation of the moraines themselves should be assigned to the next following subepoch, the Dani-glacial.

The Dani-glacial may be taken to comprise the time of repeated vigorous ice lobes pushing westward in the southwestern part of the Baltic basin. Since the Baltic ice lobes transgressed southwestern Skåne, the uncovering of this region is best included in the Dani-glacial subepoch which may then comprise the time of the uncovering of the belt from and including the Pommerania-East Jylland moraine to an ice border position running about diagonally across Skåne (Antevs, 1925, page 89). At this date we may begin the Gothi-glacial. The division between the Gothi-glacial and the following subepoch is now set by De Geer (1926, page 260) at the drainage of the Baltic ice lake at the northern end of mount Billingen. The end of the later subepoch and the beginning of the post-glacial epoch was long ago set at the giving way of the narrowest part of the hour-glass-like ice remnant in Jämtland as a result of the pressure of the large lake on its west side, resulting in a catastrophical drainage of the lake and bisection of the ice (De Geer, 1926, page 258, Plate 3).

The last Late-Glacial subepoch, during which central Sweden and almost all Finland were uncovered, was by De Geer (1910, page 1146) called the "fini-glacial", "fini" being derived from the Latin word "finis". Because of being linguistically unsatisfactory the term was soon replaced by "Scandi-glacial" (De Geer, 1911, page 466), but shortly thereafter was revived (De Geer, 1912, page 254). The term "fini-glacial" as Munthe (1912, page 446) pointed out, is improper since it was devised in accordance with a principle different from that adopted in the case of the sister terms Dani-glacial and Gothi-glacial, with reference to time and not to regions, and because the subepoch does not represent the final stage in the disappearance of the ice, there still being large ice remnants at the beginning of the Post-Glacial age. However, the term "fini-glacial" has survived, though persons, unfamiliar with its origin, take it to be monk-Latin for "Fenni-glacial". The term *"Fenni-glacial"* after Fennia or Finland may be suggested to supersede "fini-glacial". "Fenni-glacial" is proper as referring to a region, as do the sister terms, a region whose uncovering largely fell within the subepoch, and in which the contemporaneous ice retreat is, by far, best known. Thus Germany, Denmark, Götaland (southern Sweden), and Finland, that is the regions of the successive zones of ice release and in which the glacial phenomena have been most fully studied, would be indicated in the names of the subepochs.

The rate of uncovering of northern Germany is not known through chronological studies, since there is little or no varved clay there, but it is indicated by the carefully mapped moraines and the oscillations of the ice border (Wahnschaffe and Schucht, 1921, Plate 21; Woldstedt, 1929, page 208; Antevs, 1928, page 155). The terminal moraines may represent a few hundred or more years. Although the retreat just inside them probably was slow, it seems to have proceeded without marked interruptions to far beyond the site of the Frankfurt (Poznań) moraines. These may mark a fairly long halt after a considerable advance of the ice. North of the Frankfurt moraine there are numerous, less continuous, morainic ridges in eastern Germany and Poland, suggesting many halts and reversals in the ice recession. The relationship of the Pommerania-East Jylland moraine to the other morainic systems, above all its location about 125 miles (200 km.) inside the terminal moraine on the meridian of Bromberg, and its practical merging with it in Schleswig-Holstein make it probable that the ice border, especially in the western Baltic region, first withdrew far inside the site of the Pommerania-East Jylland moraine and then pushed forward to form it. A fossiliferous bed with birch and fir suggesting a subarctic climate, occurring beneath till near Vejle on the east coast of Jylland, is additional evidence of this (Madsen, 1928, page 114). The time represented by the Germani-glacial may, therefore, be long, certainly several thousand years.

Nor in Denmark is the rate of uncovering from the ice determined by clay studies. Here glacial clays do occur locally and have been studied by De Geer, Antevs, Andersen, and Hansen; but no agreement has been reached as to what constitutes the annual deposit, the varve, in these clays. The clays are largely thinly laminated. Most of the thin laminæ are regarded by De Geer as varves, while by Andersen (1928, 1929a, 1929b) and Hansen (1929) they are thought to be deposits of the daily melting, "day-laminæ". The varves according to Andersen and Hansen consist of a number, sometimes as many as 100, of such "day-laminæ", and are in many cases up to 2 feet thick. The varve limits are marked by dark layers of stiff winter clay which stands out clearly on samples that are moistened and cut after having thoroughly dried, but is very seldom distinct enough to permit measuring in the field. Less commonly the varves are formed of layers of poorly assorted sand and clay and of a lamina of stiff winter clay. This is a typical varve.

The so-called day-laminæ are in many cases marked by a great difference in the grain size between the lower and the upper part. Could such a differential settling of the glacier mud take place in the course of 24 hours, or in a few days if melt-periods of that length be assumed? If the lakes were real glacial lakes, i.e. had low temperature and were normal in other respects, such a rapid separation and settling is hardly probable from the experiments and observations that have been made (Johnston, 1922; Antevs, 1925, page 32; Kindle, 1929). To be sure, deflocculated varved clay was found by Fraser (1929, page 56) to separate and settle, in the course of 24 hours, in ice-cold water a little over a foot deep, with the production of a lamination, but the reason for the rapid deposition of the fine material was probably that the clay was not fully deflocculated. If the lakes were warmer in summer, if they in other words were ordinary temperate lakes, it is not clear how the sediments could be so distinctly laminated. The peculiar structure of the clay was perhaps due to fluctuations in the depth of the (ice-dammed?) lakes. The problem has clearly to be settled by studies of sedimentation, and for the time being the Danish clays had better be left out of consideration. This is also Nordmann's (1929) view.

The rate of the ice release of the Danish Isles is illuminated by moraines and oscillations of ice front and climate. As the Pommerania moraine is by far the biggest moraine in Germany and its correlative, the East Jylland moraine, also is prominent, they may represent many hundred years. The ice retreat that set in after this halt uncovered entire Fyn and at any rate large parts of Sjælland (Madsen, 1928, page 116). It possibly released all the Danish Isles and part of Skåne besides (Antevs, 1928, page 158). However, the retreat was interrupted by a new advance. During this the ice expanded depositing moraines and wash plains at its extreme position, forming a strongly curved line running through southern Jylland, Fyn, and Samsö to northern Sjælland. In Skåne the ice probably advanced to the region northeast of Romeleåsen (Munthe, 1920, pages 65, 108). Subsequently, according to Munthe's dating of the fossiliferous beds at Robertsdal in southern Skåne, the southern and central parts of this province were released from ice, as the temperature rose to equal that of north-central Sweden at present. It follows that the Danish Isles would also become entirely ice free. If, however, the Robertsdal deposit be younger, be an Alleröd bed, only a comparatively short ice retreat took place. In any case the recession was followed by a new advance during which the ice expanded over southern Skåne, the coastal regions on both sides of the Öresund, and over the small islands south of Sjælland including Langeland. This climax seems to have been a rather marked stage. The subsequent ice retreat probably was fairly rapid, for this seems to have been the age of the temperate continental Alleröd beds of Denmark and Skåne. The ice may have withdrawn from the western Baltic including the island of Bornholm. During a later reversal of the temperature, recorded by an arctic (Dryas) flora in the massive clays capping the Alleröd beds, the ice readvanced over southern Skåne, if the Robertsdal bed is of Alleröd age, whereupon it slowly withdrew to the northwest-southeast diagonal of the province.

These several and very marked oscillations of the climate and of the ice border, together with the formation of the numerous moraines, clearly represent a very long time. The Dani-glacial may have comprised several thousand years.

In southern Sweden the ice retreat has been studied by De Geer and his pupils by means of the varved clays. The rate increased gradually averaging about 120 m. (400 feet) a year. The retreat up to the Fenno-Scandian moraines took more than 2,000 years. The large morainic system mentioned, on the whole consists of two marked moraines, though it locally is subdivided into several. In Finland the two halts and the intervening recession together represent 660 years. The end of the Gothi-glacial, set at the drainage of the Baltic ice lake, came about 100 years after the ice had left Second Salpausselkä (Sauramo, 1929, pages 52-54, 71). The Gothi-glacial, then, represents at least 3,000 years (more than 2,000 plus 760).

The decay of the ice in central Sweden and in Finland north of the Salpausselkäs was on the whole very rapid, generally averaging some 250 m. (800 feet) a year and locally reaching as much as 500 m. annually

(Sauramo, 1929, page 55). However, the retreat was by no means uniformly rapid. Some 500 to 600 years after the beginning of the subepoch a marked readvance representing at least 200 to 300 years appears to have occurred in the region of Gävle (Sandegren, 1929, 1929a). The actual length of the Fenni-glacial subepoch is, therefore, not exactly known, but probably was about 1,500 years as held by Sauramo (1929, page 53).

The Late-Glacial subepochs that have been distinguished consequently seem to comprise the following time lengths.

North America	Europe
Younger, 6,000 (?) years	Fenni-glacial, 1,500 years
	Gothi-glacial, 3,000 years
Middle, 10,000 (?) years	Dani-glacial, several thousand years
Early, 11,500 (?) years	Germani-glacial, several thousand years

Because of great gaps in the geochronologies both in North America and in Europe, and especially because of the impossibility of making fairly reliable estimates of the lengths of the Germani-glacial and the Dani-glacial subepochs, no positive correlation of the Late-Glacial stages on the two sides of the Atlantic can be made, but only suggestions. Those here discussed were also presented some years ago (Antevs, 1925, 1928). Since rising temperature was the main cause of the cessation of growth and of waning of the ice-sheets, and since other marked temperature changes, notably those of the Post-Glacial epoch, seem to have made themselves felt contemporaneously in North America and in Europe, it is more likely than not that the last ice-sheets began to wane at about the same time. If they did not, the American ice-sheet may have been the earlier to commence waning, not vice versa. Differences in the supply of ice could cause one of the ice-sheets to react more readily to the temperature rise than did the other. The difference in size might, but scarcely could, have made the European ice-sheet begin withdrawing earlier than the American.

Since the distance from the ice border at New York to the ice centre in Labrador was about 1,000 miles (1,600 km.), whereas the distance from the margin of the European ice-sheet in Germany to the centre in Jämtland was about 750 miles (1,200 km.), since, in other words, the total distances of retreat are as 4 to 3, the recession of the American ice-sheet must have proceeded at a considerably greater rate unless it lasted much longer. It may be that it did not last much longer and, as the European ice-sheet first disappeared some 7,000 years ago, the American ice-sheet may have terminated shortly thereafter, it having withdrawn more rapidly. Each of the subepochs may, therefore, be expected to be represented by a wider belt in America than in Europe.

Perhaps the best starting point for an attempt at correlation is the change from long ages of halts and slow uncovering to that of rapid ice retreat that characterized the transition between the Middle and the Younger Late-Glacials in North America, and the Dani-glacial and the

Gothi-glacial in Europe. General considerations and present knowedge make it probable that these stages correspond to one another. The probability is strengthened by very pronounced reversals that took place on both continents after a few thousand years of rapid withdrawal, reversals resulting in the formation of the Fenno-Scandian moraines in Europe and in the oscillations at Iroquois Falls and Cochrane in America. The main halts marked by the Salpausselkäs in Finland represent 225 and 183 years respectively, and the intervening retreat 251 years, i.e. the whole morainic belt represents 659 years (Sauramo, 1918, pages 23, 35). A third halt in Finland was perhaps largely due to decrease in calving, as the water depth diminished by 27 m. at the drainage of the Baltic ice lake (Sauramo, 1929, pages 55, 57). The first of the three oscillations of the ice border in Cochrane region, Ontario, represents more than 670 years and probably much more, and the three oscillations perhaps equal about 2,000 years (Antevs, 1928, pages 149, 167). This much greater length of the period of oscillation in Canada, apparently, weakens the suggested correlation. It seems possible, though, that the difference in length of time may have been due to a heavier snowfall and greater ice supply in Canada than in Europe during the oscillations, and to the sudden drainage of the deep Lake Ojibway-Barlow that probably occurred just before the first readvance of the ice border. The high, steep ice front that had withdrawn largely through calving surely came to a halt with the sudden disappearance of the lake, and may even have pushed forward, since ablation alone probably was insufficient to counterbalance the ice flow.

The Middle Late-Glacial probably corresponds to the Dani-glacial, both being represented by a belt in corresponding positions between the terminal moraines and the ice centre. During both subepochs the ice recession was slow, because of halts and oscillations. The Early Late-Glacial probably corresponds to the Germani-glacial. To carry the discussion beyond these general suggestions would hardly be profitable.

If the correlations here made are correct, and the time estimates are of the right order, the climax of the last glaciation was reached some 35,000 years ago.

