CANADA
DEPARTMENT OF MINES
HON. CHARLES STEWART, MINISTER; CHARLES CAMSELL, DEPUTY MINISTER

GEOLOGICAL SURVEY
W. H. COLLINS, DIRECTOR

MEMOIR 146

No. 126, GEOLOGICAL SERIES

Retreat of the Last Ice-sheet in Eastern Canada

BY

Ernst Antevs

OTTAWA
F. A. ACLAND
PRINTER TO THE KING'S MOST EXCELLENT MAJESTY
1925

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CHAPTER I

INTRODUCTION

The geological evidence relating to recession of the last ice-sheet from New England was studied by the present writer in 1921, and the results of the work were published by the American Geographical Society of New York in 1922. The present memoir is the result of an extension of these studies northward into Canada in 1923. The method of studying the ice-retreat is that of Baron Gerard De Geer of Stockholm. It is based on the annual layers of silt and clay deposited in lakes which bordered the ice-sheet during the period of retreat or melting back. The work in New England showed that the recession of the ice-sheet from Hartford, Conn., to the region of St. Johnsbury, Vt., occupied a period of approximately 4,400 years. This period of time, however, is not connected with the chronology of our time, and represents only a fraction of the time that has elapsed since the disappearance of the ice-sheet from New England. The work in Canada was undertaken in the hope that a fairly complete record of the ice-retreat in northeastern America might be obtained. The present report deals with some problems of a general nature in connexion with the retreat of the ice-sheet and particularly the characteristics and mode of origin of the varved[1] glacial clay and the conditions controlling ice recession, as well as the rates of ice recession and the time involved. The relationship between the waning of the North American and European ice-sheets and the probable correlation of these events are briefly discussed. Although no connexion has been established, as yet, with the chronology of our time, the view is advanced that the remarkable interruption in the retreat of the ice, in the region between lake Ontario and Mattawa river northeast of Georgian bay, corresponds to the equally noteworthy interruption of the ice recession in the Danish islands and southwestern Scania, and that the halts and readvances indicated by moraines and overridden varved glacial clays at Porquis Junction in northern Ontario correspond to the halts during which the Fenno-Scandian moraines were formed.

[1] De Geer (1912, p. 253) proposed the use of the word *varve, n.*, pl. -s, English and French—*Warw*, pl. -e, German—as an international term for the distinctly marked *annual deposit* of a sediment regardless of its origin. A varve usually consists of two, but might consist of more, different layers, and in the varved late-glacial clay mostly of one layer of silt and one layer of stiff clay. The Swedish word *varv* (old spellings *hwarf* and *hvarf*, Icelandic *hwerf*) means turn, round, revolution (of a body), lap (sport term), time (wind a band once, four times round), row, tier, course, and layer. (Cf. the English words warve, *n.*, same as wherve, *n.* meaning a whirl of a spindle, and wharve, *n.* same as whorl, *n.* meaning the fly of a spinning-wheel, etc.) The present use of the word in geology is limited to what is stated above. The earlier use was wider: *varv* meant bed (of diabase—Emanuel Swedenborg, 1719), and layer (of sand, clay).

Varved, adj. (Swedish *varvig*) meaning stratified, the layers forming varves (Swedish årsskiktad). To my knowledge first used by Professor Charles Schuchert (Amer. Jour. Sci., vol. 4, 1922, page 417). Approved by Professor George L. Kittredge of Harvard University.

Varvity, n. (Swedish *varvighet*) meaning "year-stratification" (Swedish årsskiktning). Kindly proposed by Dr. Kittredge.

According to this view, therefore, the time that has elapsed since the ice retreated from the Mattawa region northeast of Georgian bay is approximately 13,500 years. A nearly complete record, amounting to 1,900 years, for the withdrawal of the ice from the Timiskaming basin in northern Ontario, has been obtained, the average rate of retreat being 454 feet annually. In the region north and east of Georgian bay several long varve series have been measured, but these have not been connected, so that no estimate of the time involved can be made from the varve series. A rough estimate, however, is possible from J. W. Spencer's and F. B. Taylor's studies of the rate of retreat of Niagara falls.

The work was done under the auspices of the Geological Survey, Canada, and the writer has great pleasure in thanking Dr. Charles Camsell and Dr. W. H. Collins for making the investigation possible and for their courtesy. During the preparation of the memoir the author has had the benefit of the advice and information of Dr. A. P. Coleman. Members of the staff of the Geological Survey have also aided by supplying information, advice, and friendly criticism. Acknowledgments are also due to Mr. A. A. Cole, Mining Engineer, of the Temiskaming and Northern Ontario railway, and to Mr. A. L. McDougall, Divisional Engineer, for many courtesies shown during the field work.

With kind permission of the Division of Geology of Harvard University material collected in 1924 with the support of the Shaler Memorial Fund has been used for connecting and dating some of the varve series in northern Ontario.

CHAPTER II

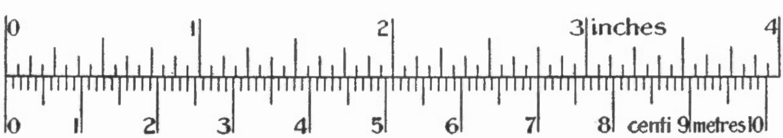
VARVED SEDIMENTS

Varved sediments are those in which the year-deposit is distinctly set off against the deposits of the preceding and following years. This condition is as a rule attained by difference in the size of the grains in the lower and upper parts of the annual deposit and usually also by darker colour of the finer-grained part. A necessary condition for annual lamination evidently is periodic supply, a large supply in summer and a small or no supply in winter. If the material is supplied at a uniform rate throughout the year and the conditions of sedimentation are uniform, a homogeneous deposit will result.

Since there are different kinds of stratified deposits somewhat resembling varved sediments, but whose lamination has no reference to the year, it is often difficult to determine whether a stratification is seasonal. If the lamination be distinct and persistent, the layers of approximately the same thickness and structure, the amount of supposed annual deposit reasonable, and if sufficient be known regarding the mode of origin of the sediments to indicate a connexion with seasonal phenomena, the lamination may with a fair amount of certainty be regarded as seasonal.

Annual lamination is of particular interest because it accurately records the rate of deposition, and the time involved, enables the establishing of a chronology, and in certain cases gives valuable data on climatic conditions. It is known to occur in greatly different sediments, and will probably be found in others. The conditions of formation are only partly known.

The varved glacial clay (Plates I, II A) is the best known and most important of the seasonally stratified deposits. Its properties and mode of origin are described in Chapters IV and V. It is here briefly referred to along with some other varved sediments which are also of considerable interest.



Conversion Tables

——	Inches to cm.	Feet to m.	Miles to km.	Cm. to inches	M. to feet	Km. to miles
1=	2·540	0·3048006	1·609347	0·39370	3·28083	0·62137
2=	5·080	0·609601	3·21869	0·78740	6·56167	1·24274
3=	7·620	0·914402	4·82804	1·18110	9·84250	1·86411
4=	10·160	1·219202	6·43739	1·57480	13·12333	2·48548
5=	12·700	1·524003	8·04674	1·96850	16·40417	3·10685
6=	15·240	1·828804	9·65608	2·36220	19·68500	3·72822
7=	17·780	2·133604	11·26543	2·75590	22·96583	4·34959
8=	20·320	2·438405	12·87478	3·14960	26·24667	4·97096
9=	22·860	2·743205	14·48412	3·54330	29·52750	5·59233
10=	25·400					
11=	27·940					
12=	30·480					

1 yard = 3 feet; $C° = \frac{5}{9}(F° - 32°)$; $F° = \frac{9}{5}C° + 32°$

VARVED GLACIAL CLAY

The seasonal character of the lamination of the glacial clay may be fully shown by the distinct, persistent, and regular stratification with upward decreasing coarseness of the material in a layer (cf. analyses by Odén, 1920, page 333, and Sauramo, 1923[1]), which indicate a regular, short, and very marked periodicity in the supply. This period must be the year with ice melting in summer and none, or little ice melting in winter, for the year is the only regular and marked period in arctic and sub-arctic regions. The seasonal character is also proved by the higher percentage of lime in the coarse-grained (summer) layers than in the fine-grained (winter) layers (Högbom, 1889, page 265; 1892, page 291), which shows that the fine material remained longer in suspension than did the coarse material. The more complete solution, doubtless, was partly due to the relatively larger surface of the fine particles. The annual periodicity of the clay now forming in lake Louise, a glacial lake in the Canadian Rockies, was directly proved by Johnston (1922a, page 383).

The varvity of the glacial clay is dependent on a series of conditions, and consequently presents all gradations from perfection to obscurity. Occasionally, as in southernmost Sweden and in Denmark, the lamination appears to be partly seasonal, partly of unknown origin.

It is on this clay that De Geer's well-known Swedish late-glacial geochronology is based (*See* Antevs, 1925 b).

late of Gold Series

VARVED GLACIAL SLATES

Varved glacial slates are known from different glaciations back to the Upper Huronian (Precambrian). Agreement with the varved glacial clays and occurrences of old till beds or tillites associated with the slates are the criteria that establish the glacial origin of these rocks. Descriptions of argillites are to be found in various papers (Sayles, 1919, pages 52-59; 1922; 1922a, 1922b, page 458; 1923; Sussmilch and David, 1919; David, 1922; Hovey, 1924; also Reeds, 1923). In Plate III two specimens of varved slate from the Upper Huronian (Collins, 1914, page 21) at Cobalt, Ontario, are shown. The slate is associated with tillite, as was pointed out by Coleman (1908). Collins (1913, pages 49, 50, 56; 1917, pages 67, 83) showed that the slate resembles perfecty the late glacial clay, and evidently is of the same origin. Not only is the structure the same, but the slate contains cobbles up to 6 inches in diameter, which, as pointed out by Collins, must have been transported by drifting icebergs. The specimens here figured are fairly thinly varved and represent deposition far away from the mouth of the glacial river. The specimen, Plate IIIA, is oxidized to a red-brown colour.

The thicknesses attained by the varved slates are remarkable. The recorded thicknesses of 1,000 feet of the late Proterozoic or Lower Cambrian slates in Australia (David, 1922) and of at least 800 feet of the Permo-Carboniferous Squantum slate at Boston (Sayles, 1919, page 43) greatly surpass the known thicknesses of late-Pleistocene clays.

[1] For publications cited see list of references at end of memoir.

VARVED POST-GLACIAL CLAY OF NORTHERN SWEDEN

Next in importance to the varved late-glacial clay comes the varved post-glacial clay-silt of central Norrland.

The formation of this clay-silt followed the deposition of the glacial varved clay, and thus it continues without break the year-registration of the glacial clay.

The post-glacial clay-silt is very thinly laminated, and shows but slight difference between the two zones that mark the year.

The fine-grained, clayey, dark grey zone, the equivalent of the winter layer in the varved glacial clay, according to Lidén (1911, page 273) is essentially deposited when the rivers are in flood during the melting of the snow in spring. It seems just as probable, however, that the coarse layer is deposited during the spring flood (*See also* Caldenius, 1924, pages 40, 86).

Central Norrland, when released from the ice, was depressed about 920 feet below the level of the Baltic and the sea. In post-glacial time it has undergone an unbroken elevation which is still in progress. Thus the mouths of the rivers and the areas of sedimentation have steadily moved outwards, and the deposits have been gradually raised above the sea, and trenched by the rivers. The clay-silt referred to forms the finest part of these fiord deposits. Since the oldest deposits are far up in the present river valleys, and the younger layers are downstream, the clay records the whole of post-glacial time except that the deposits of the last few hundred years have not been raised above sea-level. The clay forms the basis for the post-glacial geochronology which has been worked out, though not yet published, by Lidén (*See* Antevs, 1925b).

VARVED PLAYA CLAYS

Playa clays, according to Walther (1912, page 233), are in many cases annually laminated. Playa clays from the Permian of northern China are distinctly varved (Norin, 1924, pages 39-42). The varves are from 3 to 6 inches thick. The lower part of each varve consists of red-brown claystone, the upper part of greenish-yellow claystone. The transition from red to yellow is gradual and the transition zone is sandy. The lamination is due to the fact that the oxide of iron differs in character in the two zones.

VARVED GYTTJAS, ETC.

Varved gyttjas[1] are other important sediments. A seasonally stratified gyttja is now forming in Mackay lake at Ottawa (Whittaker, 1922, pages 146, 151, 152, 153). The lamination is caused by alternation of greyish-white laminæ with dark reddish layers. The grey layers consist largely of calcium carbonate in a finely divided state. The red laminæ consist of organic material with traces of calcium carbonate. The laminæ are very thin; the average thickness in the compressed sediments of one grey and one

[1] Gyttja is a sediment largely consisting of remains of plants and animals, as vascular water plants, algæ, crustaceans, molluscs, etc., and of excrements.

red layer, which may represent the annual deposition, is only 0·017 inch. The grey layer is deposited in spring, when the waters come down into the lake by many miniature torrents bringing along material from the marl beds bordering the lake. The red layer is formed during summer, autumn, and winter from algæ growth on the bottom, and from organic wind-carried detritus such as pine pollen and decomposed leaves. Only part of the deposit shows distinct lamination, so that it does not afford a means of determining the length of post-glacial time.

Varved gyttjas are not common because the bottom faunas usually thoroughly rework the surface layers (Osvald, 1922, page 36), but are found in different lakes in Sweden (Lundqvist, 1924a).

The Tertiary so called molasse at Oeningen on Bodensee presents an alternation of layers of freshwater lime, $\frac{1}{4}$ mm. to several millimetres thick, and of clay, $\frac{1}{10}$ mm. to 1 mm. thick, which as Oswald Heer could show by means of the fossils is seasonal (Heim, 1909, page 331; also Andrée, 1915, pages 371, 380). Analogous deposits are now being formed in many lakes. In Zürich lake during the cool and moist season mud is brought by the streams and deposited as clay, while in late summer, when the temperature of the water is high, a layer of fine lime is precipitated.

VARVED MARINE SHALES

In the fish-bearing shales of the Oligocene Flysch formation of Switzerland there is a bed containing a few hundred layers, each 2 to 30 mm. thick, whose lower limit consists of hard lime and upper limit of soft clay. The lamination is probably annual (Heim, 1909, pages 331, 332). The stratification of the Silurian shales of Brittany (Heim, 1909, page 332), and of the Miocene (sarmatisch) brackish-water clays of Steiermark (Winkler, 1913, page 577) are likely also of seasonal character.

VARVED EOLIAN SAND

At Ragunda (63° N.) in Sweden dunes with varvity caused by the formation of a layer of decayed vegetable matter every autumn have been observed by Gerard De Geer (*See* Caldenius, 1924, pages 81, 89). In a dune, at most 114 years old, there were 93 varves.

VARVED CHEMICAL DEPOSITS

Varved chemical deposits are also known. The regular alternation of anhydrite and halite in one of the lower salt beds of upper Permian age at Stassfurt in Germany is considered to be of seasonal origin, the anhydrite representing summers and the common salt representing winters (Görgey, 1911; Arrhenius and Lachman, 1912; Jänecke, 1923, pages 71, 89). Similar lamination occurs in Miocene salts in Alsace (Gale, 1920, page 47). The stratification of an anhydrite in Texas is probably annual (Udden, 1924).

Not unfrequently layers of clay alternate with layers of salt. The conditions of deposition are known from what is now going on in Elton lake in southeastern Russia, in the Dead Sea, and other lakes. In Elton lake in spring some clay is deposited, whereas in summer, as the concentration proceeds, gypsum, and in autumn common salt are precipitated (Heim, 1909, page 331). In the Dead Sea the process is somewhat different (*See* Geikie, 1903, p. 530). The alternation of red shale and white gypsum described by Shimer from the Triassic of Arizona may be, according to Sayles (1922b, page 458), of seasonal character.

LAMINATION OF THE SEDIMENTS AT THE MOUTHS OF THE MISSISSIPPI NOT SEASONAL

The lamination of the sediments at the mouths of the Mississippi, suggested by Shaw (1913, page 17) to be seasonal, are by Trowbridge (1923, page 58) found not to be so, but to be due to ever changing conditions including changes in velocity, direction of the currents, amount of load, texture of load, depth of current, nature and amount of electrolytes, etc., connected with the eddying and "boiling," scouring and filling, accompanying the flood stage.

HOMOGENEOUS FRESHWATER CLAYS IN TEMPERATE REGIONS

Two examples will be given to call attention to the fact, often overlooked, that in ordinary temperate freshwater lakes, fed by rivers bringing material from unconsolidated glacial deposits, settling of the material in suspension takes place too rapidly to permit lamination of the deposits, which thus become almost or entirely homogeneous.

At La Sarre, Pontiac county, Que., on the Canadian National railways, there occurs on the top of the varved glacial clay a grey, greasy clay with occasional thin layers of somewhat coarser material or of organic remains, but largely homogeneous (Plate I f). It contains shells of *Spherium striatinum* Lamarck. It was deposited in lake Abitibi after the complete withdrawal of the ice from the region and occurs practically up to an abandoned post-glacial shore-line which is cut in glacial clay on the road just south of the village. It is simply glacial varved clay redeposited by brooks and wave-action. The physical conditions of the lake are not well known; but the water is probably fairly cold, and certainly is frozen over for many months. The actual cause of the subsequent subsidence of lake Abitibi is not known, but it may be due either to greater upheaval of the northern end, or to cutting down of the outlet channel, or both. The recession of the shore has been considerable, for in spite of raising of the lake level of 7½ feet in 1913 (Hopkins, 1918, page 201) it now stands between 10 and 20 feet below the post-glacial shore.

The clay at La Sarre is the only post-glacial clay seen by the writer in Canada, but it can fairly certainly serve as a type of the clays being deposited in most lakes.

In Sweden, the varved glacial clay is overlain for the most part by homogeneous post-glacial clays (cf. page 21). There are a few of these clays which were deposited under different conditions. Attention is here directed only to the Ancylus clay, deposited in the Baltic during its freshwater stage after the disappearance of the ice. The Ancylus clay, earlier called the lower grey-clay, is light grey to dark grey and entirely homogeneous or only with faint traces of stratification. It was formed of varved glacial clay and till eroded by stream and wave-action and redeposited. The Ancylus clay at Upsala, where it has been physically and chemically analysed by Odén and Reuterskiöld (1919a), is grey, perfectly homogeneous, and very fine grained and dense. Leaving some small islands out of consideration, the clay at Upsala was deposited some 40 miles from the shore and at a depth of about 375 feet, facts which explain the fine-grained character of the clay.

The physical conditions of Ancylus lake are well known through Munthe's studies. Its temperature was about the same as that of the Baltic today. The temperature of the air of July was about 59°F. (15°C.) (Munthe, 1910, page 189).

CHAPTER III

METHOD OF STUDYING THE ICE RETREAT BY MEANS OF THE VARVED GLACIAL CLAY

The varved glacial clay is the most important seasonally banded sediment and, together with a Swedish post-glacial clay, forms the basis for the only existing exact geochronology (*See* Antevs, 1925b).

De Geer's method of investigation of the ice recession by means of the varved glacial clay is based on certain assumptions, and may be described as follows:

The varves were deposited in front of the ice edge which formed their proximal limitation, and thus they cover each other like shingles on a roof (Figures 1 and 2). The varve thickness, as determined by the amount of

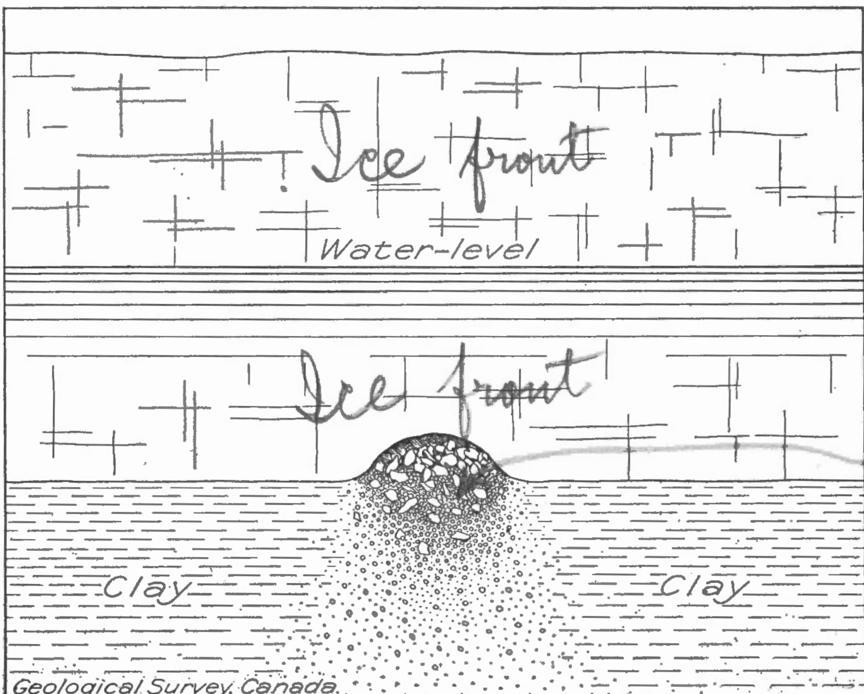


Figure 1. Mode of formation of glaci-fluvial deposits in fresh water.

ice melting, usually varied from year to year, but since the chief factor controlling melting was summer temperature the thickness, as a rule, was relatively the same each year in the whole area that was climatically uniform. The varves from widely separated localities can be recognized, therefore, by means of their relative thicknesses, and, therefore, also the position of the ice border each year and the rate of recession can be determined.

For obtaining of varve series exposures in clay pits, bluffs, or special excavations are smoothed with spade and brick trowel and carefully

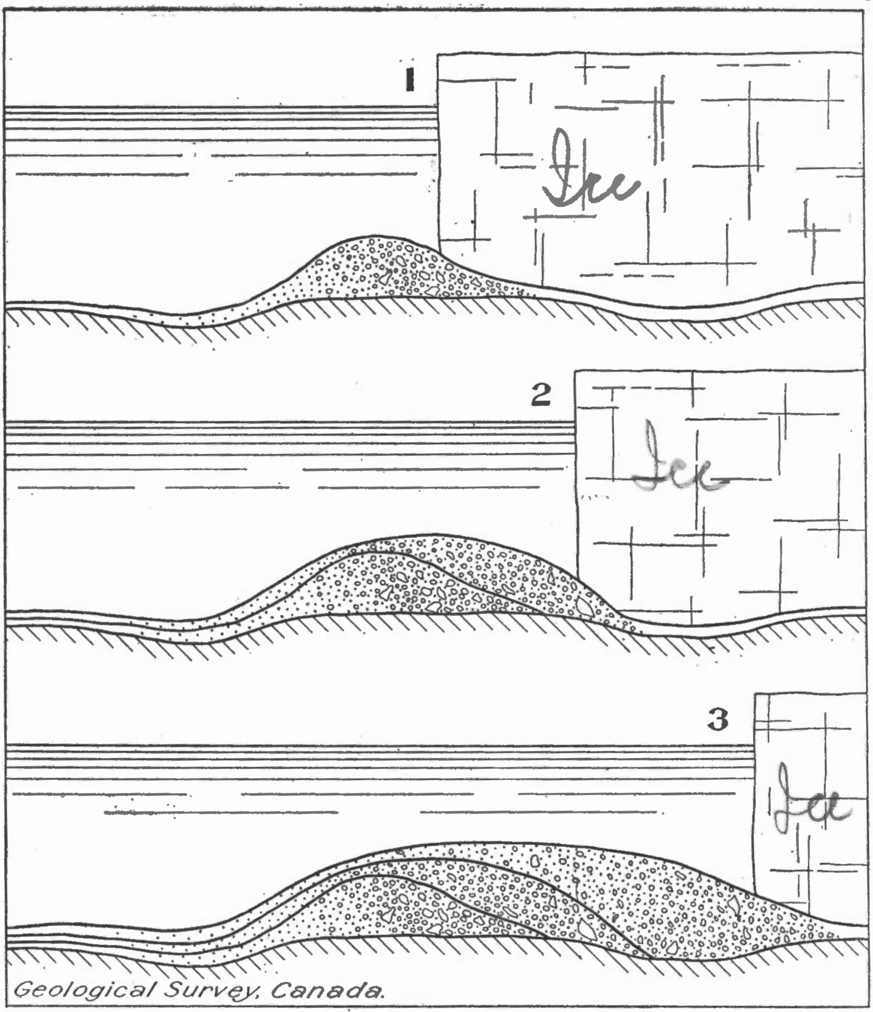


Figure 2. Mode of formation of glaci-fluvial deposits (esker gravel, sand, silt, and varved clay) in fresh water off the receding ice front during three successive years. Section through centre of glacier orifice.

examined. The varves must be perfectly undisturbed, not contorted, compressed, or faulted. Undisturbed beds between contorted zones or in faulted blocks, of course, can be measured. Disturbances, though, are

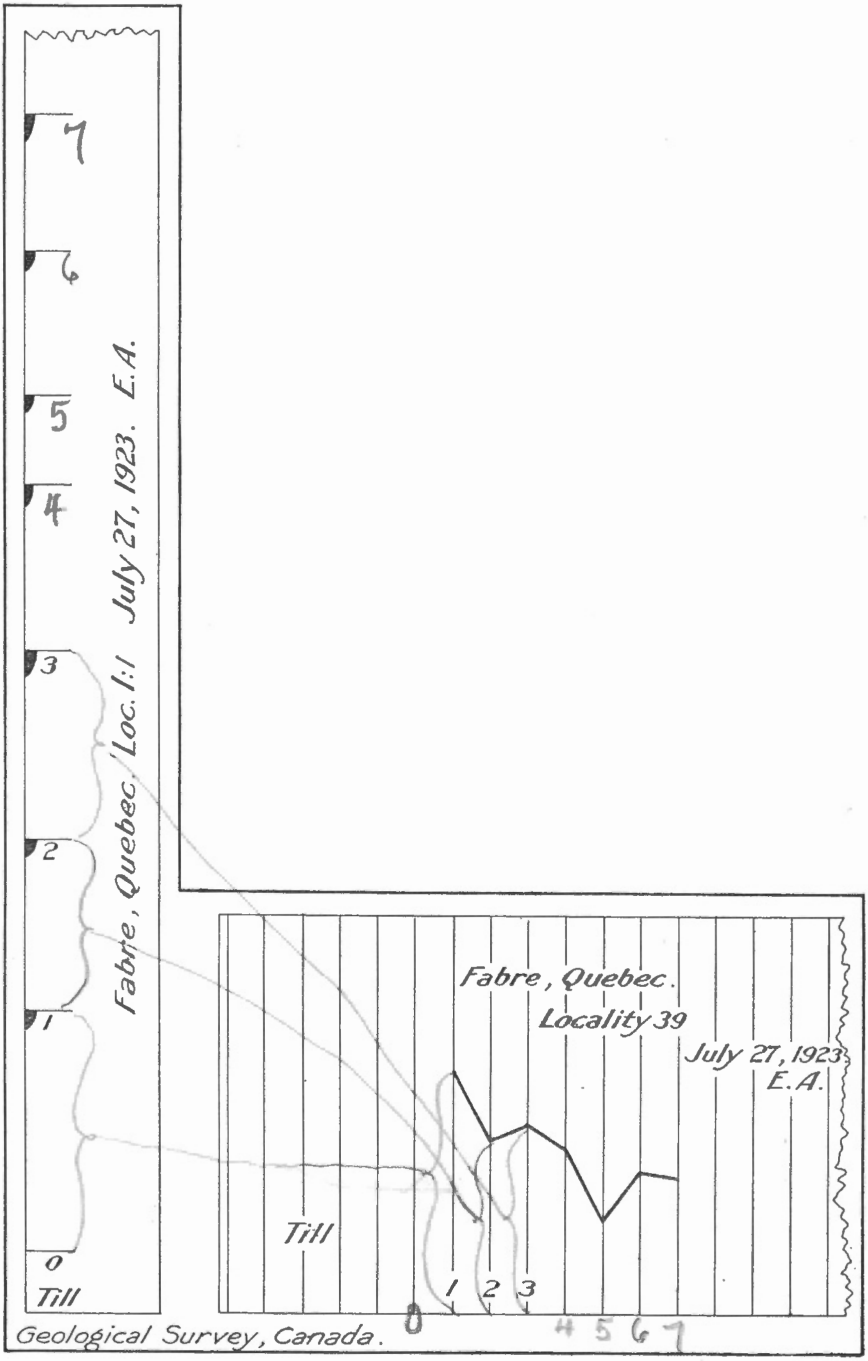


Figure 3. Sample showing method of measurement of annual layers (varves) in the field (cf. Plate Ie) and curve constructed from it. Actual scale.

always treacherous. Estimation of the number of the varves that are disturbed is difficult (cf. Antevs, 1922, page 72). The varve limits are carefully marked off on a narrow strip of strong paper (Plate I e). For exact determination of the ice retreat, series from the very bottom—that is, from rock or till—are necessary. In reality, however, the bottom in many regions can seldom be reached. The longer the series, the better; for connexion over short distances, series of twenty varves usually are enough.

The field measurements are transformed into curves or graphs on paper ruled with lines spaced exactly 5 millimetres, for this spacing has proved to be the best (Figure 3). The thickness of the lowest varve is set off on a vertical line on the left side of the paper, the second varve is set off on the next line to the right, and so on. The points marking the thicknesses of the varves are connected. The curve obtained in this way is compared with a curve from another locality in the vicinity. The curves are moved in relation to each other, until they fit, which they do if they have a sufficient part in common and are characteristic. If the bottom varve at one locality is found to correspond to varve number 16 at another locality the latter place evidently was uncovered fifteen years earlier than the former one. By sufficient number of measurements, not only determination of the rate of the ice retreat, but also mapping of the ice border every year, can be made. Such a detailed study has actually been carried out by De Geer (1912, Plate 2) in the Stockholm region.

Of poorly varved clays samples can be taken for measuring at home. The samples are taken in tight troughs of zinc plate, conveniently $19\frac{5}{8}$ inches long, 2 inches wide, and $\frac{3}{4}$ inch high. The face of the clay bank is carefully smoothed and the trough is cautiously pressed in, a knife being used to cut away the clay just outside the edges, until the trough is entirely filled with clay. The samples are taken so that they overlap one another 2 inches. The troughs are then cut out from the bank, and part of the projecting clay is removed. The samples can be conserved in good condition for some time by wrapping them in wax paper or keeping them in a damp place. If the sample is for museum purposes or for teaching, the clay is carefully cut down to the edges of the trough. If the clay has dried it must gradually be moistened and brought back to its natural consistency before it is cut. It is then treated with glycerine so diluted that it can be absorbed by the clay. When all the water is replaced the clay retains its natural appearance unchanged.

The method used for construction of the normal curves is described in Chapter XII, page 120.

CHAPTER IV

PROPERTIES AND TYPES OF THE VARVED GLACIAL CLAY

IMPORTANT TYPES

There are many different types of varved glacial clay the formation of which is dependent on a series of factors mostly little known.

The most important distinguishing feature of one varve in relation to another is the distribution and relative mass of coarse and fine material. The varve may present two distinctly limited layers of very different grain sizes, a gradation from coarse silt at the bottom of the varve to very fine clay at the top, or a mixture of coarse and fine material throughout the varve though with an increase of fine material upwards. It may consist of silt and clay in various proportions and may even consist almost exclusively of silt, or almost exclusively of fine clay.

Exceptional flocculation of fine material in summer being disregarded, the relative thicknesses of summer and winter layers are essentially dependent upon the properties of the mud. If the mud is coarse the summer layer will be relatively thick; if it is fine-grained the winter layer will be relatively thick. The winter layers as a rule remain almost equally thick from year to year, so that the variations in thickness of the varves are largely due to variations in thickness of the summer layers. Since, probably, about the same proportions of silt and clay were discharged into the lakes every year when recession took place over a uniform area, the thick varves must have relatively too thin winter layers. Thus, particularly during years of great melting, a large part of the fine material must have gone through the lakes or, in the case of very large lakes, have been transported very far out.

The distinctness of the lamination is primarily determined by the difference in grain and colour of the winter and summer layers. The distinctness can show all degrees. The varves may stand out very distinctly or be almost indiscernible.

In many cases the fineness of the material visibly increases upwards by decrease in the number of coarse grains. Sauramo (1923, page 78) calls a varve with this assortment of the grains a diatactic varve.

In many cases there is no apparent upward decrease in coarseness, but the winter layer shows sharp lower as well as upper limitation. In some the lower limit of the winter layer is more distinct than the upper limit, which may occasionally be somewhat diffuse. This is the case in the clay at locality[1] 29, Sturgeon Falls, Ont., shown in Plate I b and in some varves at locality 1, Black rapids. This fact indicates that the great bulk of the fine material settled after all the coarse material had gone down and that part of it sank only after melting began in the following spring.

Fairly often the coarse and fine material is mixed throughout the varve or is mixed to a greater extent in the lower part than in the upper part. This is because considerable quantities of the fine material went

[1] For descriptions of the sections at the localities studied see Chapter XI, page 95.

own along with the coarse material during the summer. Sauramo (1923, pages 82, 98, 110) calls this varve structure symminct. The degree of symmixis is markedly variable, and this type merges into the other two mentioned. Usually at least a number of varves are symminct, but in some cases single symminct varves occur in a series of otherwise diatactic structure (Sauramo, 1923, page 84; Plate 5). It is obvious that symmixis is due to relatively rapid flocculation of the fine particles caused by the presence of a flocculator, easily aggregating material, or other physical conditions favouring flocculation (*See* Searle, 1924, pages 243, 244). To this type belong all varved brackish-water clays and many freshwater clays. The great percentage of colloidal and fine material gives the symminct freshwater clay great plasticity (cf. Odén, 1916, pages 178, 186; Lyon, etc., 1919, page 172). Clays deposited in strongly brackish water and, still more so, clays formed in the sea, are physically different, for, though stiff, they are not very plastic. They are extremely hard to work and have uneven fracture (Plate II B).

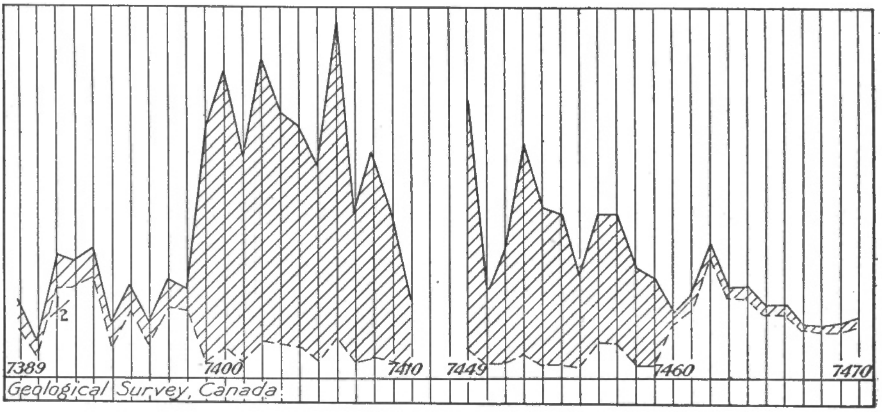


Figure 4. Part of varve diagram from Inwood, Vermont. Varves 7399–7459 are symminct though deposited in fresh water.

A very instructive case of freshwater symmixis was observed at Inwood, Vermont, 6 miles south of St. Johnsbury, as illustrated in Figure 4 (Antevs, 1922, page 36). A series of clay-silt varves ends abruptly with varve 7398 and is followed by very fat, lead-coloured clay with varves about three times as thick. This clay (at about varve 7460) is rather abruptly followed by a fairly lean silt-clay. The thickness of the varves was probably due in part to increase in the supply of material but also to flocculation and settling of the fine mud, much of which remained in suspension and was transported farther down the valley when the lean clays above and below were deposited. Part of the clays at localities 103, Iroquois Falls, 104, La Sarre, and at other places north of the height of land, are symminct. These occurrences indicate that clays of symminct structure were deposited in fresh water, in lakes to which the sea never had access, as well as in brackish water. The single symminct varves also are evidence against salinity as the only cause of summer flocculation. Sauramo's (1923, page 110) view that symminct structure is an indicator of sea water is, therefore, too narrow, and some of his conclusions relating to the history of the Baltic, conclusions which are not borne out by observations in Sweden, appear to have no foundation. It should be remembered that although the late-glacial Baltic was, from time to time, connected with the sea, particularly through straits across central Sweden, it was only locally somewhat brackish (*See* Antevs, 1922a, page 609). In Finland its waters were not even brackish from a biological point of view, judging from the absence of brackish-water fossils in the varved clays (Sederholm, 1911, page 10).

The causes of the occasional rapid flocculation in glacial freshwater lakes are not clear. Strong concentration of fine particles, and sinking of the particles into the lower level of the water because of poor conditions of transportation, quiet water, or other reasons, may have played their rôle. The physical properties of the mud may be the decisive factor in some instances, the chemical properties in other cases. Clays rich in lime concretions are, as a rule, at least partly symminct, calcium oxide and calcium hydrate being very active flocculators and calcium carbonate a somewhat active flocculator (Lyon, etc., 1919, pages 159, 193). The conditions controlling ice retreat seem to be of importance, for symmixis appears to be most common in connexion with halts of the recession. The single symminct varves are most peculiar.

CHANGES IN CHARACTER OF THE CLAYS

In some cases, as already mentioned, there occur rapid changes in the physical properties of the varves. Some of these changes were due to coming in of sea water and will not be specially considered. Changes in thickness and texture were also due to diversion of the main current bringing in mud, to increase or decrease of the drainage area, to rising or sinking of the lake level, to increased or decreased ice melting, and to changes in the properties of the material, or of the river and lake water, and consequent changes in the rate of flocculation. Original, not secondary, changes in colour may essentially be due to changes of material.

At locality 21, Trout Creek, a fairly fat, thinly varved clay (year 314) abruptly passes into silt-sand with thin salmon-coloured winter layers (Plate IX). The abrupt change may have been due to sinking of the level of Lake Algonquin[1] through drainage (cf. page 60).

At locality 22, Powassan, a lean silt (varve about 300) is followed by a sandy silt with much thicker varves because of shallowing (Plate IX).

At locality 92, Belleek, a brown, medium fat clay gradually changes into a chocolate-brown, exceedingly stiff clay during a gradual increase in the thickness of the varves. Somewhat higher in the series, about 1850 to 1880, the clay changes in the opposite way (Plate VIII). The changes may have been caused, almost entirely, by changes in rate of ice melting and of flocculation.

In the sections north of the height of land a marked increase in the thickness of the varves occurs beginning with year 1528 (Figure 5). The

[1]For description of the glacial lakes *see* Chapter VIII, page 59.

consistency of the clay usually does not undergo any essential change. The increased deposition may have been due to increased ice melting (cf. page 82).

Probably the same increase in melting is recorded at locality 81, Englehart, by the change from a medium fat, grey-brown clay to an exceedingly fat, chocolate-brown clay with thicker layers (Figure 6). The lower part of the upper clay is almost homogeneous, and indicates increased flocculation.

At locality 30, Sudbury, varve 66 abruptly begins a series of thin and fine-grained clay varves. This was due probably to diversion of the main current (Plates I d, and IX). The change in coarseness is considerable,

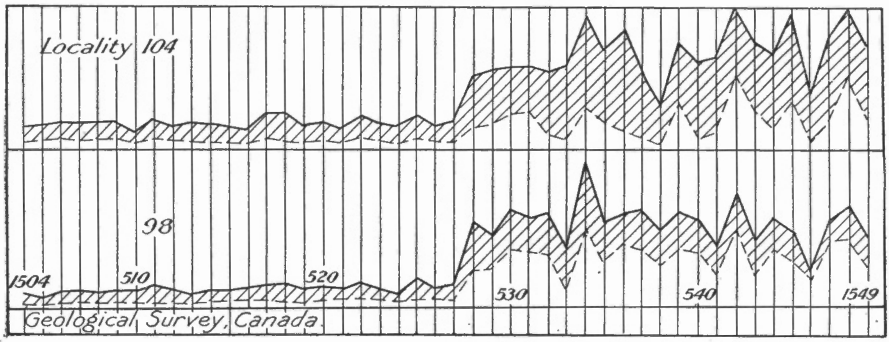


Figure 5. Curves showing marked increase in deposition due to increased ice melting.

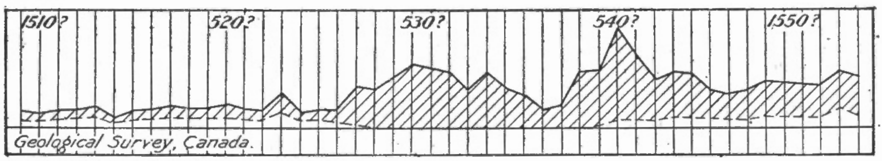


Figure 6. Increase in thickness of varves and fineness of material at locality 81, Englehart. Probably marking the increased ice melting indicated in Figure 5.

though the photograph does not show it well because of the light colour of the fine-grained material.

About year 375, in the Timiskaming series (Plate VI), a lean silt-clay gradually goes over into a fat clay, probably because the ice edge had reached the areas of Silurian rocks at the northern end of lake Timiskaming. The actual thicknesses of the varves increase but slightly.

At locality 55, Ville-Marie, situated high up on a hill side, a very marked change in the sediments occurs at varve 206, which begins a beautifully varved silt-clay (Figure 7). Varves 143 to 205 are local. Varves 143 to 182 consist of silt deposited close to the ice edge and show good lamination. Varves 183 to 205 consist of silt or, in the case of the thick varves, of fine sand. They are lens-shaped, as if influenced by currents or

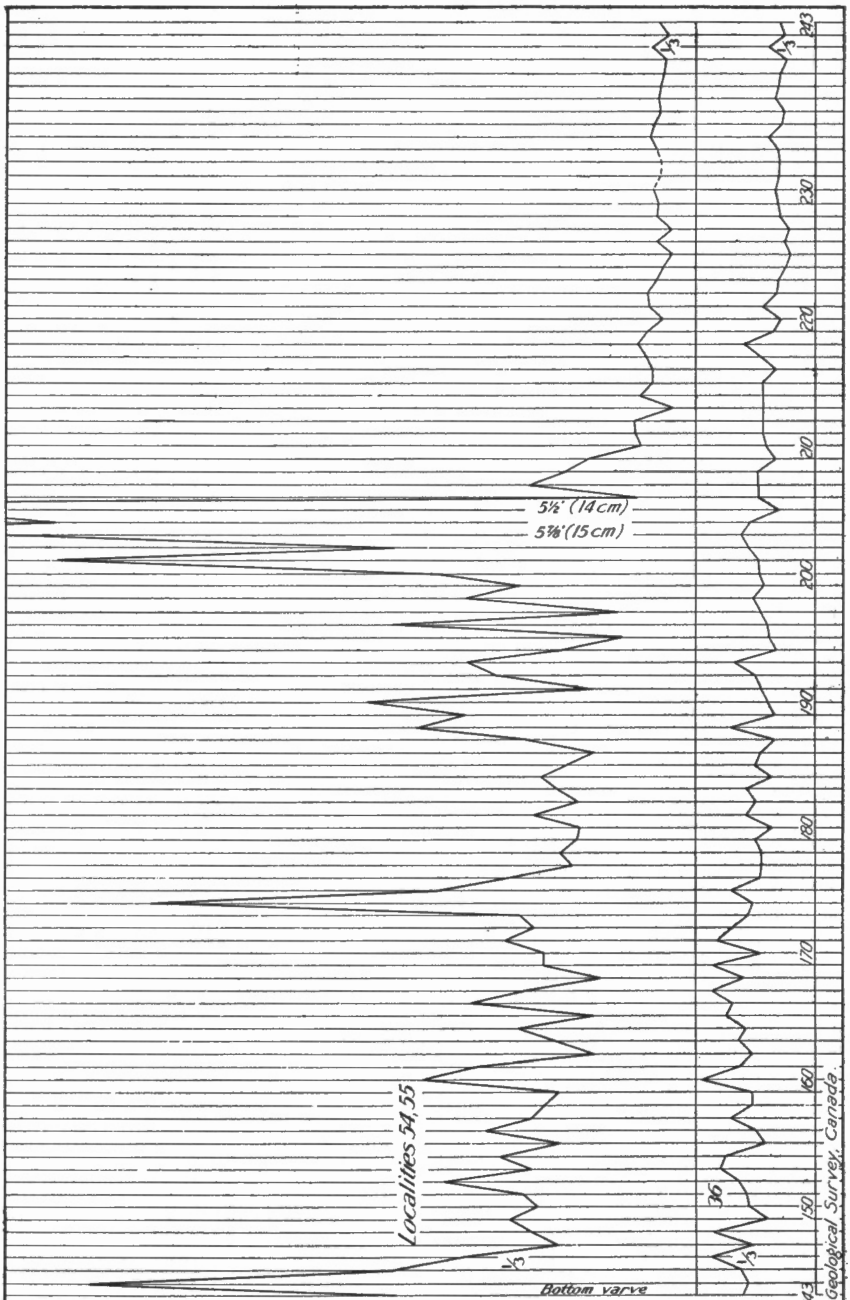


Figure 7. Curves showing marked difference in deposition at localities 55 and 36.

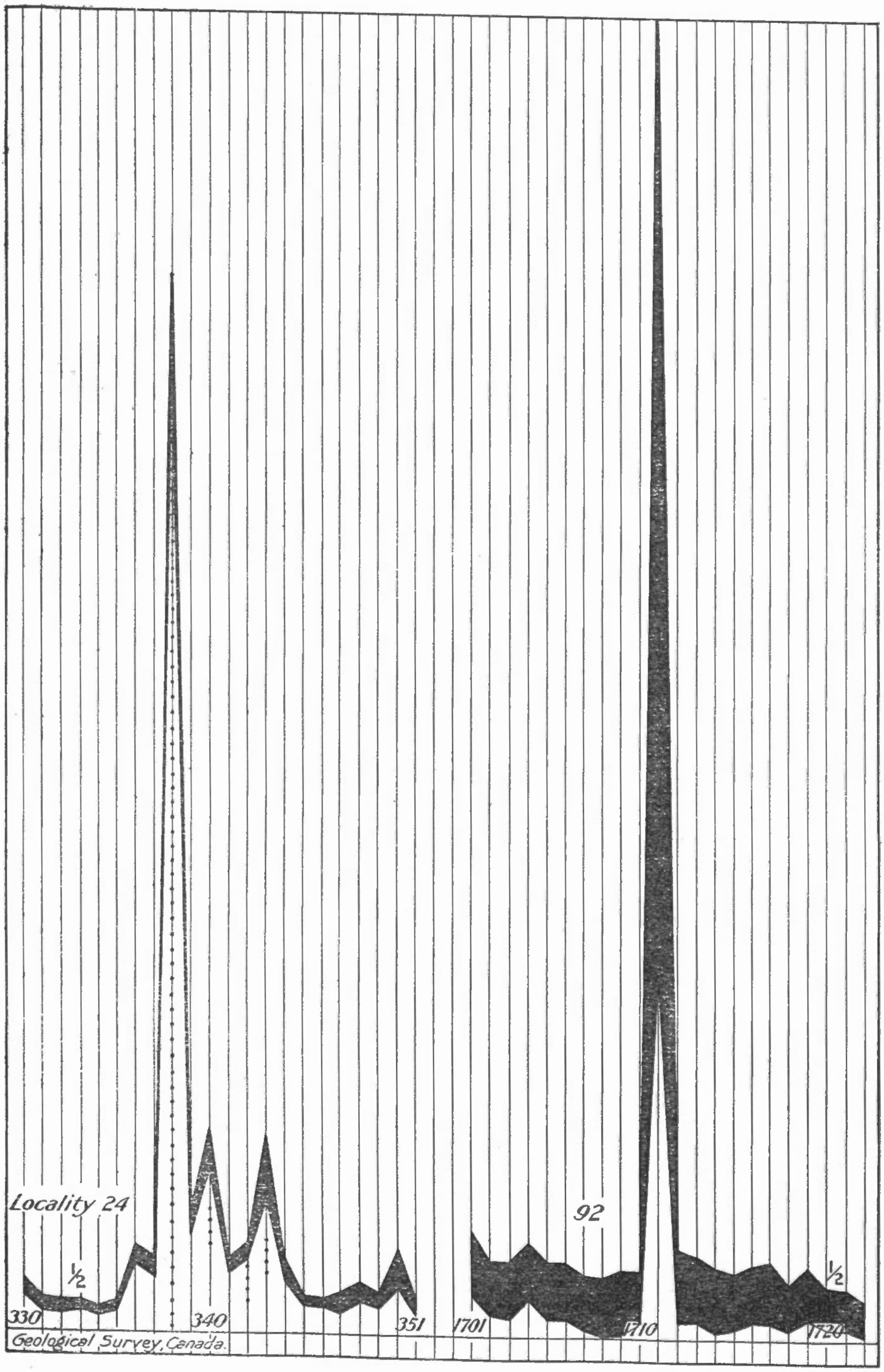


Figure 8. Curves showing that drainage occurred in the immediate neighbourhood of locality 24 and fairly far away from locality 92. Solid black indicates clay. Dots indicate fine sand.

waves. The change at varve 206 appears to be due to diversion of a current. It cannot, probably, be due to increase of the water depth because it is not evident how so rapid a rise of the water-level could occur, though this in itself would be the most natural explanation. The thick varves cannot be due to drainages, for at locality 36, lying $7\frac{1}{3}$ miles to the south, and at a lower level, no trace of drainage is to be found.

DRAINAGE VARVES

In regions with strong relief where lakes could be dammed between the ice edge and higher land, or by barriers of glacial deposits, thick varves marking the sudden drainage of such lakes are not infrequent in the clay deposits in the larger basins (cf. Antevs, 1922, page 69). The extra material in the varves was picked up by the vigorous drainage river along its course, and may be of any quantity up to a bed many feet in thickness. The

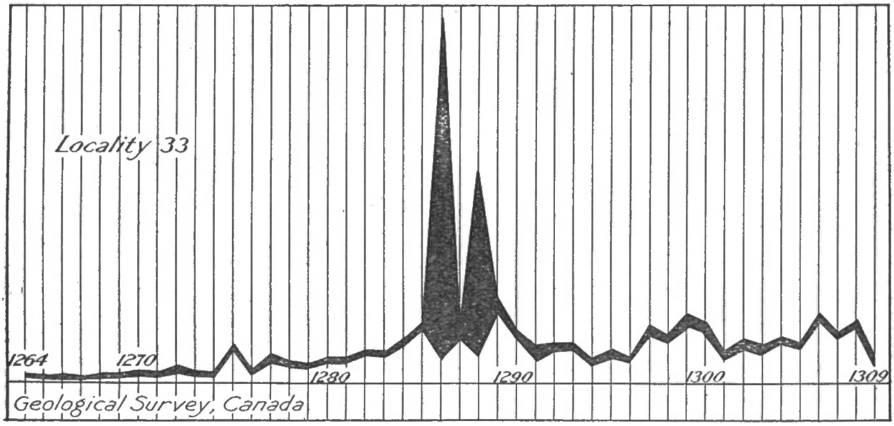


Figure 9. Curve showing drainages that occurred a considerable distance from locality 33, (cf. Plate Ic). Solid black indicates stiff clay.

thickest drainage varve known to this writer is a layer 12 feet thick at Hanover in New Hampshire. The thickest layer observed in Canada is one at locality 35, Sault Ste. Marie, measuring 3 feet. Most drainages took place during a single year, but a great many of them occurred during a small number of years. The drainage layers, of course, decrease in thickness and the material decreases in coarseness away from the mouth of the river by which the dammed lake emptied, and thus the position of this lake can be approximately determined.

Figures 8 and 9 show three types of drainages. The thick drainage layer at locality 24, Powassan, consists of $10\frac{1}{2}$ inches fine sand and $\frac{3}{4}$ inch clay and evidently marks a drainage in the immediate neighbourhood. Varves 340, 342, and 343 contain sand, and perhaps mark less pronounced drainages. The drainage varve at locality 92, Belleek, consists of 4 inches quicksilt and $9\frac{7}{8}$ inches greasy clay. The drainage apparently took place fairly far away. The two drainage varves at locality 33, Espanola, Figure

9, have no surplus of coarse material, but have large amounts of exceedingly fine-grained clay. The drainages, therefore, must have occurred far away. They evidently were of great importance for the deposition at Espanola, but unfortunately the circumstances are not known (cf. pages 123,124).

COLOURS OF VARVED GLACIAL CLAYS

An interesting feature of the glacial clays is their colour. The original colours of most glacial clays, and the colours which they still have in places where oxidation has been prevented by poor drainage, etc., are grey-white, grey, or grey-blue. The sandy and silty parts of the varves are usually white to grey. The finer the material, the stronger, as a rule, are the colours, which range from dark-grey and grey-blue to black. The exceedingly fine-grained winter layers in many cases are deep black with a strong lustre. There are, however, exceptions. The finest material is in some cases the least coloured. Thus, in Hudson valley many of the winter layers are greyish white with strong lustre which, however, fades almost instantaneously on exposure to the atmosphere. Some of the winter layers at locality 1, Black rapids, also are grey-white. The light colour of silt and coarse clay is due largely to the presence of quartz and other light-coloured minerals. The origin of the dark grey to black colours of the finest material is not so evident. The coarser fractions of brown to black winter layers were found by Odén (1920, page 342) to be white-grey, from which the conclusion is drawn that the colour depends on the size of the particles and, also, perhaps, on the chemical constitution, varying with the grain size. The dark colour of the winter layers is ascribed by Sauramo (1923, page 20) chiefly to the minuteness of the particles, this view being based upon the known fact that the whiteness of paints is greatest when the diameters of the pigment particles are between 5 and 1 μ, whereas smaller particles appear greyish. Other conditions may be equally important. The fact that the darker a clay is in its original condition, the redder it will become, as a rule, after oxidation, appears to indicate that iron compounds play an important rôle. The black colour may, therefore, be due partly to ferrous sulphide and ferric sulphide. The presence of ferrous compounds is explainable, as the iron in minerals and rocks frequently occurs in the form of soluble ferrous salts. Organic matter which gives the dark colour to most soils can scarcely have played any rôle, at least for those clays deposited in typical glacial lakes, for these practically lacked organic life (*See* page 21). (For theory of colours, etc., see Ridway, 1912, Mervin, 1917, and von Bichowsky, 1918.)

Aeration of the clay causes oxidation of the ferrous compounds, and the finer parts of the clays become yellow to red. The silty summer layers mostly maintain their white-grey appearance, but in places are yellowish after oxidation.

The red colour presents very different intensities and shades which perhaps are due to the amount of iron oxide present, to the degree of hydration of the iron oxide, and to the size and distribution of the colouring particles (*See* Lyon, etc., 1919, pages 76, 77; Searle, 1924, page 97). Oxidized clays containing lime are deep red. The fine-grained winter layers of clays deposited in brackish water assume a red or pink tint after oxidation. The winter layers in the upper, well-oxidized brackish-water clay at Ottawa are brilliantly salmon coloured. Deeper down they are red-brown.

In some cases the colour is determined by the secondary colour of the mother rock. Thus the glacial clays at New Haven, Connecticut, are red because they are mainly derived from the red Connecticut sandstone.

A feature, often of great help in connecting clay measurements, is the abrupt changes that the colour not unfrequently undergoes in a deposit. An horizon may consist of clay with brilliantly red winter layers underlain and overlain without transition by clay with brown or red-brown winter layers. Even single winter layers may stand forth because of particular colouring. In many cases a series starts abruptly with a strong red colour that gradually fades upward and changes into red-brown.

The strength of the colour of the glacial clay is a remarkable thing when the grey, dull colours of clays of the cold temperate regions are considered. The only post-glacial clay seen by this writer in Canada is grey and occurs at La Sarre (cf. page 7). The post-glacial clays of the Baltic region—the Ancylus clay and the Littorina clay—are so distinctly grey that they were originally designated the Lower and the Upper grey-clays. These clays are homogeneous, and the material is derived largely from the glacial clays (cf. page 8). The fact that they have not been oxidized, as have the glacial clays, or if they have been oxidized have not been coloured by the ferric oxides, is possibly due to the mixture of coarse and fine particles, and to rapid and great flocculation of the material during deposition. The grey colour, a mixture of white and black, suggests that oxidation of the iron compounds has not taken place.

LACK OR SCANTINESS OF ORGANIC REMAINS IN VARVED GLACIAL CLAYS

In the typical varved glacial clays, deposited in fresh water, no macroscopic fossils of any kind except land plants blown out into the lakes have been found. If molluscs, fishes, and other animals with conspicuous shells or skeletons had lived in the lakes, remains of them would certainly have been found, even although shells of freshwater molluscs are not very resistible. Shells or skeletons of crustaceans, insects, etc., may, on the other hand, occur rarely in the clays and have been overlooked because they are so small or so obscurely placed as to escape the eye. The glacial lakes, therefore, may have been devoid of macroscopic animals or at least of animals with conspicuous skeletons. In glacial lakes in northeast Greenland, Frits Johansen (1911) found only a few species of copepods.

The chief cause of the lack or scarcity of animal life in the glacial lakes may have been that the perpetual and rapid deposition of mineral matter all over the lakes prevented the development of a flora on which animals could exist. The low temperature may also have been an important cause. Freshwater molluscs cannot exist in water that is ice-cold the year round, but thrive in water of a maximum temperature of 39° to 41° F. (4° to 5° C.) (A. C. Johansen, 1904, pages 44, 49). The Great Lakes, during the latter part of their glacial history, were only semi-glacial. Therefore, their central and southern parts contained the same mollusc fauna as they do at present, while their northern shores were formed by the ice (Coleman, 1922, page 32; cf. our page 43). No animal remains are known to have been found in the glacial clays deposited in these lakes.

On the other hand, varved clays deposited in brackish water contain remains of a fairly rich fauna of molluscs, crustaceans, fishes, etc. (Jägerskiöld, 1912; Högbom, 1915; Johnston, 1917, page 25; Goldring, 1922; Kindle, 1923, etc.).

CHAPTER V

MODE OF ORIGIN OF THE VARVED GLACIAL CLAY

GENERAL ACCOUNT

In lake Louise, in the Canadian Rockies, the formation of one type of varved glacial clay was studied by Johnston (1922a). Lake Louise is fed by a little stream coming from Victoria glacier situated a mile distant. The river brings most material in spring and considerable quantities in summer, but very little in winter. As far as observed the river current does not follow the bottom of the lake, but the river water tends to diffuse through the lake water, both having the same, the maximum, density. The coarse silt is quickly deposited, whereas the finest material remains suspended, so that by the end of summer the water is slightly turbid. During winter this fine material settles, and at the arrival of spring the water is perfectly clear. A sample brought up from nearly the central, deepest part of the lake, which is 1¼ miles long, ¼ mile wide, and 230 feet deep, showed very fine-grained clay of a greyish white colour with faint but definite stratification. Each layer consists of a coarse, light-coloured lower part which passes upward into a fine, dark-coloured upper part. The layers vary in thickness, averaging 5 to 6 to an inch. The thickness of the layers corresponds to Johnston's approximate estimate of the thickness of the annual deposit in the lake, and the character of the lamination is in accord with the conditions of sedimentation. The coarse layer, therefore, represents summer deposition and the fine-grained layer winter deposition. Together they constitute a varve.

The varved late-glacial clays were formed in lakes off the receding edges of the last ice-sheets (Figures 1, 2). Most of the material was supplied by glacial streams which, flowing under hydrostatic pressure in tunnels below the ice, carried masses of material of all sizes from minute particles to boulders several feet in diameter. The rivers and the hydrostatic pressure postulate that there were crevasses in the ice through which melt-water from the surface could work down. Cracks were largely formed where the substratum was very uneven. The Greenlandic ice-sheet is much crevassed, so that comparatively little water can continue to the border in surface streams (Hobbs, 1911, page 170). Glaciers in the Alps are also much cracked (Hess, 1904, pages 221-223). In glaciers in Alaska, however, Gilbert (1904, page 198) found that whatever the distance downward to which the cracks originally extended, the resulting permanent crevasses had only moderate depth, the ice being welded into a practically continuous mass beneath. The crevasses in the waning ice-sheets may essentially have been located somewhat inside the ice margin. The organization of the melt-water into streams probably took place quite near the ice edge. There could not have been any large, empty fissures above the streams, which must have been flowing through confined tunnels. There could not have been any basins at or below the level of the mouths of the rivers.

These conditions seem necessary to explain the extraordinary erosive and transporting power and swiftness of the glacial streams. The rivers had to dig channels in ice and frozen till, often go up hill, and remove enormous quantities of material which together with the moving ice tended to obstruct their course. Subglacial streams of glaciers in the Alps behave like surface streams, but have usually about one-half as rapid flow as surface brooks of the same size and grade (G. Vallot and J. Vallot, 1900, pages 32-34; Hess, 1904, page 223; Mercanton, 1916, page 103).

The glacial streams seem to have been smaller in the United States than in Fenno-Scandia, for eskers are relatively rare in the first-mentioned region, whereas large eskers with coarse material form the commonest type of outwash in northern Europe. To a large extent this was due probably to the greater rate of ice recession in Fenno-Scandia and partly also to the more pronounced topography which favoured the opening of cracks in the ice. When the subglacial river reached the mouth of the tunnel, the pressure gave out, and assortment of the material took place. Boulders and gravel were deposited in the tunnel itself or just outside, forming eskers and outwash. If the discharge occurred directly in water, sand was carried a few to several hundred yards out, and silt and clay were transported long distances. The silt and coarse clay sank to the bottom rather soon, but the finest clay particles remained in suspension

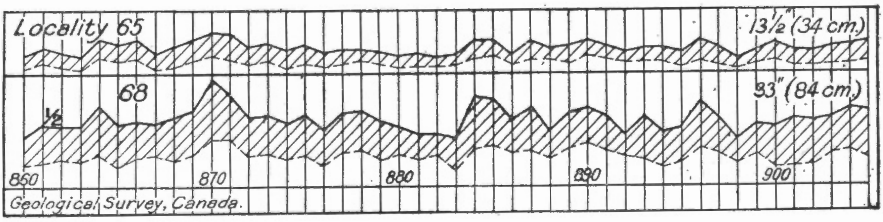


Figure 10. Curves showing greater deposition in a glacial lake in deep water than in shallow water. Winter layers shown by inclined ruling.

for a long time and settled largely first during the autumn and winter. During winter no melting occurred. Thus, the amount and coarseness of the material deposited decreased with increasing distance from the river mouth, and the typical annual deposit—the varve—shows a gradual upward transition from silt to fine clay. The water depth evidently influenced the character of the clay and of the varves, but beautifully varved clays and silts were deposited from the surface of the lakes down to their greatest depths. The deposits in very shallow water usually consist of silt and fine sand, largely derived through wave erosion, and of thin to very thin laminæ of greasy clay. The clay layer was in many cases eroded away by wave-action during the spring following its deposition. If no unconsolidated material was at the disposal of the waves the varves became very thin in shallow water at a distance from the ice. The thickness of the varves increased with the water depth towards the middle of the lake, provided the lake was not too large (Figure 10). The homogeneous character of one horizon at locality 33, Espanola, was due to insignificant deposition, not directly to great depth (*See* page 123).

TRANSPORTATION OF MUD IN GLACIAL LAKES

In temperate lakes the muddy and consequently heavy river water usually sinks down in the lake water to the cold and heavy strata below the thermocline (*See* page 38) or to the very bottom (Heim, 1900, page 165; Forel, 1901a, page 34; Halbfass, 1923, page 81). Occasionally—the Rhone in lake Geneva and the Rhine in Bodensee for example—the river water, on account of its weight, flows as a sub-water river between embankments straight down to the deepest part of the basin. In any case the great bulk of the fine mud is deposited in the deep basins whose bottoms finally become quite even. If, however, the river water contains but little mud and is much warmer than the lake water below the thermocline, or is mixed by storms with the surface water, the mud may be transported by the surface layer and become deposited in any part of the lake.

In glacial lakes, on the other hand, transportation may largely have taken place in the upper water strata. This is evident from the spreading of the fine glacier mud over the entire area of large lakes and its deposition far away from the ice edge where the water was shallow as well as where it was deep. Thus, in glacial lake Timiskaming in northern Ontario, clay was deposited on the sides of the deep basin high above the present lake, while the glacier rivers discharged in the low terrain north of the existing lake. Bottom currents evidently cannot account for this transportation, for such currents could have consisted only of downward flow of heavy river water or of secondary currents caused by the wind-induced currents in the upper compartment. Bottom currents worth mentioning may not have existed, judging from the lack of erosional phenomena in the deposits. The water of the glacier rivers, if discharged at the bottom of the lakes, evidently rose to the upper strata, bringing along part of or most of the suspended mud. This was possible because the pure melt-water at, or slightly above, the freezing point, was lighter than the bulk of the pure lake water which was near the greatest density (Figure 21, page 41).

Transportation in the upper water layers was made possible by the strong anticyclonic winds which may have set the surface waters almost permanently in motion in a direction away from the ice. This motion was supported by the escape of enormous water masses through the glacier vaults, and their rise to the upper water strata. The flow must have been very considerable, for enormous water masses went through the lakes. The moving water body was thickest in the deepest and most central parts of the basin. If, as usually was the case, the outlet was situated at the distal end of the lake, the return current may have been comparatively insignificant and too weak to induce real currents in the lower compartment of the lake. The fine material which was brought with the river water up into the upper water strata most probably diffused through the whole moving water mass. Diffusion, as mentioned, is found to take place in lake Louise. The suspended material settled largely in order of size of individual grains and of aggregations, but the original position in the moving water layer was of great importance. Coarse grains brought up to the very surface could be transported long distances before they had sunk down into the stagnant lower compartment, whereas small particles in the lower part of the upper compartment settled after a short journey.

Thus, the material was spread all over the lake and deposited almost everywhere where the lake bottom was fairly even, though it later may have been washed away from large parts. The transportation was dependent upon a series of conditions such as the shape of the lake, the amount of water masses going through it, the physical and chemical properties of

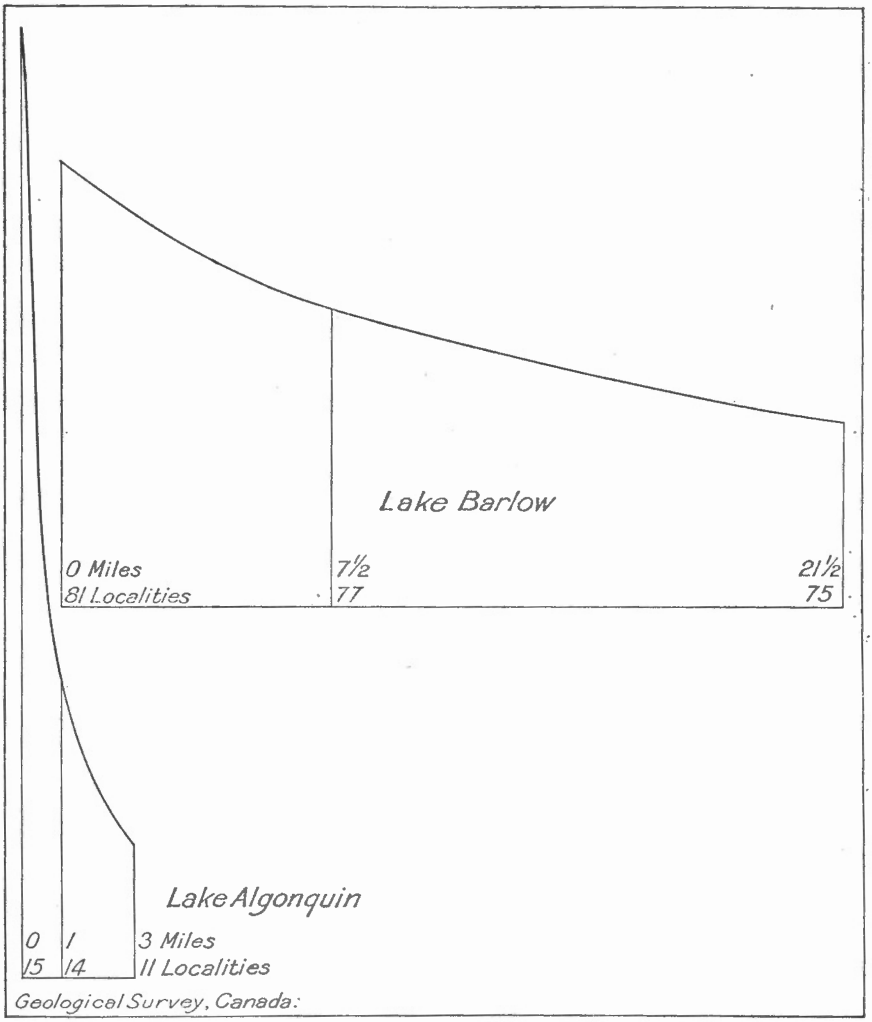


Figure 11. Curves showing the gradual decrease in deposition in a typical glacial lake (Barlow) away from the ice margin, and in a semi-glacial lake (Algonquin). Curves plotted on the same scale.

the lake water, of the river water, and of the material; it was dependent, too, upon the strength of the wind, of the hydrostatic pressure of the subglacial rivers, etc.

The transportation and distribution of material in Lake Barlow or in glacial Lake Timiskaming were very remarkable (Figures 11 to 13). The series shown in Figure 12 represents ice-recession just north of the height of land, and when it was deposited the ice edge was about 85, 80, 75, 67, 62, 47, and 40 miles north of the localities 37, 45, 59, 68, 75, 77, and 81 respectively. The thicknesses of the series figured are given at the right ends of the curves. The gradual decrease of deposition with increasing distance from the ice edge is best shown by localities 81, 77, and 75, which may have

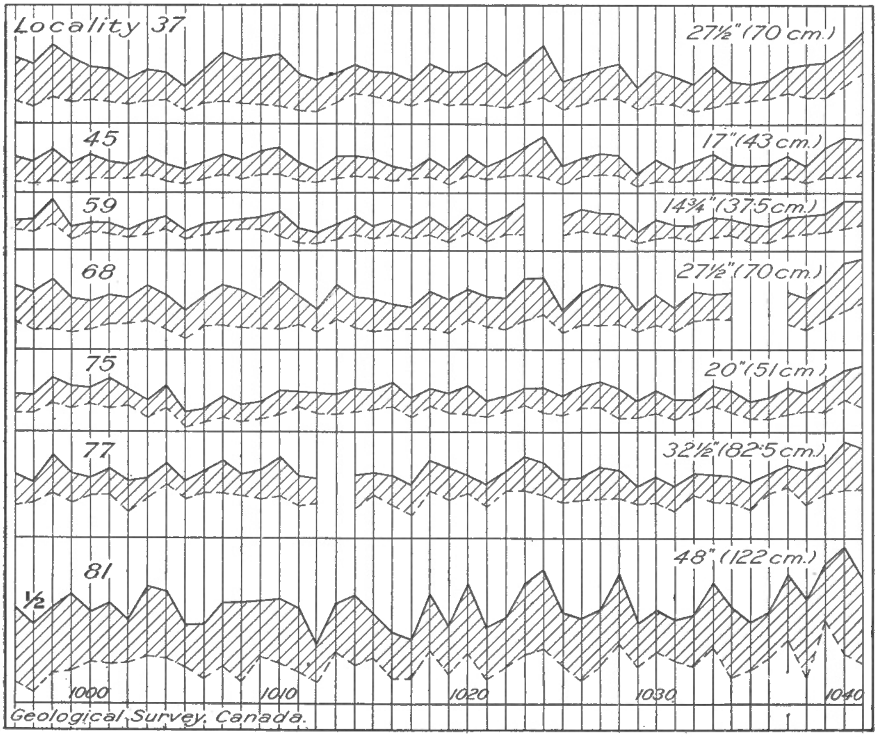


Figure 12. Irregular distribution of glacial lake clays in Lake Barlow, as shown by variations in the thicknesses of the same varves at several localities. Winter layers shown by inclined ruling.

been similarly situated with reference to the main water current and at approximately the same levels. The decrease is graphically illustrated in Figure 11.

The great thicknesses at localities 68 and 37, as compared with the lesser thicknesses at localities 75, 59, and 45, are due to the fact that localities 68 and 37 were near the deeper part of the basin and at low elevations. Here the current was stronger and the water column was higher.

The greater deposition in deep water than in shallow water, observed in every glacial lake, is well illustrated at Haileybury (Figure 10). The series figured was deposited when the ice front stood somewhat south of

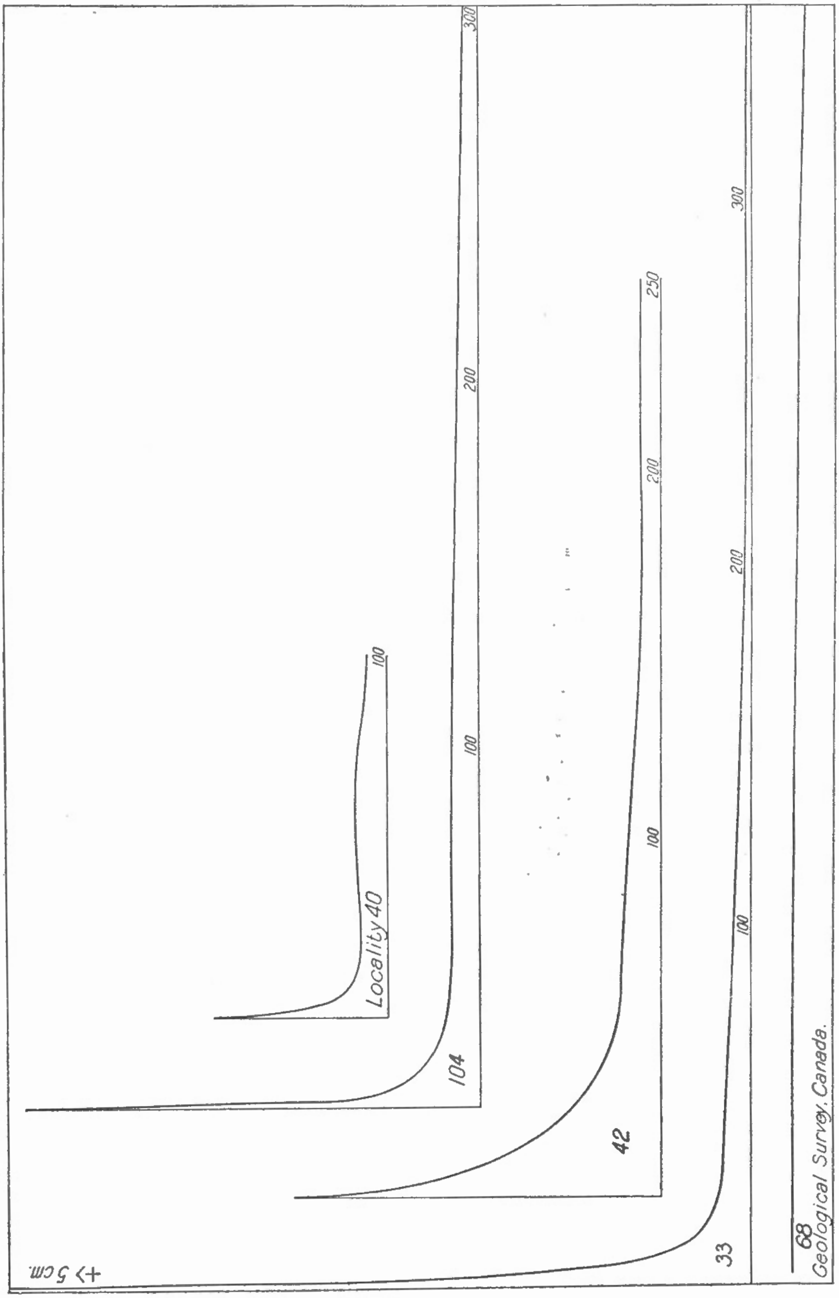


Figure 13. Curves showing decrease upwards in thickness of varves at certain localities.

the height of land, or some 40 miles away. At locality 68, Haileybury, near the shore of lake Timiskaming, the series lies approximately 25 feet above the lake whose water-level is about at 580 feet. At locality 65, North Cobalt, 1⅓ miles to the southwest, the series is somewhat lower than the railway station, which has an elevation of 841 feet. Thus, the series at locality 68 was deposited in about 200 feet deeper water than at locality 65. The deposition at the former locality was two and a half times as great as at the latter. The relative amounts of coarse and fine material are about the same in both deposits. Transportation of the material is also illustrated by the upward decrease in the thicknesses of the varves in a series. The decrease as a rule is very rapid at the bottom, but it soon becomes gradual and after some time very slow. Finally the varves wedge out. The rate of thinning out can be very different. The curves shown in Figure 13 illustrate the wedging out at some representative localities. Curve 68 starts some varves above the bottom, curves 40, 42, and 104 begin at the bottom, and curve 33 practically at the bottom. The varves were plotted with a spacing of 2 millimeters, smooth average curves were drawn, and the curves in reproduction were reduced three-fourths. The thinning out was very slow at localities 42, 68, and 104 in lake Timiskaming and Lake Ojibway. At 68, Haileybury, there is hardly noticeable thinning out, though the ice edge during the deposition of the series retired at a fairly great rate.