VARVE CORRELATIONS

The writer's correlation between the Late-Glacial of North America and of Europe is based on long-ranging climatic changes recorded by the varying rate of waning of the last ice-sheets as known from glacial studies in general and from moraines and clay varves in particular. Whereas our correlation is based on the most marked and longest climatic changes, attempts at correlation by means of the shortest climatic cycle, the year, have been made by De Geer. De Geer compares graphs of varved clay from Europe and North America, and where he finds similarities that are judged to be good and persistent, he correlates the year, since he believes varve graphs to be so highly distinctive that only one correlation is possible. By such correlations, stages and events in North America are held to be dated by years in relation to one another and to the Swedish varve chronology (not published) and almost precisely in relation to the present year A.D. The accumulated observations relating to moraines, striæ, and other features that shed light on the rate of retreat and on the trend of the ice edge, are left out of consideration. As a result varves from Denmark and Skåne have been correlated with varves from Hackensack, N.J., i.e. varves formed during the second stage of ice waning in Europe have been correlated with varves produced at the very beginning of ice retreat in North America. Similarly, as a result of varve correlations, the Ottawa region is stated to have become ice-free contemporaneously with the height of land between the Timiskaming and the Abitibi regions, though field observations of various kinds indicate a time difference of several thousand years.

Varve correlations over distances of 50 km. (30 miles) along the direction of the ice border had been made by De Geer previous to 1915, but were thought to have been possible as a result of particularly favourable conditions of transportation and sedimentation of the mud (De Geer, 1916). Agreement between curves that represented localities situated 100 km. apart and that, accidentally, were compared in 1915 suggested to De Geer a common and climatic cause of the fluctuations in the thickness of the varves. Further comparisons revealed agreements over still greater distances in Sweden and led to comparison and correlation of clay graphs from Sweden and Finland, Sweden and Norway, and Sweden and North America. The variations in thickness of the varves were thought to be due to annual variations in the total solar heat received by the earth. De Geer's procedure of teleconnexion in 1915, as well as now, is, in short: graphs are compared, agreement is found, and the matched varves are assumed to mark the same years. Other conditions are not considered.

The thicknesses of varves in a clay profile, the bottom varves being disregarded, are determined by the quantity of mud brought into the lake. This in turn is determined by the amount of the ice melting, which, finally, is determined by the total summer heat. Since the matching of graphs is based upon agreement between the relative thicknesses exhibited, it is obviously necessary to know beforehand that the factor determining the variations of thickness, i.e. the summer temperature, underwent the same and contemporaneous variations in the two regions from whence the varve curves originate. If warm summers in one region corresponded to cold summers in another region, thick varves in the former area should be compared with thin varves in the latter. And if warm summers in one region corresponded, now with warm, now with cold summers in another region (as is the case between North America and northern Europe at present), thick varves in the former area should be compared, now with thick, now with thin, varves in the latter. In other words, correlations between graphs can be made only for areas that are known, or with reasonable certainty can be assumed, to have experienced about the same annual fluctuations in the total summer temperature. Since it is not known and should not be assumed that the summer temperature at the Pleistocene ice borders in North America and in Europe underwent the same variations, De Geer in his demonstration reasons in a circle, for, ignoring other lines of evidence, he assumes that the summer temperatures were the same, matches the curves, and then concludes that the agreement between curves indicates that the summer temperatures were the same. Or, in his own words, "In this paper I have confined myself to give some examples of the long series of conclusively teleconnected varves from both sides of the Atlantic, which show at the same time that correlations only by help of seemingly analogous formations even such as the most marked moraines are not reliable, but still that evidently the deglaciations of at least the whole of the northern hemisphere was synchronous and can be followed up and correlated in a very striking detail" (De Geer, 1926, page 275).

Agreement between curves is determined by the trend of lines connecting points denoting, in this case, the thicknesses of the annual clay deposit. The transportation and sedimentation of mud in a glacial lake being influenced by a number of conditions, such as shape of the basin, islands, currents, and the chemical and physical properties of the glacial mud and of the lake water, a perfect agreement cannot be expected between varve graphs. The degree of correspondence that shall be deemed necessary for correlation is, therefore, left to the correlator's discretion. Because there is no simple way to determine or objectively to estimate the degree of actual correspondence between two graphs, it follows that the appreciation of the similarity is also subject to personal judgment. Varve graphs are individually only moderately distinctive, and, in addition, being records of the summer heat, they present periodic, more or less similar fluctuations.

These conditions make it necessary to exercise considerable conservatism in assuming varve correlations. Correlations cannot be made on agreement of curves alone. It is necessary to know from striæ, moraines, eskers, drumlins, and other features, that the compared varve series are of approximately the same age.

There is no practical method of expressing the actual degree of correspondence between two varve curves not identical. A method of determining approximately the degree of correspondence is to note the number of times the corresponding lines joining successive points are both horizontal, inclined upward, or inclined downward, and to express the number as a percentage of the whole number of pairs of compared lines. However, since this procedure leaves out of consideration the variations in degree of inclination of the sloping lines, it does not give a true valuation of the relationship. This procedure has been employed by De Geer to compare the graphs he has teleconnected.

The varve curves in De Geer's first (1921, page 72) published varve connexion between Sweden and North America are on too small a scale for analysis. However, the agreement in the uppermost varve series presented seems to be about 70 per cent, though De Geer states it to be 86 per cent. In a later paper on the subject (De Geer, 1926, page 281) such parts of the curves as seem to De Geer to be connected across the Atlantic are marked with colour; no percentages of agreement are given. In Plate I, diagram 3, showing correlation of varved series from Alleröd in Denmark and Hackensack, N.J., about 40 more agreements have been marked by colour than really exist. If 18 varves marked by dotted lines be omitted, the curves show about 67 per cent agreement. In Plate I, diagram 4, curves representing series from Merlöse and Vedde in Denmark are compared with a curve representing a varved series from Hackensack. In the case of the Merlöse curve about 13 more agreements with the Hackensack curve have been indicated than can be detected; and in the case of the Vedde curve, after varve 6096, about 12. The former graph presents about 54 and the latter about 51 per cent correspondence with the curve from Hackensack; but taken together they show about 69 per cent agreement with it, since in various parts the Merlöse and Vedde curves are not in agreement and, therefore, the chances of the Hackensack curve agreeing with one or other of them is greatly increased. In Plate II, diagram 3, a curve from Lina in Sweden is compared with curves from Toronto, Ont., Waterbury, Vt., and Wells River, Vt. In the parts comprising the varves 1182-1258 the following agreements present themselves:

	Per cent
Lina—Toronto	67
Lina—Waterbury	58
Lina—Wells River	49
Lina—{Toronto, Waterbury}	78
Lina—{Toronto, Wells River}	78
Lina—{Waterbury, Wells River}	74
Lina—{Toronto, Waterbury, Wells River}	86

According to De Geer's principle the percentage of agreement, of what he regards as normal or "solar" varves, is in this case 86. It is evident that by comparing a group of curves with a single curve, the percentage of agreement rises as the number of curves in a group increases. The fallacy of the method is clearly apparent when it is realized that the smaller the percentage of agreement between the curves of a group, the greater will be the percentage of agreement between the group and any other curve. Thus if three American curves that show no agreement amongst thenselves be compared with *any* Swedish curve, the percentage of agreement will be 100.

This procedure of comparing two, three, or more curves with one curve recurs in practically all of De Geer's teleconnexions, and largely accounts for the high values of agreement which he claims to exist. Of course, only two curves should be compared at a time. In the following table are presented the recalculated percentages of agreement between pairs of curves that have been correlated by De Geer. The series selected are believed to be representative.

De Geer's publication	Curves	Percentage of agreement
1921, page 72	1. Vallby, Sweden, and Woodsville, N.H., varves 1421-1525	70
1926, Pl. I, diagram 3	2. Alleröd, Denmark, and Hackensack, N.J., varves 6289-6578	67
Pl. I, diagram 4	3. Merlose, Denmark, and Hackensack, varves 6096-6174	54
	4. Vedde, Denmark, and Hackensack, varves 6096-6174	51
Pl. II, diagram 1	5. Hedehusene, Denmark, and Hackensack, varves 5541-5629	61

De Geer's publication	Curves	Percentage of agreement
	6. Hebehusene, Denmark, and Connecticut Valley varves 5541-5629	50
	7. Töllöse, Denmark, and Hackensack, varves 5541-5629	43
	8. Töllöse, Denmark, and Connecticut valley, varves 5541-5629	41
	9. Kabusa, Sweden, and Dutchess Jc., N.Y., varves 5736-5869	54
Diagram 2	10. Vallby, Sweden, and Beach Ridge, N.Y., varves 1484-1579	53
Diagram 3	11. Lina, Sweden, and Toronto, varves 1182-1258	67
	12. Lina and Waterbury, Vt., varves 1182-1258	58
	13. Lina and Wells River, Vt., varves 1182-1258	49
1929, Pl. II	14. Sweden and Connecticut valley, varves 1800-2064	55
	15. Sweden and Connecticut valley, varves 2336-2600	47
	Average agreement	54

By the nature of the curves and the method of calculating the degree of correspondence, 33 per cent agreement may be expected between any two curves and, of course, has no significance. It follows that agreements of 40 to 50 per cent are worthless, and that even agreements of 55 to 65 per cent are not of much value. As indicating what may be expected in the way of agreement between curves, the following table is presented. In this are given analyses of curves representing correlated deposits occurring in separate but not too distant valleys or in the area of a single glacial lake.

Antevs' publication	Curves	Percentage of agreement
1922, Pl. III	13 N.H. and 13 N.Y., varves 5501-5600, Brattleboro and Hudson, 80 miles apart	84
	14 N.H. and 14 N.Y., varves 5709-5749, 5771-5800, Concord and Albany, 120 miles apart	75
	14 VT. and 14 N.Y., varves 3713-5800, Brattleboro and Albany, 60 miles apart	82
Pl. IV	The two curves 16 Vt., N.H., varves 6001-6200, Concord and Brattleboro, 50 miles apart	82
1925, Pl. VI	Timiskaming A4 and C4, varves 601-800, Lake Timiskaming region, Glacial Lake Barlow	77
Pl. VII	Timiskaming A5 and C5, varves 801-1000, Lake Timiskaming region, Glacial Lake Barlow	79
	Timiskaming A6 and C6, varves 1001-1200, Lake Timiskaming region, Glacial Lake Barlow	82
Pl. VIII	Timiskaming A8 and C8, varves 1401-1585, Nighthawk lake and La Sarre, 80 miles apart, Glacial Lake Ojibway	70
1928, Pl. II	New Haven B2 and C2, B3 and C3, varves 201-544, Haverstraw and New Haven, 57 miles apart	81
Pl. III	Hartford A3 and C3, varves 3002-3170, Newburg and Hartford, 70 miles apart	79
	Average agreement	79

The agreements were in most cases labelled "fairly good" to "good" (Antevs, 1922, pages 55, 56; 1925, pages 126, 127). According to this standard the average agreement of 54 per cent existing between curves tele-correlated by De Geer is very slender; in two or three cases there is no agreement worth mentioning; in others the correspondence is smaller than is necessary to establish connexion between adjacent clay localities; and in no case is the degree of agreement greater than can be explained by recurrence of periods and by chance.

A third way to test the validity of De Geer's tele-correlations of clay varves in North America, in Europe, and across the Atlantic and the far-reaching conclusions drawn from them, is to consider them in the light of geological knowledge. Three important cases may be taken.

(1) De Geer (1926, Plate 3) believes that by transatlantic varve correlations he has determined that the border of the last North American ice-sheet stood at the outermost moraines at New York city some 18,000 years ago.

From our geological and geochronological knowledge the Early, Middle, and Younger Late-Glacial subepochs seem to represent some 11,500, 10,000 and, 6,000 years respectively (*See* pages 8, 10, and 18). The Post-Glacial age may comprise about 9,000 years. The last ice border thus may have stood at New York some 35,000 years ago, or twice as long ago as De Geer holds.

(2) De Geer (1926, page 267; Pl. 2) correlates a varve series obtained by Reeds (1926) at Hackensack-Little Ferry, N.J., with one measured by Antevs (1922) at Hartford, Conn., and determines the time interval between the deposition of the lowest measured varve at Hackensack and the lowest measured varve at Hartford to be 1,300 years.

The incorrectness of this varve correlation has already been indirectly touched upon by this writer (1928, pages 109-119) and has been dealt with by Reeds (1929, page 624). The Hackensack and the Hartford, Connecticut, curves show hardly any correspondence, the percentage of agreement being only 48. The curves not only fail to agree but there are valid reasons for believing that the two series of varved clays differ markedly in age.

Reeds' series from Hackensack comprises Reeds' varves + 201 to 345. These are the same as Antevs' (1928) varves 1261 to 1405 from Hackensack, with which they for the most part agree fairly well. The varves average only 6 to 7 mm. in thickness, and some of them are difficult to distinguish. As the unavoidable inaccuracy in the measuring of varves so thin as these is very great, and as such thin varves give very flat graphs, the Hackensack curves are not distinctive enough to warrant long-distance correlations. Furthermore, the whole clay bed at Hackensack, which seems to comprise about 1,950 varves (2,500 according to Reeds), may have been deposited before the retreating ice border had reached the northern limit of the Hackensack drainage area formed by a ridge west of Haverstraw, N.Y., i.e. before clay began to deposit at Haverstraw.

At Haverstraw a series of 730 varves has been measured (Antevs, 1928, page 226). The series matches well with another series obtained near New Haven, Conn., in a district that, according to the direction of the striæ, became ice-free at about the same time as that at Haverstraw.

The Haverstraw clays do not extend in age up to a series measured at Newburg farther north in Hudson valley, and the New Haven series do not extend in age up to the series measured at Hartford farther north in Connecticut. The Newburg series and the Hartford series are correlated and extend the original New England series 300 years backward (Antevs, 1928, page 227). Slow ice recession in the zone between Haverstraw-New Haven and Newburg-Hartford is also suggested by push moraines and overridden clays.

There is, consequently, a gap between the Hackensack and the Haverstraw series, and another gap between the Haverstraw-New Haven series and the Newburg-Hartford series. The lowest varve at Hartford correlated by De Geer is younger than the lowest varve at Hackensack by 2,000 years (represented by the Hackensack series), plus an unknown number of years (represented by the gap between the Hackensack and the Haverstraw series), plus 730 years (represented by the Haverstraw-New Haven series), plus an unknown number of years (represented by the gap between the Haverstraw-New Haven series and the Newburg-Hartford series), plus 300 years (representing backward extension of the last-mentioned series). That is, the time interval between the deposition of the lowest measured varve at Hackensack and a low varve at Hartford is greater, perhaps much greater, than 3,030 years, whereas by De Geer's correlation the interval is only 1,300 years.

(3) De Geer (1926, Plate 3) places the ice front at the beginning of the Fenni-glacial age, in the region north of lake Ontario, in about the same position as it took at the opening of the Trent valley or Fenelon Falls outlet of Lake Algonquin. And he places the ice border at the beginning of the Post-Glacial age at 80 degrees west, 48 degrees north, i.e. north of Englehart. Both positions are determined by transatlantic varve correlations. The time occupied by the retreat of the ice edge from the former position to the latter is given as 1,073 years.

The Trent Valley outlet was opened, and the Fenelon Falls stage of Lake Algonquin was inaugurated, when Stony lake, situated 90 miles northeast of Toronto, was uncovered during the retreat of the ice-sheet. The ice border left the valley between North Bay and Mattawa after the differential uplift that put an end to Lake Algonquin and to the Second Champlain stage in the Ottawa lowland had been nearly accomplished. The ice retreat from Stony lake to beyond Mattawa river[1] thus corresponds to the First Fenelon Falls stage and the latter stages of Lake Algonquin. It also corresponds to Lake Iroquois and Lake Frontenac in the Ontario basin and to three marine stages in the Ottawa lowland. The time involved represents several thousand years.

The time occupied by the ice retreat from Mattawa valley to the mouth of Montreal river (79° 27½′ west, 47° 8′ north) on lake Timiskaming, a distance of 60 miles (96 km.), may, according to a fair estimate, represent between 700 and 1,000 years. The recession from the mouth of Montreal

[1]The course of the ice front north of Ottawa and in Timiskaming region was about east and west, judging from striæ, wash plains, and clay records (Antevs, 1925, Fig. 27, page 74), not as shown on De Geer's map.

river to Englehart, a distance of 51 miles (82 km.), took 593 years, and that up to the 48th parallel, about 700 years, as determined by checked varve measurements (Antevs, 1925, pages 81, 106).

Thus the time occupied by the ice recession from Stony lake to the 48th parallel in Timiskaming region was several thousand, plus 700 to 1,000, plus 700 years, and not 1,073 years as claimed by De Geer. From this it follows that at least one of the transatlantic connexions is wrong. The southern correlation, as Coleman (1929) points out, cannot be right, because the Toronto series must be thousands of years older than the Swedish series. The northern correlation cannot be right, for if so the very marked climatic fluctuations which took place when the ice border stood in the region of Cochrane, Ont. (Antevs, 1928, pages 148-150), would lack correlatives in Sweden, in spite of the fact that transatlantic varve connexions postulate yearly agreement.

CHAPTER V

CONDITIONS IN NORTHERN MANITOBA DURING LATE-GLACIAL TIME

LOCATION AND SUCCESSIVE CULMINATION OF ICE CENTRES

Conditions and events in northern Manitoba during the waning of the last ice-sheet are intimately connected with the history of the centres of outflow of the ice. Undoubtedly the last glaciation was introduced by the accumulation of snow and ice at many places in central Canada. As the glaciers grew and coalesced, the small, original centres became subordinated and ultimately only a few main centres of outflow persisted.

A principal centre of outflow was Labrador. The accumulation of ice probably began in the mountainous eastern part, but as the ice grew its centre migrated westward, for the border of continent to deep sea is also the approximate border of an ice-sheet to the sea. Too close proximity to the ocean is almost as unfavourable for continental glaciers as is too great a distance inland. This may be due partly to excessive depletion by calving and partly to low temperature of the ice the year round in the exceptionally foggy climate that develops when cold ice meets the sea. Ice of low temperature has small plasticity and tends to develop a steep marginal belt that acts as a barrier on the ice farther inward, whereas the ice of high temperature flows with much greater ease, and besides in summer time attenuates much more by thawing by day and freezing at night.

Another main ice centre called by Tyrrell the Keewatin, developed west of Hudson bay. During its greatest extent this ice reached from Iowa in the south to Coronation and Amundsen gulfs on the Arctic coast, where striæ trending north-northwest are to be found (O'Neill, 1924). Its centre, or rather its axis, may have lain along a line running east of Great Slave and Athabaska lakes. Somewhat later, when the ice had begun to shrink, the axis may have passed through Great Slave lake towards which young, northerly striæ on Coronation gulf and striæ on the Mackenzie between Great Slave and Great Bear lakes converge. The last stand of the ice seems to have been in the region of Dubawnt lake (102 degrees west, 63 degrees north) (Tyrrell, 1896, page 178).

The region west of James bay and south of Hudson bay, the district of Patricia, rises inland gradually to about 1,500 feet at the sources of the rivers. It, presumably, was also an early and important gathering ground of ice, as pointed out by Tyrrell (1913, 1913a). From here ice may have spread in all directions, for striæ trending due north occur at several places near the junction of Severn and Fawn rivers, and old striæ running northwestward are to be seen on Trout lake (90° 20′ west, 53° 45′ north) and on Fawn river northeast of this lake (Tyrrell, 1913, page 205; 1913a, page 528). As the ice grew in extent, the area of outflow widened, so that it may finally have comprised the entire region between lake Superior, lake Winnipeg, Nelson river, and James bay. The ice that radiated towards the north, northeast, and northwest from central Patricia was met by ice from Labrador and from Lake Athabaska region flowing in opposite directions. The three ice-sheets coalesced, but because of the intervening, extensive lowlands of Hudson and James bays and of Nelson River region they may have formed three buckles separated by depressions, just as the two centres of the Greenland ice are separated, across the island, by a depression where the land is lowest (Koch, 1928, pages 440-446). After the welding had been established, there may have been little or no ice flow northward from the Patricia centre. The Keewatin ice and especially the Labrador sheet must have exerted a great pressure, contributing considerably to the expansion of the ice in the Great Lakes region of the United States.

During the stage of waning of the ice, a zone of weakness developed between the Patricia and the Keewatin sheets, along the axis of lake Winnipeg and in the Nelson lowland (*See* page 46). The Patricia and the Labrador sheets, on the other hand, became more and more closely united, as the ice grew thinner by depletion in the marginal belt and by outflow from the centres. When the Patricia sheet had shrunk to the level of the Labrador ice in Hudson and James bays its centre ceased to function, and the ice in Patricia and adjacent Manitoba became part of the Labrador sheet. This probably occurred when the retreating ice border had retired north of lake Nipigon and east of the north-south reach of Nelson river, that is at the time of the drainage of Lake Agassiz. The Labrador ice could now push the incorporated Patricia ice in front of it and expand southwestward, as is testified by young southwesterly striæ on Severn, Fawn, and other rivers and on Trout lake, i.e. up to the site of the previous Patricia centre (Tyrrell, 1913, page 205; 1913a, page 528).

Morainic lines and the overlapping of till sheets in the Middle West of the United States suggest that the ice centres in Labrador, Patricia, and Keewatin culminated successively (Leverett, 1929). It seems, that when the Labrador ice reached its climax in New England, New York, and New Jersey, it was still expanding in the Middle West, and that when the Patricia ice had attained its greatest extent in the Great Lakes region, the Labrador ice was retreating in the east, while the Keewatin ice was still growing in the west. When the outermost moraines in Iowa and the west were deposited by the Keewatin sheet, the ice front in the Great Lakes region and in the east appears to have been in retreat and to have stood at the Port Huron-Alden-Catskills moraines and at Northampton. The time interval between the different culminations may, therefore, represent a few thousand years.

BOUNDARIES OF THE ICE-SHEETS

In southern Manitoba the boundary between the Keewatin and the Patricia ice-sheets may have run northwesterly from Lake of the Woods to lake Winnipeg and thence along the axis of this lake, for the striæ to the east of this line have a southwesterly direction, whereas those to the west of it have a south to southeasterly trend and the preserved striæ may, almost exclusively, have been produced not long before the disappearance of the ice. In the border belt mentioned the ice erosion was insignificant. On islands in the southern part of lake Winnipeg not even the loose material was removed, the rock being deeply weathered (J. F. Wright). Along the border belt to the west of it, the ice retreat was much faster than in the region to the east, probably because of greater supply of ice to the northeast, and the ice front withdrew to the position of The Pas moraine.

The Pas moraine (*See* Figure 3) runs westward from Long point on lake Winnipeg as a high ridge which separates Cedar lake and lake Winnipegosis and, at a point some 35 miles south-southeast of The Pas, loses itself in low and swampy country (Dowling, 1900, page 93; Plunkett, 1913-14). Farther northwest the moraine reappears on the east shore of Pasquia lake, which it follows for more than 10 miles to The Pas (Plunkett, 1913-14). In the other direction from Long point the moraine probably continues through Sandy and George islands in lake Winnipeg and Big Stone point on the east shore of the lake. The islands consist of gravel and sand (Dowling, 1900, page 96), and Big Stone point of boulders (Tyrrell and Dowling, 1900, page 29).

At The Pas the moraine splits. A hummocky moraine was observed on Carrot river 5 miles west-southwest of The Pas; knolls form islands in Saskeram lake 9 miles west of The Pas; a morainic hill was observed between Saskeram lake and Saskatchewan river 10 miles northwest of The Pas; and a ridge crosses Saskatchewan river below Tearing river some 30 miles west of The Pas (Dowling, 1900, page 16). Moraines as well as striæ indicate that the ice front trended about due west from The Pas.

A branch of The Pas moraine follows, as a more or less pronounced ridge, the north bank of the Saskatchewan from The Pas for 4 miles, then turns north and a little east of north and extends along the east side of Reader lake and the west side of Atikameg and Cormorant lakes (Dowling, 1900, pages 17, 23-25). North from Cormorant lake another morainic ridge follows the west side of Cowan river and Yawningstone lake and seems to extend to Reed lake. Since the ice front trended about due west from The Pas, this northeasterly trending moraine may have formed in a re-entrant between the Keewatin and Patricia ice-sheets. It may have been gradually extended northward, as the two ice lobes retreated, for, considering that the ice flow in a marginal belt is in the shortest direction towards the border, the striæ suggest that the re-entrant was never deep. On Atikameg and Cormorant lakes the striæ run somewhat south of west (McInnes, 1913, page 122), that is, form an acute angle with the interlobate moraine. The re-entrant may have been due to thinness of the ice at the junction of the two sheets and to great local depletion because of fissures opened when the direction of the ice flow changed at the meeting of the two ice borders. The moraine, running north-northeast from The Pas, marks the line of equilibrium between the two ice-sheets at the time of uncovering. Varying trends of the striæ show that this line shifted its position and that the relative push of the ice-sheets changed.

Striæ, as well as the east-west direction of The Pas moraine on lakes Winnipeg and Winnipegosis, indicate that previous to the formation of this moraine, the ice flow in the districts north and west of lake Winnipeg was about due south and the line of equilibrium between the Keewatin and Patricia sheets lay along the axis of lake Winnipeg. Northeast of The Pas, old striæ that run south 45 degrees east one mile south of the first great bend on Grass river below Wekusko lake and doubtful striæ that trend south 32 degrees east at the south end of Wekusko lake perhaps originated at this time. At the time of formation of The Pas moraine the Patricia ice extended westward to The Pas, where the boundary between the two ice-sheets now became located. An encroachment of the Patricia ice upon the Keewatin ice is indicated by the shift of the direction of ice flow at the south end of Wekusko lake (at Fosler's) from south 26 degrees west to south 42 degrees west.

But the Keewatin ice seems soon to have recaptured its lost ground, for the interlobate moraine ends on Reed lake, and north of there the south-southwesterly trend of the striæ on Grass river between Wekusko lake and Whiteforest rapids 11 miles west of the south end of Setting lake, and the southwesterly to almost westerly direction of the younger striæ east of here, suggest that the line of equilibrium later was located west of Pakwa and Setting lakes. Late change in the direction of the ice flow at the south end of Wekusko lake to south 34 degrees west is additional evidence. The greater westward trend of the striæ at Whiteforest rapids than at the river loop north of Wekusko lake (south 35 degrees west as against south 20 degrees west) shows that the Keewatin ice was still deflected by the flank attack of the Patricia ice. The Patricia ice was no doubt in its turn deflected. However, after losing its contact with the Keewatin ice, the Patricia sheet, because it decayed slower than the other ice and thereby gradually developed a south-northerly ice border, shifted its direction of flow from southwest or south-southwest to nearly due west.