The enormous transportation in a glacial lake and the relative thinness of the winter layers in thick varves (cf. page 13) makes it probable that considerable quantities of fine rock-flour went through the lakes into the outlet rivers just as fine glacier mud nowadays goes through lake Geneva, Vierwaldstätter lake, the Oberengadiner lakes, etc. (Heim, 1885, page 362).

TRANSPORTATION OF MUD IN BAYS OF THE SEA AND SEMI-GLACIAL LAKES

The preceding has reference to fresh glacial lakes with conditions unfavourable for flocculation. In brackish or salt water and in fresh water in which rapid flocculation took place, somewhat different conditions prevailed. The transportation was shorter; both coarse material and fine material were deposited near the mouth of the glacial river, and the greater part of the fine material went down during the summer. As a consequence the varves were relatively thicker near the ice and tapered out rapidly, and the assortment of the material in the annual deposit was less marked.

The late-glacial clay at Upsala, analysed by Odén (1920, pages 339-341), was deposited in slightly brackish water. The increase of fine material in very thick varves of this clay, the occurrence of considerable quantities of fine particles along with the coarse ones, and decrease in distinctness between summer and winter layers with increasing varve thickness, may largely be ascribed to flocculation due to the weak salinity. In a big lake such as Lake Algonquin, which was essentially of temperate type, the river water and its load soon, or immediately, sank down into the stagnant lower compartment and fairly rapid flocculation occurred because of the physical properties of the water, etc. The conditions of transportation as far as currents are concerned also were less favourable than in typical glacial lakes. Thus, the material brought down by glacial rivers from the north, instead of spreading over the whole of Lake Algonquin, was mostly dumped near the ice or the lake shore, although part of it may have spread over a fairly large fan-shaped area. These conditions are shown by the relatively rapid upward thinning of the varves at locality 33, Espanola (Figure 13), and the very rapid horizontal decrease in the thicknesses of the annual deposits at Bracebridge. Figure 14 shows the horizontal thinning in the lower parts of the deposits at Bracebridge, when the ice edge stood in the neighbourhood. The localities lie only about one-half mile apart. Figure 15 illustrates the wedging out higher up in the deposits, when the ice front

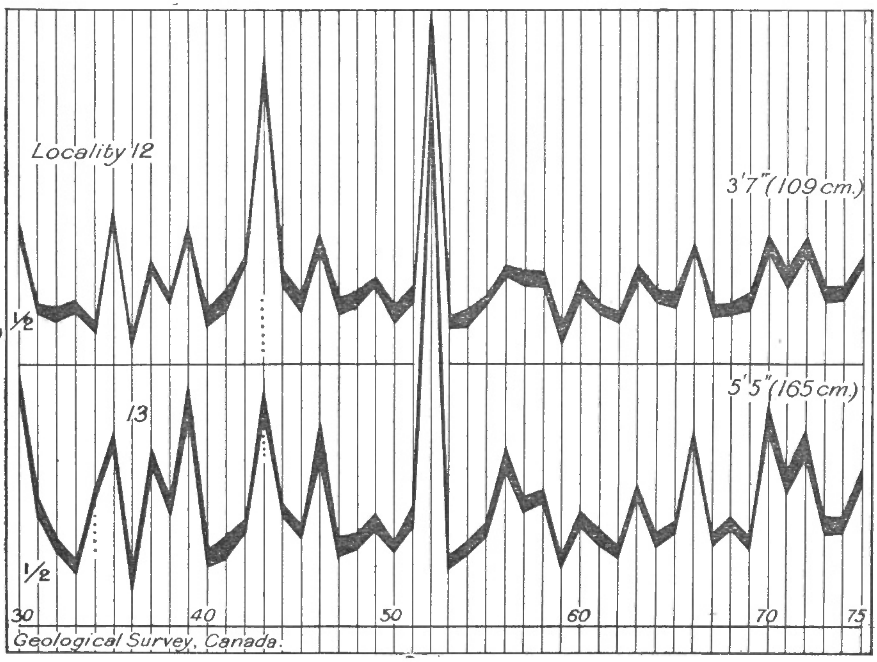


Figure 14. Horizontal wedging out of varves, near the bottom of the deposits, in semi-glacial Lake Algonquin. Winter layers shown by solid black. Dots indicate sand.

probably had retired from the immediate neighbourhood. The lake shore, though, may have been situated close to locality 15 and a glacial land river may have fallen out close by, judging from the coarseness of the material and the thicknesses of the varves. Locality 16 lies 600 yards west of 15, and localities 14 and 11 lie 1 mile and 3 miles respectively down the valley from locality 15. Locality 10 lies in another valley 3½ miles from 15. The thinning out at localities 15, 14, and 11, which may have had similar situations in the late-glacial bay, is graphically shown in Figure 11. The steepness of the curve as compared with the gentle slope of the curve for Lake Barlow or glacial Lake Timiskaming, which is plotted on the same scale, is striking. How characteristic the two curves are for the

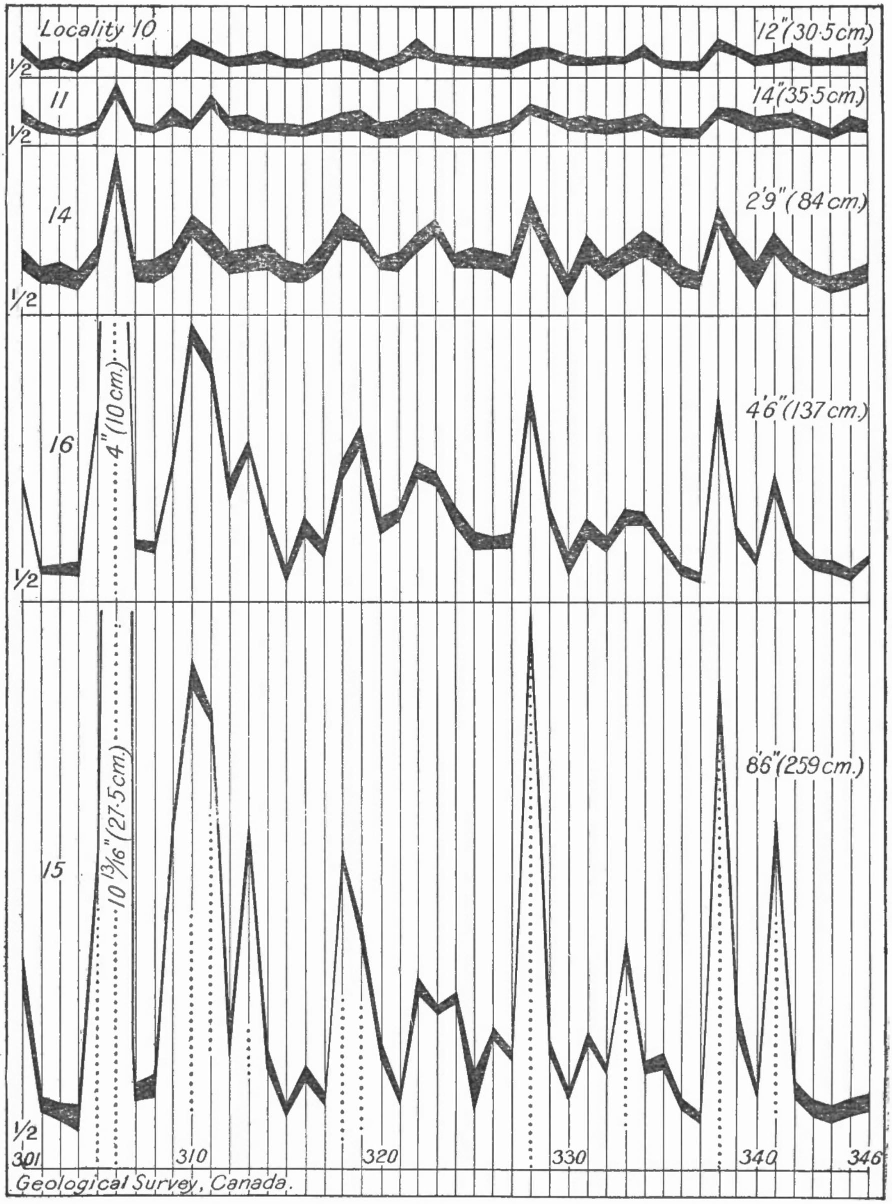


Figure 15. Horizontal wedging out of varves, in upper part of deposits, in semi-glacial Lake Algonquin. Winter layers shown by solid black. Dots indicate sand.

respective lakes and for other similar lakes is not known. Along with decrease in thickness of the varves goes diminution in grain size of the material. The thicknesses of the fine-grained winter layers at locality 14, although more distant from the ice border, are actually greater than at 15 and 16, and are relatively much greater. Still higher up in the beds the relative thicknesses of the varves at the different localities are about the same as in the part shown in Figure 15.

DEPOSITION

Deposition of clay takes place through sinking of separate grains and of flocculated particles. The rate of sedimentation of unflocculated clay of different grain size stands in direct proportion to the quantity of suspended material, in inverse proportion to the height of the water column, and is dependent on time (Odén, 1918, page 25). Since, however, a clay particle of $0 \cdot 3\mu$ needs 100 hours to fall 10 centimetres in fresh water, and 14 months to fall 10 metres (Odén, 1916, page 191), and since a large percentage of the winter layers consist of particles of this fineness (Odén, 1920, pages 337-341), it is evident that most of the small particles did not go down separately. The deposition of the fine clay only partly proceeded according to the laws of movements of minute particles (found through experiments and calculations), but flocculation occurred also in fresh water.

Thus the fundamental difference in the formation of the varved glacial clay and homogeneous marine and freshwater clays lies rather in the rate of flocculation than in the rate of settling of the small particles. The decisive thing is whether essential flocculation and subsequent deposition takes place while large quantities of material are still brought out into the basin or not. In salt water and in temperate lakes coagulation takes place immediately, or in any case during the summer. If summer flocculation had gone far enough in the glacial lakes the aggregations would have sunk and the deposits would have become homogeneous.

In typical glacial lakes, however, flocculation worth mentioning occurred only during the winter, when all or almost all the suspended material settled. Lake Louise, which is fed by a glacial river, has clear water in spring (Johnston, 1922a, page 380). At Stockholm, greasy winter layers are observed in esker gravel deposited close to the mouth of the subglacial river. In winter no melting took place. The lakes were frozen over. The quietness of the waters which distinguished winters from summers may have been the essential cause of the flocculation and subsequent settling, for the physical properties of the waters were the same. In many instances, however, fairly extensive summer flocculation took place in glacial lakes with perfectly fresh water, and occasionally so great aggregation occurred that the clay became practically homogeneous (cf. page 14). Thus, the conditions for formation of the one or the other kind of deposit were quite sensitively balanced. Flocculation of particles to form larger aggregates can hardly be imagined in nature in water free from salts or other dissolved constituents. The flocculation of fine suspensions depends on several factors, including the grain size of the dispersed substance, its concentration, as well as that of salts in solution, and the presence of protective colloids (Wells, 1923, page 50; Odén, 1916, page 192; Searle, 1924, page 243). Silt, like most insoluble substances, when suspended in water, is most easily flocculated by calcium salts, when the suspension is neutral. Soil clay behaves in an opposite manner and is precipitated from alkaline suspensions more readily than from neutral suspensions (Comber, 1920, page 436). The different behaviour of silt and of clay when acted upon by calcium hydroxide may not be due to any essential difference in the structure of the particles, but may be determined by the ratio of the emulsoid surface to the core of the particle. In silt the core dominates the system, in clay the surface dominates the system (Comber, 1921, page 471). Flocculation is accelerated by acids, and by calcium oxide and calcium hydrate, but not by other alkalis (Lyon, etc., 1919, page 159). The necessary salt content for flocculation varies inversely with the size of the particles. The amount is also different for different salts. With large excess of electrolyte the rate of flocculation is usually very great. A freshly flocculated and precipitated clay can be deflocculated, that is, the particles can be reconverted into a form in which they will indefinitely remain in suspension by addition of a suitable electrolyte (Odén, 1916, page 192; Wells, 1923; Searle, 1924, pages 243, 246). Shaking is sufficient largely to disperse the particles of a flocculated suspension, but if it is allowed to stand the particles recoagulate, and deposition the second and following times will go more rapidly than it did the first time (Searle, 1924, page 243; Trowbridge, 1923, page 57; Littlefield, 1923, page 59). Flocculation is accompanied by a large increase in the volume occupied by the particles (Pickering, 1918, page 317).

PHYSICAL CONDITIONS ESSENTIAL FOR FORMATION OF VARVED GLACIAL CLAY

MATERIAL BROUGHT BY GLACIAL RIVERS

It is obvious from what has been already stated that an essential condition for the varvity of glacial clay is the lack of, or only partial, flocculation of the fine particles during the course of the summer. The physical conditions for this, though, are only partly known. In the following, an attempt at analysis of these conditions will be made as far as this is possible, when no experiments directly bearing on the subject have been made. The most important conditions are, probably, that the material be brought by glacial rivers, and that the lakes in which the deposition occurs have fresh or almost fresh water of a temperature below 39·2° F., or only slightly above. Varved glacial clay, so far as known, is formed only of material brought by glacial rivers. However, not all material derived from melting glaciers gives rise to varved deposits. Homogeneous clay is deposited in salt water, and almost homogeneous clay is also occasionally formed in fresh water (cf. page 14). The deposits in the Pleistocene lakes in the Great Basin largely lack varvity for reasons not well known (Antevs, 1925). Why glacial origin of the stream is necessary is not exactly known. One reason may be that the glaciers produce new-ground material which differs physically from the unconsolidated materials that largely supply land rivers. The material from a glacier probably consists of a greater part of new-ground rock than any other erosional material and probably contains a large percentage of minute particles which do not flocculate very easily. It appears as if igneous rocks give origin to clays with greater distinction between the lower and the upper part of the varve than do sedimentary rocks. Fine-grained slates can give clays which are practically homogeneous and exceedingly greasy. Of course, this may be due largely to lack of coarse material, but it shows that very fine material derived from a slate partly or largely settles during the summer, whereas material of the same fineness derived from igneous rocks largely or entirely remains in suspension until the end of the mild season. This indicates that material once flocculated is not entirely deflocculated. It has also been found that a suspension of a fine-grained sediment in the presence of a constant amount of electrolyte settles at the same rate and in the same time whenever put into suspension except when the material is put into suspension for the first time, when some time is consumed by flocculation (Littlefield, 1923, page 61; Trowbridge, 1923, page 57; Kindle, 1916, page 543). Other reasons may be physical properties and conditions connected with the low temperature of the melt-water. In many cases the subglacial rivers have their mouths on the lake bottom and thus discharge directly into the lake. Most late-glacial lakes were fed in this way. Then the temperature of the river water is at, or but slightly above, the freezing point (Heim, 1885, page 254; Hess, 1904, page 229; Øyen, 1893, pages 47, 49; 1905, page 70). In other cases the subglacial rivers are continued by land rivers which finally discharge into the lakes. The temperature of such rivers, of course, will rise somewhat. The rise depends on the size of the rivers, on the temperature of the ground and of the air, on the sunshine, on aeration, etc. It is greatest near the glacier, and diminishes with increasing distance from the glacier. Louis Agassiz found the following values in the Aar, Switzerland, and in the Triftbach, one of its tributaries (*See* Hess, 1904, page 229).

Triftbach		Aar	
Distances from the glacier vault Miles	Temperatures F.°	Distances from the glacier vault Miles	Temperatures F.°
0	32·0	0	33·8
0·12	34·7	2·5	35·6
0·3	37·4	5·9	38·3
0·6	41·0	9·0	41·0
0·9	42·8	11·2	42·8
1·2	43·7—44·6	12·4	44·6
		16·8	48·2

Greim (1903, page 626, *See* Hess, 1904, page 230) determined the temperature rise in the Jambach in the Tyrol Alps with increasing distance from the orifice and found:

July 14, 1897

Distance from the glacier	Miles	0·06	1·9	3·1	4·7	6·4
Temperature of the brook	F.°	32·9	40·8	41·5	40·6	41·5
Temperature of the air	F.°	49·8	50·5	53·4	51·4	47·7

August 19, 1898

Distance from the glacier	Miles	0·09	4·7	6·4
Temperature of the brook	F.°	32·9	45·1	48·0
Temperature of the air	F.°	58·5	68·5	62·2

Greim (1903, page 633, *See* Hess, 1904, page 230) also gives seasonal mean temperatures of the Jambach at Galtür more than 6 miles from the glacier as based upon daily measurements in 1896 to 1900:

—	Winter F.°	Spring F.°	Summer F. °	Autumn F.°
Brook	33·3	37·6	42·7	39·2
Air	25·5	35·8	52·7	40·1

Further data are to be found with Øyen (1893, page 49; 1905, pages 73, 80), Westman (1899, pages 72, 73; 1910, page 25), Greim (1903, 1910), and Rabot (1909, page 88).

The stream draining the Victoria glacier into lake Louise in the Canadian Rockies, a stream one mile long, in June 1921 had a temperature of 36° to 38° F. at the mouth (Johnston, 1922a, page 378).

Distinctly varved clays can be deposited also from glacial land rivers, for example, the clay being formed in lake Louise. The upper part of profile 33, Espanola, is a good example from late-glacial time. How high the temperature of the river can rise without causing too rapid deposition of the suspended material is not known. The amount probably varies with other factors. The question will be discussed in connexion with the treatment of the hydrography of the late-glacial lakes (cf. page 44).

The low temperature of the river has perhaps some physical influence upon the material, and it gives the water the required lightness to rise in the lake water and permit the material to diffuse (cf. page 25). The lightness, of course, could be attained by high temperature, but as a matter of fact varved clay is not known to be deposited from ordinary river water except under very special conditions (cf. page 5).

DEPOSITION IN FRESH WATER

It is a well-known condition of formation of the varved glacial clay that the water body in which the deposition takes place must be fresh or almost fresh. This also applies to the varved clays in Finland, for although the Baltic over Finland during part of the late-glacial time was somewhat saline, the salinity was so low that it did not even permit the most tolerant brackish-water molluscs (Sederholm, 1911, page 10). Therefore, Sauramo's (1923) applying of the term "marine" to part of the Finnish clays is inappropriate (cf. Sayles, 1924). The minimum salt content necessary for flocculation varies inversely with the size of the particles (Odén, 1916, page 192; Wells, 1923). The rate of flocculation increases strongly with the number of particles per unit volume, and with the excess of the electrolyte. The critical degree of salinity for the glacial clays is little known. It evidently varies with the conditions just mentioned and also with the properties of the new-ground material, the temperature of the river water, and of the lake water, etc. In a general way it can be said that, when the water permits the existence of a marine to brackish-water fauna, fine-grained clay as a rule is homogeneous or almost so, whereas sand or coarse silt may show fairly distinct varves. *Portlandia* (*Yoldia*) *arctica*, however, which can endure very diluted water, is in many cases found in distinctly varved clay—as for instance at Stockholm. On the west coast of Sweden the clays are homogeneous, except at the bottom of the deposits and in deep bays where the sea-water was diluted by melt-water. The same almost certainly holds true for the Canadian and American east coast north of Boston. In the marine area of southeastern New Hampshire, the only part on the open coast visited by the writer, the clays were found to be entirely homogeneous. In Ottawa district there are, besides distinctly varved freshwater clays, deposited before the invasion of the sea, shell-bearing brackish-water and marine clays. These latter are from fairly distinctly varved to practically homogeneous; mostly they show faint, not surely measurable varves. At localities 5 and 6, Pakenham, sandy clays present distinct varves, though they contain shells abundantly.

The fine-grained bottom-set beds now forming off Fraser river, British Columbia, lack varvity because of rapid sedimentation as a consequence of flocculation in the sea-water (Johnston, 1921; 1922, page 128). Also off the Muir glacier, Alaska, the mud settles quickly in the salt water of the bay, and already a few miles from the glacier the amount of suspended material is slight (Reid, 1896), (*See also* Andrée, 1920, pages 102, 103).

THE LAKE WATER OF LOW TEMPERATURE

Fresh water has its greatest density or specific gravity at a temperature of 39·2° F. and this property largely determines the thermic stratification of the water in a lake (Forel, 1901a, pages 105-111; Murray, 1911; Needham and Lloyd, 1916; Halbfass, 1923). Water at or nearest the temperature of 39·2° forms the bottom strata. Water above 39·2° has direct stratification, that is, the warm water is at the top, and the temperature decreases downwards. Water below 39·2° has inverse stratification; the coldest water is at the top, and the temperature increases downwards. This condition divides lakes into three types, viz., lakes whose stratification is constantly direct (tropical lakes), constantly inverse (polar lakes), and direct in summer and inverse in winter (temperate lakes).

The glacial lakes off the Pleistocene ice-sheets were extraordinary in several respects. They occurred under relatively low latitudes, where summers were long and insolation was strong. Consequently, warming above the critical temperature normally should have taken place. However, contemporaneously with the breaking up of the winter ice on the lakes, melting of the ice-sheet began. Ice cold water streamed out into the lakes and being lighter than the bulk of the warmer lake water, that part which entered at the lake bottom rose to the surface. When it was warmed by insolation, contact with the air, etc., so as to become heavier than the water beneath, it sunk until it reached a layer of its own temperature with which it mixed.

In a temperate lake this process of heating to the critical temperature of 39·2° in spring, as a rule is rapid. The thickness of the practically uniform surface layer, steadily rising in temperature, increases rapidly, and, in a few days to one or two weeks, the whole water body has attained a uniform temperature of 39·2° (Richter, 1897, pages 59, 69; Forel 1901a, pages 106, 117, 121). The process is illustrated after Forel and

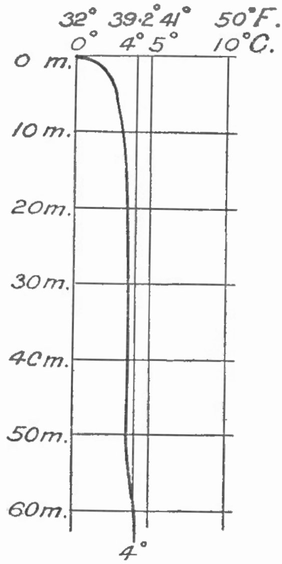


Figure 16. Temperature stratification in cold lake during cooling. (From Forel 1901a, Fig. 10, p. 116.)

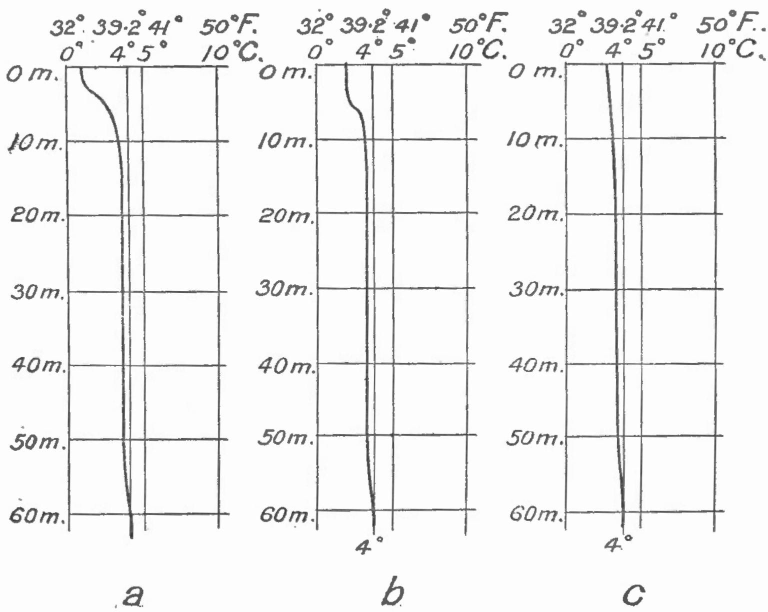


Figure 17. Equalization of temperature in water with inverse stratification in different stages. (Figure b from Forel, 1901a, Fig. 12, p. 117.)

Needham and Lloyd in Figures 16-20. Figures 16 and 19 show the temperature stratification in a cold lake during cooling. Figure 17 shows the equalization of the temperature in water with inverse stratification in different stages, and Figure 18 shows the equalization in water with direct

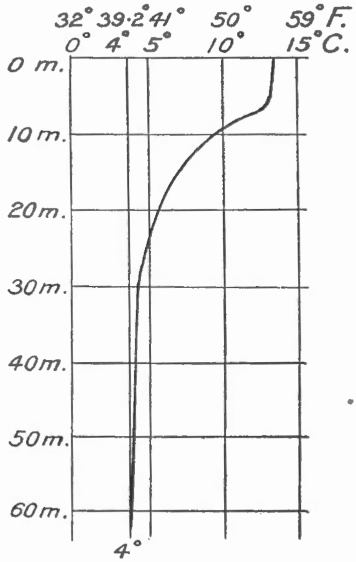


Figure 18. Equalization of temperature in water with direct stratification. (From Forel, 1901a, Fig. 13, p. 117.)

stratification. The zone of rapid temperature change, the break of the curve, at a depth of 8 metres, is called thermocline or discontinuity-layer (*See also* Figure 19). The thermocline can lie at any depth but usually is from 5 to 20 metres below the surface. It forms the lower limit of circulation due to winds and the lower limit of the vertical convection currents caused

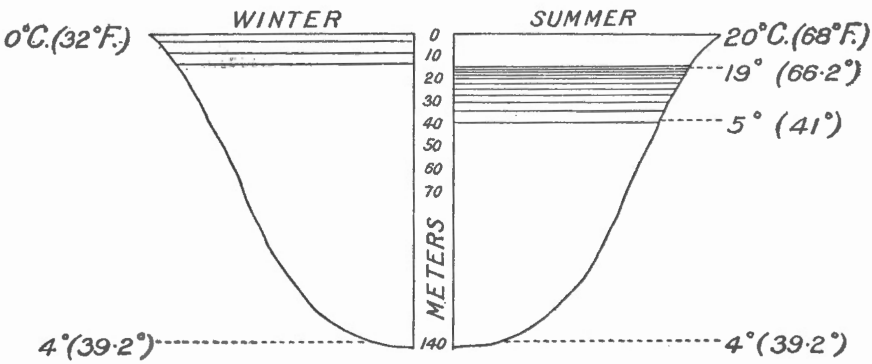


Figure 19. Diagram illustrating summer and winter temperature in Cayuga lake, New York. The spacing of the horizontal lines represents equal temperature intervals. A typical temperate lake. (From Needham and Lloyd, 1916, p. 32, Fig. 4.)

by heating by day and cooling by night (Forel, 1901a, page 118; von Hann, 1915, page 62; Needham and Lloyd, 1916; Halbfass, 1923, pages 182, 183). Above it, the waters are more or less constantly stirred; below it, they are stagnant or set in motion by the return current of the upper compartment (*See* Figure 21).

In glacial lakes the conditions differed in the first place in that the more the insolation increased and the temperature rose, the more ice melting took place and the more ice-cold water was discharged into the lakes. The muddiness of the water limited the effect of the insolation, which in clear water penetrates only a few metres (Richter, 1897, page 70; Forel, 1901a, page 122; von Hann, 1915, page 56), into the top strata. These were rather rapidly warmed, since muddy water absorbs heat more readily than does clear water (Forel, 1901a, page 104). Near the ice the stratification may have been about as shown in Figure 16. As the surface temperature rose with increasing distance from the ice the conditions may have become as shown in Figure 17, a and b. Judging from the enormous transportation of mud that took place, the inverse stratification may not have changed into direct stratification, for if it had done so, the melt-water, discharged at the bottom, would have stayed there. It would very soon have become stagnant or practically so on account of lack of force of motion, or it would have sunk into the deepest part of the basin. The current of the Rhone dies out in lake Geneva a few hundred yards from the river mouth (Forel, 1895, p. 275; Forel, 1901a, pages 61, 81). If the mud had remained in the stagnant lower compartment of the lake it probably would have settled, for the most part at least, in summer. The mobility of the water in the upper compartment prevented flocculation and sinking. Thus, typical glacial lakes may have belonged to the polar type. Glacier lakes in Jotunheimen mountains, Norway, belong to this type (Øyen, 1893, page 52; 1895; 1905, pages 72, 81). The low temperature of these lakes is found by Øyen to be mainly due to the presence of numerous icebergs. Insolation, nightly radiation, winds, and precipitation also seem to play a great rôle, but the temperature of the air only a small rôle. For the late-glacial ice lakes the almost certain presence of numerous icebergs, the constant ice-cold winds coming from the ice-sheet, and the nightly cooling, may have been the chief factors preventing warming of the water. The heat received by a lake through insolation by day is almost all lost during the night, particularly if the sky be clear (Richter, 1897, page 70; Forel, 1901a, page 113; von Hann, 1915, page 57).

In a polar lake, the thermic stratification during the warm season is gradually evened out, so that by the end of summer the whole lake has an almost uniform temperature of 39·2° F. or slightly below (Forel, 1901a, Figure 8, page 109). It seems probable that the temperature of the upper strata of the glacial lakes did not rise above 3° C.; that the stage shown in Figure 17c was not reached. A homogeneous water body has complete circulation, that is, a circulation comprising the whole water mass down to the bottom; storms stir up the bottom deposits and mix thoroughly both dissolved and suspended matters. Storms during the brief period in spring and the longer period in autumn, when the whole water body has the same specific gravity and temperature and the inversion of the stratification takes place (Figure 20), may be the cause of gaps representing up to 6,000 years found in post-glacial lake deposits in Sweden (Lundqvist, 1924, page 75). However, glacial lake deposits are remarkable in that they, if not laid down in very shallow water or directly at the mouth of a glacial river, in which instances, of course, ripple-marks and erosional phenomena are observed, never show any features or gaps referable to wave-action or currents, although such would be very easily detected. Grains and particles that had once reached the bottom remained in their original position. This makes it highly probable that a thermocline as in Figures 17 a and 17 b permanently existed and that it here, as it does in temperate lakes, limited the circulation to the upper water strata (cf. page 38).

The continuity of the deposits of the glacial lakes may be ascribed partly to the well-known effect of icebergs in preventing effective wave-action in the large and deep lakes. Stefansson (1922, page 44) tells that

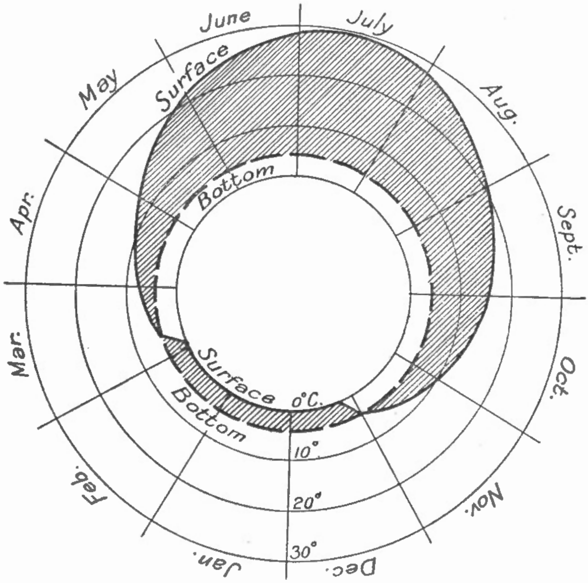


Figure 20. Diagram illustrating the distribution of temperature in Cayuga lake throughout the year. (From Needham and Lloyd, 1916, p. 34, Fig. 6.)

on the journey from the mouth of Mackenzie river to Herschel island, a route open to the Arctic ocean, they first were in a gale, but then came into smooth water in among the ice. "The ice floes were scattered. Few of them were bigger than a city block in area and there were between half-mile open patches where we sailed through smooth water though the wind was blowing stiffly." The breaking loose of icebergs can be a very spectacular phenomenon and its effects consequently are in many cases exaggerated. These effects can be far-reaching, but as a rule are local and generally cannot be compared with the effects of storm waves. The disturbances caused by calving may be more than counterbalanced by the smoothing effect of drifting icebergs (cf. Sauramo, 1923, page 118).

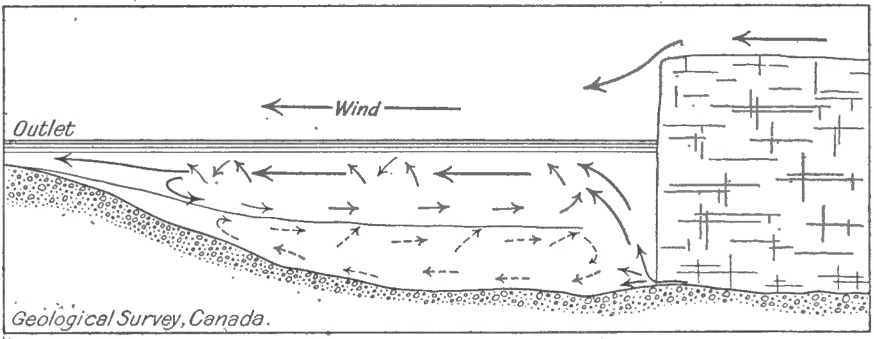


Figure 21. Circulation of water in a glacial lake. Assumed direction and strength of currents indicated by arrows.

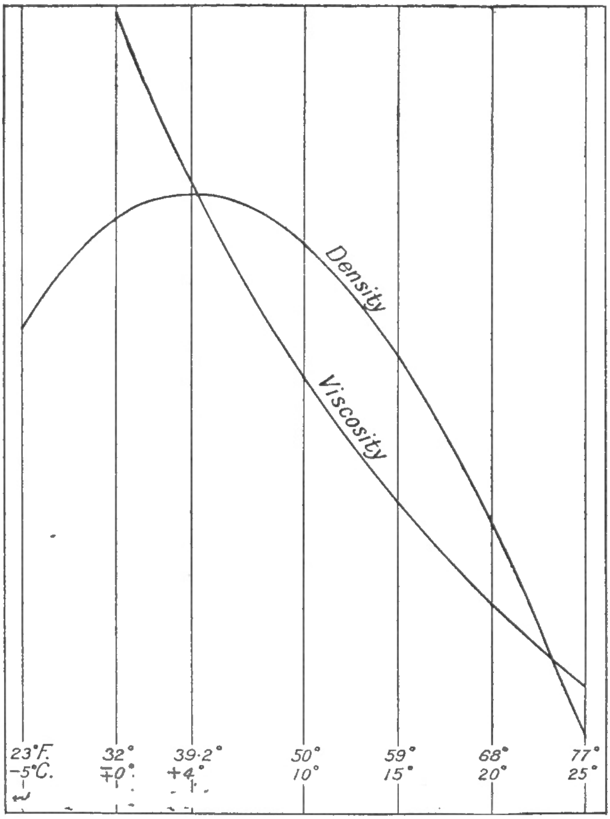


Figure 22. Density and viscosity of water. Density curve plotted on much larger vertical scale than viscosity curve.

The unchanged inverse stratification may have been an important factor not only for the transportation of mud but also for the formation of the glacial varved clay, and the inversion may be an important cause of the homogeneity of clays deposited in temperate lakes. Since the most typical varved clay known is formed in glacial lakes, and all clays deposited in typical temperate lakes with single period of supply as far as known are more or less homogeneous, the lamination probably is dependent also upon other peculiarities of the water of the glacial lakes connected with the low temperature, viz., the great density and the high viscosity. Whether temperatures of 32° to 39° F. have direct and different influence upon flocculation than have temperatures of 60° to 70° F., that is summer temperatures of ordinary temperate lakes, is not known to the writer.

Great density will cause suspended particles to fall at a slow rate. The densities of water at the respective temperatures are the following (Fowle, 1920, page 120): 23° F.—0·99930, 32°—0·99987, 39·2°—1·00000, 50°—0·99973, 59°—0·99913, 68°—0·99823, and 77°—0·99708.

As is clear from the graphic presentation in Figure 22, the density changes but little with the temperature near 39·2°, but changes rapidly as the temperature diverges from 39·2°. The density of the water of a glacial lake with a temperature just below 39·2° is consequently rather different from the density of a lake of 59°. At a temperature of 50° the density is still rather great, and this fact may have something to do with the lamination of clays in semi-glacial lakes such as Lake Algonquin.

High viscosity of the water is another property tending to decrease the rate of settling of suspended material. The viscosity of water in centipoises at the respective temperatures is as follows (Fowle, 1920, page 155): 32° F.—1·7921, 39·2°—1·5674, 50°—1·3077, 59°—1·1404, 68°—1·0050, and 77°—0·8937.

As shown in Figure 22 the viscosity undergoes a rapid and regular decrease with increasing temperature. Density and viscosity may together be the essential cause of the more rapid sinking of particles in water of high temperature than in water of low temperature. With the coarser grains temperature makes less difference. For the finest clay particles the rate of settling at different temperatures is by Hazen (1904, page 64) found to vary as $\frac{t + 10}{60}$, t, being the temperature on the Fahrenheit scale. The relative rate of settling of the same particles at different temperatures is, according to Hazen:

Temperature F.°	Temperature C.°	Relative rate of settling
32	0	70
38	3·33	80
44	6·67	90
50	10·0	100
56	13·33	110
62	16·67	120
68	20·0	130
74	23·33	140

Thus, the settling of very fine particles was found to be twice as fast at a temperature of 74° F. as at the freezing point.

Hazen's figures evidently bear only upon the conditions under which they were obtained, upon the material and water with those properties, and upon that quantity of mud in suspension. Relatively rapid settling in some cases takes place in glacial lakes, and the rate of flocculation increases rapidly with the number of particles per unit volume. The figures, however, give a good general idea of the importance of temperature.

The preceding has reference to typical glacial lakes. Varved clays, however, also were deposited in semi-glacial lakes, such as Lake Algonquin, the Baltic, etc.

Mostly the ice extended out into the lakes, and the subglacial rivers discharged at the lake bottom. During winter the lakes no doubt were entirely or largely frozen over, particularly near the ice-sheet. The stratification of the waters here was inverse; the cold and light waters were at the top. When spring came rapidly, as it does in the Arctics, and the ice broke up and melting began, the ice-cold melt-waters rose to the upper strata which probably had not yet been warmed up above 45° F., but were heavier than the melt-water. For reasons already explained the stratification may have remained inverse in the neighbourhood of the ice edge. Here the lake was a true glacial lake and the deposition of the varved clay proceeded as in such a lake, except that transportation of material did not take place over long distances. This was chiefly because the muddy surface water, before travelling far, reached a temperature of 39° F. and was drawn into complete circulation, and somewhat farther out sank into the lower compartment, for the parts of the lake which were not in immediate contact with the ice may have been of the temperate type with direct stratification in summer and inverse in winter (cf. below).