Thus the trend of the ice flow at the rapid in Grass river 1½ miles above Mitishto lake just before the uncovering changed from south 50 degrees west to south 80 degrees west. Also, between this point and Setting lake, on Setting lake, and on Wintering lake this change is recorded by intersection of striæ, whereas on Landing and Armstrong lakes all the striæ run practically due west. The Keewatin ice retreated nearly due north and the Patricia ice almost directly east.

FINAL DRAINAGE OF LAKE AGASSIZ

Almost all Manitoba west of the 95th meridian and south of the divide between Burntwood and Churchill rivers, together with adjacent lowlands, was occupied by glacial Lake Agassiz, whose northeastern boundary was formed by the ice-sheets (Upham, 1896). As the ice front retreated the lake expanded; but as lower outlets were freed, the lake also lowered. First it discharged through lake Traverse, Big Stone lake, and Minnesota and Mississippi rivers; later it drained eastward through different passages in Ontario to the Great Lakes; and ultimately it drained to Hudson bay.

Late and low stages of Lake Agassiz are recorded by beaches at 906 feet altitude 5 miles south of The Pas, at 845 feet at mile 109 of the Hudson Bay railway, and at 828 feet at mile 110 (Johnston, 1917, page 31). These are the lowest observed beaches of Lake Agassiz; and the lake cannot have undergone any large later lowering, for all the varved clays measured by the writer occur above 600 feet altitude, and the clay at locality 235 on the Nelson, the last to be uncovered, at about 650 feet.

So Lake Agassiz, standing at a level that now lies about 800 feet above the sea and that lay about 350 feet above the then sea-level, grew in extent as the Keewatin ice retired northward and as the Patricia and Labrador sheets retreated eastward. When the Patricia-Labrador ice border, trending slightly west of north, stood somewhat east of the north-south reach of the Nelson, the lake drained to the sea, for varved clays appear to be absent east of this line (McInnes, 1913, page 124, Plate). The Keewatin ice border had by this time probably retired to Indian lakes and Churchill river, though the clays and sands on Southern Indian lake (McInnes, 1913, page 86) should not be taken as evidence of this, since they extend to an altitude of about 875 feet and, therefore, probably were laid down in a separate lake standing at a higher level than did Lake Agassiz.

The final drainage of Lake Agassiz to Hudson bay must have been either over, or through, or beneath, the ice, as the lake during its closing stage was held up by the ice-sheet alone. Because of the debris it contained the ice may have been about equally as heavy as the water. The ice, therefore, probably terminated in a cliff that was at least as high as the water-level and rose back from the ice front in the direction of ice accumulation. Overflow or drainage across the ice, therefore, could hardly have been possible. The highest observed marine beach in the region, namely that $2\frac{1}{4}$ miles southwest of Kettle rapids, stands at 430 feet (Johnston 1917, page 32). *Saxicava rugosa* is represented abundantly and *Macoma* cf. *balthica* sparingly at the gravel pit at mile 329, three-fourths of a mile inside or west of the beach, but these fossils may occur in esker gravels. If they are Late-Glacial in age they indicate that the sea-level may have been somewhat higher than 430 feet, say 450 feet. The ice front in the region to the west stood at an altitude of at least 800 feet and rose towards the east. Since the ice reached considerably above sea-level at that time it must have rested with its entire weight on the ground in the area later occupied by the Late-Glacial sea, so that the lake could not have drained beneath the ice.

The lowest and weakest section of the ice barrier was undoubtedly the line of junction of the Keewatin and Patricia-Labrador ice-sheets. This point probably failed first. The catastrophe may have started by leaking of water through cracks that were quickly widened. Since the line of junction at the time probably lay on Churchill river, the beginning of the drainage of Lake Agassiz probably took place along Little Churchill and Churchill rivers. The divide north of Split lake probably lies at about 600 feet altitude, and the junction of the two named rivers at about 500 feet. The sea stood about 450 feet higher than at present. As the Patricia-Labrador ice border receded, lower regions were freed permitting the lake gradually to fall to the sea which on the Nelson extended up to Gull rapids, 25 miles west of the second railway crossing. However, drift barriers and lake sediments held many of the residual lakes at considerably higher levels than they have today.

CHAPTER VI

RATE OF THE ICE RECESSION IN GRASS RIVER REGION, MANITOBA

Since the region of Wekusko lake lies near the site of the junction of the Keewatin and the Patricia ice-sheets and since its relief is relatively strong, the striæ do not agree well in direction. However, the ice may have retreated about due north or somewhat east of north. A series of 168 varves measured at locality 175 Wekusko lake (*See* Figure 5) does not match the series measured at locality 176, 12½ miles to the north, but since the varves are very thin no significance should be attached to this fact. Locality 177 became ice-free 19 years later than locality 176, and if the distance in the direction of the ice retreat be taken to be 2 miles, the rate of the retreat was 9 years to a mile.

The thinness of the varves at localities 175, 176, and 177 may be due partly to the small amount of drift in the ice, for there is very little drift of any kind in the region. Above some 300 thin but measurable varves at locality 177, there are 4 feet 5 inches of clay consisting of varves that are so thin that they cannot be measured and that upwards gradually grew so thin that the clay in the upper half appears massive. The number of years represented by this bed is more than 400, probably much more (*See* page 57). A similar clay appears at locality 178 near the loop of Grass river 11 miles distant in the direction of the ice flow. The number of measurable varves here is 86, and the thickness of the unmeasurable bed is 2 feet 8 inches, that is somewhat more than one-half of the similar material at locality 177. If the varves are assumed to be of the same thickness as at locality 177, their number in the unmeasured part is more than 200. Since deposition of thin varves terminated simultaneously at both localities (*See below*), the number of varves at locality 177 deposited previous to this event is more than 700 and at locality 178 more than 300. The recession between them consequently seems to have taken some 400 or more years, and the rate appears to have been about 40 years to a mile, which is very slow. The relative evenness of the varves indicates a slow, continuous retreat rather than a more rapid retreat and halt. It is possible though that the slow retreat ended in a halt north of locality 178.

At both locality 177 and 178 the massive clay is abruptly overlain by clay with varves that are rather thick, thicker at locality 177 than at 178. The increase in thickness of the varves is too sudden to be due to augmented melting. It may have been caused either by a marked and sudden increase of the drainage area of the rivers feeding this part of Lake Agassiz or by a drop in the level of the lake as a result of which the sedimentation area was reduced.

During the uncovering of the reach between localities 177 and 178 the ice also withdrew north of Grass river to the east, since the ice border trended east-southeast to a point some distance west of Setting lake where it turned nearly due south.

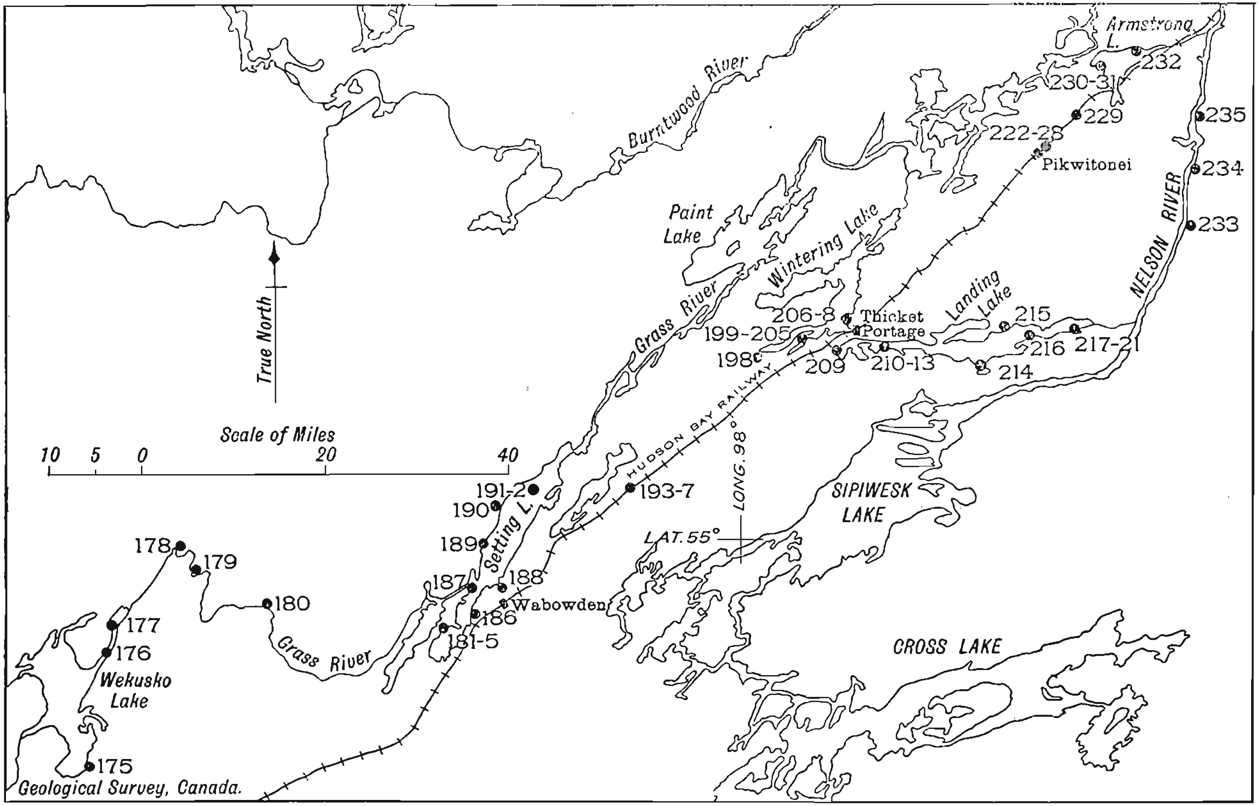

Figure 5. Localities examined in Wekusko Lake-Nelson River district, Manitoba.

In the southern part of Setting lake, where the ice front trended a little east of south, the ice withdrew at a rate of about three years to a mile. This remarkably rapid breaking up may have been favoured by the greater water depth over the basin and by fissures formed during the change in the direction of the ice flow from southwest to nearly west that occurred just previously. However, as the situation presented below indicates, this fast recession does not appear to have lasted long. A series of 276 varves at locality 188 on Setting lake does not appear to correlate with the varve series obtained at locality 193-197, at mile 154½ on the railway, some 15 miles distant in the direction of the recession, though the varves 79 and 112 in the series at locality 188 possibly correlate with varves 1 and 31, respectively, in the series at locality 193-197. On some islands located about 1 to 2 miles south of locality 191, that is about in the centre of the northern part of Setting lake, there are morainic knolls and ridges, and exposures show varved clay disturbed by ice push. One section presented:

Surface.
2 to 3 feet, sand.
6 feet, varved clay, into which gravel has been kneaded. Varve series repeated in blocks of (frozen) clay pushed over the original clay bed.
10 to 20 feet, fine sand extending to lake-level; bedrock above lake-level close by.

If, as the evidence seems to indicate, there was an oscillation of the ice front, it may have represented at least from 100 to 200 years.

Between locality 193-197, and Wintering lake (*See* Figure 6) some 16 miles distant in the direction of retreat, another halt perhaps occurred, for the series of 230 good varves from locality 193-197 does not match the series on Wintering and Setting lakes. No moraines or other features suggestive of a halt are known in the intervening zone, and it is possible that the thicknesses of the varves are local or uncharacteristic in one or both regions. If there was a retardation or halt here it may have been caused by an increased ice supply, for the clay at locality 193-197 consists of relatively thick varves such as are normally formed during considerable melting.

The rate of withdrawal of the ice from the west end of Wintering lake to beyond Nelson river is determined by a continuous varve series and bottom varves at many places. The rate was remarkably rapid, no doubt because of high summer temperature, considerable calving in Lake Agassiz, and a low supply of ice. It averaged nearly 5½ years to a mile between locality 204 on Wintering lake and locality 221 at the east end of Landing lake. It was fastest in the western part of this zone and decreased gradually. On eastern Wintering lake it amounted to 4 years to a mile, in the western half of Landing lake to a little more than 5 years to a mile, and in the eastern half of the lake to almost 6 years a mile. At Pikwitonei the rate was still relatively slow, but it increased again to 4½ years to the mile between locality 228, northeast of Pikwitonei, and locality 230 on Armstrong lake. After the drainage of Lake Agassiz which took place when the ice front stood somewhat east of the south-north reach of the Nelson, the rate of withdrawal may have diminished considerably, depletion now being limited to melting. For a short time the ice, which may have terminated in a high and steep calving bluff, perhaps even remained stationary. A few moraines marking halts are on record, namely along the railway at miles 279-280, 313, 322-324, 325-330½ (also Johnston, 1917, page 31). Of these moraines that at mile 313 consists of a prominent ridge and that at miles 325-330½ of a number of widely spaced ridges and hummocks, whereas the others seem to be small. After leaving the last moraine the ice may have increased its retreat, for calving and melting in sea water were now added to ablation as depleting agencies. The sea probably reached the vicinity of Gillam, mile 327, for marine shells are abundant at the gravel pit at mile 329, three-fourths of a mile inside the highest observed beach.

In the following table, the rates of retreat are given for stretches separating pairs of localities at which the varved clays have been correlated.

Rate of Retreat of the Ice Age

Localities	The number of the bottom varve at each of the two respective localities	Distance at right angle to ice motion		Time of retreat in years	Rate of retreat a year	
		Miles	Kilometres		Feet	Metres
From—						
176 to 177........	1, 19	2	3·2	18	575	177
182 to 186........	1, 17 (?)	4	6·4	16 (?)	1,310 (?)	400 (?)
182 to 187........	1, 15 (?)	5	8·0	14 (?)	1,900 (?)	578 (?)
182 to 188........	1, 22 (?)	7¼	11·7	21 (?)	1,825 (?)	557 (?)
182 to 189........	1,26	8	12·9	25	1,690	516
187 to 188........	15 (?), 22 (?)	2¼	3·6	7 (?)	1,685 (?)	514 (?)
187 to 189........	15 (?), 26	3	4·8	11 (?)	1,430 (?)	436 (?)
198 to 202........	-5 (?), 22 (?)	4½	7·2	26 (?)	905 (?)	276 (?)
198 to 203........	-5 (?), 25 (?)	5	8·0	29 (?)	900 (?)	275 (?)
203 to 204........	25 (?), 35	1	1·6	10 (?)	525 (?)	160 (?)
204 to 205........	35, 45 (?)	1⅓	2·2	10 (?)	720 (?)	220 (?)
204 to 206........	35, 45 (?)	2⅔	4·3	10 (?)	1,410 (?)	430 (?)
204 to 207........	35, 49	3½	5·6	14	1,310	400
204 to 209........	35, 48	3¼	5·2	13	1,310	400
204 to 214........	35, 129	18½	29·8	84	1,155	352
204 to 221........	35, 194	29½	47·5	159	980	298
204 to 234........	35, before 296	44	70·8	<261	>890	>271
207 to 208........	49,60	⅘	1·3	11	385	118
209 to 210........	48, 66	3¼	5·2	18	945	288
209 to 214........	48, 129	15½	24·9	81	1,005	307
209 to 221........	48, 194	26½	42·6	146	955	291
209 to 234........	48, before 296	41	66·0	<248	>875	>266
210 to 211........	66, 74	1¾	2·8	8	1,150	350
211 to 213........	74, 78	¾	1·2	4	985	300
213 to 214........	78, 129	about 10	16·0	51	1,030	314
214 to 215........	129, 143	2⅔	4·3	14	1,005	307
214 to 216........	129, 161	6	9·7	32	995	303
214 to 221........	129, 194	11	17·7	65	890	272
214 to 234........	129, before 296	25½	41·0	<167	>805	>245
216 to 217........	161, 177	2¼	3·6	16	740	225
217 to 218........	177, 185	1⅓	2·2	8	900	275
218 to 219........	185, 186	½	0·8	1	2,625	800
219 to 220........	186, 188	⅓	0·53	2	870	265
220 to 221........	188, 194	1	1·6	6	870	266
221 to 222........	194, 200					
221 to 234........	194, before 296	14½	23·3	<102	>750	<228
222 to 228........	200, 213	1¾	2·8	13	705	215
222 to 230........	200, 240	7¾	12·5	40	1,025	312
222 to 234........	200, before 296	15	24·1	<96	>825	>251
228 to 230........	213, 240	6	9·7	27	1,175	359
230 to 232........	240, before 266	4½	7·2	<26	>905	>276

CHAPTER VII

GLACIAL CLAYS OF GRASS RIVER REGION, MANITOBA

DISTRIBUTION

THE PAS REGION

In the region of The Pas varved glacial clay seems to be absent or inaccessible. The banks of Pasquia river southwest of The Pas are formed exclusively of recent beds deposited during high-water of the river. The same is true of the banks of Saskatchewan river northwest of The Pas, except for small parts of the north bank between the city and the river loop 4 miles upstream, which are formed of bouldery till of the The Pas moraine. Only northeast of the river loop mentioned does the moraine approach the river as a ridge. The Saskatchewan was travelled for 13 miles upstream from The Pas.

HUDSON BAY RAILWAY

The region traversed by the Hudson Bay railway for more than 100 miles north of The Pas is underlain by nearly horizontal Ordovician dolomite that in many places is bare, but for the most part is covered by till, one to a few feet thick, which in its turn is covered by muskeg (peat land) over large areas. Very locally, as at mile 20½, there is some gravel. At miles 43, 47, 54, 66, 68, 70, 73, and 75 there seems to be glacial clay, but which, at least in most places, is thin. Beyond mile 90, clay, sand, and gravel are to be seen more commonly and probably are widely distributed, though largely concealed by muskeg. At about mile 115 is the boundary between the dolomite and the Precambrian rocks. From about mile 130, i.e., opposite the south end of Setting lake, to the first crossing of Nelson river at mile 241½, glacial clay occurs in an almost continuous sheet, though, of course, overlain by muskeg in large tracts. Railway cuts, 5 or more feet high, in varved clay, were observed from the train at the following mileages: 154½ (10 feet high), 175¾ (5 feet, on till), 180 (5 feet, on bedrock), 181½ (10 feet), 182 (10 feet), 182½ (8 feet), 189¼ (7 feet), 191 (10 feet), 194¾ (5 feet, on bedrock), 204 (8 feet), 205 (10 feet, on bedrock), 210½ (10 feet), 214 (8 feet), 214¾ (8 feet), 222 (6 feet), 228½ (8 feet), and 236 (5 feet, on bedrock). The shores of Armstrong lake are largely rocky and little clay was observed except at localities 230-1, and 232.

Beyond Nelson river, clay that seemed to be glacial was observed from the train in ditches and low cuts at miles 241½ (east end of bridge across the Nelson), 243, 244½, 245⅓, 253, 264, 275½, and 279. But since it would have been nearly impossible to obtain varve series, they were not investigated and hence it is not known whether or not the clay at any or all of these places is varved. Sand deposits at miles 251, 254⅓ (25 feet high exposure), 255, 257, and 286, can hardly be taken as evidence that the region was released from the ice before the disappearance of Lake

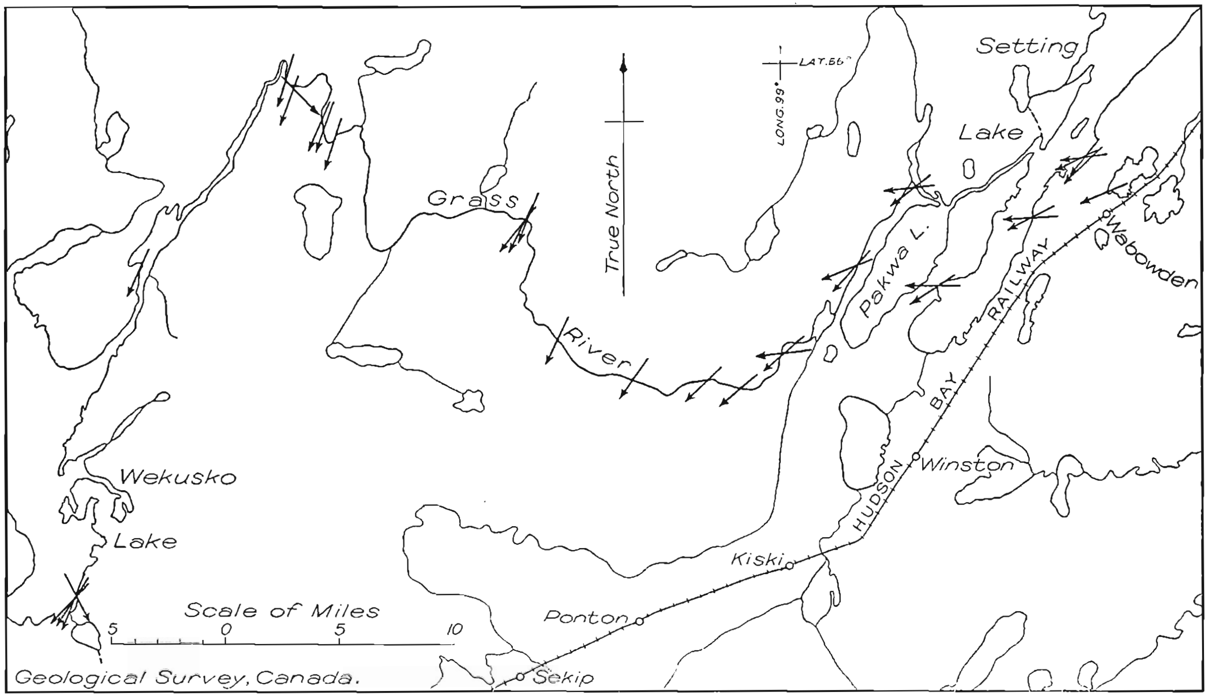

Figure 6. Glacial striæ, Wekusko Lake-Setting Lake area, Manitoba (*See* Figure 5).

Agassiz since they probably represent outwash and eskers. Consequently varved clay probably does not occur at all east of mile 265 and perhaps not east of mile 246.

WEKUSKO LAKE

The flat-lying dolomite just south of Wekusko lake has a thin cover of till. Peat and clay are first met north of the escarpment of dolomite on the border with the Precambrian at the south end of the lake. Clay occurs locally on the east shore as far north as Puella bay. The shore north of here and the southern and western shores are rocky and, for the most part, rather high.

GRASS RIVER

The clay on a small island and in which profile 176 was measured, was the only clay observed in the long, narrow bay from which Grass river flows. The high clay ridge in which section 177 was obtained, about 2 miles north of locality 176, was the only clay seen in that vicinity, the beautiful reach of the river, $6\frac{1}{2}$ miles long, to the south line of township 69, having rocky shores and islands with little soil. The change at the township line is as sudden as complete. The west river bank consists of muskeg and swamps from here to a mile south of the first abrupt turn from a north to a south course and the east river bank is made up of swamp, muskeg, till, and bedrock. The short easterly deflexion of the river a mile above the abrupt turn is due to a drift ridge on the north side. The river makes the abrupt turn to find a gap in a rock ridge. At and just southeast of this turn is varved clay (locality 178). Clay also forms banks, 5 to 10 feet high, on the two easterly reaches in township 69, range 14, whereas swamps border the river along the intervening north-south stretches. Beyond locality 179 low clay land with swamps here and there borders the south reach of the river in the eastern parts of townships 69 and 68, range 14; and largely swampy land extends along the east reach across township 68, range 13, to locality 180 at Kanisota rapids. Especially along the north-south sections, rocky ridges and knolls break the monotony.

From Kanisota rapids to Whiteforest rapids, about 10 miles southeast, Grass river flows between banks of clay and silt that average 25 feet in height, but in places reach 75 feet or more. The drift banks have slopes of 20 degrees or less and nowhere present steep bluffs. Bedrock outcrops in the first 2 miles and at one or two places farther downstream.

At Whiteforest rapids thick beds of silt and clay rest on bedrock. For 2 miles below the rapids the river is bounded by drift banks that are mostly about 30 feet high and in places are as steep as 30 degrees; but opening a cut would mean much work. For the next 2 miles, down to Skunk rapids, the land rises only 5 to 15 feet above the river. Bedrock is to be seen only just below Whiteforest rapids. At Skunk rapids there is disturbed varved silt, 7 feet thick, on bedrock. Between Skunk rapids and Whitewood falls, the land is largely low, clayey, at places even swampy. For 3 miles below Whitewood falls to the rapids, the river banks of drift vary in height, but for the most part rise some 15 feet above the river. Many outcrops of bedrock occur, especially in the first half of the reach.

Below the rapids, small exposures of bedrock occur nearly everywhere. The sediments form lower and lower banks as we proceed downstream, and rocky knolls project from swamps. Swamps with rocky ridges and hummocks border the river for the last 15 miles, from south of Mitishto lake to Setting lake.

SETTING LAKE

On the east side of Setting lake south of Wabowden portage (opposite the mouth of Grass river) clay occurs occasionally in the small bays that indent the rocky shore. The erosion bluffs are mostly moderately steep. The clay is almost everywhere disturbed by sliding. On the west side of the south part of the lake, clay forms a continuous bed on the bay where the sections 181-185 were obtained. North of here and on islands in the northern part of the lake, clay is lacking or is in small volume. The localities where varved sections were obtained are the main clay occurrences.

SETTING LAKE TO LANDING LAKE

On Grass river, between Setting lake and Paint lake, no clay, or only very thin clay, was observed. Below Lynx falls, 3½ miles below Setting lake, the river for ½ mile or less flows through a rock canyon of, probably, Post-Glacial age throughout. Clay covers part of the land crossed by the portage between Paint lake and Wintering lake. Clay caps islands in the northwestern of the two main water bodies forming Wintering lake, and forms bluffs on the east side of the hook-like termination of the peninsula dividing the northwestern and southern branches of the lake. On the southern branch varved clay occurs everywhere. It probably covers the narrow neck of land between Wintering and Landing lakes.