Finally, varved clays in some cases were deposited in cold temperate lakes, in which, in spite of the cold water coming from the ice, the temperature was at, or slightly above, 39·2 degrees, the critical point. The varved clay in lake Louise may have been formed under these conditions (*See* Johnston, 1922a, page 378). The upper part of profile 33, Espanola, was probably deposited in the Nipissing Great Lakes, which may have been temperate. The material may have largely been brought by a river coming from the ice edge far north of the lake shore. The varves are distinct and consist of silty summer layers and thick to very thin winter layers. The surface water of the lake may have been fairly cold. The surface temperature of lake Superior at the present time is rather low (Coleman, 1922, pages 63, 64). Off Ontonagon, Michigan, on August 24, 1921, it was 64·6° F. according to observation by Walter Koelz; near the north shore in July, 1899, it was by Coleman found to be 40°; a temperature of 39° was observed by Ramsay Wright. It, therefore, appears likely that the surface temperature near the north shore of the Nipissing Great Lakes kept below 50° F. and that thus the density was almost as great as in a glacial lake and the viscosity also fairly high (Figure 22). The river water, ice-cold when leaving the glacier, warmed up in spring more quickly than did the lake water, and thus may have been able, at least partly, to mix with the upper water strata of the lake and permit the suspended material to diffuse. The surface water of the lake after having reached the critical temperature of $39 \cdot 2°$ decreased in density with increasing temperature and thus remained at and near the top, and grew warmer and warmer until the beginning of the cold season. Thus, in summer, it probably reached higher temperatures than the river water which only had the brief time of its journey from the ice to the lake to get warmed up. Then the heavier river water sank somewhat in the lake water and spread upon the cold and heavy strata at or below the thermocline.

The actual conditions of settling of the fine particles of new-ground rock brought from the glacier were, therefore, not much different from those in true glacial lakes. In any case at least part of the fine material remained in suspension until the arrival of winter, so that varved deposits resulted. During the inversion of the stratification in autumn and spring, however, stirring up of the deposits could take place. Erosional phenomena occur in the beds at Espanola and form evidence of this action.

The upper part of the deposits south of Englehart, at localities 81 and 82, appears to have been formed under still different conditions. The beds consist of lean silt and thin winter layers. The varvity on the whole is good. The beds were laid down in Lake Barlow when the ice edge stood north of the height of land. Lake Barlow by this time (and in this part) probably was a temperate lake. If the events are correctly understood, the lake was entirely cut off from the ice-sheet, so that all the material may have been brought out by an ordinary river (cf. page 76).

To sum up: the rôle of each condition and factor for the formation of the varved glacial clay is not well known. In the last analysis the essential condition is that at least part of the fine-grained material remains in suspension until the end of the warm season. The physical conditions necessary for this action to go on seem to be that the water of the basin of sedimentation is fresh or practically so and of low temperature, that at least part of the material is derived from a melting glacier, and that the river water is lighter than the bulk of the lake water, so that the material can diffuse. In some instances varved clays seem to have been formed under conditions that did not differ very greatly from those in ordinary temperate lakes, though the deposits are so strikingly different from the homogeneous clays commonly formed.

CHAPTER VI

CONDITIONS CONTROLLING ICE RECESSION

The origin of ice-sheets is primarily due to low summer temperature and heavy precipitation in solid form, and vanishing of ice covers is due to rise in summer temperature and decrease of solid precipitation.

The conditions controlling the recession are nourishment and depletion. The last ice-sheets during their disappearance were nourished, and the ice moved constantly forward, so that the retreat represented the excess of disintegration over growth.

During the waning of the ice-sheets their nourishment was mostly slight. In fact this was a necessary condition of the ice retreat (*See* Antevs, 1925). At times, however, the marginal nutriment was sufficient to counterbalance the depleting agencies. Thus, the halt of the ice border, 183 years long, at the Second Salpausselkä in Finland seems to have been due to great supply during a time of considerable melting (cf. page 53).

The depletion of the land ice took place through agencies working from above, from the ice front, from below, and in the interior, that is, in fissures.

Ablation or depletion from above occurred in the form of melting and evaporation and was caused or influenced by temperature, insolation, rain, air, wind, melt-water, debris-cover, etc. Disintegration from the ice front, when it ended in water, occurred through discharge of icebergs and melting caused by the water. Depletion from below and in the interior took place through melting. Bottom melting was caused by circulating water, and internal melting by water and air working down in fissures.

Depletion from above and from the front was evidently by far the most important. The rôle played by each agency varied greatly.

The amount of disintegration in even terrain and with constant nourishment and thickness of the ice is directly recorded by the rate of ice retreat. Although these conditions in a strict sense are seldom fulfilled, the depletion generally is approximately recorded by the recession. There are, however, important exceptions. In a deep, water-filled basin in which the ice moves rapidly the ice can expand, even though strong depletion by calving and ablation takes place. An ice front reaching shallow water, after rapid recession by calving in deep water, can remain stationary for some time, although considerable ablation occurs, and so can an ice front resting against a mountain wall (Sauramo, 1923, pages 150-155).

Not all the depleting agencies are recorded by the quantity of the sediments nor by the thickness of the clay varves. Calving is scarcely recorded at all, and evaporation only partly. The material below ice detached as icebergs remains on the spot as till. Debris attached to icebergs can become partly assorted during the sinking, and contribute somewhat to the sediments, or it can go down along with an overloaded ice-mass and form a local till bed. Material concentrated through evaporation of ice may partly be washed out into the sedimentation basin by melt and rain waters, but much may remain on the spot to form till. Thus the quantity of the sediments primarily record the total melting.

The present-day depletion of glaciers furnishes, it is true, the basis of the discussion of the disintegration of the Pleistocene ice-sheets, but it should be kept in mind that the late-glacial conditions were very different in degree. The fronts of the Pleistocene ice-sheets ordinarily retired hundreds of feet a year, whereas those of the existing ice-caps since long have remained stationary or practically so, and the existing glaciers, which since about 50 years are shrinking in most parts of the earth, recede a few feet or tens of feet annually (Hess, 1904, page 290; Rabot, 1909, page 333; Supan, 1921, page 216).

MELTING

Melting of ice takes place through high temperature, insolation, rain, air, wind, and melt-water, and is influenced by the existence or lack of debris-cover and its thickness.

By far the most important factor for ablation of the glaciers in the Alps and in the Jotunheimen is warmth (Heim, 1885, page 257; Hess, 1904, page 210; Øyen, 1893, page 46). During bright, fair weather the glacier brooks carry much more water than during and after rains, and what they carry after rains is mostly rain water. In August, 1844, Desor and Dollfuss found the water quantity of the Unteraarebach in the Alps during three consecutive phases to be (Heim, 1885, page 257):

—	Mean temp. F°	Daily ablation	Water quantity per day Cub. m.
Fair	44·0	0·051	2,000,000
Snowfall	35·6	0·000	1,260,000
Rain	39·2	0·013	344,000

The importance of temperature and insolation for ice melting in the Alps is also shown by the annual depletion. In winter no ablation, or surface-melting, but only bottom-melting, occurs. The ablation after its start in spring increases gradually and reaches its maximum during the hottest month, generally July, after which it gradually decreases to zero at the arrival of winter (Heim, 1885, page 259; Hess, 1904, pages 232, 239, 240; Greim, 1903, 1910). The following table gives the monthly means of the quantities of water of two glacial brooks in the Tyrol (the Ruetzbach at Ranalt and the Jambach at Galtür), and also the precipitations (*See* Hess). The last column gives the run-off of the Rhone glacier as measured at Gletsch just below the glacier (Mercanton, 1916, page 100). Figure 23 shows the monthly quantities of water for 1900 and 1901 at Gletsch, and also the rainfall in the drainage area of the brook (Hess, 1904, page 235). At Ranalt and Galtür the greatest runs-off during July coincide with the heaviest precipitations, but as explained were essentially not a consequence

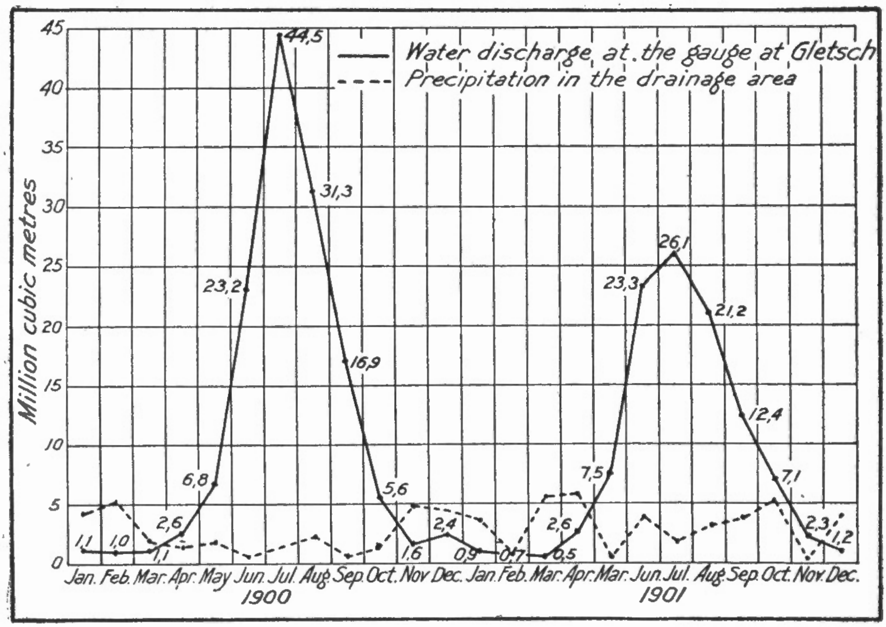


Figure 23. Curves showing water discharge in the Rhone at Gletsch and precipitation in the drainage area during 1900 and 1901. (From Hess, 1904, Fig. 48, p. 235.)

Mean Monthly Discharge of Glacial Brooks and Precipitation

—	Ranalt			Galtür		Gletsch
	Means for 1891 to 1896 Quantities of water cub. m. per second		Precipitation Means for 1892 to 1896 Mm.	Means for 1894 to 1900 Monthly quantities of water million cub. m.	Precipitation Mm.	Means for 1894 to 1903 Quantities of water cub. m. per second
	7 a.m.	5 p.m.				
January	1·76	1·76	48·2	3·28	73	0·3
February	1·46	1·48	25·9	2·95	69	0·3
March	2·20	2·54	60·5	3·78	146	0·3
April	3·65	4·22	53·8	3·94	70	0·7
May	6·07	7·29	74·7	10·25	84	1·8
June	9·13	10·24	74·1	20·21	116	6·0
July	11·60	12·55	94·5	25·57	139	9·6
August	9·48	11·72	67·5	22·03	132	8·2
September	7·90	9·52	86·0	15·69	109	5·3
October	5·34	5·67	83·2	8·33	62	1·7
November	3·40	3·45	42·3	5·52	26	0·8
December	2·75	2·75	47·1	3·94	67	0·5
Year	5·3	6·1	757·8	....	....	2·95

of the rains, the variations in the amount of which are relatively very small, for the glacier brooks swell, particularly during days and periods of fair weather. The depletion of the Rhone glacier in 1895 as recorded at Gletsch increased very abruptly on June 4, and decreased quite gradually in September and October (Mercanton, 1916, page 100, Figure 14), (*Also* Rabot, 1909, pages 89-96).

The rôle played by temperature is also illustrated by the daily course and fluctuations of ice melting and of glacier brooks (Heim, 1885, page 255; Hess, 1904, pages 213, 230-233; Mercanton, 1916, page 101). As the sun rises melting slowly begins. During a fair day the ablation reaches its maximum in the afternoon and thereafter decreases, as the temperature falls with the setting of the sun. The glacier brooks vary accordingly. The smaller the glacier, the quicker and relatively greater is the response of the brook to the differences between day and night. Small brooks whose waters have only short distances to travel to the glacier edge reach their greatest volumes early in the afternoon and run dry at midnight, whereas brooks from long ice tongues may be highest at midnight and lowest at 10 o'clock a.m. Examples of daily fluctuations of glacial rivers in Asia are given by Huntington (1907). These daily variations are practically limited to summer (cf. Mercanton, 1916, page 101, Figure 14).

The above statements bear upon the combined effects of warmth and insolation. The relative rôle played by each agency has been somewhat studied (Hess, 1904, page 210; Maurer, 1914, 1914a). At Zürich, at an elevation of 1,575 feet, on bright days in August the melting of horizontal ice blocks, about 3·6 square feet, in sunshine and in shade during nine to ten hours, was about as 6 to 5 (Maurer). The exposed blocks lost about 20 pounds or a layer of 0·79 inch through the insolation alone. It is calculated, by Maurer from his experiments, and from actinometric observations by Dorno, that at Davos in Switzerland, 5,250 feet above the sea, the ice ablation through the immediate effect of the sun's rays from May to September may amount to 8·92 feet.

Melting by high temperature and insolation may have been the most effective causes of the disappearance of the ice-sheets. That this was possible from a physical point of view was proved by Schiötz (1891, page 22). The best direct evidence bearing on the question is the exceedingly slow retreat on the Swedish west coast, which in late-glacial time probably was excessively foggy and fairly cold, and the rapid recession in the Baltic region which probably had clear, sunny summers (De Geer, 1910a, 1912, 1914, 1914a; cf. Antevs, 1922, page 85). In West Greenland the summer temperature because of cold fogs is low on the coast and increases landward (von Hann, 1911, page 666).

Rains in the Alps play a much less important rôle in the depletion of the glaciers than do warmth and insolation. This is due to the low temperature of the rain water which usually is 2° to 5° F. below that of the air (von Hann, 1915, page 309). In Arctic regions rains may play an even less important rôle, for they are generally only a few degrees above freezing point. In some parts of the Arctic, such as East Greenland, rains form only a fraction of the precipitation, but in most other arctic regions (West Greenland, Spitsbergen, Franz-Josephs-Land, and the North American archipelago) they represent a fairly large percentage (von Hann, 1911, pages 615, 623, 630, 662). Most of the precipitation, however, usually comes in autumn as snow.

The annual ablation of the glaciers about 1,300 feet below the snow-line in the Alps amounts to about 13·11 feet; in Scandinavia to 10·8 feet; and in Greenland to 6·6 feet (Hess, 1904, page 218; cf. Wright and Priestly, 1922, page 293). The small ablation of the Greenlandic ice-sheet as compared with that of glaciers in the Alps and in Scandinavia is attributed by Hess to the fogs and rains which prevent fair summer days. The intensities of insolation in the Alps and in Greenland are about 107 to 46·5.

Warm rains evidently have very great ablating effect. Warm rains connected with foehn winds in 1912 caused considerable ice melting at 70 degrees north in west Greenland (Mercanton, 1920, page 272). Such rains, though, may be rare, for foehns are almost always characterized by extreme dryness (cf. page 51).

It seems possible that rains played a relatively greater rôle in the waning of the Pleistocene ices and particularly in the depletion of their southern parts, than they do in the ablation of existing ice caps and glaciers. The summer temperature in late-glacial time was high, and the ice-sheets extended to low latitudes, so that the rains in many instances undoubtedly were fairly warm. However, the exceedingly slow ice retreat on the Swedish west coast, which certainly had much greater rainfall than the central and eastern parts of the ice front, is significant. Furthermore, on the whole the precipitation appears to have been scanty during the ice retreat.

Conditions of great importance for melting of ice are the occurrence of a debris-cover and the thickness of this layer (Heim, 1885, page 223; Hess, 1904, page 219; Hobbs, 1911, page 166; Wright and Priestly, 1922, page 283). A thin layer of fine, dark-coloured material transmits the absorbed insolation heat to the subjacent ice and greatly facilitates ablation. Small, scattered rock particles cause holes and thus give other depleting agencies larger surfaces of attack. A thick debris-layer and boulders, on the other hand, because of the slight conductivity of rock, cannot transmit their absorbed heat to the subjacent ice, and since they shut off the warm air and the sun's rays, they check ablation considerably (*Also* Philipp, 1912).

EVAPORATION

When an ice body is heated on the surface it will partly thaw and partly evaporate. Melting begins at 32° F., but evaporation takes place at any temperature. The latent heat of fusion of ice or amount of heat in calories required to melt a gramme of ice without raising its temperature is 79·6, that is practically as much heat as is required to raise the temperature of the same quantity of water from 32° F. to 175·3° F. The heat of vaporization is still greater, being 595 calories at 32° F., 590 calories at 50° F., and 585 calories at 68° F. Therefore, when an ice surface is warmed up to 32° F. thawing will greatly exceed evaporation however favourable the conditions for evaporation may be.

Evaporation depends on the temperature of the surface from which it takes place, on the conditions of saturation of the air above the surface, on the air exchange, and on the pressure.

The surface temperature of the ice, even during the hottest days, evidently does not rise above 32° F., but the melt-water can be somewhat heated.

The capacity of the air for vapour increases progressively with rise of temperature. The expansive force of the greatest quantity of vapour that can exist at 32° F., or the pressure of the saturated aqueous vapour at this temperature is 0·18 inch. The vapour pressure over ice and water at some other temperatures is as follows: −4° F. 0·03 inch, 14° F. 0·08 inch, 50° F. 0·36 inch, 68° F. 0·69 inch (Smithsonian Tables, 1918, pages 160-166).

The mean temperatures of the air above the Pasterzen glacier in the Alps were found by H. and A. Schlagintweit during several bright days in August to be (Heim, 1885, page 237; Hess, 1904, page 211):

Air Temperatures above Pasterzen Glacier

Height above the ice Feet	Mean Temperature F.°
0·5	36·7
3·3	40·8
6·6	42·3
11·2	43·3
124·7	47·7

The daily temperature variations were much less near the ice surface than at some higher elevation. The temperature gradient was greatest at about 2 p.m.

Evaporation is greatest when the temperature is highest, and when the vapour pressure difference between the ice and the air is greatest. The lower the relative humidity of the air layer in contact with the evaporating surface, the greater its capacity of absorbing vapour and the greater the evaporation. Since the lowermost air layer quickly becomes saturated with vapour, rapid exchange of the air by wind and convection currents greatly increases the rate of evaporation. Of greatest importance naturally is the wind. The amount of evaporation increases somewhat in proportion to the square of the velocity of the wind (Wright and Priestly, 1922, page 275). The rate of evaporation is indirectly proportional to the air pressure (von Hann, 1915, page 214).

Studies on evaporation of snow and ice have shown that evaporation, when the atmosphere is dry and the air exchange is rapid, plays an important rôle in ice depletion (*See* Brückner, 1921a). At Ottawa in the spring of 1924 a noticeable percentage of the thick snow cover disappeared by evaporation. On the other hand, in a region such as the central Swedish lowland, evaporation of snow is only little greater than the condensation on it (Westman, 1913).

In late-glacial time the sky at some distance from the sea may have usually been bright and the air dry, and the atmospheric absorption as a consequence small and the solar radiation strong. Thus, the summer temperature of the air over the ice may have been relatively high. During the summer probably almost permanent constant winds came from the inner parts of the ice-sheet because of the glacial anticyclones and the heating of the land off the ice front. In north Greenland and in west Greenland far inland there is in summer a constant flow of cold air from the ice plateau (von Hann, 1911, pages 669, 675). In winter when the temperature of the air over the sea is higher than that over the ice, centrifugal winds are practically permanent everywhere on the Greenlandic coast.

The winds coming from the interior of the ice-sheets gave up their moisture at low temperatures inside the marginal belt, and when they approached the ice border were far below saturation because of the higher temperature there. Mostly the winds were still ice cold when reaching the ice edge, but winds coming from great heights were dynamically heated during the descent and formed foehns.

Foehn winds are particularly characteristic of the Alps and are well known in connexion with the Greenlandic and Antartcic ice-sheets (von Hann, 1915, pages 572-581; Hobbs, 1911, pages 149, 268). They are primarily characterized by high temperature and extreme dryness. They are caused by a steep barometric gradient or by an anticyclone. Foehns occur at all seasons, but are most frequent in winter and spring. The average number of foehn days on the north side of the Alps is as follows (von Hann, 1915, page 575):

—	Winter	Spring	Summer	Autumn	Year
Switzerland (20 to 37 years, 5 localities) %	26·0	31·7	17·3	25·0	100·0
Bludenz (10 years), days....................	10·6	8·2	3·1	10·0	31·9
Innsbruck (25 years), days.................	9·5	17·0	5·0	11·1	42·6

As the rise of temperature of a descending wind is about 1° F. for every 180 feet of descent, the actual rise can amount to several tens of degrees, and foehns may cause summer heat in the middle of the winter. At Bludenz in the Alps a foehn on February 1, 1869, raised the temperature to 66·7° F., and the mean afternoon temperature of twenty foehn days in winter time was 57·2° F. In East Greenland a foehn in the winter season in a single hour changed the temperature by 43° F.

Although all winds greatly promoted evaporation foehns evidently did so particularly. Furthermore the warm winds caused melting. Besides these effects strong winds caused considerable mechanical ablation of ice projecting above the general level of the surface (Wright and Priestly, 1922, page 275).

Thus evaporation may have been a fairly important factor in the ablation of the Pleistocene ice-sheets. In comparison with melting, however, it may have played a rather subordinate rôle. Its importance cannot very well be estimated by the relative quantity of water-sorted material in the drift as tried by Visher (1922, page 123). The amount of stratified drift in a region is primarily dependent on the occurrence and size of basins which later have been laid bare. The stratified drift corresponding to the till in the regions north of the Great Lakes is to be found largely on the bottom of these lakes, and the sand, silt, and clay corresponding to the till in South Dakota may to a great part occur on the Missouri and the Mississippi, and in the delta of the latter river.

DEPLETION FROM THE ICE FRONT

Disintegration from the ice front, when the ice extended into water, took place through discharge of icebergs, and through melting. Discharge of icebergs takes place in different ways (*See* Engell, 1910; Hobbs, 1911, page 178; Wright and Priestly, 1922, page 406). Most glaciers outside the Antarctic, where floating ice fronts are a rule (Wright and Priestly, 1922, page 196), probably rest with their whole weight directly on the bottom (Gilbert, 1904, page 215), but a few examples of floating glacier extensions are known (*See* Hobbs, 1911). Floating ice tongues can be attacked by the buoyancy of the water without sub-water melting. The overriding of the upper ice layers over the lower strata and undermining through melting cause formation of crevasses and facilitate detachment of ice blocks which may shear off from higher levels or break off at or below the water-level. Erosion through wave-action at and near the water surface, so that a protruding ice-foot under the water is formed, renders detachment of bergs possible through the lifting power of the water. In this way very large icebergs can be formed. The very biggest bergs separate from the entire thickness of the ice edge. The Greenlandic fiords are almost constantly so filled with pack ice that it is impossible to get through them with a boat (*Also* Rabot, 1909, page 104).

Calving played a very great rôle in the recession of the Pleistocene ice-sheets. Its effects evidently were greatest in wide and deep waters with large and powerful waves, but they made themselves felt even in quite small basins. But basins and valleys also facilitated the motion of the ice. The result of the large supply and large wastage usually was different in the distal and proximal parts of a basin. During the uncovering of the distal part of a basin this part of the basin was occupied by a protruding ice lobe. When the ice front had retired about to the central part of the basin the edge formed a practically straight line. Thereafter it grew increasingly concave. These conditions are particularly evident in the case of large basins as the Baltic (De Geer, 1914a, page 216), but may have regularly prevailed in small also, wherever the topographic conditions were favourable, as is shown by detailed mappings of the ice front (De Geer, 1912, Plate 2; Antevs, 1915; Frödin, 1916; Sauramo, 1923, pages 130-141). Even when the ice front was stationary on land considerable retreat could take place in deep water. A zone of retreat 12 miles wide in Södertörn, the region south of Stockholm, represents the halt in shallower water and on land during which the second series of the Fenno-Scandian moraines were formed (De Geer, 1917, page 20).

The actual rôle of calving can be estimated only approximately because retreat through discharge of icebergs is not recorded in the quantity of the glaci-fluvial deposits which largely depended upon the debris-loaded melt-water. Great sedimentation, thick annual layers as a rule, records much ice melting, thin varves indicate little thawing. The rate of flocculation and the spreading of the material must also be kept in mind. It was observed long ago by De Geer that generally speaking rapid recession was represented by thick varves, slow recession and halts by thin varves. This means that melting, and, therefore, temperature, largely determined the rate of recession (*See also* Sauramo, 1918, page 37; Antevs, 1922, page 84).

The halt during which the southern belt of the Fenno-Scandian moraines was formed in southernmost Södertörn is represented by exceedingly thin varves. The rapid retreat at Stockholm is represented by thick varves. The water depths were about the same in both cases.

The halt of the ice border at the First Salpausselkä in Finland and at the corresponding moraines in southernmost Södertörn was represented by small glaci-fluvial deposition. This halt cannot, as Sauramo (1923, pages 135, 151) holds, be due primarily to topography and calving, for then the recession between this halt and the Second Salpausselkä, which took place at an increasing rate (Sauramo, 1918, Plate 3) although the water depth decreased (Sauramo, 1923, page 135, Figure 14), is unexplainable. It is particularly significant that the Swedish correlatives of the Salpausselkäs are located north of the South-Swedish highland on low land which was covered by the Baltic ice lake and by the sea. The halt at the Second Salpausselkä on the contrary was corresponded by large deposition (Sauramo, 1918, page 38; Plate 4) which, as the present writer (1922, page 87) pointed out, may be due to considerable melting. The fact that the ice edge did not retire can only be explained by large supply. The existing favourable conditions for flocculation can explain the great thicknesses of the varves in the neighbourhood of the ice, but not the increased deposition at distant points (cf. Sauramo, 1923, page 162). In this equilibrium between large supply and large wastage the topography, as Sauramo (1923, pages 135, 138) points out, played an important rôle. The halt occurred when the ice edge stood on the highest land, i.e. in the shallowest water. If, however, the supply had been sufficiently small the ice front would have slowly retired in spite of the checked calving, as it usually did in the supra-aquatic regions.

The lack of correlatives in supra-aquatic Karelia or East Finland, of the moraine lines, is put forth by Sauramo (1923, pages 143, 155) as a reason for his attributing the halts at the Salpausselkäs to topographic conditions and decreased calving. Lack of moraines, however, does not necessarily indicate ice retreat. The Fenno-Scandian moraines are not continuous across Sweden (De Geer, 1910; Munthe, 1910a, Plate 47). In New England continuous ridges corresponding to the halts seem to be lacking.

Therefore, Sauramo's (1923, pages 129-141, 150-159) opinion that calving determined by water depths was the chief factor of ice retreat and of halt may represent a great over-estimation of its rôle and an under-estimation of the importance of the climatic factors. The invalidity of the statement that the annual periodicity is the only climatic periodicity indicated by the varved clays and that all variations longer than the year may be due to physiographic factors (Sauramo, 1923, pages 161, 162) may be evident from the correspondence between the curves from the different valleys in New England (Antevs, 1922). The frequently poor agreement between the clay curves in Finland (Sauramo, 1923, page 11) may be due largely to the conditions of transportation of the glacier mud, which were unfavourable in the first place because of scarcity of distinct currents following the same course year after year.

INTERNAL DEPLETION AND WASTAGE FROM BELOW

In the marginal zone internal melting and bottom melting through air and by melt-water and rain-water working down through fissures were of some importance. Locally, melting by heat conducted from below the surface of the earth, probably mostly by springs, took place. Melting by friction in the interior of the ice perhaps did not occur because the heat was abducted to the surrounding cold ice (Schiötz, 1891, page 15), or in any case was exceedingly insignificant. Wastage at the bottom of the ice-sheet by heat produced by pressure or friction and by heat stored in the surface layers of the earth's crust, which do play or are supposed to play their part in the depletion of existing glaciers, as will be discussed, may not have had any effect on the ice-sheets. As a consequence there was practically no melting in winter, except through warm springs or foehn winds (cf. page 51). The glacier rivers then went dry, as is shown by the occurrence at Stockholm of greasy winter layers in large eskers of coarse gravel deposited immediately outside the glacier vaults.

In the Alps temperature measurements on glaciers have been made to depths of 535 feet (Hess, 1904, pages 151-154, 320; Mougin and Bernard, 1905, pages 169, 173). It has been found that the influence of the atmospheric temperature penetrates to depths of somewhat more than 26 feet and that the inner constant temperatures of glaciers practically equal the melting temperatures of ice under the respective pressures. The melting point of ice is known to be lowered under pressure, because water expands on freezing and ice expands further with sinking temperature. The temperatures of alpine glaciers are a trifle lower than the melting temperatures calculated from the weight of the ice and with 0·0075° C. as the amount of the lowering of the melt-temperature for each atmosphere of pressure.

The differences between the observed and calculated temperatures increase with the depth, but the results are by Hess considered not to be accurate enough to justify the conclusion that the differences represent side pressure caused by ice motion.

The effects of the winter's cold on the snow and ice of the Great Karajak glacier near the coast in central West Greenland is noticeable down to a depth of 60 feet or more, below which depth the temperature is very near 32° F. (Drygalski, 1897, pages 470-472; *See* Hobbs, 1911, page 160).

A pressure of 500 kilograms per square centimetre corresponds to a melting temperature of ice at −4·1° C. (Bridgman, 1912, page 537). An ice-sheet of a thickness of 1,000 metres, that is of the supposed average thickness of the Pleistocene ice-sheets, through its weight exerts a pressure of 99·98 kilogrammes per square centimetre and, therefore, if there were no side pressure, could exist under a temperature near the bottom of about −0·82° C. A land ice 2,000 metres thick under the same conditions could exist at a bottom temperature of −1·64° C. Because of additional pressure caused by the slope of the ice surface and the resulting ice motion, by obstacles to the ice flow, etc., the temperature must have been somewhat lower than the values calculated from the weight alone. Thus, the temperature of the bottom strata of the Pleistocene ice-sheets may have been 2° to 4° F. below freezing. Because of the great stability of ice against pressure, due to the enormous latent heat of fusion of ice (79·6 gramme-calories) which caused the produced heat to be too quickly consumed by the melting, it seems likely that no melting worth mentioning occurred as a consequence of pressure, at least near the ice edge during the waning, where the thickness was comparatively small. The bottom melting by pressure is by Schiötz (1891, pages 13-16) calculated to be only 1 to 2 per cent of the annual growth of the ice, an amount that is entirely negligible.

The surface layer of the earth's crust is in summer time warmed up considerably above the temperature of the air and in winter cooled off somewhat below the temperature of the atmosphere above. The mean temperature of the ground is somewhat higher than that of the air (von Hann, 1915, page 48). The annual fluctuations of the temperature at the earth's surface make themselves felt to depths corresponding to the amount of the fluctuations and to the severeness of the winters. They penetrate only slightly in the tropics, 65 to 100 feet in mid-latitudes and high latitudes (von Hann, 1915, page 21), and possibly still deeper in the Arctics. How deep they reach in arctic regions is not known. The depth of frozen ground, at Yakutsk, calculated to amount to 650 feet, does not furnish any information, for ground ice is essentially formed by burial of surface ice, by material transported by spring floods, and by development of vertical wedges of ice by growth in place (Leffingwell, 1919, pages 182, 242). The low temperatures in the coal mines in Spitsbergen (*See* Leffingwell, 1919, page 183), which furnish the only data known to this writer from the Arctics on temperature in rock far from the surface, give no indication, for the tunnels lie high up on the exposed mountain sides, where the winter cold easily penetrates along the surfaces of the layers of shales and sandstones. In Connecticut the seasonal fluctuations are believed to reach a depth of 50 to 60 feet (Palmer, 1920, page 65). In Europe as far north as Finland winter frost usually penetrates less than 3 feet (von Hann, 1915, page 54). The temperature of the surface layer of the earth below the Pleistocene ice-sheets, shut off from the sun and the air, equalled the temperature at the bottom of the ice, that is was 2° to 4° F. below freezing. Thus, the ground was permanently frozen to some depth. The ground temperature consequently was considerably higher than in unglaciated high arctic regions. Since the present year-isotherm of 28° F. runs from the easternmost part of Labrador over the southernmost end of James bay to the southern part of the Alaskan peninsula, generally speaking the ground temperature during the ice age north of this line was higher, south of it was lower, than at present. It is evident that since no solar heat in summer was stored in the ground under the ice-sheets, no bottom melting through stored heat could take place (*Also* Kayser, 1915; Supan, 1921, pages 9–11; Pohle, 1924).

Below the zone affected by seasonal variations the temperature is uniform the year round and somewhat higher than the mean air temperature at the place. From the upper limit of the constant zone the temperature increases downwards about 1° F. for each 64 feet. As a consequence there is a constant flow of warmth from the interior of the earth towards the surface. The quantity of heat delivered by the earth's surface to the atmosphere is, however, very small. The annual mean amount is by von Hann (1915, page 22; cf. Schiötz, 1891, page 12) calculated at 54 calories per square centimetre, that is a quantity of heat able to melt a layer of ice 0·29 inch thick. Normally, therefore, bottom melting of glaciers through heat from the interior of the earth is negligible, and in the case of ice-sheets entirely absent, because the heat was quite insufficient to prevent the ground beneath the ice from freezing.

In the Alps, in Scandinavia, in Canada, and in other countries, bottom melting of existing glaciers takes place all the year round (Heim, 1885, pages 259-263; Hess, 1904, pages 225, 226, 233-240; Mercanton, 1916, page 100; Johnston, 1922a, page 377). Only brooks from very small glaciers are found to dry in winter. Also in Greenland winterly bottom melting of glaciers was observed by K. J. V. Steenstrup, Fridtjof Nansen, and Erich von Drygalski (*See* Heim, 1885, page 261; Hess, 1904, page 240). The quantities of water in the glacier streams in winter, of course, are small compared with those in summer, but in some cases are rather considerable (cf. page 47). The bottom melting in exceptional cases in the Alps is found to amount to one-ninth of the surface melting. The water of the glacier brooks in winter in some cases is almost clear, but mostly, and as in Greenland, it is rather muddy owing to suspended material, though less so than in summer. The winter melting may entirely or largely take place through abnormal heat coming from the earth, for instance by means of warm springs. In the Alps the ground temperature is abnormal. Its year-isotherm of 32° F. lies at elevations of 9,000 to 9,500 feet, whereas the same year-isotherm of the air lies at elevations of 6,200 to 6,550 feet (Hess, 1904, page 224). The glacier tongues lie in beds whose temperature is always above the melting point of the ice. Beneath the Hintereisferner glacier about 20 inches below the ice a constant temperature of 33·8° F. may prevail (Hess, 1904, page 225).

In temperate regions part of the winter melting on the edges of the glaciers may take place through solar heat stored in the ground. The quantity of heat stored in summer and given off in winter is quite large. The temperature gradient in winter is considerable. The quantities given off by the ground at Tiflis and Irkutsk, respectively, are calculated by von Hann (1915, page 55) at 13·8 and 18·4 gramme-calories per day and square-centimetre.

CHAPTER VII

VARVED CLAY IN EASTERN ONTARIO, TIMISKAMING AND COCHRANE DISTRICTS

The areas studied are the environs of Ottawa and Arnprior; a belt along the railway from Toronto to North Bay via Beaverton, and from North Bay to Sault Ste. Marie; the Timiskaming and Mattagami regions from Mattawa to lake Timiskaming; and a belt along the railway past Cochrane to the second crossing of Abitibi river 44 miles north of Cochrane.

At Ottawa and Arnprior marine clay is abundant, but exposures where varves can be measured are rare. The measured clays were deposited partly in fresh water, that is in Lake Frontenac, and partly in the Champlain Sea (*See* pages 64 and 66).

Along the railway from Toronto to North Bay clay deposits are small and far between. Probably all the important deposits were visited and all important exposures were measured. Except those places where series were obtained, no detailed examination was made except at Gravenhurst, but observations from the train gave a fair idea of the character of the sediments and of the possibilities of good sections. At no place except those visited did the prospects appear favourable. The country is largely either rocky or swampy. In both cases it is forested and the ground is mostly covered by till, gravel, and sand. Varved clay and silt were observed in places, for example near lake Simcoe, but the deposits were either very thin, poorly drained, or without exposures, so that good profiles seemed out of the question.

West of North Bay stops were made at Sturgeon Falls, Sudbury, Espanola, Massey, Cutler, and Dean Lake. From Sudbury trips were made to Larchwood and Capreol. This region north of lake Huron presents nearly the same conditions as that east of Georgian bay. Clay is very rare. It occurs in places, but for different reasons, as disturbances, insignificant thickness, poor drainage, etc., it is mostly valueless for geochronological studies.

Between North Bay and Cobalt there is probably no varved clay. Locally there is gravel and sand, but mostly the rocky ground is bare or covered by till.

On the Ottawa north of Mattawa no material finer than sand may occur. Gravels, partly very coarse, are abundant.

Lake Timiskaming from its southern end at Timiskaming station to the mouth of Montreal river has, with few exceptions, high, rocky shores. At McLaren point, as observed from the steamer, there may be sediments high above the lake level. Just south of McMartin point some bluffs cut in silt or till were observed, and at Martel point sediments some 60 to 80 feet above the lake; otherwise, nothing promising was seen.

From the mouth of Montreal river and northward clay is widely distributed, particularly on the east side of the lake (Wilson, 1910). Clay of great thickness covers the whole lowland north of lake Timiskaming and extends northward to and across the height of land in westernmost Quebec. Along the railway, clay occurs to north of Krugersdorf. At Boston creek the ground is above the level of Lake Barlow, as the glacial Lake Timiskaming is called, and clay is absent or nearly so. Some miles south of Ramore the railway descends on the clays of Lakes Ojibway and Barlow-Ojibway. North from here clays and silt are again widely distributed as far as Porquis Junction, Iroquois Falls, and La Sarre. North of the two first-mentioned places no clay was found along the Temiskaming and Northern Ontario railway, nor along the new line between Iroquois Falls and Stimson, which largely follows eskers.

By courtesy of Mr. A. L. McDougall, Division Engineer of the Temiskaming and Northern Ontario railway, the writer was enabled to follow the new railway from Cochrane north to the second crossing of the Abitibi at mileage 44·4. For the first 6 miles, and particularly north of mileage 2½, there are several cuts in gravel, sand, and stony clay, or in mixtures of all three. At mileage 3½ to 4 there are some thin deposits of varved clay containing 20 to 25 varves, and underlain by till or gravel. At mileages 5 and 7 there are deposits of about 4 varves and 15 varves respectively, both being underlain by gravel. Just south of mileage 20 there is another clay bed with, at most, 10 varves. These are the only occurrences in the railway cuts up to 44 miles north of Cochrane. The clays, evidently, were laid down in small depressions, not in a large lake. For the rest, the ground consists of unstratified stony clay with scattered small pebbles and an occasional boulder. The ground is even or gently rolling.

CHAPTER VIII

CONDITIONS IN SOUTHEASTERN CANADA DURING THE DEPOSITION OF THE VARVED CLAY

LAKE ALGONQUIN

Lake Algonquin is the water-body which in late-glacial time occupied the basins of lakes Superior, Michigan, and Huron. It has been studied particularly by Taylor and Leverett (1915, page 409), Coleman (1922, page 19), Goldthwait (1910), and Johnston (1916), from whose writings the following account is largely taken. During its early stage in the Huron basin the lake discharged at Port Huron. When the ice front had retired to the northern end of Michigan peninsula, and Lake Chicago in the Michigan basin became part of Lake Algonquin, it probably was drained

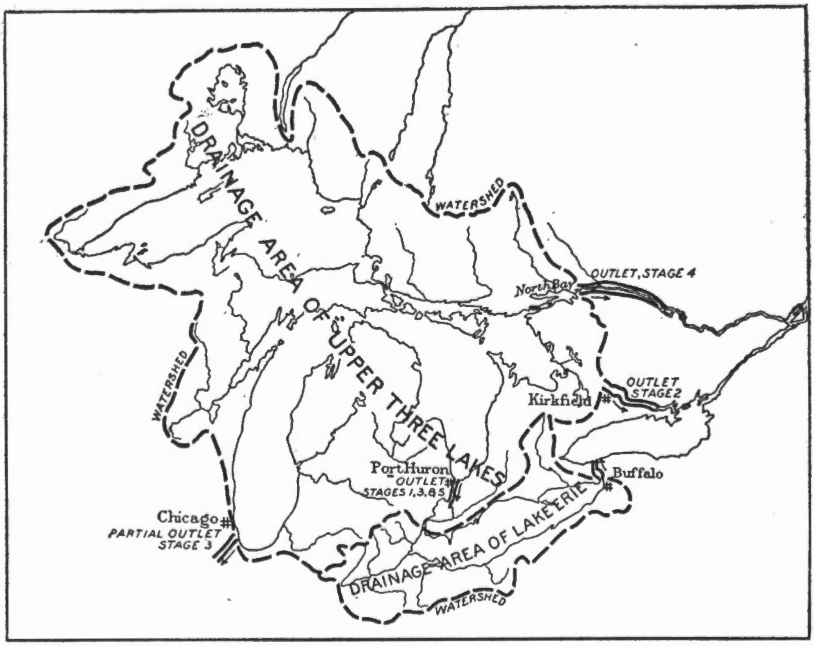


Figure 24. Sketch map showing drainage areas tributary to Niagara river, during five lake stages, and various outlets of the upper three Great Lakes during these stages. 1, Early Lake Algonquin; large discharge divided between Niagara river and four other spillways over Niagara escarpment. 2, Kirkfield stage of Lake Algonquin; 15 per cent of present volume through Niagara river and large discharge through outlet at Kirkfield. 3, Port Huron-Chicago stage of Lake Algonquin; first 110 per cent, then 90 per cent, of present volume through Niagara river and small discharge through outlet at Chicago. 4, Nipissing Great lakes; 15 per cent of present volume through Niagara river and large discharge through outlet at North Bay. 5, Present Great Lakes; full discharge (100 per cent) of the upper four lakes through Niagara river. (From Taylor, 1913a, p. 21, Figs. 12 and 13.)

both past Port Huron to the Hudson system and past Chicago to the Mississippi. These introductory stages were of short duration. During this time the clays on lake Simcoe, at Bracebridge, Sault Ste. Marie, and Massey may have been deposited. When Stony lake, 88 miles northeast of Toronto, was uncovered, an outlet, which was 50 to 100 feet lower than those at Chicago and Port Huron, was opened through Trent valley to the Ontario basin. This outlet, sometimes called the Kirkfield outlet, soon took the whole overflow, leaving the older ones dry. As melting of the ice went on and the load diminished, the Trent Valley region slowly rose, so that the previously abandoned outlets at Port Huron and Chicago, which were not rising, were again taken into use. The strong beach developed during the transition stage with three outlets appears to indicate either a pause or a very slow rise, or more probably first a slow rising and then a slow falling (Taylor, 1915, page 411). For some time the region of Trenton stood higher in relation to the level of Admiralty Lake, as that stage is called, than it does now to lake Ontario, for the base of the Algonquin River channel at the mouth seems to lie below the level of lake Ontario (Coleman, 1922, pages 31, 51). This was due to the relatively lower position of the outlet of Admiralty Lake. The Trenton region was not necessarily higher in relation to the sea, for lake Ontario has an elevation of 246 feet. Finally the Kirkfield outlet was abandoned.

The Kirkfield stage, as shown by the strength of the beaches, lasted for a long time, probably about 10,000 years, judging from the contemporaneous recession of Niagara falls. During this stage the Old Narrow gorge, 1⅛ miles long, was formed, and this may have taken around 10,000 years, which can be determined from data furnished by Taylor, if the cutting of the whole Niagara gorge is assumed to have taken 25,000 years as Taylor (1913c, page 24; 1913a, page 53) and Coleman (1922, page 71) consider likely.

After the abandoning of the Kirkfield outlet, the increased water masses rapidly cut down the channel in the clay at Port Huron, so that the Chicago outlet, which was limited by a rock sill, was also left dry. Thus, the drainage of the entire area of the upper lakes went through Niagara river. During this phase Lake Algonquin reached its largest size and extended about 50 miles north of the present shore of lake Huron (*See* Figure 27). This stage continued until Lake Algonquin came to an end and Mattawa valley was uncovered. Lake Algonquin came to a close largely through differential uplift and emptying out before the opening of the North Bay-Mattawa outlet channel. This is evident from the fact that the group of Algonquin beaches not far from the isobase of North Bay extends down to levels not much above the Nipissing beach (Taylor, 1915, Plate 23, page 430). The elevation of the lowest Algonquin beach at North Bay above the outlet there evidently would give the amount of lowering by drainage. The isobase of North Bay for this stage runs about 80 miles north of Sault Ste. Marie (Taylor, 1915, pages 439, 461). If the 700-foot beach at Sault Ste. Marie (Taylor, 1915, Plate 23) is taken as the lowest and last Algonquin beach and its plane and that of the Nipissing beach are extended for 80 miles northward the interval between them will be about 75 feet, which gives roughly the amount of lowering by drainage when the North Bay outlet was opened. If, however, the 675-foot beach at Sault Ste. Marie is an Algonquin

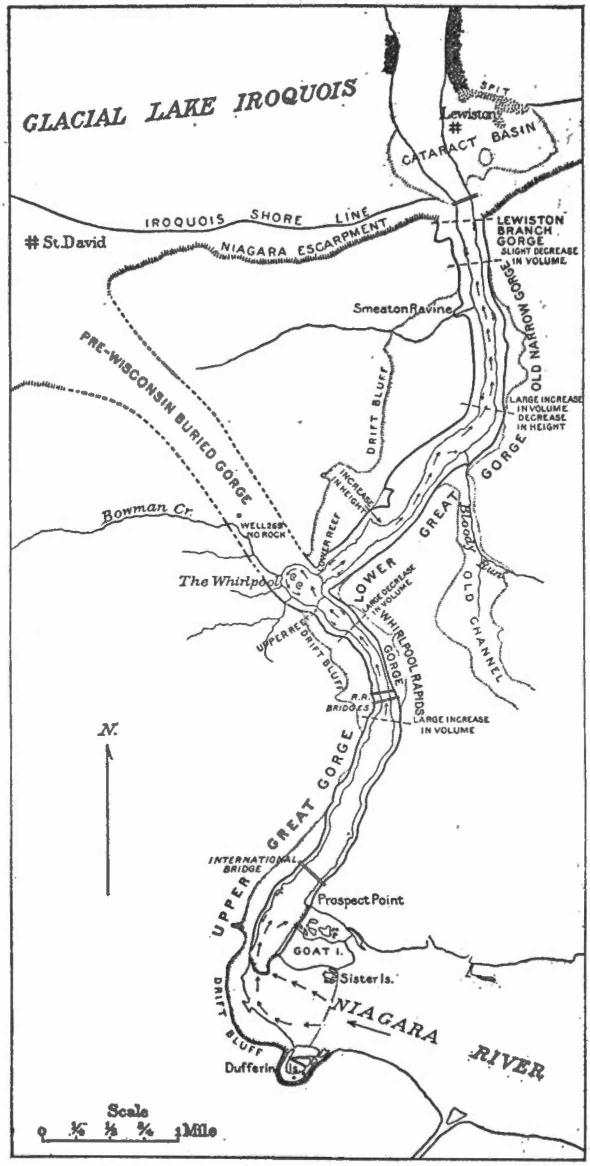


Figure 25. Sketch map of Niagara gorge showing five divisions of the gorge corresponding to changes in volume of the river and height of falls during the five lake stages shown in Figure 24. 1, Lewiston Branch gorge, cut during the first lake stage; small volume; age approximately 2,000 years. 2, Old Narrow gorge, second lake stage; small volume; about 10,000 years. 3, Lower Great gorge, third lake stage; large volume; 2,500 to 3,000 years. 4, Whirlpool Rapids gorge, fourth lake stage; small volume; about 7,000 years. 5, Upper Great gorge, fifth lake stage; full volume of upper four lakes; 3,000 to 3,500 years. (From Taylor, 1913c, p. 21, Fig. 14, and data given on pages 24 and 25.)

beach the lowering was only about 25 feet. The drainage may be recorded at locality 21, Trout Creek, where at varve 314 there is an abrupt change from rather stiff, thinly varved clay to sandy silt with thick varves (Plate IX).

The lake could not have come to an end through a long-continued drainage to the sea, at the same time as the land rose, especially in the north, for this postulates that the lake stood at the same level as the sea, whereas the fact that the Algonquin shore-line everywhere in the southern parts of lakes Michigan and Huron lies at exactly the same level, 607 feet above the present sea (Goldthwait, 1910, page 45), indicates that these regions have not changed their vertical position since the registering of the Algonquin beach, and that the level of Lake Algonquin stood more than 607 feet above the late-glacial sea-level, which was somewhat lower than the present one. If the beach at 1,221 feet above the sea near Trout Creek, 20 miles south of North Bay, is the highest Algonquin shore-line, the uplift may have begun just after the uncovering of this tract, for the highest beach in North Bay region, recorded 5 miles northeast of the city, lies at an elevation of only 1,178 feet (Goldthwait, 1910, page 33). In any case, if uplift took place before Trout Creek became ice-free, it must have been insignificant. The greater part of the rise occurred after the ice front had retired beyond North Bay, but still blocked Mattawa valley. Ice damming of the valley is evident, for the lake when highest stood about 488 feet above its pass floor, which now lies at an elevation of about 690 feet (Goldthwait, 1910, p. 46).

The thus resulting stage in the Upper Great Lakes is called the Nipissing Great Lakes. The last stage of Lake Algonquin, with outlet at Port Huron, may have lasted for 2,500 to 3,000 years according to Taylor's (1913c, pages 24, 25) estimate of the time of the contemporaneous cutting of the Lower Great Gorge of Niagara river.

Thus, according to the estimates of the rate of the recession of Niagara falls, the ice retreat from Trent valley to Mattawa may have taken 12,500 to 13,000 years.

During the Port Huron stage the clay deposits north of Huntsville and north and east of Sudbury region were formed.

NIPISSING GREAT LAKES

The Nipissing Great Lakes occupied the basins of lakes Superior, Michigan, and Huron after warping and drainage past North Bay had brought Lake Algonquin to a close (Taylor, 1915, page 447). The Nipissing Great Lakes had a slightly larger area than the present Great Lakes. They were largely fed by rivers from the ice front which during the initial stage was about 50 miles north of the shore in Huron basin. Later they were largely nourished by melt-water from Lake Ojibway ponded north of the watershed.

The levels of the lakes in the initial stage evidently stood below the former outlet at Port Huron by an amount equal to that of the drainage at North Bay, that is perhaps 75 (or 25) feet, and during the life of the lakes again rose to the Port Huron threshold because of uplift in the northern parts, whereas the Port Huron region kept a stationary level. The final stage of the lakes is marked by a strong beach connecting the North Bay and the Port Huron outlets. This shore-line at North Bay lies at a height of 698 feet and on southern lake Huron at 596 feet or 15 feet above the level of the present lakes Michigan and Huron (Goldthwait, 1910, Figure 4, page 40). Thus North Bay at the end of the Nipissing stage lay 102 feet and at the beginning of the stage perhaps about 175 (or 125) feet lower than it does today.

Nipissing Great Lakes may have lasted for about 7,000 years, which is the length of the time of the cutting of the corresponding Whirlpool Rapids gorge of Niagara river as calculated from data given by Taylor (1913c, pages 24, 25) under the assumption that the whole gorge represents 25,000 years. The present stage of the Great Lakes began 3,000 to 3,500 years ago (Taylor, 1913c, page 24). The Nipissing Great Lakes accordingly came into being about 10,000 or 10,500 years ago.

LAKE IROQUOIS

The first stage in Ontario basin after its waters had become separated from those in Erie basin is called Lake Dawson (Fairchild, 1909, Plate 41). At this time Niagara falls were initiated. The next stage, according to Fairchild, was Lake Iroquois which was established when a low outlet at Rome, through the Mohawk and the Hudson, was uncovered from the ice (Fairchild, 1909, page 53, Plate 42; 1912, Plate 17). Before the establishment of Lake Iroquois there occurred a series of slight oscillations of the ice front, causing rises and falls of the waterlevels (Taylor, 1915, page 444).

Lake Iroquois existed until the ice border had retired beyond Covey Hill west of lake Champlain and just north of the International Boundary. The lowering of the lake began when a pass at an elevation of between 1,000 and 1,020 feet was uncovered southwest of Covey Hill (Goldthwait, 1913a, page 126; Fairchild, 1918, Plate 5). The chief drainage took place when the ice front had receded 1½ miles to 2 miles farther and uncovered the north side of Covey Hill. Here, Lake Iroquois was emptied into Lake Vermont as the late-glacial Lake Champlain is called. The water escaped southward through the Hudson. The old shoreline of Lake Iroquois 1 or 2 miles south of the International Boundary now stands at an elevation of 1,030 feet, and the abandoned shore of Lake Vermont, southeast of Covey Hill, has an altitude of 740 feet (Fairchild, 1918, page 62, Plate 5) or possibly of 720 feet (Woodworth, 1905, page 172, Plate 28). Thus, the lowering of Lake Iroquois amounted to 290 feet. The subsequent water-body west of Covey Hill, at the same level as Lake Vermont, is called Lake Frontenac. A supposed post-Iroquois stage with a water-level between that of Lake Iroquois and Lake Vermont is by Chadwick (1917, page 138; 1923, page 504) distinguished as Lake Emmons. If this stage existed it may have been due to readvance of the ice front at Covey Hill, for a stationary or receding ice border should have been quickly eroded until Lake Iroquois came to the level of Lake Vermont.

The position of the ice edge, when Lake Iroquois came to an end, is difficult to reconstruct, as the Iroquois beach in the northeast does not extend beyond Havelock, situated 30 miles northwest of Belleville and just north of Trent river (Coleman, 1904, page 357), and an ice border reconstructed by help from the striæ and with Covey Hill as starting point will run much farther north. Whether the lake reached farther than Havelock is not known (cf. Taylor, 1924). The ice front on land north of Trent valley was perhaps practically stationary during the greater part of the Iroquois stage, while the ice border where it ended in deep water in St. Lawrence valley was gradually eaten back by calving, though it was probably slowly supplied by ice, flowing, as the striæ show, due south or somewhat west of south. Also in the Adirondack region the ice seems to have been stagnant (Cook, 1924).

Lake Iroquois lasted for about 8,000 years according to Coleman (1914, page 441; 1922, page 70). The estimation is primarily based upon the fact that the Iroquois shore is about as mature as the shore of lake Ontario, which, as calculated from the rate of abrasion of Scarborough bluffs and the formation of Toronto island, has existed for about 8,000 years. The long duration of Lake Iroquois is also shown by extensive wave-cut terraces, by large gravel bars near Hamilton and at Toronto, and by filling of former depressions in terraces. The bay of the Humber in West Toronto was filled with sand, silt, and clay to a depth of 100 feet (Coleman, 1913a, page 32), an enormous quantity of material considering the insignificant character of the creek whose two branches rise respectively only 23 and 35 miles from the mouth.

LAKE FRONTENAC

The post-Iroquois lake covering Ontario basin and the lowlands of, and between, St. Lawrence and Ottawa valleys, is called Lake Frontenac (Taylor, 1915, page 445). It formed the greater part of the water-sheet of which the younger stage of Lake Vermont was a part. It emptied through Lake Vermont and the Hudson. Its shore-line is marked by gravel bars north and northwest of the Adirondacks (Fairchild, 1918, Plate 5). Near Covey Hill the beaches occur at an altitude of 740 feet.

Lake Frontenac and Lake Vermont came to an end through drainage to sea-level by way of the St. Lawrence. The relative position of land and sea in the region at that time is not exactly known, but judging from the conditions at Ottawa the land may not have been fully submerged to the upper marine limit. Thus, lowering may have taken place from the present altitude of 740 feet at Covey Hill to below that of 523 feet, which is the height of the upper marine limit at Covey Hill (Goldthwait, 1913a, page 125). It may consequently have amounted to more than 217 feet.

The drainage could have occurred in two ways, viz., by retreat of the ice front to the vicinity of Quebec city or by overflowing of the water and breaking up of the thinned ice on the submerged St. Lawrence valley. The last-mentioned alternative, as will be shown, seems more probable.

The northward extension of Lake Frontenac, or the position of the ice front, when the lake came to a close, is not exactly known. The lake may have extended at least to Ottawa, as is probable from the occurrence of freshwater clays beneath the marine deposits, although these may have been deposited in the Champlain sea when the water in the upper part of the estuary was still fresh. It probably extended little farther, for the number of the freshwater varves at Ottawa seems to be less than 100 and on Gatineau river, 5½ miles north-northwest of the city, freshwater deposits appear to be lacking. This latter condition, however, could be due to subsequent erosion. Furthermore, the highest beach at Kingsmere, 8 miles northwest of Ottawa, of an altitude of 690 feet, probably was registered by the Champlain Sea (Johnston, 1917, page 17). So was probably, also, the highest shore-line near Rigaud determined at 671 feet (Johnston, 1916a, page 6). The drainage and the invasion of the sea, consequently, may have occurred shortly after the uncovering of Ottawa. The ice front at that time, reconstructed by help of the striæ, runs somewhat south of the junction of Ottawa and St. Lawrence rivers and crosses southeastern Quebec in a west-easterly direction just north of Sherbrooke.

The uncovering of Quebec southeast of the St. Lawrence seems to have taken a long time. Thus, on Clifton river, 16 miles east-southeast of Sherbrooke and on the railway near Angus, 15 miles northeast of Sherbrooke, varved clay is overlain by till (Chalmers, 1898, page 44; Keele, 1915, page 93). Near the mouth of Rivière-du-Loup, 58 miles south-southeast of Quebec, two beds of clay and sand are separated by till, and the deposit is capped by boulder clay (Chalmers, 1898, page 43; Keele, 1915, pages 44, 93). In the region of Beauceville, 45 to 50 miles south-southeast of Quebec, clay at a number of localities is covered by till (MacKay, 1921, page 54).

In the Champlain basin in Vermont thick beds of late-glacial freshwater clays occur at least up to Fairfax, 18 miles north-northeast of Burlington, where a section with about 250 varves was measured by the present writer. West of St. Albans near and on lake Champlain thinly varved clays are to be found. The deposits are small and at most 10 feet thick. Clay was observed only at and slightly above the lake-level. At places sand with marine shells occurs, but on the whole waterlain deposits are scarce in the district. In the region of Swanton, 32 miles north of Burlington, the exposed clay and sand deposits, which are more than 30 feet thick, are marine, shell-bearing. No exposure down to substratum was observed, so that the deposits might be underlain by freshwater sediments. If so these cannot be thick. The material of the marine beds may have been brought out into the basin by Missisquoi river after the departure of the ice from the region. The important thing in this connexion is the insignificant quantity of the glacial freshwater clays, which shows that the ice damming cannot have lasted very long. For if there had been an ice dam in the north, until the ice edge had reached the vicinity of Quebec, the Champlain depression, being a part of the basin of sedimentation for the western half of Vermont, for the greater part of Quebec southeast of the St. Lawrence, and for the region beyond Montreal, ought to contain large quantities of fine-grained sediments.

Consequently, the conditions at Ottawa and the direction of the striæ seem to show that the ice front at the time of the drainage of Lake Frontenac and Lake Vermont stood south of Montreal with an east-west trend. The till interbedded with varved clays in southeastern Quebec, indicates that the uncovering of that region took a long time. The small quantity of freshwater clay in the Champlain basin in northern Vermont indicates early drainage.

All the known facts, therefore, point to drainage across the ice of Lake Frontenac and Lake Vermont and breaking up of the ice in the submerged St. Lawrence valley changing it into a marine gulf. It does not appear likely that the ice front retired to the vicinity of Quebec before the commencement of the marine stage.

The conditions in St. Lawrence valley which made this possible were as follows. During its greatest extension the Labrador ice-sheet crossed the lower part of the St. Lawrence and impinged on the northern shore of Gaspe peninsula, but left no traces higher than 100 feet above the present sea-level (Coleman, 1922a, page 12). Its thickness may have been 2,000 feet (Coleman, 1920, page 325) plus the depth of the St. Lawrence which is 900 to 1,200 feet; that is the thickness may have amounted to about 3,000 feet.

Probably long before the uncovering of southern Quebec the ice in the estuary of the St. Lawrence was broken up by the lifting power and attack of the water, so that the local ice centre in Gaspe became cut off from the main ice-sheet. Since ice but slightly below thawing is only little lighter than diluted water just above freezing, and the ice, furthermore, was more or less loaded with debris, this may have happened shortly before the ice surface had melted down to the sea-level. The sea-level then may have stood about 400 feet higher in relation to land than it does now, for the highest beach south of Ste. Flavie, 185 miles northeast of the city of Quebec, stands at 434 feet (Coleman, 1922a, page 14). When the wastage of the ice had well started it probably proceeded rapidly because of the insignificant nourishment and the probably great tidal range in the enormous estuary.

From Gaspe the late-glacial submergence first increased toward the southwest. From the city of Quebec it again gradually decreased (Goldthwait, 1913a). The highest marine shore-line at Quebec, Montreal, and Covey Hill seems to lie at 632, 617, and 523 feet respectively (*See* page 70). The exact amount of submergence at the time of the drainage of Lake Frontenac and Lake Vermont is not known, but it probably was less than the figures given.

The water of the lakes, if the conditions are correctly interpreted, began to flow over when passes in the ice had been ablated to below the lakes' surface which stood at least 217 feet above the sea at that time and probably a good deal more. The ice barrier that dammed Lake Iroquois and reached an altitude of more than 1,030 feet above the present sea-level had then shrunk at least 290 feet in thickness. The overrunning water masses cut a trench in the ice and hastened its ablation, and before long the ice could be successfully attacked by the sea.

Through these probable events a large ice mass was isolated southeast of the St. Lawrence.

CHAMPLAIN SEA

The marine gulf that extended over Lake Champlain, over the St. Lawrence and Ottawa valleys, and the lowland between, when the lower St. Lawrence became free from ice, is called the Champlain Sea. The discussion of the complicated history of this stage may be begun with descriptions of columnar sections at Ottawa.

The sections at localities 1 to 3, described on pages 95 and 97, are of interest in this connexion.

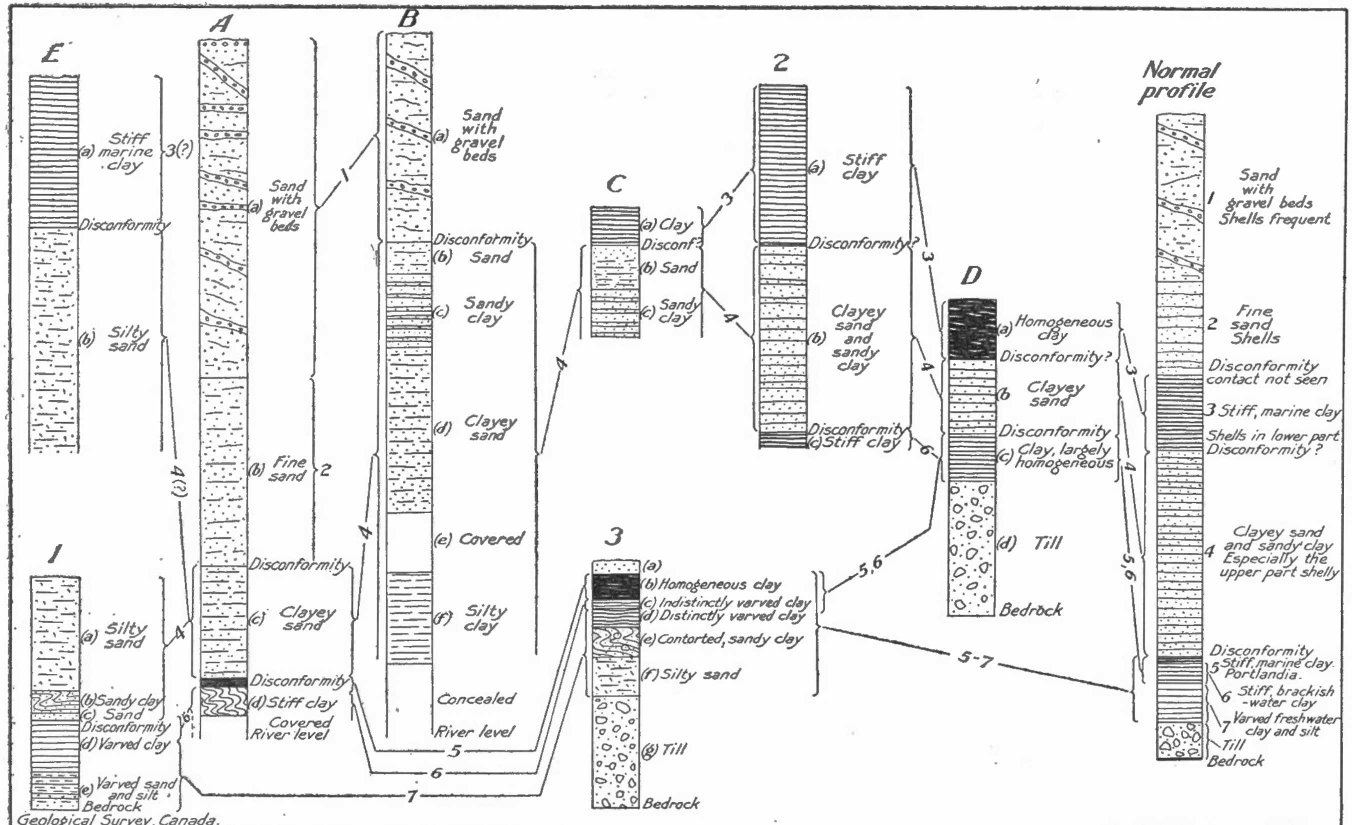


Figure 26. Correlation of sections of varved clay and marine deposits in Ottawa area, Ontario (*cf.* Figure 33, page 96).

Section A. 6½ miles south of Ottawa, ¾ mile south of the railway bridge across the Rideau, sand-pit on the east river bank (Figures 26 and 33).

(a) 45 feet diagonally and crossbedded, coarse sand with layers of fine gravel. Abundant in shells of *Macoma* sp., *Mytilus edulis*, *Saxicava rugosa*, and *Balanus crenatus*.
(b) 25 feet fine, loose, evenly stratified sand, in the lower part somewhat clayey. Poor in shells.
Disconformity. The contact in a fresh cut appears fairly gradual, for the uppermost part of the underlying clayey sand is rather lean, but in an old exposure (in spring) it is marked, for the clayey sand is water soaked, whereas the overlying sand is quite dry. As a matter of fact the upper marine clay is entirely missing and one-half or more of the original clayey sand is also wanting.
(c) 15 feet rather coarse, clayey sand, hard packed. The percentage of clay is fairly high in the lowest foot and in the uppermost 5 feet. The clay largely occurs in irregularly-shaped pieces, some of which may have been rolled masses, but in some places it forms well-defined layers. The bed is stratified in layers 3 to 8 inches thick separated by laminæ of practically clay-free sand. The lamination is particularly distinct on an old exposure. The bed contains scattered shells of *Saxicava rugosa* and *Macoma* sp. The latter was found 10 feet from the bottom of the bed.
Disconformity. The contact is distinct. In a few inches the underlying stiff clay goes over into strongly sandy clay. One foot above the contact the upper bed consists of clayey sand. There evidently must have been a break in deposition.
(d) 5 feet stiff, grey-blue clay, mostly homogeneous because of disturbance and, at the top, probably also by being deposited in salt water. Some feet below the top it presents rather distinct, though disturbed varves. The lower part of the varves is grey-blue, the upper part is brownish, and a thin lamina at the top is reddish. The character of the varves seems to indicate brackish water; but no shells could be found in the varves. In the uppermost homogeneous clay down to 1 foot below the top tiny shells of *Portlandia arctica* occur, so that the uppermost part of the clay is marine.
(e) 3 feet covered to river-level.

Section B. The section described by Johnston (1917, page 26) from the Rideau sand-pits, profile B in Figure 26, was measured 200 yards north of section A, from which it differs somewhat. Thus, it lacks beds *b* and *d* in section A, the latter bed not being exposed. On the other hand the correlatives of bed *c* are much thicker and are by Johnston divided into several beds given the numbers *b* to *f* in his description.

Section C. Brick-yard 3¼ miles south-southwest of Union station, Ottawa, three-quarters mile south of Experimental Farm, and just west of Rideau canal.

(a) 5 feet fat, grey-brown, weathered clay with very indistinct lamination. *Portlandia arctica* up to 1 foot above the base.
Disconformity?
(b) 6 feet sand, rather fine, clayey in the lower part, pure in the upper part. No shells observed.
(c) 6 feet sandy clay. Lamination, particularly in the upper part, distinctly shown by thin layers of sand. Shells of molluscs and *Balani* abundant. *Axinus flexuosus* not previously recorded in the Champlain Sea, and *Nucula* sp., not *N. tenuis*, represented by solitary specimens.
Bottom of clay pit. Depth to substratum unknown.

The coarseness of the material increases continually from the lower part of the section to the top of the sand. Here in 8 inches, there is a perfectly gradual transition from sand of medium coarseness to fine-grained clay.

Section D. 250 yards north of locality 3, on the west side of Greens creek and just north of the highway.

(a) 8 feet homogeneous, yellow-brown clay.
Disconformity?
(b) 10 feet clayey sand. In some layers almost pure sand, in others as much clay as sand.
Disconformity.
(c) 6 feet grey-brown clay, largely homogeneous. At least one horizon, though, with varves one-half cm. thick.
(d) 18 feet till resting on bedrock.

Section E. 5½ miles north-northwest of Ottawa, 100 yards north of the highway bridge across the Gatineau, on the west side of the river.

(a) 20 feet stiff, dark-grey clay, indistinctly laminated to almost homogeneous. Containing *Portlandia arctica*.
Sharp limit.
(b) 30 feet rather coarse sand with layers of clayey silt. No shells observed.
Highwater level of the river. Covered. Probably 5 to 6 feet to bedrock.

The connexion of the beds in the different sections is evident from their similar character and fauna or lack of fauna. It is shown in Figure 26. The sequence of strata is compiled in the normal section, as follows:

(1) Crossbedded, coarse sand with layers of thin gravel. Containing a great abundance of shells of *Macoma* sp., *Mytilus edulis* L., *Saxicava rugosa* L., and *Balanus crenatus* Brug. (Sections A and B).
(2) Fine, loose, evenly stratified sand, in the lower part somewhat clayey. Sparse in shells. (Section A).
Disconformity. Contact not seen.
(3) Stiff marine clay, the upper marine clay, with sparse shells in the lower part, especially of *Portlandia arctica* Gray and *Saxicava rugosa* L. (Sections 2, C, D, and probably E).
Disconformity? (Sections 2, C, D, and E).
(4) Clayey sand and sandy clay, of which particularly the upper part is abundant in shells. Forms found: *Portlandia arctica* Gray, *Saxicava rugosa* L., *Macoma calcaria* Chemn., *M. baltica L.* var. *grönlandica* Beck, *M.* sp., *Astarte compressa* Mont., *Nucula tenuis* Mont., *N.* sp., *Axinus flexuosus* Mont., *Natica affinis* Gmel., *Neptunea despecta* L., *Cylichna alba* Brown, and *Balanus crenatus* Brug. (Sections 1, 2, A, B, C, D, and probably E).
Disconformity. (Sections 1, 2, A, and D).
(5) Stiff marine clay. With certainty observed only at locality A, where it contains *Portlandia arctica*, and at Black rapids on the west side of the river just north of the lock. The homogeneous clay bed at locality 3 may be a correlative and so may be part of bed c in section D.
(6) Stiff, indistinctly to rather distinctly varved clay without shells; probably brackish water clay. (Sections 1, 2, 3, A, and D).
(7) Varved freshwater clay and silt. (Sections 1, 3).
(8) Till and bedrock.

The events recorded by these deposits appear to be as follows:

Until the ice front had retired just beyond Ottawa the uncovered valley was occupied by Lake Frontenac in which the freshwater clay resting directly on rock or till was laid down (cf. page 64). After the drainage of this lake the waters at Ottawa changed to brackish and soon became salt enough to accommodate *Portlandia arctica.* Although the water-level fell through the drainage more than 217 feet the water was still deep in Ottawa valley, for the brackish water clay and the lower marine clay are very fine grained. It is even probable that the highest marine shore-line in the region was recorded during this stage. The upper marine limit at Kingsmere, 8 miles northwest of Ottawa, lies at 690 feet (Johnston, 1917, page 17; cf. De Geer, 1892), at Rigaud, 35 miles west of Montreal, at 671 feet (Johnston, 1916a, page 6), and at Dalesville, 50 miles west-northwest of Montreal, at 735 feet (Wilson, 1924, page 15). These figures at the first glance appear somewhat high in comparison with those at Montreal and in the region east of that city (Goldthwait, 1913a, page 120, Map), but then the amount of submergence increased westward. Besides, the marine limit at Montreal may lie at 617 feet up to which level shell-bearing gravel occurs (Stansfield, 1915, page 29). The registering of the highest shore-line at this time is probable from the fact that marine shells in Ottawa region are not found at greater height than 510 feet (Johnston, 1917, page 27). If the greatest submergence had occurred during the deposition of the upper marine clay, when the molluscs had immigrated, they probably would have occurred in the highest shore deposits, since they largely are shallow water forms. It is noteworthy, however, that shells have rarely been discovered at or near the line of the maximum submergence of the Champlain Sea (Goldthwait, 1913a, page 123).

The deepwater stage may not have lasted long. Rapid upheaval probably had been in progress for a time. The uplift proceeded until the region stood so high that the clay previously deposited at localities 1, 2, 3, A, and D at a present elevation of from 200 to 250 feet was partly eroded away. Then followed slight sinking as indicated by the clayey sand and sandy clay, bed 4. This bed, which on the Rideau reaches a thickness of more than 55 feet (Johnston, 1917, page 26), especially in the upper part contains a great abundance of shells. The species found in the Ottawa region are *Portlandia arctica, Saxicava rugosa, Macoma calcaria, M. baltica, M.* sp., *Astarte compressa, Nucula tenuis, N.* sp., *Axinus flexuosus, Natica affinis, Neptunea despecta, Cylichna alba,* and *Balanus crenatus. Portlandia arctica* and *Saxicava rugosa* are frequent, whereas the others generally are scarce. A small *Macoma* which may be *M. baltica* occurs occasionally, as in section 2, on the transition to the upper marine clay. Solitary shells of a *Macoma* which probably is the same as that occurring in the uppermost sand, bed 1, were found in sections A and B (cf. page 72).

With the exception of *Portlandia arctica,* whose southern limit of present distribution is the strait of Belle Isle, all the species are known to occur in the colder parts of the gulf of St. Lawrence (Whiteaves, 1901). *Portlandia* is a high arctic form, which on the European side does not go south of the White sea. In Sweden it occurs as fossil only in deposits that were formed when the ice edge stood in the vicinity (Antevs, 1917, page 418). *Macoma baltica* is the only species that does not go up into high arctic regions, its range being from the White sea to the Mediterranean. It is not found in the oldest late-glacial shell beds in Sweden. However, it is reported by Dall (1919, page 5) from the north coast of Alaska. All the other forms are distributed from the polar regions down into the boreal or even the warm belts (Nordgaard, 1913; Hägg, 1904; Odhner, 1915). Thus, only *Portlandia arctica* and *Macoma baltica* give any clue as to the temperature of the Champlain Sea at that time. The temperature may have been comparable to that in the White sea at the present time.

Although the fauna usually prefers insignificant water depth its vertical range of distribution is very great, so that it is an unreliable bathymetrical indicator. *Portlandia arctica* is found down to a depth of 722 feet (Hägg, 1904, page 17; Odhner, 1915, page 60) and the others are found to still much greater depths (Nordgaard, 1913).

After the deposition of the clayey sand, bed 4, possibly some uplift again took place, for in some sections sand goes over into stiff clay in less than a foot, which seems to indicate break in deposition. However this may be, transgression soon set in as is evident from the stiff upper marine clay, bed 3. The amount of the transgression is not known, but since *Macoma baltica* and *Mytilus edulis* occur up to elevations of 470 and 450 feet, respectively, at Kingsmere 8 miles northwest of Ottawa (Johnston, 1917, pages 27, 30), and the same large—24 mm. long—convex *Macoma* sp. as that which occurs in bed 1 is abundant at about 450 feet 5 miles northwest of Ottawa, the shore-line must have reached at least the 470-foot level, for the forms mentioned could probably not have immigrated during the previous low stand of the land. *Macoma baltica*, as mentioned, does not go north of the White sea, but occurs on the northern coast of Alaska. *Mytilus edulis* occurs west of the mouth of Mackenzie river, up to Labrador, on the western and southeast coasts of Greenland and up to Nova Zembla. Full-grown specimens occasionally found in great depths in water of a constantly very low temperature most probably have been transported there from their natural habitat by means of seaweed and ice-floes (Jensen, 1912, page 48). That the transgression was considerable is also indicated by the exceedingly fine texture of the clay.