LANDING LAKE

On the western part of Landing lake, clay sufficiently deep for varved profiles is scarce and was seen only where the sections were obtained (localities 209-213). Clay seems to occur for a reach of about 2 miles on the north shore west of the Principal meridian. East of here clay is again scarce to a place about a mile west of the large point projecting south from the northern shore about 5 miles west of the fork of the lake. West of and on this point, the clay proved to be generally contorted. On the south shore and on islands in the eastern part of the lake, clay is widely distributed, but frequently contorted. The disturbances, which at places are very pronounced, may have been caused by grounding icebergs, by frost, and perhaps also by other agencies. Landing river (draining the lake to Nelson river), above the first rapid, the only part passable by a canoe, flows through low or swampy land.

NELSON RIVER

Nelson river has, for the first few miles below Cross portage, at the lower end of Sipiwesk lake, low or moderately high banks of drift resting on bedrock at or below the high-water mark. At 4 and 6 miles below the portage there are, on the south side, exposures, 50 and 35 feet high respectively, showing fine sand and silt with ice-rafted boulders. Below

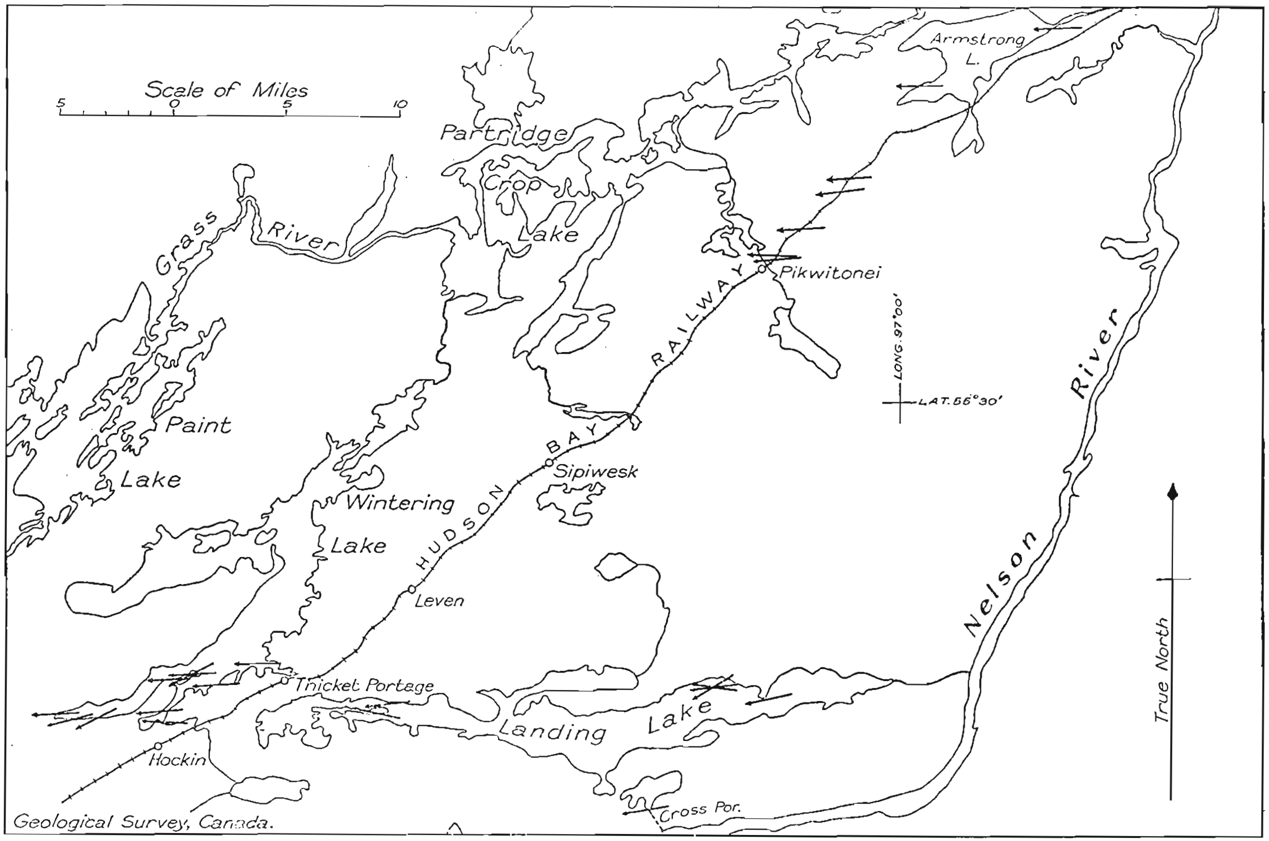

Figure 7. Glacial striæ, Landing Lake-Armstrong Lake area, Manitoba (*See* Figure 5).

here the bedrock rises higher on both sides of the river, rising up to some 100 feet. It is largely bare, but in many places is covered by a thin soil which nourishes a poor forest growth. Clayey silt is to be found only here and there, in depressions. Below the bend of the river there seem to be no sediments except that one-half mile north of the latitude line 55° 17′ north, there are, on the west side of the river, bluffs, 100 or more feet high, exposing sandy silt which in the upper part contains a great many boulders and cobbles. The boulders may either have been ice-rafted or directly deposited by the ice during a re-advance over the silt. Except for these bluffs, high, rocky banks of the pre-Glacial channel border the river onward to the mouth of Landing river, the stream draining Landing lake to the Nelson.

There, Nelson river completely changes its aspect. The west bank now consists of sediments through which the bedrock frequently protrudes. The east bank is formed by sediments with rare rock outcrops. This latter bank is in many cases steep and 30 to 40 feet high. Three miles below Landing river, a bluff on the east side proved to consist of silt all the way to the top. Five miles below Landing river, bluffs, 30 feet high, are made up of silt capped with a few very thick clay varves. North of there down to locality 233, the east bank is low and swampy with occasional rock outcrops. The west bank is a little higher consisting of sediments and till, or of bedrock. Between localities 233 and 234 the west river bank, some 20 feet high, consists largely of bedrock with or without till. Where forest fires have raged, the bedrock is practically naked showing that the soil cover is very thin. Sediments occur in depressions, but nowhere form bluffs. The east river bank is largely of sediments forming a low bank with occasional high bluffs. A bluff 2½ miles below locality 233, showed 25 feet of gravel, 25 feet of silt, and a thin top layer of clay. Another bluff 2 miles farther down stream, presented 40 feet of silt with a thin layer of brown, weathered clay at the surface.

Between locality 234 and Small Devil rapids, 2½ miles above the first railway crossing, the west bank is mostly formed of bedrock rising 50 or more feet above the river. The east bank consists largely of sediments and includes several bluffs, 20 to 40 feet high. These are essentially composed of silt and sand, but occasionally with gravel at bottom and clay at top.

DESCRIPTIONS OF SECTIONS STUDIED

For the sake of clarity the clay localities in Canada described in the author's earlier papers have been given running numbers, as have those in the United States (Antevs, 1925, page 95; 1928, pages 172, 203). The number of the Canadian sections that have been treated previously, is 174, all situated in Quebec and Ontario. The southwesternmost clay profile here described, on Wekusko lake, has, therefore, been given number 175; the northeasternmost, on Nelson river, has been numbered 235.

The obtaining of clay profiles in northern Manitoba encountered difficulties connected with ground-frost. In the middle of July the men who dug the foundation for the station house at Wabowden, mile 137 from The Pas, reached frozen ground at a depth of 3 feet, while they were scorched by the sun at a temperature of nearly 90 degrees F. in the shade. The place is open, level, and relatively high. The ground consists of varved clay silt. The frozen ground was so hard that the picks made very slow progress. In level ground, frost was encountered throughout the summer at depths that varied with the advancement of the season, the exposure to the sun, the nature and moisture of the ground, and other conditions, but was everywhere less than 8 feet. When it was necessary to obtain clay profiles in frozen ground, this had to be opened up with an ax.

The most important consequence of the frozen condition of the ground, from the point of view of chronological clay studies, is, however, the absence of high, shear erosion bluffs in undisturbed clay. Bluffs do occur here and there, but instead of presenting undisturbed strata in the face or under a thin cover of talus, they invariably have a thick talus cover and under this slightly slumped clay extending inward to permanently frozen ground. The reason for this condition is that the contact between thawed and frozen ground forms a shear zone sloping outwards. In many cases the ground behind the bluffs is traversed by one or more deep, partly open cracks, probably caused by the combined action of slumping and frost. On account of these conditions to obtain a clay section in a bluff may mean digging a trench so deep and long that in spite of the frozen condition of the ground it frequently is less work to dig down from the level clay plain. Probably because of the frost which causes the unconsolidated strata to move even on very small gradients, the clay is more disturbed than in any other region in the writer's experience.

In each of the following sections, the beds are enumerated in order of occurrence from the top downward.

(175) East shore of Wekusko lake, 50 yards west of Fosler's house.

1½ feet leached clay.
1½ feet clay with varves too thin and indistinct to be measured.
4 feet lead-coloured clay consisting of thin, not very distinct, varves, 168 in number. Occasional varves that are abnormally thick contain grey-white silt. Varve No. 1 is bottom varve.
Till.

(176) Small island in Grass river where it leaves Wekusko lake.

½ foot leached clay with cobbles. Overrun by ice?
2¾ feet leached clay without cobbles.
2¼ feet lead-coloured, thinly varved clay with slightly disturbed zone. The thick varves contain white silt. Series correlated with that of locality 177, varves 1-57, 86-124. Varve No. 1 is bottom varve.
Boulder clay.

(177) About 2 miles north of locality 176, bluff on west side of Grass river, 200 yards south of point.

2½ feet leached clay.
2 feet somewhat disturbed clay, 44 varves, which thus are relatively thick, averaging 1·4 centimetres. Varves consist of about equally thick layers of white silt and chocolate-brown clay.
4 feet 5 inches (135 centimetres) chocolate-brown clay. The varves even in the lower part are too thin to be measurable; and they become thinner and thinner upwards,

so that the upper half of the bed is massive. The number of varves being greater than 30 in 10 cm. of clay, as is evident from the bed just below, the total number of years represented by this thinly bedded, massive clay is more than 400, probably much more.

6 feet chocolate-coloured clay with occasional white silt layers, marking drainages. Varves distinct though thin. The topmost 4 inches (10 cm.) of clay contain 26 varves; the topmost 10 inches (25 cm.) contain 70 varves; and the topmost 20 inches (50 cm.) contain 134 varves. The series is correlated with that of locality 176. Varves measured, 19-256, 261-313, a disturbed zone, 2 cm. thick, being arbitrarily estimated to represent 4 varves. Varve No. 19 is bottom varve.

½ cm. till.

Horizontal bedrock.

(178) 350 yards south-southeast of sharp turn of Grass river, 25 yards east of left river bank.

1½ feet silty soil.

½ foot brown, stiff, massive clay.

⅔ foot silty clay with contorted varves.

8 inches silty clay, disturbed, so that thickness of varves not good, varves 23-34.

5 inches stiff, brown clay, also disturbed, varves 6-22.

1⅓ feet (40 cm.) chocolate-coloured, practically massive clay.

1⅓ feet chocolate-brown clay. Varves too thin and disturbed to be measurable.

2 feet fine-grained chocolate-brown clay, with several layers of white silt. Varve limits frequently difficult to distinguish from other sedimentation limits. Series not correlated with any other. Varves 15-86.

Sharp limit, caused by very marked reduction in supply of material.

1¼ feet varved silt, varves 1-14. Varve No. 1 is bottom varve.

Bedrock.

(179) 75 yards east of rapid, a few hundred yards north of the sharp eastward turn of Grass river.

3 feet covered.

1 foot stiff, brown, massive clay.

1½ feet silty, coarse clay, contorted.

4 feet brown, silty, hard clay, with disturbed and indistinct varvity, 27 varves. The thicknesses in millimetres of the varves in order of deposition are: 73, 17, 61, 124, 50, 101 (?), 33 (?), 56, 68, 27, 40, 53, 57, 47, 61, 63, 41, 19, 29, 33, 22, 31, 48, 37, 17, 17, 15.

Contorted clay. Depth to substratum unknown.

(180) North bank of grass river 100 yards below Kanisota rapids.

Hill rising gently 10 feet higher.

1½ feet leached, stiff, brown clay soil.

3 feet silty clay, disturbed.

2⅔ feet silty clay, 19 varves. The thicknesses in millimetres of the varves in order of deposition are: 64, 127 (?), 62, 50, 45, 39, 43, 54, 40, 25, 32, 18, 17, 20, 17, 19, 16, 17, 15.

2 feet massive, stiff, chocolate-brown clay with scattered cobbles at top. Eventually varved clay overridden by advancing ice.

10 feet covered to river-level.

(181) 9 miles southwest of Wabowden, west side of Setting lake, ¼ mile north of township line 68, bluff in deep bay.

4 feet leached clay.