The transgression may largely have been due to sinking of the land in relation to the sea. Some rise of the sea-level because of melting of ice evidently occurred, but this process cannot account for the rapid and great change, especially as ice melting was comparatively small at this time.

The upper marine clay everywhere in the lower part of the valley is exceedingly fine grained, as is also the lower marine clay, bed 5, and the brackish-water clay, bed 6. It shows lamination which, however, is so indistinct, that no sure measurement of a whole section is possible. The summer layers are grey-blue, the winter layers are red-brown, or, when more strongly oxidized, pink. The varves are quite thick, at locality 2 up to 6 inches, and on the Gatineau, 3½ miles north-northwest of Ottawa, 4 to 12 inches. The clay on the Gatineau is more than 35 feet thick. The clay contains numerous shells at the base and for a few feet up, but for the rest it appears generally to be barren of fossils (Johnston, 1917, page 26). At locality 2, *Portlandia arctica* was found 6 feet above the base. The scarcity of shells in the clay may be due to the fact that, while it was deposited in deep water, the molluscs that frequented the marine gulf preferred shallow water.

The events appear to have developed rather rapidly and soon subsidence was followed by upheaval of the land. This uplift proceeded so far that the upper marine clay, which must have been deposited everywhere in the lower parts of the valley, was entirely eroded away at the sand-pits on the Rideau, at profiles A and B. Besides the clay, an unknown part of the underlying clayey sand, at Section A at least 40 feet, was cut away. The erosion surface is very uneven. At section A it lies some 40 feet lower than at profile B situated 200 yards to the north. The erosion may be attributed to the meandering Rideau of that time. Tidal currents could not have caused such an erosion. Furthermore, there may not have been any tidal currents worth mentioning in the Champlain Sea, for this was too large in relation to the narrow sound at Quebec city. Consequently, the base of erosion of the Rideau at the sand-pits was not more than 30 to 40 feet above and perhaps as low as the present base which is determined by a rock sill at Hogsback, 4 miles south of Ottawa. The marine shore-line must at least have retired below the 270-foot level which is the present altitude of the erosion surface in section A.

This low position of the marine shore-line was followed by a rise, as is evident from the sand beds at the Rideau sand-pits, at localities A and B. To an insignificant part the transgression may have been due to rise of the sea-level, for now fairly rapid melting of the ice, whose edge stood north of Mattawa valley, may have occurred, but largely it may have been due to slight sinking of the land. First a fine, in the lower part somewhat clayey, sand indicating some water depth, was deposited. This sand, which is observed only in section A, goes gradually over into coarse, gravelly, crossbedded sand indicating decreasing water depth which may have been due both to rise of land and to deposition. At section B where the fine sand is missing it may have been partly laid down, but subsequently eroded away before the deposition of the top bed.

The uppermost bed contains a great abundance of shells of the species *Mytilus edulis*, *Macoma* sp., *Saxicava rugosa*, and *Balanus crenatus*. All the forms are frequent, but the two first-mentioned occur in the greatest numbers. *Mytilus edulis* is small, only attaining a length of 45 millimetres. *Saxicava* is represented by thin, at most 20 millimetres long, shells. *Balanus* is about 25 millimetres long. *Macoma* sp., which also occurs in bed 4, is rather convex and up to 23 millimetres long. It has thick, strong shells. It has been referred to *Macoma baltica*, but almost surely is another species, for a small, thin-shelled, fairly flat *Macoma baltica* var. *grönlandica* occurs in the Champlain Sea deposits, viz., in the sandy clay, bed 4. The fauna indicates a low-arctic or high-boreal climate.

The uplift recorded by the uppermost gravelly sand proceeded until the shore-line reached the level of about 250 feet, where it remained for a long time, as indicated by a well-developed shore-line (Johnston, 1916a, page 8). The resulting stage is called Lake Ottawa.

The Champlain Sea probably extended somewhat west of Pembroke. Marine fossils have been found as far up Ottawa valley as Fort Coulonge 60 miles northwest of Ottawa (Goldring, 1922, page 165).

The upheaval between the two marine deepwater stages probably was the same as that which closed the Kirkfield outlet, and the uplift that put an end to the marine epoch may have been the same as that which ended Lake Algonquin. If this be so, the part of the marine stage from the time of the erosion of the lower marine clay to the end lasted some 2,500 to 3,000 years, which is the probable length of time of the cutting of the Lower Great Gorge of Niagara river corresponding to the last stage of Lake Algonquin (Taylor, 1913c, page 24).

Thus, when the region south of Ottawa was freed from the ice it became occupied by Lake Frontenac, in which freshwater clays (bed 7) resting on till or bedrock were deposited. After the drainage of the lake, as the ice front had just passed Ottawa, the sea entered through the St. Lawrence (beds 6 and 5). The land was deeply submerged; the marine limit at 690 feet probably was registered at this time. Uplift probably was in progress, and before long the shore had retired at least to the 200 or 250-foot level. Slight sinking followed, and shell-bearing clayey sand and sandy clay were laid down (bed 4). After this marine shallow water stage probably some uplift took place. Soon, however, land sinking set in—the second marine deep-water stage (bed 3). The shore-line reached at least the 470-foot level. Then again uplift followed, during which the shore retired at least below the 270-foot level. Small transgression then set in (bed 2) and was followed by uplift (bed 1). The rise proceeded until the shore-line reached the level of about 250 feet, where it remained for a long time, and Lake Ottawa came into existence.

LAKE OTTAWA

The uplift of the Ottawa region during which the top sand containing a low-arctic to high-boreal fauna was deposited faded out, and equilibrium was attained long before the present stand was reached as is shown by a very marked beach that occurs over the greater part of the valley. This beach, as Johnston (1916a, page 7) has found, marks the outline of a lake. The lake may be called Lake Ottawa. The following account is exclusively based upon studies by Johnston. Material obtained after the publication of the paper quoted has been kindly put at the writer's disposal.

Lake Ottawa may have been held by a dam 6 miles east of Hawkesbury, or 60 miles east of Ottawa, where there was an outlet channel to the sea 1 or 2 miles long. It probably extended westward to the region of Pembroke, some 80 miles northwest of Ottawa. It was mostly from a few to several miles wide. It contained a number of small and large islands. In places there were narrow passages, and here the water perhaps most nearly had the character of a large river. This seems probable because the water masses going through the valley must have been enormous, for it formed the outlet channel not only for the basins of the contemporaneous Nipissing Great Lakes (cf. page 62) and of the region eastward to the divide running due north from the mouth of Ottawa river, but also at least for a long time, for Lake Ojibway lying north of the height of land. The run-off from this area must have been much greater than it is today because of the rather rapid ice melting, even if the rainfall in the area at that time may have been fairly small.

The shore-line of Lake Ottawa now has an altitude of 264 feet near Breckenridge, 12 miles west-northwest of Ottawa, of 252 feet at Aylmer, 5½ miles southeast of the former point, of 242 feet southwest of lake Deschênes, of 234 feet 4½ miles southeast of Ottawa, of 240 feet at Eastview, 2 miles east of Ottawa, of 252 feet 5 miles north of Ottawa, and of 225 to 230 feet southwest of Hawkesbury and 40 to 50 miles east of Ottawa. The beach consequently is tilted at a rate of about 3 feet per mile in a north and south direction, but only slightly in a west and east direction.

An exceptionally strong marine beach that probably is contemporaneous with Lake Ottawa has, near Rigaud, been determined at 207 and 212 feet or about 15 feet below the Ottawa beach (Johnston, 1916a, page 6). Since the sea-level must have risen during the life-time of Lake Ottawa because of return of water stored in the continental glaciers, and since perhaps also the Ottawa valley was uplifted after the establishment of the lake, the conditions are quite complicated.

Lake Ottawa may have been contemporaneous with the Nipissing Great Lakes. The much greater final uplift of Ottawa valley than of the North Bay region, where it is 102 feet, may be due to local conditions. The much greater trend to the south of east of the isobases of the Nipissing Great Lakes than of those of Lake Algonquin (Taylor, 1915, Figure 9, page 461; Figure 8, page 439) shows that the uplift of the region to the east lagged behind. The northward tilt of the Ottawa beach of 3 feet a mile which, considering the small amount of later uplift, is very great, may be regarded as a local anomaly. Both the amount of uplift and of tilt makes Ottawa valley stand in contrast with the Nipissing Great Lakes basins whose water-plane is remarkably even over the whole area over which it has been uplifted and tilted. The conditions, though, are not unparalleled. Thus, the beach of the post-glacial Great Vänern in Sweden shows up to 30 feet difference in elevation on different sides of fault-lines. In other districts it is almost horizontal for long distances, but in places suddenly shows a fairly great tilt (Sandegren, 1916; 1922, page 45). The anomalies are due to differential movements of the long blocks, separated by faults, that form the earth's crust in the region. It is particularly noteworthy that the central parts of the lake basin have lagged behind in the general upheaval. The causes of the anomalies at Ottawa are not known.

LAKES BARLOW, OJIBWAY, BARLOW-OJIBWAY

As touched upon on pages 60 and 62 the breaking down of the ice barrier in the Mattawa valley and the resulting opening of a lower outlet here caused Lake Algonquin to adjust itself to the passpoint at North Bay. The waterlevel probably fell about 75 (or 25) feet, and the North Bay region then probably stood about 175 (or 125) feet below its present stand. Mattawa at that time stood more than 175 (or 125) feet lower than at present, for it lies northeast of North Bay and the uplift increases in that direction. The Champlain Sea was at an end. The lower part of Ottawa valley held Lake Ottawa standing slightly above sea-level. The upper part of the valley was traversed by a large river carrying the whole discharge of the Nipissing Great Lakes and of Timiskaming region.

Thus, there was free run-off in the upper Ottawa River valley until the ice had retired to Timiskaming where a lake became ponded between the ice barrier on the north and the broad drift barrier that now dams lake Timiskaming (*See* Wilson, 1918, page 40). The lake was much larger than the present lake because the region was depressed to a greater extent in the north than in the south. This late-glacial lake has by Wilson (1918, page 143) been called Lake Barlow. As the ice edge withdrew Lake Barlow extended across the height of land in western Quebec. When the ice front was somewhere between the divide and Ramore station Lake Barlow merged with Lake Ojibway. The name Lake Ojibway was proposed by Coleman (1909; 1922, page 40) to signify the extensive late-glacial water-body that was dammed between the height of land and the receding ice front. Lake Ojibway originated north of the divide directly north of lake Huron, and as the ice front retired expanded over large areas. The lake during its early stages may have emptied through different passes to the Nipissing Great Lakes (Coleman, 1909, page 292). The pass at Missinaibi, 45 miles northeast of Michipicoten bay, now has an altitude of 1,090 feet (Coleman, 1922, page 40). The merging of this lake with Lake Barlow has not left any record so far known in the sediments. The resulting huge lake, which may be called Lake Barlow-Ojibway, seems to have lasted for many hundred years. The relationships of the lakes to one another is little known; no certain distinction can be made between their limits and deposits.

Over the drift barrier at Timiskaming and through the gorge of Ottawa river the great water masses from Lakes Barlow and Barlow-Ojibway sought their way to the sea. The threshold now stands slightly less than 575 feet above the sea, the figure being the elevation of the lowwater level of lake Timiskaming. At the time of withdrawal of the ice, the outlet stood lower, probably by 225 to 275 feet. The barrier may have been higher than at present, for it has probably been cut down somewhat. The highwater level of Lake Barlow, because of the long, fiord-like character of the southern part, may have risen considerably, as the water-masses grew with the increase of drainage area and rate of the ice melting, especially after the merging of Lakes Barlow and Ojibway. Thus the levels of Lakes Barlow and Barlow-Ojibway just above the outlet probably exceeded the present highwater level of 592 feet, and may have stood above 600 feet. If the outlet has been cut down the level may have been still higher. This seems probable from the occurrence of sand and clay terraces up to 811 feet near the mouth of Montreal river 30 miles north of Timiskaming (Coleman, 1922, page 43). For if the level of Lake Barlow at Timiskaming stood at 600 feet the warping would be 7 feet to a mile, which is a rather high rate.

No beaches of Lakes Barlow and Barlow-Ojibway are determined on the watershed, so that the amount of tilt can not be definitely determined, but the fact, as pointed out by Cooke (1923, page 66), that clays on Opasatika lake about 100 miles north of the south end of lake Timiskaming occur up to elevations of 1,000 to 1,050 feet shows that the tilt in this distance may be 400 to 500 feet, that is, 4 to 5 feet per mile. Thus also, if Timiskaming is assumed to have stood 250 feet lower than it does today the height of land when uncovered may have stood 650 to 750 feet lower than at present.

A lowering of the water-level probably took place at year 1185, for that year begins long series of varves with exceedingly stiff clay, indicating trapping of the silt north of the divide.

About year 1650 another important event appears to have occurred as indicated by the change at locality 81, Englehart, from greasy clay to lean clay and silt with thick varves. The ice front at this time stood north of Iroquois Falls, so that the silt could not possibly have come directly from the ice. It must have been derived from the neighbouring hills. Since no silt whatever was brought out when the stiff clay was deposited at a very slightly lower level the deposition of silt may not be due to shallowing through deposition. It might be due to a river which through changed course came to fall out in the vicinity, but this appears little likely. Probably a lowering of the lake occurred. This subsidence perhaps was the same as that found by Cooke in Opasatika area, on the height of land in westernmost Quebec.

In Opasatika area, according to Cooke (1922, pages 25, 67), clays lie at two levels, at 1,000 to 1,050 feet, and at 900 feet. The high-level clays occur essentially north of the watershed, but in small areas also south of it. The low-level clays were only observed south of the divide, where they form the bulk of the sediments. Cooke explains the conditions by assuming that a lake which extended across the height of land underwent a rapid subsidence by drainage of the waters from the area north of the divide, and that the new low-level lake south of the watershed thereafter was lowered very gradually, so that the waves eroded away all clay as the lake fell. By the drainage the water-level may have been lowered from about 1,075 or 1,100 feet to about 975 feet from where a gradual sinking took place to near the 900-foot level. The drainage may have taken place eastward or southwestward when a lower pass was uncovered, or, if the drainage were other than that at locality 81, or if the interpretation of profile 81 is incorrect, northward to James bay. In the latter case it probably was the final draining of Lake Barlow-Ojibway. In any case, by the drainage recorded at Opasatika lake, the southern part of Lake Barlow-Ojibway seems to have fallen to the level of the lowest pass in Opasatika district, which now stands at 935 to 940 feet (Coleman, 1922, page 43). The resulting lake south of the height of land was entirely cut off from connexion with the ice. It may have stood below the sill of the outlet channel at Timiskaming and may have entirely discharged northward, across the divide. The slow fall of the lake subsequent upon the drainage may have been due to cutting down of the outlet in drift. The main subsequent event may have been that differential uplift again shifted the outlet to the south and emptied the lake until the present state of affairs was attained. Terraces at different levels on lake Timiskaming indicate halts in the general fall of the water-level (Coleman, 1922, page 43; Hume, 1920, page 298).

The extent of the large late-glacial lakes in northern Ontario and Quebec is yet only partly known (cf. Wilson, 1918, Figure 6, page 140). Eastward, they may have extended to Bell river. Westward, Lakes Ojibway and Barlow-Ojibway extended all the way to the region north of lake Nipigon (Coleman, 1922, Figure 7, page 41).

The northern limit of extensive deposits of varved clay in the eastern part is north of the National Transcontinental railway. At Cochrane the superficial deposits largely consist of sand and silt mixed with gravel, but also of clay. On the western kettle lake, in the town, and at other places in the vicinity, there occur in the fine material very large boulders dumped from drifting icebergs or from the ice front. The material is largely glaci-fluvial and may have been deposited during a halt of the ice edge or during a very slow retreat. No extensive deposits of varved clay

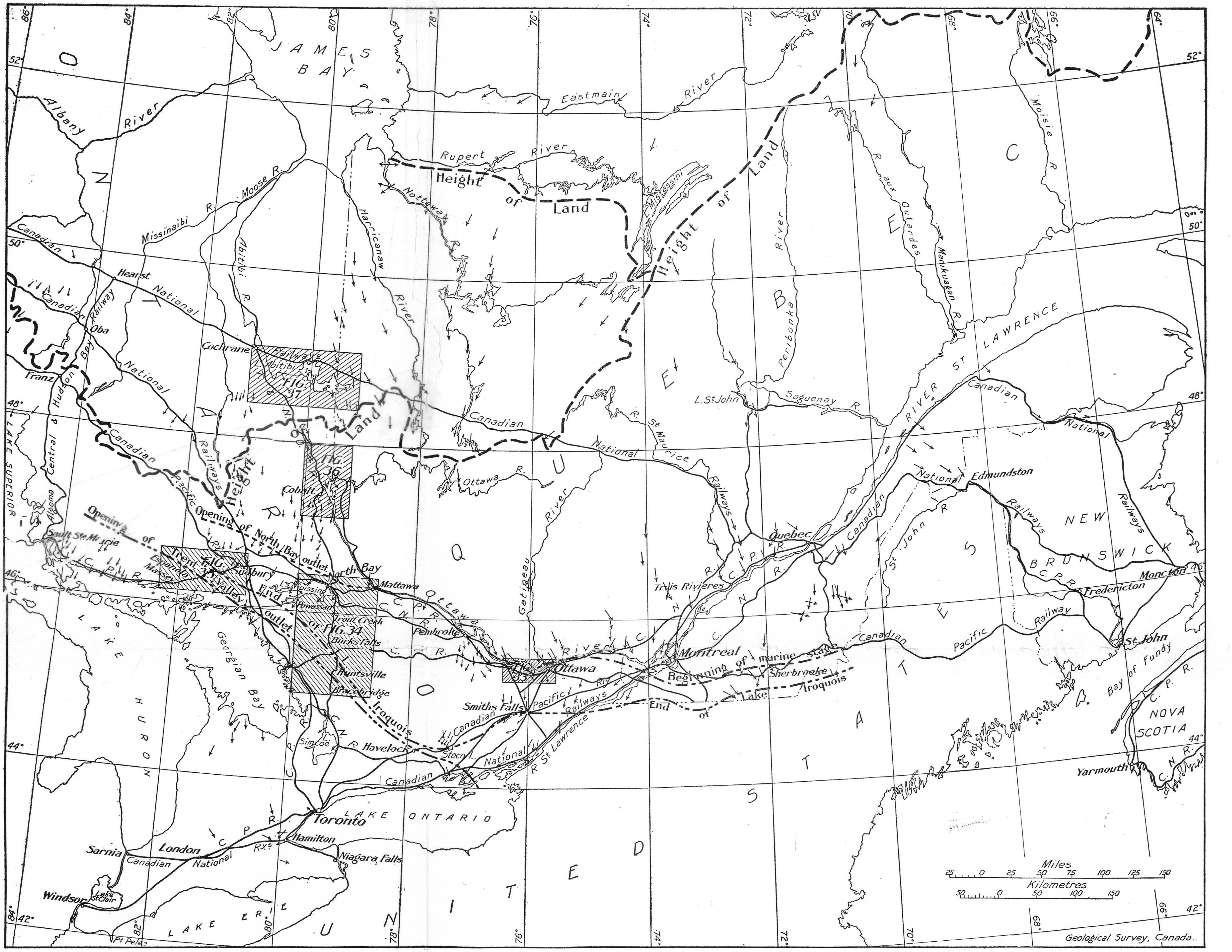


Figure 27. Glacial striæ in eastern Canada and probable positions of the ice edge at various stages. Index to location of areas shown by Figures 33-37. (*See footnote on page 79.*)

occur north of Cochrane (cf. page 58). The railway cuts between Cochrane and Mattagami river are in boulder clay, so far as could be determined from the train. On Mattagami river at the railway crossing 30 miles west of Cochrane, the surface deposits are largely boulder clay with few and small pebbles. The only observed varved clay in the region was 3 to 4 varves on the top of a gravel deposit at Jacksonboro. Near Groundhog river, 50 miles west of Cochrane, 26 feet of varved clay was exposed in a railway cut (M. B. Baker, 1911, page 231). On Missinaibi river at the railway crossing 110 miles west-northwest of Cochrane the soil largely is boulder clay. Occasionally depressions contain deposits of varved clay up to 15 feet thick (Keele, 1922, page 172). Farther down the river no varved clay worth mentioning is to be found. The region extending from the National Transcontinental railway to James bay is an even to somewhat rolling plain with gentle northward slope. The bulk of the surface deposits consist of thick, almost pebble-free boulder clay derived from the Palæozoic and Mesozoic rocks bordering the southern part of Hudson Bay basin (Keele, 1920; Tyrrell, 1914, page 530).

Thus, Lake Barlow-Ojibway or a subsequent Lake Ojibway came to an end when the ice front in the eastern part stood north of, and in the western part somewhat south of, the National Transcontinental railways. The drainage was complete, so that later only small lakes existed in depressions of the largely level ground north of the divide. The lake evidently must have been emptied to James bay. Most likely the drainage was northwest, along the present Albany or Moose rivers. Since striæ northwest of Hearst Junction indicate that the ice front stood about east and west, the drainage probably took place across the ice, when this had been thinned so as to offer lower passes than those on the watershed. The assumed events explain the complete drainage, for trenching of the ice would have proceeded until the river was held by a barrier of rock, or of coarse drift, or until it had reached the level of the sea. The average thickness of the ice south of James bay at the time of the drainage may have been several hundred feet.

When the region of James bay became free from ice the tract on the southwest side was submerged beneath the sea to a maximum depth of 400 to 500 feet, so that it was flooded for more than 100 miles from the present southwest shore of the bay (*See* Coleman, 1909, page 293; Miller, 1912, page 139).

CORRELATION OF THE EVENTS

As suggested by this writer in a previous paper (1922, page 99) Lake Warren in Erie basin and the Niagara Falls moraine, which probably marks the ice barrier of the lake, may correspond to the ice readvance at lake Winnepesaukee and at Claremont, in New Hampshire. The ice retreat in New York through which the outlet of the Great Lakes for the second time was shifted to Mohawk valley, thus establishing Lake Lundy (Dana), may be correlated with the subsequent rapid recession in New England. The uncovering of lower outlet channels, owing to which the waters of the Ontario basin became separated from those in the Erie basin initiating Niagara falls, also may correspond to the rapid retreat south of St. Johnsbury, Vermont. The long Iroquois stage may be represented by the repeated halts and readvances of the ice border in northern Vermont.

Lake Iroquois probably came into existence at about the same time as Lake Algonquin, and largely corresponded to the Kirkfield stage of Lake Algonquin. Lake Frontenac and the first marine deep water stage also may have corresponded to the Kirkfield stage which may have lasted for approximately 10,000 years, for not until after deposition of the lower marine clay had taken place did the region lie high enough in relation to the water-level in the Ontario basin and to the sea to permit of erosion of the Algonquin River channel below the level of lake Ontario. Thus the uplift recorded by the sands (bed 4), etc., between the marine clay beds at Ottawa, may be the same as that which put an end to the Kirkfield stage. The southward emptying out of Lake Algonquin and recession of the northern shore of the lake that must have been a consequence of the differential uplift may not be recorded by preserved beaches, but should be registered in the sediments at fairly high levels. The amount of the change was probably less than at Ottawa. The later subsidence during which the upper marine clay at Ottawa was laid down is not observed in the Algonquin basin as a direct landsinking, though one such must have taken place, for the maximum submergence at Ottawa most probably corresponded to the low position of the northern part of the Algonquin basin which permitted its waters to extend about 50 miles north of the present shore of lake Huron. The uplift which put an end to the marine stage no doubt was the same as that which emptied out Lake Algonquin about 10,000 to 10,500 years ago, according to the Niagara chronology. The resulting stages, Lake Ottawa and the Nipissing Great Lakes, also most certainly were correlatives. They existed until upheaval about 3,000 to 3,500 years ago emptied Lake Ottawa and closed the North Bay outlet. Lake Ottawa and the Nipissing Great Lakes during parts of their life were correlatives of Lakes Barlow, Ojibway, and Barlow-Ojibway.

On the lower St. Lawrence there is a great terrace and sea-cliff called the Micmac beach (Goldthwait, 1911, 1913). It continues for 300 to 400 miles northeastward from the city of Quebec, is approximately horizontal, and lies at an average elevation of 20 feet. It possibly marks a stage of stability of the land. Since the sea-level slowly rose as the ice-sheets disappeared, it may not be necessary to assume land sinking to explain its development. The agreement in strength between this shore-line and the Nipissing beach was pointed out by Taylor (1915, page 463). The contemporaneous development of these shore-lines and of the Ottawa beach is possible. However, Daly (1920) thinks that the beach was formed after the region had attained its present vertical position and during a 20 feet higher stand of sea-level during the post-glacial temperature maximum.

Geochronological studies in Sweden and in Finland have shown that the ice edge can be reconstructed by drawing lines at right angles to the striæ, because the majority of the preserved scratches were formed near the ice edge shortly before the ultimate uncovering. This is particularly well illustrated by Sauramo's (1923, page 150; Plate 8) detailed studies in Finland. Some uncertainty, though, arises when part of the ice edge ends in deep water-filled depressions which both promote supply and wastage of ice. Figure 27[1] shows the approximate positions of the ice edge during some important stages, viz., at opening of the Trent Valley outlet, at the time of drainage of Lake Iroquois, at the time of drainage of Lake Frontenac or the beginning of the marine stage, and at the opening of the North Bay outlet.

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[1]Notes on the map, Figure 27, page 74

The following is a list of the sources from which the data on the striæ have been obtained. Publications issued by the Geological Survey, Canada, are mostly only indicated by the number of publication, as they can be easily found by the help of Ferrier's "Annotated Catalogue of Geological Survey Publications." Maps issued by the Ontario Department of Mines have the same numbers as the volumes.

Ontario

Geological Survey, Canada: Pub. 343; 514, pp. 890-91; 523, p. 12; 536, p. 27; 570, 606, 660, 789, 820, 852, 903, 1066, 1662, 1697, 1739, 1964.

Ontario Department of Mines: 1896, p. 181; 1899, pp. 157, 175; 1900, p. 178; vol. X, 1901, pp. 211, 220; vol. XI, p. 185; vol. XII, pp. 183, 186; vol. XIV, pt. 1, pp. 185, 288; pt. 3, p. 101; vol. XV, pt. 1, p. 135; vol. XIX, pt. 2, pp. 111, 159, 168; vol. XXI, pt. 1, p. 208; Maps 22e, 24a, 24e, 24d, 25d, 27a, 28b, 29a, 29g, 30d, 31a.

Quebec

Geological Survey, Canada: Pub. 240, pp. 46-48; 268, p. 99; 350, 375, 514, pp. 891-92; 571, 585, 599, 660, 665, 667, 702, 750, 903, 918, 1063, 1066, 1112, 1154, 1184, 1180, 1202, 1300, 1630, 1634, 1662, 1680, 1691, 1739 (Mem. 136), 1795 (Mem. 136), 1835 (Mem. 127).

CHAPTER IX

RATE OF THE ICE RECESSION

On account of lack of clay and of exposures, and because of the very long time involved, varve records have been obtained for probably only a fraction of the time occupied by the ice retreat in the regions east and north of Georgian bay in spite of the long series measured (Plates IV, V, and VI). On the bases of the clay studies not even an estimate of the length of the time can be made. The varve series obtained at the different localities do not seem to overlap each other, or if they actually overlap cannot be connected because they are local and uncharacteristic. It is evident, however, from the great number of varves in the sections, for example, 800 at Bracebridge alone, from the thinness of the varves, from indications of halts and readvances of the ice front, etc., that the uncovering took a very long time. At Beaverton on the east side of lake Simcoe a readvance of the ice border is indicated by a stony clay with scattered pebbles, the clay being essentially an overridden varve clay with varves 2 to 8 mm. in thickness. Crumpled clay at locality 26, North Bay, also indicates readvance of the ice front.

However, Spencer's (1907) and Taylor's (1913a, 1913c) studies of the retreat of Niagara falls furnish a rough estimate of the length of the time of the ice recession from Trent valley to Mattawa valley. This ice retreat corresponds in time to the Kirkfield and Port Huron-Chicago stages of Lake Algonquin, which together may represent some 12,500 to 13,000 years (cf. page 62). Coleman's (1914, 1922; cf. this report, page 64) estimate of the life of Lake Iroquois at about 8,000 years is a corroborating evidence of the great length of the time involved. During the greater part of the Iroquois stage the ice front was stationary just north of Trent valley (cf. page 63). During the last part of the Algonquin stage the ice edge was stationary in Mattawa valley for a long time, probably more than 1,000 years, as is evident from the great uplift of land which occurred between the registering of the Algonquin shore-line at 1,178 feet 5 miles northeast of North Bay and the opening of the North Bay-Mattawa valley.

Between North Bay and Cobalt and between Mattawa and the mouth of Montreal river no clay was observed, and no conditions that can throw light upon the rate of ice retreat are known.

In Timiskaming region the ice recession was moderately rapid, averaging 454 feet annually. The actual or approximate position of the ice edge every 100 years is indicated in Figure 36. The ice front retired practically at right angles to the basin of lake Timiskaming and changed from east and west to somewhat north of east and south of west at the bend at Miron. This may have been largely due to greater calving in the lower region east of the present lake than in the hilly supra-aquatic upland on the west side. The rate of retreat where it has been determined between adjacent localities is fairly or almost even. The gradually decreasing or almost even thicknesses of the varves also indicate even retreat where this has not been directly determined.

Rate of Retreat of the Ice Edge

Localities	No. of bottom varves	Distance	Time of retreat in years	Rate of retreat a year
From—		Miles		Feet
63 to 39	1, 48	3¼	47	364
63 to 53	1, 142	11	141	410
63 to 82	1, 594	51	593	454
63 to 95	1, 1212 (?)	109	1,211	475
39 to 40	48, 57	1	9	587
40 to 42	57, 86	3	29	545
42 to 53	86, 142	4	56	377
42 to 62	96, 185	12	99	640
53 to 62	142, 185	8	43	681
53 to 82	142, 594	40	452	467
82 to 95	594, 1212 (?)	58	618	496
95 to 98	1212 (?), about 1400	15	About 188	About 421

The sediments below varve 375 are very lean, consisting almost entirely of silt. The material above gradually becomes much more fine-grained. At about varve 440 the clay is fat. At the change, the ice edge had just reached the north end of lake Timiskaming where there are large areas of Silurian rocks. The coarser sediments were derived largely from the hard igneous and Precambrian rocks, the fine-grained material mainly from Silurian limestone and shales.

Above varve 498 there is a disturbed zone everywhere in the numerous and long railway cuts at and north of Béarn, at all localities examined except 62b. The disturbances of this zone as well as of other zones is very thorough. The clay is faulted, tilted, squeezed, and contorted. In some sections, many feet high, the clay is disturbed all through. In other cuts part of the clay is thoroughly contorted, whereas blocks of clay which must have been frozen at the time of the disturbance show entirely undisturbed varves. In some cases, as at locality 58, blocks of clay from the lower strata are placed on the top of the deposits. In spite of the thorough disturbance of the clay, readvance of the ice edge evidently did not take place. The ice edge at the time of the deposition of the contorted zone above varve 498 stood far north of lake Timiskaming, and when the disturbances happened stood still farther north or had entirely disappeared. The disturbances, therefore, are probably due either to landslides in the deeply trenched clay, to grounding of icebergs, or to melting of buried ice (cf. Antevs, 1922, page 72).

By means of clay measurements carried out in 1924 under the auspices of Harvard University the varve series from Timiskaming region has been connected with that from the region north of the divide, and the average rate of the retreat between locality 82, Englehart, and 95, Matheson, has been determined at 496 feet a year.

The rate of the ice recession between the height of land and Matheson, was rapid. North of Matheson it slowed down somewhat, averaging from this town to locality 98 about 421 feet annually. In the region of Porquis Junction it may have gradually decreased so as to become quite insignificant, judging from the tapering varve curves and the very thin varves ending at varve 1527. The time of little melting seems to have lasted for about 75 years.

Beginning with varve 1528 an extraordinarily sudden and great increase in deposition occurred. The increase cannot be due to flocculation, for that played an equally great rôle before and after. Neither can it be due to decreased water depth and subsequent shore erosion as is evident from the uniform character of the clay. Therefore, the increased deposition indicates an increase in the amount of mud brought into the lake. Since the increase is equally as great at Nighthawk lake as at La Sarre, situated 84 miles apart and in entirely independently fed parts of the ancient lake, it cannot be due to sudden enlargement of the drainage area, for this would mean greater increase in deposition in the adjacent parts than in the distant parts. It was probably caused, therefore, by increased ice melting. Considerable melting lasted for about 200 years as is evident from the sedimentation at localities 91 and 92. Then decrease set in for about 100 years. This was followed by an increase lasting 30 years, and this in its turn by decrease during at least 60 years. Thus, on the whole the sedimentation seems to indicate considerable melting. The actual rate of the ice retreat, however, is not known. Neither is it known how far the ice edge retired before the halt and readvance that occurred near Iroquois Falls, but the ice border stood in the vicinity of Iroquois Falls, when the rapid retreat began following varve year 1528. The region of Monteith was probably covered by an isolated ice block after the departure of the continuous ice border, for the bottom varves at localities 99 and 100 are 1494 and 1493, respectively. The varves are also very thin.

The upper part of the clay at locality 103, Iroquois Falls, above varve 1904, is contorted. With the exception of a few clay blocks the clay in the long railway cutting from Iroquois Falls to ¾ mile north of the Abitibi, and in other exposures, is so contorted as to be almost homogeneous. These facts show that the ice edge halted, readvanced, and overrode locality 103. The ice readvanced to more than 1 mile south of Iroquois Falls, for in the cuttings on the south side of the creek south of the town the clay is completely contorted. But it did not go 2 miles south of the town, for there the clay is undisturbed. The time of the return of the ice edge to Iroquois Falls is not known, but it was after year 1900, that is more than 350 years—probably much more—after it left for the first time.

The enormous gravel and sand deposits at Nellie Lake, 4 miles north of Porquis Junction, may have been largely formed during the time of readvance of the ice (*See* Knight, etc., 1919, page 39; Map). The highest ridges rise 250 feet above the surrounding plain. The gravel and sandplain contains a number of large and small kettle lakes.

The rate of the ultimate ice retreat north of Iroquois Falls is not known, since no clay profiles worth mentioning were obtained (cf. pages 58, 78).

CHAPTER X

RELATIONSHIP BETWEEN THE RETREAT OF THE LAST ICE-SHEETS IN NORTH AMERICA AND IN EUROPE

VALIDITY OF VARVE CONNEXIONS

Correlations between varve curves are based upon agreement among them. The relative thicknesses of the varves, as has been shown (page 46), are largely dependent upon the amount of ice melting. Melting in its turn is primarily determined by the summer temperature. Thus, agreement of the summer temperatures of the regions is the first condition for similar clay sedimentation and for connexions of varve curves.

The transportation and deposition of glacial mud was much influenced by the topographic conditions of the lake basins. Islands and other obstructions could cause the main current to change its course from year to year. Sedimentation, therefore, does not always correctly record the ice melting. This is especially true in the case of very large lakes such as the late-glacial Baltic, for such lakes had less stable surface currents than had small and medium-sized lakes, particularly if these had outlets at the distal ends. These conditions make several measurements from different parts of the same lake and from different lakes necessary for elimination of local features and obtaining of characteristic curves.

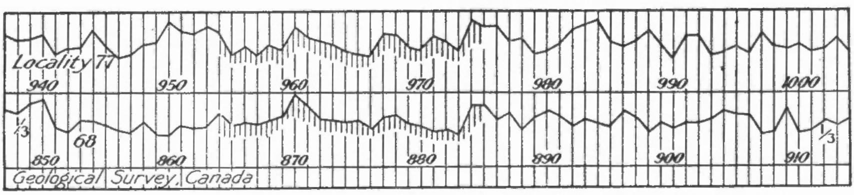

Figure 28. Erroneous connexion of varve curves to illustrate recurrence.

Transportation and sedimentation were also greatly influenced by the physical and chemical properties of the glacial mud, of the river water, and of the water in which deposition took place (cf. pages 23-33). A large part of the mud could be transported long distances or most of it could go down near the ice edge. Other factors, also, could so change conditions that sedimentation curves from a climatically uniform area can differ to such an extent as to defy connexion.

The properties of curves are of importance in this connexion. In the course of a curve there are three alternatives between the breaking points, viz., the horizontal, the upward, and the downward course. The upward and downward courses theoretically can have all variations, but in the case of most curves, and in the case of the varve curves, even these vary within fairly narrow limits. In very flat curves the same general course regardless of the degree of the fluctuations is generally considered enough for agreement. It is evident that two such curves, even if they have nothing to do with each other, are bound to show 33 per cent correspondence. In ordinary varve curves, of course, much less than 33 per cent of the agreement is due to the fact mentioned.

Another important thing concerning the varve curves as well as many other curves is the periodicity through which series recur almost or perfectly alike. Figure 28 shows an erroneous connexion. The numbers of the varves indicate the actual relationship. The parts that agree more or less are marked by shading. The correspondence extending over a time of twenty-two years would justify connexion, if only these parts were considered and not also the continuations that disprove it. Thus, agreement does not necessarily mean contemporaneity. Furthermore, fairly long series of varves are requisite for sure connexions.

In the study of these varves the personal factor must be taken into account, for a connexion is, after all, based upon personal opinion both as to the proportion of agreement between the curves and the degree of correspondence that is deemed necessary. There is often a tendency to overestimate the degree of correspondence.

From what has been said it follows that, in widely separated regions, a mere agreement between the curves is not sufficient to prove the validity of a connexion. It is necessary to know that the regions were uncovered at about the same time, and that the conditions of temperature, transport, and deposition, etc., were the same or similar. Thus, the varve curves

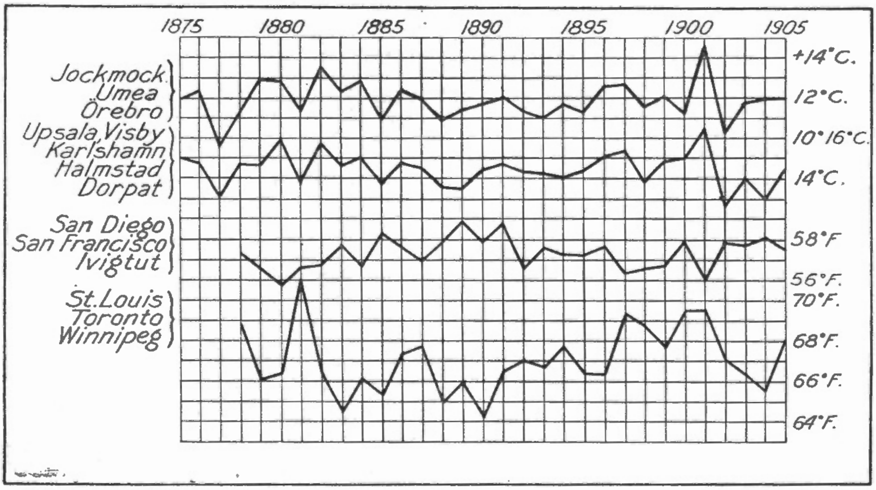

Figure 29. Curves showing the temperature during July-September in Sweden and in the northern and southern parts of North America. (From Hildebrandsson.)

really furnish a means of making the rough correlations obtained through moraines, rate of retreat, long-range periodic phenomena, etc., more accurate.

It is evident, therefore, in attempting to connect the late-glacial history of distant regions by means of the varved clays, that it is necessary to proceed from large features to smaller features and so eventually to details.

COMPARISON OF CLIMATIC CONDITIONS IN NORTH AMERICA AND IN EUROPE

A comparison of the best estimates of the time that has elapsed since the last uncovering from the ice of Niagara Falls region—20,000 to possibly

35,000 years (Taylor, 1913c, page 24); of the Toronto region—25,000 years (Coleman, 1922, page 71; 1914, page 442); of southern Sweden—13,500 to 14,000 years (De Geer, 1914, Plate I = Figure 31 of this report; Sandegren, 1924; Lidén, 1911) and of the Alps—about 20,000 years (Penck and Brückner, 1909, page 1169) indicates almost certainly that the ice-sheets disappeared at the same time on both sides of the Atlantic and mainly because of a rise in temperature (*See* Antevs, 1925).

That fairly well-marked climatic changes have taken place contemporaneously in North America and in North Europe seems also to be proved. The warmth period in North America a few thousand years ago almost certainly was contemporaneous with the temperature maximum in Europe, dated by Sandegren (1924) in Sweden to have prevailed from 7,000 to 3,500 or 2,500 years ago. The large growth in thickness of the big tree (*Sequoia washingtoniana*) in California during the fourteenth century most probably was connected with the contemporaneous cold and stormy spells in the Old World (Antevs, 1925, 1925a). The nature of the relationship, however, in the cases mentioned may have been different.

At present there is an intimate relationship between the atmospheric centres of action, as was especially shown by Hildebrandsson's (1897, 1899, 1909, 1910, and particularly 1914, page 4) studies. During summer, which is the time of particular interest in this connexion, the meteorological conditions in California and in Canada north of the Great Lakes greatly differ from those in the belt from Toronto and Winnipeg to the gulf of Mexico. Meteorological records show differences in pressure, temperature, and precipitation.

The summer temperature in northern Europe between North cape in Norway and northern Germany is different from that in southern Europe from Paris and Switzerland southward. The summer temperature in Scandinavia, the Baltic, and northern Germany is determined by the air temperature during the preceding winter above the sea between Norway and Iceland and by the length of the period that the ground was covered by snow.

A comparison between the summer conditions of North America and Europe according to Hildebrandsson (1914, page 4) is hardly possible. However, Figure 29 gives after Hildebrandsson (1910, Plates 4 and 6) curves of the mean temperatures of the months July to September at representative stations in North America and in North Europe. The northern part of North America is represented by San Diego and San Francisco in California and by Ivigtut in West Greenland. The south-central part of North America is represented by St. Louis, Toronto, and Winnipeg. The European stations, except Dorpat in Esthonia, all lie in Sweden and are well distributed over the country. The Ivigtut curve and the Swedish curve largely have an opposite course, whereas the St. Louis curve and the Swedish curve in many instances follow each other.

Figure 30 gives after Köppen (1914) and Mielke (1913) curves showing the annual departures from the normal temperatures in the regions that were glaciated. They are reproduced chiefly because an assumed correspondence between the annual temperature means in North America and in northern Europe has been used as foundation for correlation of clay varves in the two continents (De Geer, 1921, page 70). It seems doubtful, however, whether the annual temperatures give even a hint of the relationship of the conditions that determined the ice melting in North America

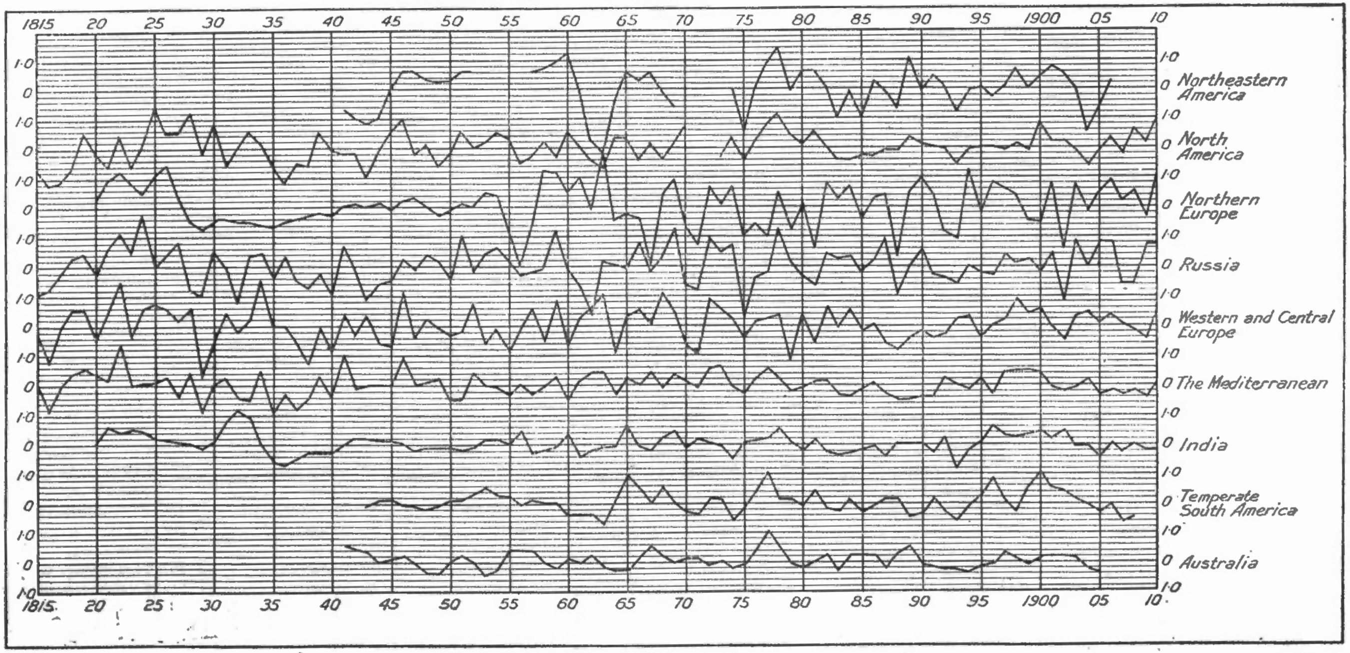

Figure 30. Curves showing annual departures from the normal temperatures in different regions of the earth. (From Köppen and Mielke.)

and in Europe, for, while the chief factor of the melting was the summer temperature, the annual temperature in mid-latitudes and high latitudes is determined by the winter temperature (von Hann, 1915, pages 139, 140).

The curves for North America, Russia, western and central Europe, and the Mediterranean have been taken from Köppen's plate; the remainder have been taken from Mielke's plate except the parts 1855 to 1879 which have been plotted from data given by Köppen in his Table I, page 301.

The curve for northeastern America is based upon 16 stations from lakes Huron and Erie to Newfoundland. The curve for North America comprises Greenland and the whole of North America except the southern states—133 stations. The curve for northern Europe comprises Denmark, Sweden, Norway, and Finland—24 stations. The curve for Russia comprises European Russia and Caucasus—31 stations. The curve for western and central Europe comprises the area from Ireland to Siebenbürgen and from Scotland and Schleswig to Switzerland—228 stations. The curve for the Mediterranean comprises all lands around the Mediterranean south of the Alps and Hungary—46 stations. The curve for India comprises India from the Himalayas to Ceylon—14 stations. The curve for temperate South America comprises South America south of the Tropical Circle —10 stations. The curve for Australia comprises both the Australian mainland and the islands as far as Hawaii—15 stations.

The curves for northeastern America and northern Europe show correspondence, particularly during the years 1845-1853 and 1882-1889, and the curves for North America and northern Europe during 1848-1859 and 1903-1910. On the whole, however, the annual temperatures in the two continents show little relationship. This result is corroborated by Behler's (1922; *See* Köppen and Wegener, 1924, page 237) recent study.

It may be concluded, therefore, almost surely, that there was correspondence between the disappearances of the North American and the European ice-sheets in the large features and probably relationship and possibly also correspondence in smaller features down to seasonal ones. The meteorological conditions in late-glacial time, however, were so different from those prevailing today that, from a study of the present climatic conditions, whether there is any agreement or not between European and North American conditions, probably no direct conclusions relating to late-glacial time can be drawn. Take, for instance, the Icelandic low which now plays a great rôle in the temperature of northern Europe and is also a factor in that of North America. It drives the warm water of the Gulf Stream towards the north and northeast, and distributes the oceanic heat over the North European continent. By contemporaneously bringing cold northwest winds over Greenland, Canada, and the United States, it causes temperature differences in winter in northern Europe and North America (Hildebrandsson, 1914, page 3; von Hann, 1915, pages 638, 640; Helland-Hansen and Nansen, 1917). In late-glacial time this Icelandic low may have been situated far to the southwest. Moreover, the Gulf Stream most probably did not cross the Wyville Thomson ridge north of Scotland until during the very last stage of the ice retreat in Fenno-Scandia, until the ice left the Fenno-Scandian moraines (Enquist, 1918, page 107; cf. Nordenskjöld, 1916, page 45; cf. Ahlmann and Helland-Hansen, 1918). Consequently, it is necessary to study the ice retreat in both continents and by going from large features to small ones find out in what details correspondence exists.

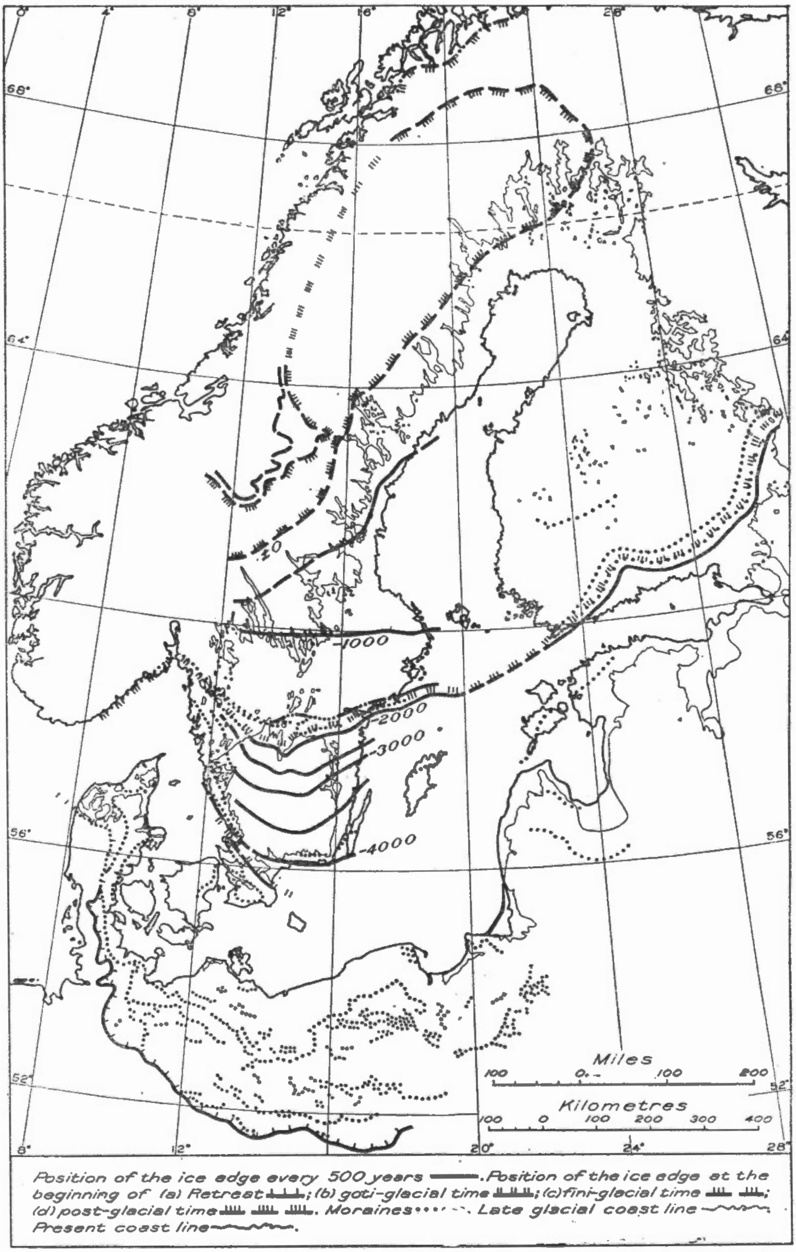

Figure 31. Retreat of the last ice-sheet in north Europe. (After Gerard De Geer, 1914, Pl. 1, except the position of the ice edge at the beginning of gotiglacial time and except the moraines in Germany which are after Wahnschaffe and Schucht, 1921, Pl. 29.)

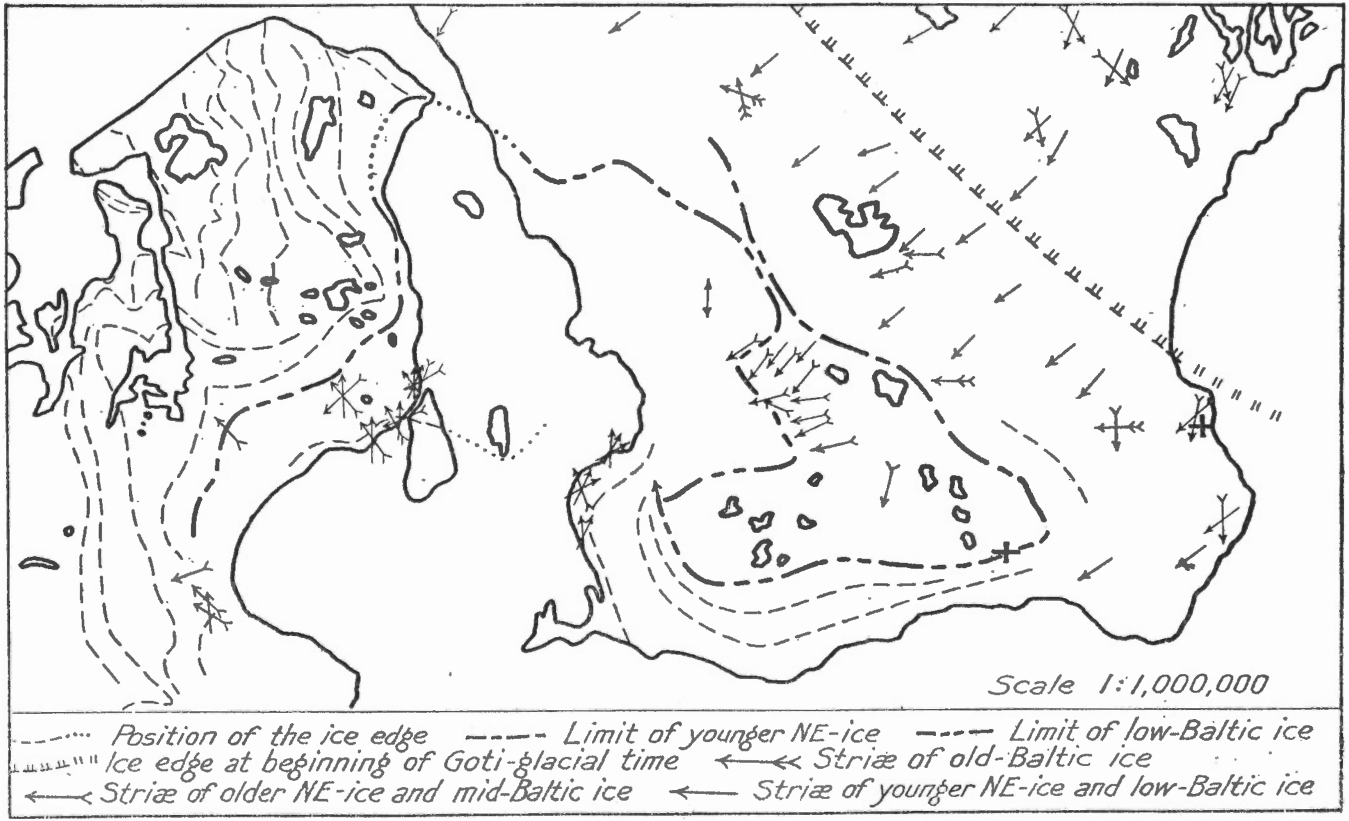

Figure 32. Positions of the ice edge and striæ in northeastern Zealand and southern Scania. The crosses in southern and eastern Scania mark Robertsdal and Rörum respectively. (From Holmström, 1904, Munthe, 1920, and Milthers, 1922.)

PROBABLE CORRELATION BETWEEN THE ICE RETREAT IN NORTH AMERICA AND IN EUROPE

If the North American and the European ice-sheets disappeared at the same time and for the same reason, New England should correspond to northern Germany in respect to ice uncovering, and Canada to the Scandinavian countries. The belt across lake Champlain and between lake Ontario and Mattawa river should approximately correspond to the Danish islands.

The uncovering of northern Germany has not been chronologically studied, but series of morainic lines, summarily reproduced from Wahnschaffe and Schucht (1921) on Figure 31 indicate repeatedly interrupted ice recession not dissimilar to that which took place south of the Great Lakes (Leverett and Taylor, 1915, Plate 5, page 62) and in New England (Antevs, 1922).

In the area north of lake Ontario and in Denmark as far as known the two longest interruptions in the otherwise fairly rapid ice retreat occurred. These regions, therefore, seem to form the natural starting points for attempts to connect the events in the two continents. The uncovering of Mattawa valley and the beginning of the Nipissing Great Lakes, which approximately mark the end of the relative stagnancy of the ice edge and the beginning of rapid retreat, took place, according to the time estimates based upon the cutting of the Niagara gorge, 10,000 to 10,500 years ago in round figures (Taylor, 1913c, pages 24, 25; cf. this report, page 63, and Figure 25). The uncovering of central Scania, the southernmost province of Sweden, which initiated rapid retreat in Europe, happened about 13,500 to 14,000 years ago according to the geochronological studies in Sweden. The figures, and particularly the American one, being only approximate, the lack of good agreement is explainable. The beginning of the final rapid retreat of the ice seems, therefore, to have been contemporaneous in North America and in Europe. The uncovering of the zone under discussion in Ontario has been touched upon on page 80. The ice retreat from the Danish islands and southwestern Scania, which was very complicated, will be briefly discussed.

To make the events clear it is necessary to describe the ice streams. Zealand, at least its northeastern half, during the last glaciation was first covered by ice coming from the northeast (Milthers, 1918, 1922). The retreat of this ice, in the northeastern part of the island, is recorded by a series of stadial moraines running north and south (Figure 32). Later, part of the area earlier occupied by this northeast ice was overrun by an ice lobe which pushed northward in the Öresund depression and expanded over the greater part of the island. Another lobe following the Great Belt spread over the southwestern part. In Copenhagen region there are two sets of striæ caused by Baltic ice, one from the southeast, and one from the south. The south set is attributed to the last advance of the ice (Milthers, 1922, page 50). The limits of the expansion of the Baltic ice lobes and stages of retreat are recorded by moraines, etc.

In southwestern Scania the following ice streams, all belonging, as far as known, to the last glacial epoch, are distinguished by Munthe (1920, pages 65, 108).

(1) The *old-Baltic ice* stream, moving in northwesterly direction and, in the southern part of the province, in westerly direction. This ice, which covered all Scania, without retreat passed over into
(2) The *older northeast ice* which during its retreat was divided into
(2a) The waning older northeast ice and
(2b) The *mid-Baltic ice,* coming from the east and later moving southwest and finally northeast. This ice at first advanced over southernmost Scania to the region northeast of Romeleåsen and then retired.

Finally, at least the southern and central parts of Scania were uncovered. The climate in southern Scania, as shown by a fairly rich flora and fauna in a deposit between two till beds at Robertsdal, 7 miles north of Ystad, improved, and the temperature, when highest, may have corresponded to that now prevailing in the northern part of central Sweden, that is the July temperature may have been about 59° F. (Munthe, 1920, pages 127, 137).

After this period of northerly temperate climate the ice again advanced as
(3a) The *younger northeast ice* which extended to the northwest-southeast line marked in Figure 32 and
(3b) The contemporaneous *low-Baltic ice* which expanded over southernmost and westernmost Scania (Figure 32).

The retreat of both ice-sheets in the beginning was interrupted several times.

Thus, the old-Baltic ice is not recorded in Denmark. The older northeast ice is the same in the two regions. The mid-Baltic and the low-Baltic ice-sheets in Scania may be the southeast and the south ice-sheets respectively in Zealand. A serious difficulty, however, is that so far no correlatives of the interstadial bed at Robertsdal in Scania have been found in Denmark. This may, of course, be due to lack of exposures, but the suspicion cannot be suppressed that the deposit at Robertsdal may be wrongly dated. It may be younger than the low-Baltic ice and may be an Alleröd bed (cf. below). Munthe's (1920, pages 133, 137) correlation of the beds resting on the top of the upper till at Robertsdal with the Alleröd beds and the superjacent covering arctic deposits seems to have weak foundation. If this be so the mid-Baltic and the low-Baltic ice streams may have been separated only by a comparatively slight oscillation.

When the recession of the low-Baltic ice in southeastern Scania had proceeded for some time it may have become fairly rapid, judging from the rather thick clay varves in the region 6 miles east of Ystad. Probably at about this time the rise in temperature occurred which is recorded by the so-called Alleröd beds at about twenty-five localities from the Kaiser Wilhelm canal in Holstein and the district of Silkeborg in Jutland across Fyn and Zealand to southern Scania and Bornholm. The Alleröd beds are freshwater deposits, consisting essentially of marl or calcareous gyttja with a fauna and flora indicative of a temperate continental climate (*See* Nordmann, 1910, pages 315-319, 325; 1915). They rest upon varved clays containing arctic plants. They are also constantly overlain by clays with an arctic flora indicating that the warmer time was followed by a relapse to arctic conditions. These upper clays in distinction from the lower clays are not varved; they were not deposited directly off the ice.[1]

The positions of the ice border during the deposition of the Alleröd beds and the overlying clay have been much discussed. In the writer's opinion the ice during the Alleröd period must have stood north and east of the area in which the Alleröd beds occur, and most probably south and west of northeastern Scania, where, and north of where, the ice retreat is determined by clay studies. It, consequently, may have stood on the northwest-southeast diagonal of the province. This view that the ice did not stand very far from the region is supported by the fact that the Alleröd beds in Bornholm occasionally contain *Dryas octopetata* and *Betula nana* probably suggesting that the climatic oscillation was less pronounced here than in Zealand, etc. (Milthers, 1916, page 232). At the present time a similar, typically continental climate with very dry and abnormally warm—54° to 59° F.—summers exists in West Greenland between the 64th and 69th parallels in the immediate neighbourhood of the land ice, but about 80 miles from the open sea (Nordenskjöld, 1916, page 39).

If the Robertsdal deposit is an Alleröd bed the ice readvanced to the spot during the following period when arctic conditions prevailed, but probably not much farther. It did not return to Bornholm. If, however, the Robertsdal deposit is older than the low-Baltic ice, as believed by Munthe, the cold climate subsequent to the Alleröd period may not have been accompanied by any great readvance of the ice edge. The ice may have remained on the diagonal of Scania. On the east coast it may have stood at Rörum 6 miles north-northwest of Simrishamn, for here a local clay deposit containing about 1,500 varves postulates ice damming and stationary ice border for at least that time. But more probably this deposit is a later one. The deposit is similar to the clays formed during a halt of the ice border caused by little melting.

Thus, there may have been one large oscillation of the ice front between the old northeast ice and the mid-Baltic-southeast ice, a somewhat less fluctuation between this latter and the low-Baltic-south ice, and a third, probably large, fluctuation recorded in the Alleröd beds. There were also several minor oscillations registered by moraines, etc. Although probably insignificant as far as the movements of the ice edge go several of these oscillations represent very long periods of time. In the practically flat, northeast Zealand there occur deposits of varved glacial clay up to 30 feet thick formed in small lakes dammed off the ice front (Milthers, 1922, pages 83, 84). The annual layers are very thin. The present writer found some of these deposits to represent several hundred—perhaps more than 1,000—years. Thin wedges of till are interbedded with the clays at different levels near the margins of the clay deposits, showing that the ice edge all the time stood close by. Similar clay deposits are frequent in southwestern Scania.

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[1] The upper clays at Stenstrup in Fyn as Nordmann (1918, 1922) and Milthers (1918a) maintain, and as this writer observed in 1919, on the whole show only obscure lamination, as may any water-lain deposit, not a lamination comparable to that of the varved late-glacial clays. Only in one of the clay pits a zone about 1½ feet thick showed fairly distinct lamination, which probably is seasonal. Our imperfect knowledge of the formation of varved deposits, however, may not justify the conclusion that this clay is glacial, for the fact that in no instance is there more than one Alleröd bed, as Nordmann points out, together with other evidence, seems to show that all the deposits date from the same time and that the ice edge may have been far from Stenstrup when the clay was formed.

Many deposits of varved clay in the flat, south Scania, Zealand, and Prussia occur on the top of hillocks and form the highest parts, which rise 30 to 50 feet above the surroundings (Westergård, 1906; Munthe, 1907; 1920, pages 97, 107; Milthers, 1922, page 86). As pointed out by different students this clay must have been deposited in lakes ponded on all sides by ice. The clays are up to 20 feet thick; and many of these true glacial lakes must have persisted for a few hundred years. How the lakes came into existence, and why they were situated just on the top of hillocks is not quite clear. It seems probable that they were results of differential melting of the ice. It is well known that a thick debris cover prevents ablation and a thin and dark cover promotes it by transmitting the absorbed solar heat to the subjacent ice. Particularly good examples are known from the Antarctic where depressions formed in this way are frequent (Wright and Priestly, 1922, pages 278-286, 290). Oblong depressions formed as moraines have sunk in the ice may reach more than 100 yards in width. Also, clear ice is more readily ablated than is thickly buried ice. To this fact the interior flats or crescent-shaped depressions somewhat inside the lower ends of some glaciers in Alaska may be due (Tarr and Martin, 1914, pages 76, 118, etc.).

The hillock clays in Munthe's opinion were largely deposited in ordinary glacial lakes extending over the region. The ice cover disappeared first from the knolls because it was thinnest there. While clay was being laid down on the hillocks the depressions were still occupied by debris-covered dead ice which prevented deposition of clay in the depressions. Since this view postulates that the dead ice in the depressions had melted down to the level of the hillocks, clay evidently would have been deposited on the top of the ice and would have sunk to the ground, when the buried ice finally disappeared. Varved clay, therefore, should also occur in the depressions, but it does not. On the hypothesis of differential melting, the dead ice extended up to the water surface, and the varved clay could be formed in lakes off the ice front. Normally, however, these clay deposits may have been formed inside the ice edge.

All the evidence tends to show that the disappearance of the ice from Zealand and southwestern Scania was very complicated and took a considerable length of time. The time involved cannot be estimated with accuracy, but it probably amounted to 10,000 to 15,000 years. Regarding the length of time there seems to have been agreement between the uncovering of southern Ontario and Zealand-Scania. In what further respects correspondence is to be found is as yet little known. No beds corresponding to the Alleröd beds appear to be known in North America, but then these beds in Europe appear to be limited to the southwestern Baltic region north of the German mainland, and furthermore peat bogs, etc., have been given little attention in North America. As for the temperature conditions in North America during the time in question the shellfish and trees observed in the Algonquin beds all still live in the waters or on the shores of lakes Superior and Huron, and do not suggest a subarctic climate (Coleman, 1922, page 34; F. C. Baker, 1920). It should be noted, however, that the molluscs which form the majority of the remains may be doubtful indicators of the atmospheric temperature, since the water of the lakes, particularly the water of lake Superior, is very cold (Coleman, 1922, page 64; cf. this report, page 43).

The next most marked interruption in the ice retreat in Europe north of Germany is that during which the Fenno-Scandian moraines—that is the Ras in Norway, the finiglacial moraines in Sweden, and the Salpausselkäs in Finland—were formed[1] (Figure 31). These halts are also of great importance in this discussion. In Finland, where the moraines and the time of their formation have been studied in greatest detail, there occurred two marked halts separated by a time of slow recession (Sauramo, 1918, pages 23, 35). The first halt represents 225 years, the intervening retreat 251 years, and the second halt 183 years, making in all 659 years. In the western part of Finland, but not farther east, the ice front also halted or retired but slowly for about 50 years, beginning about 125 years after it left the second moraine line (Sauramo, 1923, Plate 8). The halts according to De Geer (1914, Pl. I; this report Figure 31) began about 2,500 years after the ice retreat had started in earnest in central Scania.

In Canada, the ice edge readvanced an unknown distance and reached Iroquois Falls more than 350 years after it first left the place somewhat before year 1550 (*See* page 82). The figure indicates the time of the retreat from the mouth of Montreal river on lake Timiskaming. The time occupied by the ice recession from Mattawa valley to Montreal river, a distance of 55 miles, is unknown, but it seems as if the readvance at Iroquois Falls occurred about as long after the ice border left Mattawa valley as did the Fenno-Scandian halts after the beginning of the rapid retreat in central Scania. The readvance and the subsequent events in Canada are too little known to permit a profitable discussion of their agreement or disagreement with the halts in Fenno-Scandia. Exact agreement is not to be expected, for even if the cause of a halt were the same its effects might have been influenced by local conditions. Furthermore, halts may be due to regional conditions, especially to rich nourishment. Probably only halts caused by marked fall in temperature were fairly similar on the different sides of the Atlantic. For instance, the halt at the Second Salpausselkä which, as discussed on page 53, may essentially have been due to increased nourishment, probably cannot be expected to be faithfully recorded in Canada.

The correlation outlined appears to be the only one that with the present knowledge of the waning of the ice-sheets can be considered.

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[1] The correlation of these moraines has recently been doubted by Sauramo (1923, pages 86, 117, 123, 124, 158). The breaking in of the sea across central Sweden into the Baltic is placed at the end of the First Salpausselkä period, instead of at the end of the Second Salpausselkä stage, on the ground that flocculated clays, believed to be indicators of salt water, appear in southern Finland at the end of this period. Since, however, flocculation sometimes occurred in fresh water (page 14), there is no reason to doubt the old correlation between the Fenno-Scandian moraines. On the contrary, since climatic conditions played a dominating or very important rôle in the stages both in Sweden and in Finland (cf. page 53), their contemporaneity seems to be directly proved. Most of the supposed discrepancies in the history of the Baltic (Sauramo, 1923, pages 125-129), consequently, may not exist. It is certainly peculiar that no marks of lowering of the water-level, according to Sauramo (1923, page 128), are observed in the clays in Finland. However, Ramsay (1924, page 28) thinks that Sauramo's varve 0 marking the transition from stiff clay to more lean clay indicates the year when the ice lake was drained, not the year when the ice border left the Second Salpausselkä. Also, the sudden decrease in the thickness of the varves and change from stiff clay to silt and fine sand in profile 52, Mukkala (Sauramo, 1918, Plate 4; 1923, page 69) at year +30 probably indicates a sudden shallowing.

CHAPTER XI

DESCRIPTION OF THE SECTIONS AT THE LOCALITIES STUDIED

In the following a description is given of the sections at the localities studied. Their locations are indicated on the maps as follows: localities 1-7, A-E, Figure 33; localities 9-29, Figure 34; localities 30-34, Figure 35; localities 36-84, Figure 36; localities 85-104, Figure 37. The layers in the sections are enumerated from the top downward. The numbering of the varves is explained on page 120.

Localities in the Ottawa Region

(Figure 33)

(1) 7 miles south of Ottawa, Black rapids, east side of Rideau river, 200 yards north of the dam[1]

15 feet sand, somewhat silty

3 feet contorted sandy clay

1 foot sand, somewhat clayey

Abrupt limit. The underlying clay stiff

7 feet silt and clay. 54 varves. Varves 9-21 consist largely of sandy silt; varves 22-43, of silt and clay. Lamination very distinct. Varves 44-53 consist of stiff, almost homogeneous clay. Summer layers are grey-brown, and the thin winter layers are red-brown. Varves 54-62 are similar the underlying ones, though fairly distinct. The lower part of the bed was deposited in fresh water, the upper part probably in brackish water. No shells were found

2 inches slidden zone

$4\frac{3}{4}$ feet sand and silt, 6 varves. Winter layers grey-white

Bedrock at river level

(2) $2\frac{3}{4}$ miles south of Ottawa, Billings Bridge, brick-yard 500 yards south of the highway bridge across the Rideau

6 feet leached grey clay

1 foot clay, a few indistinct varves

$6\frac{1}{3}$ feet stiff clay, 12 varves, fairly distinct to distinct. Summer layers grey-brown, winter layers red-brown or pink

3 feet stiff clay. Varvity indistinct. With *Portlandia arctica*

5 feet stiff, dark-grey clay. Varves not surely distinguishable. The lowest part with numerous shells

Probably disconformity. 2 inches above the contact, silty clay; 1 foot above, stiff clay. 2 inches below the contact, fine clayey sand

25 feet clayey sand and sandy clay. No varves distinguishable. With masses of shells

Disconformity

2 +x feet stiff, grey-blue clay with rather distinct varves, exactly as at localities 1 and A (page 68). Probably deposited in brackish water. Said to be 12 to 15 feet thick

[1] Distances are from the main railway stations.

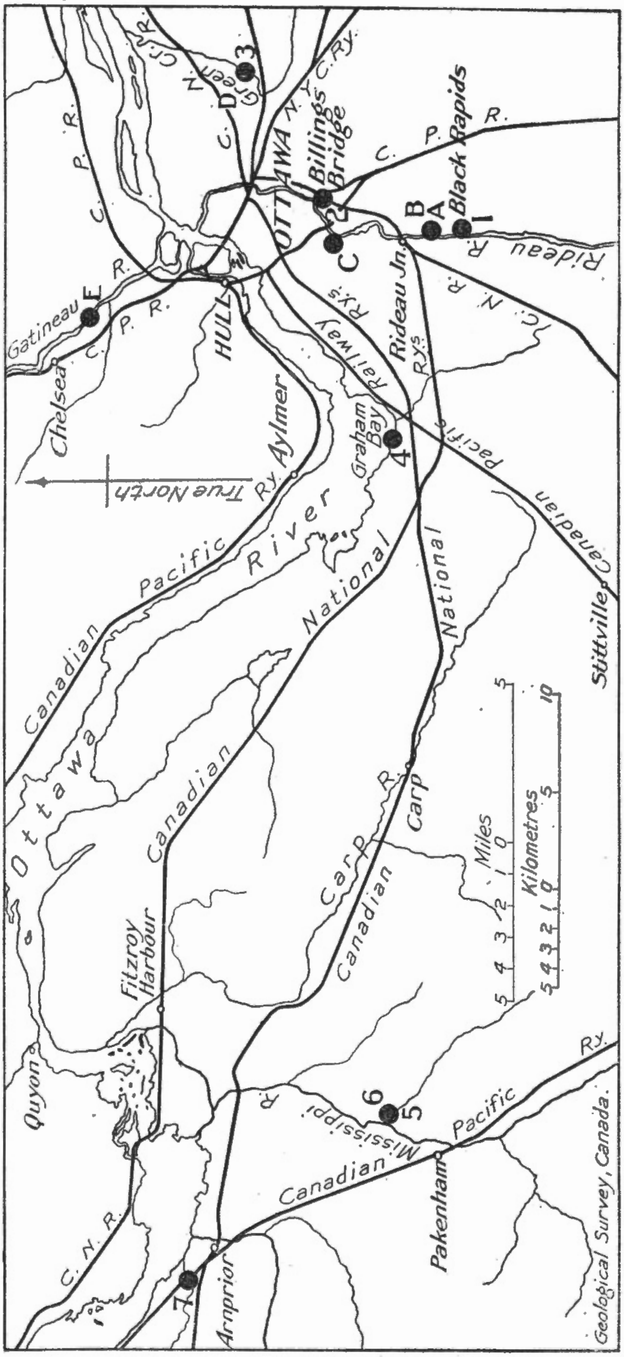

Figure 33. Localities examined in Ottawa area (1–7, A–E).

(3) $4\frac{3}{4}$ miles east of Ottawa and $1\frac{1}{3}$ miles west-northwest of Blackburn station, Greens creek, bluff on the east side 250 yards south of the highway bridge

$1\frac{1}{2}$ feet fine sand

$2\frac{1}{2}$ feet homogeneous clay

8 inches clay. Varvity indistinct. Probably deposited in brackish water

14 inches clay, 29 distinct varves, deposited in fresh water. Summer layers light, silty; winter layers, grey-brown (Plate IX)

$2\frac{1}{2}$ feet sandy clay with scattered boulders, 20 to 30 varves. Lamination distinct but disturbed. The zone indicates readvance of the ice edge or grounding of an iceberg

4 feet silty sand

5 feet sand

15 feet till

Slate

(4) $8\frac{1}{2}$ miles southwest of Ottawa, bluff on the west side of Graham bay, 250 yards southeast of Rocky point

7 feet clayey and silty sand with laminæ of clay and some moraine material. Zone somewhat disturbed. A few varves.