3½ feet well varved clay, varves 10-83. Varves consist of cream-coloured silt and of brown clay layers.

2 feet, more or less disturbed, silty clay.

Bedrock.

(182) 100 yards northeast of locality 181, shore bluff.

5 feet leached clay.
4 feet distinctly varved clay, varves 1-61. Varve No. 1 is bottom varve.
Bedrock.

(183) 700 yards northeast of locality 181, on north shore of bay, 200 yards west of point.

5 feet leached clay.
11 inches varved clay, varves 37-54.
2 inches disturbed zone.
2½ feet silty clay, 14 varves, not surely connected with the normal curve. The thicknesses in millimetres of the varves in order of deposition are: 80, 87, 65, 67, 92, 36, 57, 56, 47, 41, 35, 41, 42, 23.
1½ feet disturbed silt, probably 1 or 2 varves.
Bedrock.

(184) 150 yards east of locality 183.

6 feet, more or less disturbed and leached varved clay.
1⅓ feet clay, varves 27-54.
5 inches disturbed zone.
1½ feet silty clay, 7 varves, correlated with the bottom varves at locality 183, but not with the normal curve. The thicknesses in millimetres of the varves in order of deposition are: 95, 67, 60, 62, 80, 27, 52.
> 8 inches silt, perhaps bottom varve, since bedrock outcrops close by.

(185) 300 yards north-northeast of locality 184, 250 yards north of point mentioned under locality 183, bluff in little bay.

5 feet, more or less disturbed and leached clay.
1⅔ feet clay, varves 33-68.
8 inches disturbed clay.
1 foot till.
Bedrock.

(186) 4½ miles west-southwest of Wabowden, east shore of Setting lake, bluff at end of small bay.

7 feet stiff, red-brown clay. Varvity disturbed.
3 feet silty clay. Varves consist of yellow silt and of red-brown clay layers, probably varves 17-33. The thicknesses in millimetres of the varves in order of deposition are: 70, 231, 71, 70, 68, 89, 36, 59, 65, 56, 46, 31, 29, 22, 18, 17, 22.
Practically bottom of the clay.
Frozen ground.

(187) 4½ miles west-northwest of Wabowden, west shore of Setting lake near end of peninsula south of mouth of Grass river.

6 feet disturbed clay.
2¼ feet well-varved, silty clay, probably varves 16-35.
1+ feet silt, perhaps bottom varve. Bedrock outcrops close by. The thicknesses in millimetres of the varves in order of deposition are: 44, 60, 64, 54, 37, 30, 38, 39, 40, 41, 35, 37, 23, 26, 21, 15, 22, 23, 15, 19.

(188) $2\frac{1}{2}$ miles north-northwest of Wabowden, bluff on Setting lake 200 yards southwest of Wabowden portage.

3 feet leached and disturbed clay.
6 feet red-brown clay, mostly rather stiff, varves 77-276. The clay is cut through by several small faults; but with exception of one or two places the curve here represented is probably correct.
3 feet clay, varves 28-76. The varves are undisturbed and very distinct, consisting of cream-coloured silt and red-brown clay layers. The lowest varve is probably No. 7 from bottom.
1 foot disturbed silt with remains of winter layers, probably 5 varves.
5+ inches sandy silt, hard packed, probably varve No. 2 and the bottom varve. Bedrock outcrops close by.

(189) $5\frac{1}{2}$ miles north-northwest of Wabowden, west shore of Setting lake, north end of little island at an Indian cabin.

3 feet leached clay.
8 inches clay, varves 144-160.
6 inches disturbed zone.
5 feet clay, beautifully varved by alternating layers of cream-coloured silt and brown clay, varves 26-104. Varve 26 is certainly the bottom varve.
Bedrock.

(190) About 10 miles north of Wabowden, and 4 miles north-northeast of locality 189, bluff on west shore of Setting lake.

5 feet disturbed and weathered clay.
5 inches clay, varves 70-175.
2 inches disturbed clay.
1 foot clay, varves 142-168. Thickness not fully characteristic.
4 inches disturbed zone, representing varves 119-141.
2 feet clay, varves 83-118. Varves 95-103 are repeated by sliding—a fact hardly noticeable in the exposure.
4 inches disturbed zone. No varve missing.
4 feet well-varved clay, varves 42-81. Curve agrees well with the others, but the varves are somewhat thicker.

(191) About $12\frac{1}{2}$ miles north-northeast of Wabowden, bluff at northwest corner of island on the south side of the northernmost strait connecting western part of Landing lake with its eastern part.

3 feet weathered and disturbed clay.
2 feet 3 inches well-varved clay, 33 varves, not correlated with any other series. The thicknesses in millimetres of the varves in order of deposition are: 26, 20, 18, 22, 26, 30, 23, 24, 16, 17, 26, 30, 26, 40, 28, 28, 29, 25, 19, 26, 15, 13, 20, 16, 14, 17, 15, 22, 16, 15, 9, 8, 10.
Discordance.
2 feet 2 inches well-varved clay, varves 62-106.
Discordance.
7 feet fine sand.
Level of Setting lake.

(192) $\frac{1}{2}$ mile north of locality 191, bluff on west side of island.

$1\frac{1}{2}$ feet of weathered and contorted clay.
1 foot clay, 18 varves, not correlated with any others. Thicknesses of varves in millimetres in order of deposition are: 15, 18, 17, 13, 12, 13, 12, 19, 37, 32, 12, 15, 28, 16, 15, 13, 8, 12.
2 inches contorted clay.
Sand.

(193–197) Long, high railway cut ½ mile south of Tooth lake, at mileage 154½. Sections were measured at points 200, 150, 135, 125, and 115 yards south of the telephone post marking mile 154½.

(193) 200 yards south of mile 154½.

1½ feet leached clay.
3½ feet well-varved clay. Varves consisting of white silt and red-brown clay layers, varves 136-233.
2 inches slidden zone, representing 3 varves.
4 feet clay, distinctly varved by alternate layers of white silt and brown clay, varves 31-132.
4 to 6 inches contorted clay and till.
1 foot 8 inches well-varved silty clay, varves 4-25.
1 inch disturbed clay, probably 1 varve.
7½ inches sandy silt, 2 varves.
Sand, probably bottom varves.

(194) 150 yards south of mile 154½.

2 feet leached clay.
3 feet well-varved clay, varves 145-221.
3 inches slidden zone.
1¾ feet distinctly varved clay, varves 89-134.
Covered. Far to bottom.

(195) 135 yards south of mile 154½.

2½ feet leached and disturbed clay.
2½ feet well-varved clay, varves 110-164.
Covered. Far to bottom.
Series measured: 110-164, 127-159.

(196) 125 yards south of mile 154½.

2½ feet leached and contorted clay.
1 foot clay, varves 174-196.
Shear zone, hardly noticeable, but representing varves 128-173.
4½ feet clay with good varvity, varves 44-127.
1 foot 10 inches pressed, varved clay.
1+ feet silt, sand, and pebbles. Probably a few disturbed bottom varves.

(197) 115 yards south of mile 154½.

2½ feet leached clay.
2 feet 2 inches silty clay, varves 179-218.
Disturbed zone.
Covered. Far to substratum.

(198) Bluff on the south side of the southwesternmost end of Wintering lake, 10 miles west-southwest of Thicket portage.

4 feet leached clay.
10 feet clay, distinctly varved by alternating layers of white silt and red-brown clay, varves 1-121.
1½ feet quicksilt, a few contorted varves.
Till.

(199) 3 miles east of locality 198, bluff at point on Wintering lake.

4 feet leached clay.
3¼ feet well-varved clay, varves 23-72. A few bottom varves have perhaps slidden away.
Bedrock.

(200) Bay on Wintering lake; at a slide 100 yards west of mouth of McLaren creek, 6 miles west-southwest of Thicket portage.

7½ feet leached and disturbed clay.
4 feet well-varved clay of light silt and brown clay, varves 47-92. The clay is cut through by small faults. The lowest varve measured is sandy and the distance to bottom may not be great.

(201) 350 yards northwest of locality 200, south shore of long peninsula in Wintering lake.

2 feet leached clay.
4 feet clay. The curve does not correspond very well with other series, but probably represents varves 34-92. A few varves probably washed away.
Till on bedrock.

(202) South shore of Wintering lake, 5½ miles west of Thicket portage.

2½ feet leached clay.
6 feet distinctly varved clay, varves 22-93.
Bedrock, which, however, is uneven, so that there may be one or two varves below No. 22.
Series measured: 22-93, 37-63, 39-56.

(203) ½ mile northeast of locality 202, 1½ miles west of point of long peninsula, Wintering lake.

1½ feet leached clay.
2½ feet clay with good varves of white silt and brown clay, varves 27-65.
4 inches clay, representing one or two slidden varves.
Bedrock.

(204) 1 mile east of locality 203, and ½ mile west of point of long peninsula, north shore of peninsula, Wintering lake.

2 feet leached clay.
4 feet clay, varves 35-93. Varve 35 is bottom varve.
Till.

(205) Bluff at little bay on southeast shore of Wintering lake, 3 miles west of Thicket portage.

1 foot leached clay.
1¾ feet clay, varves 100-134.
2 inches disturbed clay; 6 varves missing.
3½ feet clay, varves 47-93.
3 inches disturbed silt, one or two varves.
Sand. 8 feet covered to shore-line.

(206) 1¾ miles northeast of locality 205, Wintering lake, bluff on southwest part of big island, off inlet to Thicket portage.

4 feet weathered and contorted clay.
9 inches clay, probably varves 47-57. The thicknesses in millimetres of the varves in order of deposition are: 55, 31, 23, 23, 12, 18, 8, 11, 21, 11, 13.
7 inches silt, one or two varves.
Bedrock.

(207) $1\frac{1}{4}$ miles west-northwest of Thicket portage, Wintering lake, bluff on west side of island north of inlet to the portage.

3 feet leached and disturbed clay.
7 inches clay, varves 49-59. Varve 49 is bottom varve. The thicknesses in millimetres of the varves in order of deposition are: 29, 15, 9, 10, 12, 12, 17, 11, 10, 14, 16.
Till and bedrock.

(208) $\frac{3}{4}$ mile north of Thicket Portage station, bluff on point in bay, Wintering lake.

4 feet leached clay.
$1\frac{1}{3}$ feet clay, varves 61-78.
14 inches silt and sand, bottom varve, No. 60.
10 feet gravel to lake shore. The thicknesses in millimetres of the varves in order of deposition are: 48, 21, 31, 25, 17, 25, 20, 28, 29, 22, 30, 17, 24, 20, 17, 13, 14, 20, 22.

(209) Bluff at west end of Landing lake $2\frac{1}{3}$ miles south-southwest from Thicket portage.

$1\frac{1}{2}$ feet leached clay.
3 feet 8 inches distinctly varved clay, layers of light silt and brown clay, varves 48-94. Varve No. 48 is bottom varve.
$\frac{1}{2}$ inch gravel on bedrock.

(210) $3\frac{1}{2}$ miles east of locality 209, west side of island, 400 yards from west end of bay $2\frac{1}{2}$ miles long, Landing lake.

3 to 4 feet leached and slidden clay.
$2\frac{1}{2}$ feet well-varved clay, varves 66-107. Varve No. 66 is bottom varve.
Till.

(211) $1\frac{1}{2}$ miles east of locality 210, bluff on the south shore and $\frac{1}{2}$ mile from end of long and narrow peninsula, Landing lake.

4 feet weathered clay.
$4\frac{1}{2}$ feet distinctly varved, yellow-white clay with relatively thin winter layers, varves 81-154.
1 foot yellow-brown clay with relatively thick winter layers, varves 74-80. Varve 74 is bottom varve.
Till.

(212) 5 miles east-southeast of Thicket portage, $\frac{3}{4}$ mile east of locality 211, bluff on island 400 yards east of end of the long peninsula, Landing lake.

4 feet leached clay.
1 foot clay, varves 130-154.
2 inches disturbed zone.
10 inches clay, varves 107-123.
1 inch contorted clay.
1 foot clay, varves 89-99.
2 feet disturbed, coarse, varved silt.
Lake-level.

(213) Bluff 50 yards east of locality 212, Landing lake.

3 feet leached clay.
$2\frac{1}{2}$ feet well-varved clay, varves 78-128. At this place, as well as at locality 212, the silt is white up to varve 113, but brown upward from varve 114. Varve 78 is bottom varve.
Thin till and bedrock.

(214) About 15 miles east-southeast of Thicket portage, south side of Landing lake, bluff on small island just off end of peninsula around which leads the route to the small lakes and to Cross portage.

2½ feet leached clay.
2 feet stiff, brown clay, varves not measurable.
4 feet clay distinctly varved by alternating layers of white silt and brown clay, varves 129-200. Varve 129 is bottom varve.
Till.

(215) North shore of Landing lake, about 17 miles east of Thicket portage.

1½ feet leached clay.
2½ feet clay, probably varves 143-185. Varve No. 143 is bottom varve.
Till.

(216) Small point on south shore of Landing lake, one mile west-southwest of the large point on the north shore 5½ miles west of foot of lake; in forest, 15 yards inland. The section was, on August 18, frozen almost throughout.

4½ feet leached and disturbed clay.
8 inches distinctly varved clay, varves 187-201. Both silt and clay brown.
1¾ feet rather well-varved clay, varves 161-186. Silt white, clay brown. Varve 161 is bottom varve.
Till.

(217) About 2¼ miles east of locality 216, island 400 yards from south shore, Landing lake.

2 feet leached clay.
1¾ feet well-varved clay, varves 177-203. Varve No. 177 is bottom varve.
Till and bedrock.

(218) About 1⅓ miles east of locality 217, 3 miles west-southwest of east end of Landing lake, 100 yards inland from the south shore of the lake. August 15.

3 feet weathered and thawed clay.
1¼ feet brown, frozen clay, varves 184-211.
Till.

(219) ½ mile east of locality 218, peninsula on the south shore of Landing lake, 25 yards inland.

1½ feet leached clay.
2½ feet clay, varves 186-215.
Till.

(220) ⅓ mile east-northeast of locality 219 and about 2 miles west of east end of Landing lake, small island near south shore.

3 feet leached clay.
11 inches stiff, brown clay, varves difficult to distinguish, numbers 226-255.
3½ feet clay, distinctly varved by alternating layers of white silt and brown clay, varves 188-225.
Gravel and bedrock.

(221) 1 mile east of locality 220, and $1\frac{1}{4}$ miles west of east end of Landing lake, centre of a small island.

2 feet leached clay.
$1\frac{1}{2}$ feet stiff, brown clay, varves very indistinct, numbers 227 to about 275.
$3\frac{1}{4}$ feet clay, well varved by layers of white silt and brown clay, varves 194-226.
Till.

(222–226) Mile 214 on Hudson Bay railway, just northeast of Pikwitonei village and river, a long railway cut. The mile post 214 stands 50 yards northeast of the river. Varve series were measured:

30 yards southwest of the mile post and west of track, locality 222.
15 yards south of mile post and east of track, locality 223.
30 yards west of mile post and west side of track, locality 224.
50 yards north of mile post and east side of track, locality 225.
250 yards north of mile post and west side of track, locality 226.

(222)

$1\frac{1}{2}$ feet leached clay.
2 feet clay. The varvity looks satisfactory, but a comparison with other series shows that it is slightly disturbed.
$5\frac{1}{2}$ feet clay, distinctly varved by alternating layers of white silt and red-brown clay, varves 200-286. Varve 200 is more than 1 foot thick and is probably bottom varve.
Sand.

(223)

$1\frac{1}{2}$ feet leached clay.
$4\frac{1}{2}$ feet clay with slightly disturbed varvity, as at locality 222.
$2\frac{1}{4}$ feet well-varved clay, varves 201-252.
Sand.

(224)

$3\frac{1}{2}$ feet leached and disturbed clay.
$3\frac{3}{4}$ feet well-varved clay, varves 201-260.
$1\frac{1}{2}+$ feet sand.

(225)

1 foot leached clay.
$1\frac{3}{4}$ feet pressed, varved clay.
$2\frac{1}{4}$ feet clay with good varvity, varves 202-249.
Sand.

(226)

$1\frac{1}{2}$ feet leached clay.
2 feet varved clay, somewhat disturbed.
2 feet clay, varves 203-228. Varve 203, the bottom varve, is 4 inches thick and contains gravel and sand.
Till.

(227) 1 mile north-northeast of Pikwitonei, railway cut at mile $214\frac{3}{4}$, west of track.

$3\frac{1}{2}$ feet leached clay.
$3\frac{1}{2}$ feet varved clay, more or less disturbed.
3 feet clay with rather good varvity, varves 206-268.
A few contorted varves.
Till.

(228) $2\frac{1}{2}$ miles northeast of Pikwitonei, railway cut north of telephone post marking mile $216\frac{1}{2}$, south side of the track.

$4\frac{1}{2}$ feet disturbed clay.
$2\frac{1}{4}$ feet red-brown clay, varvity somewhat disturbed by frost, varves 213-244.
Till.

(229) $5\frac{1}{2}$ miles northeast of Pikwitonei, mile $219\frac{1}{2}$, ditch 75 yards southeast of the railway track.

3 feet leached clay.
$1\frac{2}{3}$ feet stiff, red-brown clay, disturbed by frost.
$1\frac{1}{3}$ feet similar clay, only little disturbed by frost, probably varves 217-238.
Gravelly till.

(230) Small island off south point of northernmost large bay, west side of Armstrong lake, at centre of the island.

$1\frac{1}{2}$ feet leached clay.
$4\frac{2}{3}$ feet clay, beautifully varved by layers of white silt and brown clay, varves 240-297. Varve 240 is bottom varve.
Bedrock.

(231) Same island as locality 230, at the shore.

$1\frac{1}{2}$ feet leached and disturbed clay.
$2\frac{1}{4}$ feet clay, varves 240-263.
Bedrock.

(232) Armstrong lake, bluff on mainland $\frac{1}{2}$ mile west-northwest of the outlet of the lake.

2 feet leached clay.
9 feet clay, distinctly varved by alternating layers of white silt and brown clay. The clay is somewhat pressed in one or two horizons, but on the whole the thicknesses are good. Varves 266-374.
A little below lake-level. Depth to bottom unknown, but probably not great.

(233) Bluff on east side of Nelson river, north of mouth of a small tributary, about latitude 55° $27\frac{1}{2}'$ north.

6 inches silt, sand, and peat.
6 inches brown, weathered clay.
10 inches clay, varved by layers of white silt and brown clay, 18 varves, not connected with the normal curve. Thicknesses of varves in millimetres in order of deposition are: 21, 10, 13, 14, 9, 10, 18, 17, 10, 7, 9, 14, 13, 14, 14, 21, 17, 15.
6 inches contorted clay.
50 feet fine sand and silt. No varves observed.
River-level.

(234) About 5 miles north of locality 233, bluff on the east bank of the Nelson.

3 feet disturbed and leached brown clay.
$2\frac{3}{4}$ feet clay, distinctly varved by layers of yellow-brown silt and red-brown clay. Varves 295-335.
3+ feet white silt. No varve limits observed.
20 feet concealed.
River-level.

(235) A few hundred yards north of south line of township 77, bluff on the east side of Nelson river.

1 foot leached clay.
1 foot silty clay, probably varves 312-333. The thicknesses in millimetres of the varves in order of deposition are: 18, 30, 11, 16, 19, 25, 8, 6, 13, 8, 5, 6, 5, 9, 12, 16, 7, 18, 22, 12, 21, 13, 20.
4 feet silt, gravel, and cobbles—probably till.
50 to 60 feet sediments mostly covered, but, where exposed, silt.
River-level.

NOTES ON THE VARVE CURVES

The varve curves here presented (Plate I) were constructed in the same way as those previously published (Antevs, 1922, page 47; 1925, page 120; 1928, page 224). The individual graphs of the sections as measured in the field were matched and corrected for numbers of varves. Then the curves or the parts that represented varves of normal variation and thickness were selected for calculating the "normal" thickness of each single varve, and the obtained values were used in plotting the normal or average curves—the curves here presented. Graphs of the same varve series that differed in thickness or in other important respects are also presented; abnormal varves have been indicated by dashed lines.

Wekusko Lake, Locality 175 (Graph No. 1, Plate I)

Many of the varves are so thin that they were difficult to measure. Since, furthermore, those indicated by dashed lines contain silt and are abnormally thick, the curve is not good.

Grass River, Localities 176 and 177 (Graph No. 2, Plate I)

1-18. Curve probably good, though varves thin.

19-57. Aside from excessively thick drainage varves, the curve is good, since the two measurements on which it is based agree well.

58-85. Several varves indicated by dashed lines contain silt and are too thick to be considered as "normal" records.

86-124. Curve good, since the measurements match perfectly. A few varves, though, are silty and abnormally thick.

125-313. The curve may be fairly good, as the varves are distinct though very thin, and though only one measurement is at hand. A disturbed zone of 2 cm. after varve 256 has arbitrarily been estimated to contain 4 varves.

Grass River, Localities 177 and 178 (Graphs Nos. 3 and 3A, Plate I)

These graphs represent the varves above the massive clay at each locality. Each graph represents the average of two measurements. The graphs should probably be correlated as shown, though their correspondence is only moderate. The varves are disturbed at both localities.

Grass River, Locality 178 (Graph No. 4, Plate I)

This series is not correlated with any other.

1-14. Curve probably fair. The relatively thick varves indicate moderate supply of material from the ice.

15-86. Although some varves are abnormally thick, the varves on the whole are too thin to be good records of the summer temperature. The striking change in thickness beginning varve 15, is too abrupt to be due to change in the amount of ice melting; it may have been caused by diversion of the glacier mud.

Setting Lake, Localities 181-190 (Graphs Nos. 5 and 5A, Plate I)

1-25. The curve may be good.

26-83. The different measurements within the groups agree well with each other. The two curves also correspond well and may faithfully record melting and summer temperature.

No. 44 should perhaps be two varves, since it shows a dark zone that in places is so distinct as to give the impression of being a true winter layer.

84-118. The different measurements correspond from fairly well to well; and the normal curve may, consequently, be good.

119-143. Only one measurement which may be fairly good, except that varves 128 and 129 may be one varve.

144-160. Curve good as the two measurements composing it agree very well.

161-174. The two curves show fairly good agreement. The varves 168-172 are, though, somewhat pressed.

175-210. Only one section, but it is fairly good. Varves 175, 178, and 213 contain silt and may be too thick.

211-276. Not very good. Nos. 224-226, 238, and 269 are silty and too thick. Nos. 254, 255, and 256 perhaps represent a single varve.

Hudson Bay Railway, Mile 154½, Localities 193-197 (Graph No. 6, Plate I)

This varve series is not correlated with either that from Setting lake or that from Wintering-Landing lakes.

1 and 2. Abnormally thick, having been deposited directly after the uncovering of the ice.

3. Slidden.

4-25. Good. The gap is of unknown magnitude. The series above it has been arbitrarily begun with No. 31.

31-233. The curve may be a good record of the rates of ice melting. The varvity of the clay is very distinct. The different measurements, but which were made close to one another, show excellent agreement. A few varves, marked by lines of dashes, contain silt and probably record drainages. About varve 135 there is a slidden zone in most of the sections, but it is bridged in section 195. Between varve 155 and varve 185 a few varve limits are locally difficult to distinguish, and other winter layers are indistinguishable as a result of pressing.

Localities 198-234

Localities 198-205, Wintering Lake (Graph No. 7, Plate I)

Localities 209-221, Landing Lake (Graph No. 7A, Plate I)

Localities 222-227, Pikwitonei; Locality 230, Armstrong Lake; Locality 232, Armstrong Lake; Locality 234, Nelson River (Graph No. 7B, Plate I)

1-47. The different measurements correspond well with one another. Some of the varves are rather thick, but since in them the proportion between summer layers and winter layers is about the same as in the thinner varves, they may be normal and record warm summers.

48-61. The series from Wintering lake may be good; the series from Landing lake, consisting of varves deposited close to the ice border, is less dependable.

62-134. The Wintering Lake and the Landing Lake series show from good to excellent correspondence and may thus faithfully record summer temperature and ice melting. Because of the rapid upward thinning out of the varves in the sections on Landing lake, four separate curves have been constructed from the graphs obtained there.

Nos. 80, 121, 128, and 134 contain silt, but since they do so in both series, they probably record considerable melting.

135-154. The two series correspond well, though the series from locality 214 was deposited 10 miles closer to the ice edge than was the series from localities 211, 212, and 213.

152-172. Probably good, although based on a single locality (No. 214). Profile representing locality 215 is not so good and does not agree well with that of 214.

173-187. Good.

188-200. Although the varve series from localities 217 and 218 are very short, the two curves agree fairly well and may record the ice melting satisfactorily.

201-239. The different curves show fair correspondence, in spite of the fact that the distance between the east end of Landing lake (Graph No. 7A continued) and Pikwitonei (Graph No. 7B, localities 222-227) is 20 miles. Since the two sedimentation areas were supplied independently, the thick varves that are in common mean summers of considerable ice melting.

240-297. As on Landing lake so at Pikwitonei (Graph No. 7B, localities 222-227) and on Armstrong lake (Graph No. 7B, localities 230 and 232) the varves decrease rapidly in thickness upward in the sections. Three separate curves have, therefore, been made. The Armstrong Lake graphs agree satisfactorily with one another and with the Pikwitonei series, 12 to 16 miles distant.

298-336. The curves (Graph No. 7B) match fairly well, though they represent localities 232 on Armstrong lake and 234 on Nelson river, 15 miles apart. The correspondence between graphs from widely separated localities on Landing lake, Pikwitonei, Armstrong lake, and Nelson river, that is regions which, though all were a part of Lake Agassiz, yet surely were fed by independent glacial rivers, suggests that nearly all the thick varves are normal and record the ice melting. Varves influenced by drainages are as a rule only thick in a limited area.

337-374. Fairly good. Slight difference between measurements made in the same exposure shows that the clay has been somewhat disturbed by pressure.

CHAPTER VIII

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