$4\frac{3}{4}$ feet clayey and silty sand with morainic material and gravel; 9 varves. Each varve consists of some to several layers of yellow sand in alternation with thin laminæ of dark-brown silty clay. Winter layers not very marked

2 feet layers of gravel and sand, and fine laminæ of clay in alternation. Part of a varve

River level. Depth to substratum unknown

(5) 29 miles west of Ottawa, or 7 miles southeast of Arnprior and $1\frac{3}{4}$ miles north-northeast of Pakenham, bluff on the south side of the tributary to the Mississippi 200 yards below the bridge

8 feet clay. Varves not distinguishable

15 feet sand and clay, 45 varves. Each varve consists of a sand layer and a clay layer. Varves are in many cases distinct, but some are indistinct. The sand layers contain numerous shells, especially *Macoma baltica* (Plate IX)

3 feet covered to creek level. Depth to substratum unknown

(6) 175 yards east of locality 5, 75 yards below the bridge, bluff on the north side of the creek

Many feet covered

9 feet sand and clay. 26 varves, similar to those at locality 5 and containing shells. One varve consists of 19 inches sand and $7\frac{1}{2}$ inches silt-clay. It records a drainage (Plate IX)

8 feet covered to creek level. Depth to bottom unknown

(7) $1\frac{1}{3}$ miles northwest of Arnprior, brick-yard on the creek and Canadian Pacific railway

(A) Close to the railway track

3 feet sand

3 feet weathered clay

10 feet sand and clay, about 15 varves. Possibly deposited in fresh water

Probably not deep to substratum

(B) 200 yards west of the track

15 feet stiff, dark-brown clay, varves partly distinguishable with difficulty; average thickness 2 to 3 inches. *Portlandia arctica*

Localities in Lake Simcoe-North Bay Region

(Figure 34; Plates IV, V, IX; cf. pages 121, 122)

(8) Beaverton, brick-yard and bluff on the creek at the eastern edge of the town

(A) Clay pit

2 feet sand
3 feet clay with layers of sand. Varves not distinguishable
4 feet stiff, yellow-brown clay, about 18 varves. Lamination very faint
Depth to substratum unknown

(B) Bluff on the brook, 100 yards from A

2 feet sand
5 feet clay, not measured
1¾ feet yellow-brown clay with some layers of silt and sand. Lamination indistinct. 27 varves
4 feet stiff, dark-grey clay containing scattered pebbles and contorted so as to be mostly homogeneous. One zone contains minute white lime concretions so that it appears spotted. Average thickness of undisturbed varves 1/16 to ⅛ inch. The zone probably marks a readvance of the ice edge
5¼ inches stiff clay, 27 varves
3 feet boulder clay, likely varved clay overridden by ice
Brook level. Depth to substratum unknown

(9) 1 mile south of Bracebridge, west of the railway, slide on the north side of the river

Many feet covered
4 feet clay, 112 varves, not connected with the other series in the district
18 feet covered to river level. Depth to substratum unknown

(10) 1½ miles east of Bracebridge, brick-yard north of the road and west of the creek

1 foot sand
2 feet leached clay
20 feet clay, 784 varves. Summer layers of grey, silty clay, winter layers red-brown
Bottom of clay pit. According to information more than 10 feet to substratum

Series measured: 61-446, 61-141, 167-844, 403-743

(11) ¾ mile northwest of Bracebridge, old brick-yard at the crossroads

2 feet leached clay
1¾ feet clay, 75 varves. Thicknesses not good
2¾ feet clay, varves 490-581
2 feet clay. Thicknesses of varves not characteristic
8 feet grey-brown clay with a thin, disturbed zone, varves 82-242, 262-360
Depth to substratum unknown

Series measured: 82-223, 262-333, 295-360, 305-360, 490-581, 490-535

(12) 1½ miles north of Bracebridge, bluff 200 yards northwest of the fall

Many feet covered
12⅓ feet silt, in the lower part sandy, varves 8-157. Winter layers fairly thin
4½ feet varved sand, varves 1-7
12 to 15 feet covered to bedrock

Series measured: 1-157

(13) 2 miles north of Bracebridge, bluff on the west side of the river directly below the end of the east-westerly road

7 feet sand
6 feet disturbed varve clay
38 feet grey, sandy, varved silt with thin red winter layers. Towards the top grey silt and rather thick, red winter layers. Lamination very distinct
6 feet covered to river level. Depth to substratum unknown

Series measured: 144-461, 401-481 (27-143)

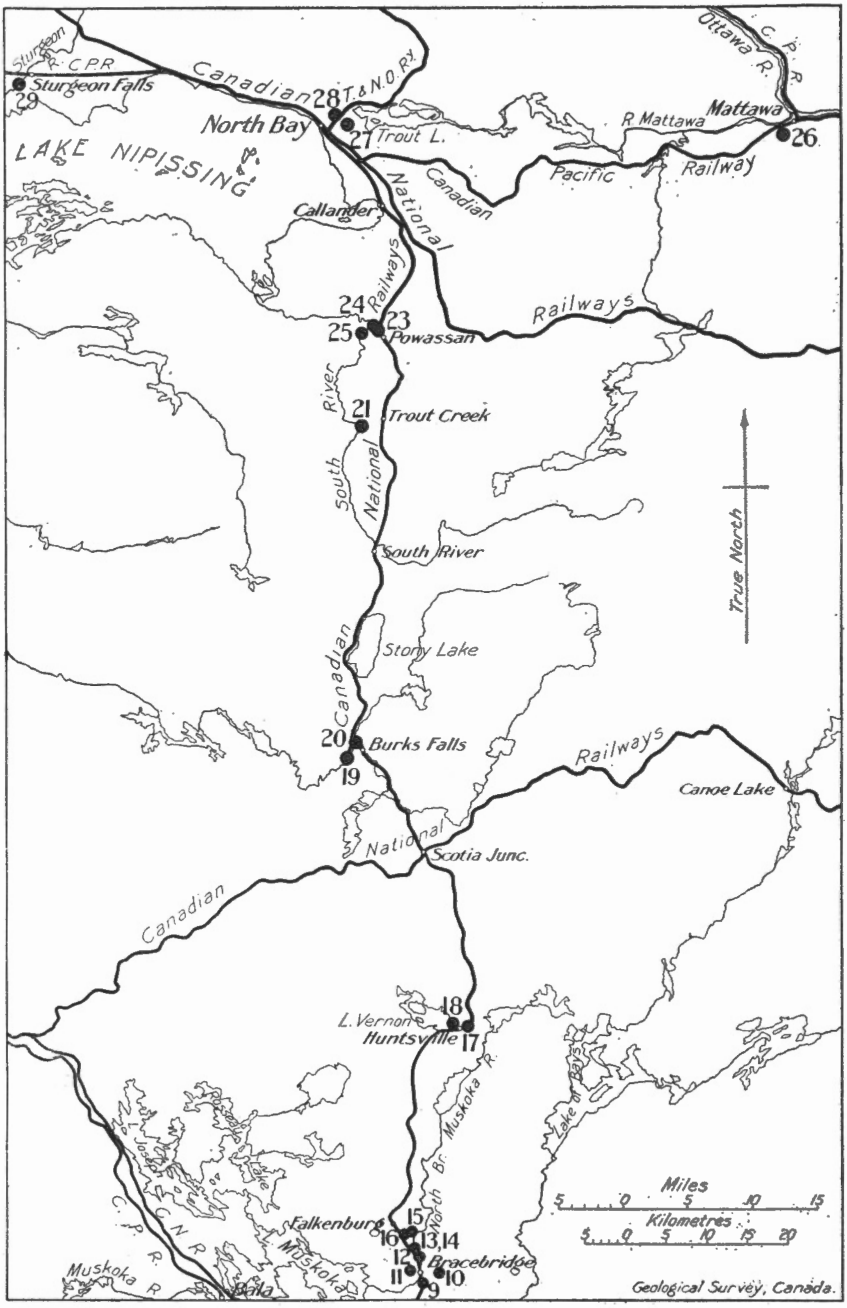

Figure 34. Localities examined in Bracebridge-North Bay area (9–29).

(14) Bluff 100 yards north of locality 13
Surface of sedimentation
9 feet sand
9 feet stiff, slidden clay
16 feet beautifully varved clay-silt, varves 262-581
20-25 feet covered to river level. Depth to substratum unknown
Series measured: 262-581

(15) 3 miles north-northwest of Bracebridge, roadcut 300 yards north of the big bend of Muskoka river
Plain of sedimentation
4 feet sand
4 feet varved silt. Lamination and thicknesses good. Varves 342-409
10½ feet varved sand and silt. Winter layers thin. Thicknesses not good. Varves 276-341
3 inches slidden zone. No varve missing
2¾ feet varved sand and silt, varves 266-275
1½ feet slidden zone
9 feet varved sand and silt, varves 208-251
Series measured: (208-251, 266-409)

(16) 600 yards west of locality 15, roadcut at the brook 300 yards east of the railway track
Plain of sedimentation
2 feet unvarved sand
1 foot varved sand
18¾ feet silty, dark-grey sand and thin, red clay laminæ—winter layers. Some varves contain rather coarse sand. Thicknesses partly not good. Varves 232-557
Depth to substratum unknown
Series measured: (232-557)

(17) 1 mile west-southwest of Huntsville, brick-yard just south of the railway crossing
2 feet leached clay
9½ feet silty, grey-brown clay. Lamination distinct. Varves 1-427
Depth to substratum unknown, but probably not great
Series measured: 1-226, 42-145, 82-400, 159-279 (401-427)

(18) 1¾ miles west of Huntsville, bluff on the eastern shore of lake Vernon, opposite the island
5 feet clay, 500-600 indistinct varves
1¾ feet clay, about 175 partly indistinct varves
10½ feet grey-brown clay, silty in the lower part, stiff in the upper part. Lamination distinct. Varves 26-397
1½ feet sand, varves 17-25
Lake level. Not far to substratum
Series measured: 17-182, 90-179, 204-290, 101-397, 101-218, 149-228, 165-288, 204-329

(19) 1½ miles west of Burks Falls railway station, old brick-yard at the southwestern edge of the village
5 feet clay and sand, mostly covered
5¼ feet fairly stiff clay, varves 311-660. Varves indistinct in the lower part, but becoming better upwards
1½ inches slidden zone
8 inches clay, about 50 varves. Varve limits very indistinct
5½ feet clay, silty at the base, but becoming fat upwards, varves 1-252. Varves distinct except at the top
2 inches slidden zone
7 inches silty clay, 15 varves
Disturbed zone. Depth to substratum unknown
Series measured: 1-252, 1-193, 311-660

(20) 1 mile northwest of Burks Falls railway station, roadcut west of the junction of the rivers

2½ feet silt and fine sand, 21 varves. The series evidently represents a lower horizon than profile 19. Depth to substratum unknown

(21) 1¾ miles west-southwest of Trout Creek, roadcut on the east side of South river (Plate IX)

Surface of sedimentation
3 feet sand
2 feet sand-silt
10 feet sandy silt, 31 varves
About 25 feet sandy silt with thin bands of clay. Varve limits not surely distinguishable
3½ feet sandy silt with very thin salmon-coloured winter layers, varves 314-341. Varve limits difficult to distinguish (cf. page 15).
Sharp limit, but no discontinuity
17 inches rather stiff, thinly varved clay, varves 261-313
4 inches slightly disturbed zone
9¼ inches clay, varves 203-237. Varves consist of grey silt and thick red-brown winter layers
Probably gap
13 inches clay, varves 178-200
4 inches disturbed zone
2½ feet clay, varves 114-172
8 inches disturbed zone
3¼ feet clay, varves 34-105
8 inches disturbed zone
1⅓ feet beautifully varved clay, varves 1-25
4 feet above river level. Depth to substratum unknown

(22) 2 miles south-southeast of Powassan, roadcut 200 yards northeast of the crossroads west of the railway

Plain of sedimentation
8 feet silt and sand
2⅔ feet coarse silt to sandy silt with thin winter layers, varves 284-316. Varves increase in thickness upwards because of shallowing
2½ inches clay, varves 273-283. Varves difficult to measure
5 inches distinctly varved clay, varves 256-272
10 inches silty clay, varves 221-255. Varve limits exceedingly difficult to distinguish. Thicknesses valueless
1 foot, partly lean, partly fairly greasy clay, varves 162-220. Varve limits very difficult to distinguish
12½ feet beautifully varved clay, varves 8-162. Varves consist of light-grey silt and brown clay
10½ feet sand, varves 1-7
12 feet above brook level. Probably not far to bottom

Series measured: Powassan 1-316, 116-162

(23) ½ mile north-northwest of Powassan, bluff on the north side of the brook

2 feet leached clay
5 feet yellow silt and brown winter layers. Varves 104-148. Fine
Brook level. Depth to bottom unknown

Series measured: Powassan 104–148

(24) 1 mile northwest of Powassan, bluff on the west side of the brook

3 feet sandy, disturbed, and leached clay
2½ feet clay, silt, and sand, distinctly varved, varves about 330 to about 351
1¾ feet fat, very indistinctly varved clay. Varves 228 to about 329
9 inches indistinctly varved clay, varves 186-227
3½ feet distinctly varved clay-silt, varves 28-185
3 feet below brook level. Depth to substratum unknown

Series measured: Powassan 24 to 227 (24-64, 228 to about 351)

(25) $1\frac{1}{2}$ miles west-northwest of Powassan, slide on the west of South river just west of the highway bridge

Several feet weathered and covered
11 inches greasy, red clay, varves 173-227
$6\frac{1}{4}$ feet yellow silt and brown winter layers, varves 110-172. Fine
Covered 4 feet down to river level
Depth to bottom unknown, but probably not great

Series measured: Powassan 120–227, 149-167, 180-194 (110-119)

(26) $1\frac{1}{2}$ miles southwest of Mattawa, bluff on the brook $\frac{1}{2}$ mile south of the railway track

Several feet covered
8 feet grey-brown silt, 21 varves. Winter zones generally consisting of a lower blue-grey layer and an upper red-brown layer. This is perhaps due to unsynchronous flocculation of two kinds of material (Plate IX)
Brook level. Unknown depth to bottom

(27) Brick-yard 2 miles east of North Bay

Clay pit 6 feet deep. Depth to substratum unknown. The clay is thoroughly contorted so that only a series of 12 varves—15 inches—could be measured at one place. Since the surface is very even the contorsion may have been caused by readvancing ice. No till was observed on top of the clay, though

(28) $1\frac{1}{4}$ miles northeast of North Bay, old brick-yard on the brook west of the Trout Mills road (Plate IX)

2 feet leached clay
5 feet yellow silt with salmon-coloured winter layers, 44 varves. Lamination distin ct. Three sandy layers indicate drainages
Clay pit filled with water. Depth to substratum unknown

(29) 1 mile southwest of Sturgeon Falls, bluff on the west side of Sturgeon river. Two series measured

(A) (Plate I b)

Surface of erosion
2 feet pressed and leached silt with thin winter layers
Sharp limit, though no discontinuity
$2\frac{1}{2}$ feet stiff clay, 42 varves. Summer layers silty, cream coloured; winter layers, thick, brown. Lamination very distinct, but thicknesses not good, since the clay is somewhat pressed
River level. Depth to substratum unknown

(B) 200 yards below A

5 feet weathered clay-silt, partly covered
$4\frac{1}{2}$ feet stiff, grey-brown clay, somewhat pressed, so that thicknesses not good, 105 varves
River level. Depth to bottom unknown

Localities in Sudbury-Sault Ste. Marie Area

(Figure 35; Plates V, VI, IX; cf. pages 123–125)

(30) $2\frac{1}{2}$ miles southwest of Sudbury, bluff on the west side of Junction creek, 600 yards south of the railway track (Plates I d, IX; page 16)

2 feet leached clay
5 inches stiff, yellow-brown clay, varves 66-77
Sharp limit, but no discontinuity
$9\frac{3}{4}$ feet silt or sandy silt and thin, brown winter layers, varves 1-65
About 6 feet covered to substratum. Underlying gravel bed exposed close by

Series measured: 1-77, 25-41

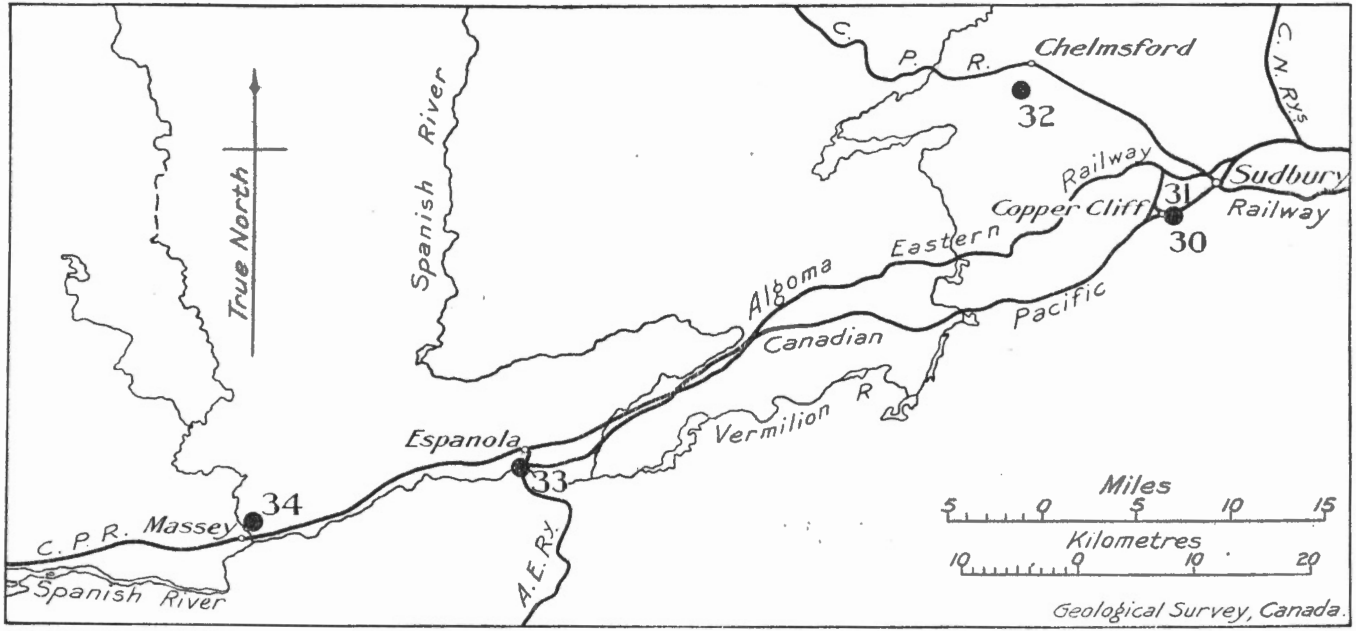

Figure 35. Localities examined in Sudbury area (30-34).

(31) 500 yards north of locality 30, on the brook 75 yards south of the railway track (Plate IX)

4 feet leached clay
3¾ feet very fat, yellow-brown clay, varves 101-183
Brook level. Depth to bottom unknown

Series measured: 1-83, 1-24, 15-69, 25-53, 66-83

(32) 12 miles northwest of Sudbury and 1½ miles south of Chelmsford, bluff on Whitson river, just south of the highway bridge (Plate IX)

6 feet leached silt
7 feet coarse, hard-packed, light-grey silt and thin winter layers, 91 varves
River level. Depth to bottom unknown

(33) 40 miles west-southwest of Sudbury, 1¼ miles south-southwest of Espanola station, bluff on the north side of Spanish river a few hundred yards west of the railway bridge (Plate I c, e)

1 foot silt-sand
8 inches granular clay
26 feet sandy, grey-white silt and red-brown winter layers, varves 1575-1802. Some winter layers are very thin and were locally eroded away during the spring following their deposition. Towards the bottom the winter layers are quite thick
2 to 12 inches slidden zone. Number of disturbed varves not known, but probably not great
1½ feet silty clay, varves 1555-1570
4 inches silty clay, somewhat disturbed. Likely a few varves
12 feet clay-silt, varves 1289-1551. Summer layers of white silt. Winter layers rather thick, brown. Lamination very distinct
6½ inches, varves 1286-1288. Varves 1286 and 1288 are exceedingly fat and mark drainages so far away that only very fine material was transported to this place
13½ inches stiff clay, varves 1201-1285
2 feet 7 inches stiff, almost homogeneous, brown clay. Varve limits not surely distinguishable. Probably more than 600 varves
10 inches stiff, grey-brown clay, 145 varves
16 feet, varves 1-412. At the bottom, sand which upwards gradually goes over into silt and stiff clay. Varve limits very distinct except in the uppermost part
A few feet sand to bedrock which outcrops close by

Series measured: 1-412, 1-269, 1201-1551, 1219-1533, 1219-1285, 1555-1570, 1575-1802, 1575-1637, 1726-1785

(34) 55 miles west-southwest of Sudbury and 1½ miles north-northeast of Massey roadcut at old brick-yard

5 feet disturbed and leached clay and silt
6½ feet clay-silt, varves 11-221. Some varves contain sand. Lamination good
5¾ feet sand, varves 1-10
Probably not deep to substratum

Series measured: 1-211, 32-211

(35) 2 miles north-northwest of Sault Ste. Marie, bluff just south of the railway bridge across a brook. Above the Nipissing shore-line

3 feet slidden silt-clay
1½ inches sand, drainage layer
2½ feet stiff, brown clay, varves not measurable
2 inches sand, drainage layer
1½ feet stiff, brown clay. Varve limits only partly distinguishable
4¾ inches grey sand, drainage layer, varve 324
6¾ feet clay-silt, varves 122-323. Some varves sandy. Varve limits partly difficult to distinguish
16½ feet silty clay, silt, and sand, varves 1-121. Several drainage layers. Varve limits partly faint. Winter layers red-brown in varves 1-16 and from 101 and upwards and red in varves 17-100

Fault
4 feet 10 inches, 23 varves. Varves consisting of a few millimetres of silt, a thick layer of fat red clay, and thin greyish-white winter layers. In some varves there are layers of silt
2 feet covered to brook level. Depth to substratum unknown
Series measured: 1-323, 58-115, 121-237

Localities on the Eastern Side of Lake Timiskaming

(Figure 36; Plates VI, VII; cf. pages 126, 127)

(36) 20 miles south-southeast of Haileybury, bluff on rivière Lavallée at the sawmill near the river mouth
Many feet clay, partly disturbed and faulted, not measured
19 feet light-grey silt and thin, dark-grey winter layers, varves 96-308. The silt is very coarse in the lower part, but becomes gradually finer upwards. Lamination very distinct. Thicknesses good
Creek level. Depth to substratum unknown
Series measured: 96-243, 214-308

(37) 1 mile north of locality 36, ½ mile northeast of Fabre wharf, bluff on lake Timiskaming
2½ feet leached clay
3 feet quite stiff clay, varves 996-1057
Lake level. Depth to substratum evidently very great
Series measured: 996-1057

(38) 2¼ miles south of Fabre station, railway cut a few hundred yards northwest of the railway bridge across the Lavallée
4 feet silt, partly disturbed
2½ feet light-grey silt with thin winter layers, varves 35-62
Big boulder; bottom or practically bottom
Series measured: (35-62)

(39) 1 mile south of Fabre station, railway cut
2 feet leached silt
1½ feet silt with thin winter layers, varves 48-78. Varve 48 is bottom varve
Till
Series measured: (48-78)

(40) Fabre station, excavation for the foundation of the station house. The till underlying the clay has a very uneven surface, so that the profile given is a composite of several measurements
2 feet leached clay
2 feet 2 inches clay, gradually becoming fat upwards, varves 331-447
10 feet light-grey silt, with thin winter layers, varves 57-330. Varve limits distinct in the lower part, but somewhat obscure in upper part. Varve 57 is bottom varve
Till
Series measured: 61-130, 61-112, 68-126, 80-249, 141-167, 200-265 (57-60, 266-334, 325-447)

(41) 3 miles north of Fabre station, railway cut 500 yards south of the railway bridge across Young brook
2 feet leached clay
3¾ feet rather stiff, brown clay, varves 560-782. Thicknesses not characteristic
Probably far to bottom
Series measured: (560-702, 648-782)

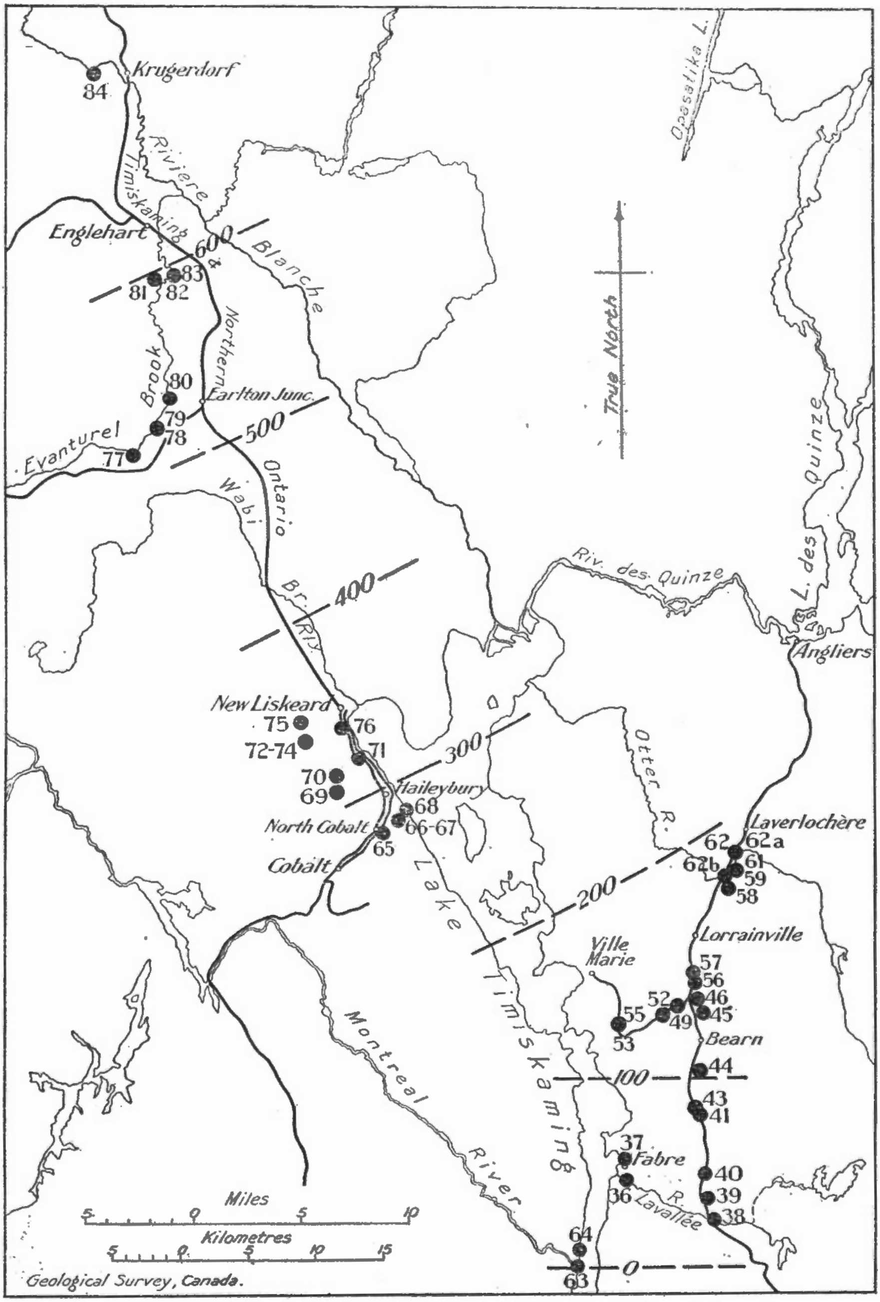

Figure 36. Localities examined in Timiskaming area and position of ice edge every 100 years (localities 36-84).

(42) 500 yards northwest of locality 41, bluff on the north side of Young brook 200 yards west of the railway bridge
10 feet stiff, brown clay with faint varve limits, not measured
4½ inches clay, gradually becoming fat and brown upwards, varves 331-344
19 feet grey silt with thin winter layers, varves 86-330. Varve limits very distinct. The silt is coarse in the lower part, but grows gradually finer upwards. Varve 86 is bottom varve
8 feet till to brook level
Series measured: 107-344 (86-106)

(43) 400 yards northeast of locality 42, railway cut 300 yards north of the bridge across Young brook
2 feet leached clay
4½ feet grey-brown clay-silt, varves 172-344. Fine
Far to bottom
Series measured: 172-344

(44) Railway cut 1¼ miles south of Béarn station
4 feet disturbed clay
5 feet stiff, brown, distinctly varved clay, varves 627-785
Far to bottom
Series measured: 627-720 (721-785)

(45) Railway cut 400 yards south of Gaboury
1½ feet leached clay
5½ feet stiff, brown, beautifully varved clay, varves 913-1088
Very far to bottom
Series measured: 913-1088, 980-1081, 980-1081

(46) Railway cut 100 yards north of Gaboury
3 feet disturbed and leached clay
10 feet stiff, brown, well-varved clay, varves 506-748
Fault. Far to bottom
The clays much faulted in the long section
Series measured: 506-748, 652-740

(47) Railway cut ½ mile west of Gaboury
8 feet sand
17 feet fat, brown, distinctly varved clay, varves 588-943. Thicknesses mostly good, but in some zones locally deformed by pressure
Far to bottom
Series measured: 652-748, 738-851, 785-929 (588-651, 749-862, 929-943, 852-909)

(48) Railway cut 25 yards west of locality 47
3 feet leached clay
4 feet clay, somewhat pressed
Fault
3½ feet light-grey silt with thin, brown winter layers, varves 270-342. Fine
Fault
8 feet sand and silt, 21 varves. Not connected with the normal curve
Not deep to till, which outcrops close by
Series measured: 270-342

(49) Railway cut 50 yards west of locality 47
2 feet leached clay
2¼ feet rather fat, brown clay, varves 401-469
2¼ feet light-grey silt with thin, brown winter layers gradually passing into fat layers in the upper part of the zone, varves 336-400

Fault
1⅝ feet light-grey silt with thin winter layers, varves 200-248
Fault
4½ feet silt with thin winter layers, varves 160-200
Depth to substratum unknown
Series measured: 336-400 (160-200, 214-248, 401-469)

(50) Railway cut ¾ mile west of Gaboury
Several feet leached clay
2 feet 10 inches fairly fat, distinctly varved clay, varves 357-464
Several feet somewhat disturbed varved silt
Depth to bottom unknown
Series measured: 357-400 (401-464)

(51) Railway cut 1 mile west of Gaboury
Several feet somewhat disturbed clay
16 feet sand gradually passing upwards into silt, varves 131-220
Probably not far to bottom
Series measured: (131-183, 159-220)

(52) Railway cut 1¼ miles west of Gaboury
(A)
2 feet leached clay
7½ feet, varves 266-498. Silt with thin winter layers gradually passing upwards into rather fat, brown clay. Good
Far to bottom
Series measured: 266-400, 309-400 (401-498)
(B)
2 feet leached silt
9 feet distinctly varved silt, varves 142-210
Probably not far to bottom
Series measured: 142-210

(53) 3½ miles west of Gaboury, or 2½ miles south-southeast of Ville-Marie, railway cut 100 yards east of the sharp bend of the track
1 foot leached silt
6½ feet, somewhat sandy, light silt, varves 142-165. Varve 142 is bottom varve
Till
Series measured: (142-165)

(54) Railway cut 100 yards west of section 53, at the sharp curve
4 feet leached silt
10 feet sandy, light silt, varves 143-183. Varve 143 is bottom varve
Series measured: (143-183)

(55) Railway cut 100 yards north of section 54
3 feet weathered clay
2 feet silty clay, varves 302-373. Fine
Fault
3 feet light silt with thin winter layers, varves 206-271
9 feet sand and silt, varves 183-205. The thick varves contain lenses of sand
2⅓ feet light-grey silt, varves 174-182. Good
Several feet to bottom
Series measured: 212-265, 244-271, 302-373 (174-211)

(56) Railway cut 1¼ miles north of Gaboury, or 2 miles south of Lorrainville station
2 feet leached clay
4 feet stiff, brown clay, varves 785-936. Fine
Far to bottom
Series measured: 785-887, 806-936

(57) Railway cut 1½ miles south of Lorrainville station

The clay in the greater part of the long section thoroughly contorted and faulted

3 feet leached and disturbed clay
4 feet rather stiff, grey-brown clay, varves 905-1017
Disturbed clay. Far to bottom

Series measured: 905-1011, 920-1017

(58) 3 miles north of Lorrainville station, railway cut 600 yards south of the railway bridge across Otter river

In the long and high cuts south and north of the bridge the clay is practically everywhere thoroughly contorted from top to bottom. The following section, therefore, is compiled from measurements of several undisturbed clay blocks or horizons

3 feet contorted clay-silt with rather thick varves; evidently coming from a considerably lower horizon
5 feet clay-silt, not measurable
3½ feet fairly stiff, brown clay, varves 1059-1226
Contorted zone
8½ feet rather stiff, distinctly varved clay, varves 749-979.
3 feet clay, 65 varves. Thicknesses deformed by pressure
Far to bottom

Series measured: 749-979, 749-979, 749-936, 749-828, 749-814, 798-936, 1059-1157, 1097-1195, 1097-1195, 1106-1174 (1196-1226)

(59) Railway cut 400 yards north of locality 58, 200 yards south of the bridge across Otter river

3 feet leached clay
10 feet rather stiff, grey-brown, well-varved clay, varves 870-1217. Varve 1024 is crumpled up—2 inches. Above 1109 is a slidden zone of 3 inches, but instead of varves missing 8 varves overlap.
Disturbed clay. Substratum at considerable depth

Series measured: 870-1195 (1196-1217)

(60) Railway cut 150 yards north of locality 59, 50 yards south of the bridge across Otter river

3 feet leached clay
4 feet stiff, brown clay, varves 1029-1157. Good. At varve 1081 a fault, but no varve missing
2 feet slidden zone, varves 990-1028 missing
2 feet distinctly varved, brown clay, varves 930-989. At varve 1040 a slidden zone, but no varve missing
2 inches contorted clay, varves 920-929 missing
1¾ feet distinctly varved, brown clay, varves 870-919
Disturbance. Far to bottom

Series measured: 870-919, 930-989, 1029-1157

(61) Railway cut 250 yards north of the railway bridge across Otter river

The clay is very much contorted, so that the profile is compiled from measurements in three undisturbed blocks of clay

4 feet leached and contorted clay
4½ feet stiff, distinctly varved, brown clay, varves 1036-1194
Contorted clay. Far to substratum

Series measured: 1036-1087, 1088-1005, 1106-1194

(62) Railway cut $\frac{3}{4}$ mile north of the railway bridge across Otter river, or $1\frac{1}{2}$ miles south of Laverlochère station

The clay is considerably faulted, so that the following profile is compiled from measurements of four different clay blocks

3 feet leached and contorted clay
3 feet 2 inches clay, somewhat silty in the lower part, but passing into stiff brown clay upwards, varves 360-464.
9 feet light silt, coarse at the bottom but becoming fine upwards, and thin winter layers, varves 185-359. Varve 185 is bottom varve
Bedrock

Series measured: 222-335, 295-400, 295-335, 302-344, 309-354, 374-400 (185-217, 196-221, 236-274, 240-279, 291-313, 302-354, 400-420, 401-434, 421-464)

(62a) Clay pit on the east side of the track, 200 yards north of locality 62
Many feet clay, largely covered
$2\frac{1}{2}$ feet stiff, grey-brown clay with distinct lamination, varves 687-744
$1\frac{1}{4}$ feet clay with three slidden zones, varves 664-686
3 feet stiff, grey-brown clay, varves 601-663
7 inches fairly coarse sand, drainage varve, varve 600
Disturbed zone, covered. Far to substratum

Series measured: 600-661, 644-663 (687-744)

(62b) $2\frac{1}{2}$ miles south-southwest of Laverlochère station, slide in the north bank of Otter river, just east of the highway bridge
3 feet weathered and disturbed clay
11 feet rather fat clay, beautifully laminated by alternation of layers of light silt and brown clay, varves 440-640.
$1\frac{1}{2}$ feet clay, at the bottom lean, but becoming greasy upwards, varves 395-439
10 feet light silt with very thin layers of clay, varves 203-394. Good
2 feet sandy, lens-bedded silt, varves 192-202
10 feet sandy silt. Varves indistinct
12 feet till to river level

Series measured: 220-640, 328-480, 401-453, 433-544, 604-627 (192-219)

Localities on the Western Side of Lake Timiskaming

(Figure 36; Plates VI, VII; cf. pages 126-128)

(63) 24 miles south-southeast of Haileybury, bluff at the mouth of Montreal river, on the west side
8 inches sand
1 foot leached clay
$1\frac{1}{2}$ feet stiff, brown, distinctly varved clay, 50 varves. Not connected with the normal curve
$1\frac{3}{4}$ feet fairly stiff clay, 80 varves. Varve limits difficult to distinguish. Not connected with the normal curve
2 feet sand-silt in lenses, some 30 varves. Varve limits not surely distinguishable
8 inches clay, 18 varves. Thicknesses not good
21 feet light-grey silt, very coarse at the bottom but becoming finer upwards, and thin winter layers, varves 1-191. Good. Varve 1 is bottom varve
25 feet till down to river level

Series measured: 1-191

(64) $\frac{1}{2}$ mile north-northeast of profile 63, slide above the north farmhouse
Many feet silt and clay, mostly covered
10 feet light silt, with thin winter layers, varves 110-271. Good
About 20 feet to till

Series measured: 110-271

(65) 1¾ miles south of Haileybury, roadcut 300 yards southeast of North Cobalt station

4 feet stiff, brown clay, varves 702-905. Good
Far to bottom

Series measured: 702-905

(66) 1½ miles south-southeast of Haileybury, bluff on Farr creek ½ mile from its mouth and 50 yards above the waterfall

Many feet covered and disturbed
4 feet clay-silt becoming rather fat upwards, 68 varves. Not connected with the normal curve
1 foot till
Bedrock

(67) 150 yards east of locality 66, slide on the south side of the creek, below the waterfall

Many feet covered
7½ feet rather stiff, by light silt and brown winter layers distinctly varved clay, varves 601-717
Covered. Far to substratum

Series measured: 601-717

(68) 1 mile south of Haileybury (station), old brick-yard near the mouth of the brooklet at the south edge of the town

1 foot leached clay
30 feet rather stiff clay, varves 698-1135. Varves consist of light silt and brown clay. Fine
3 inches slidden zone, varves 672-697 missing
5½ feet fat, blue clay, varves 580-671. Fine
Depth to substratum unknown

Series measured: 601-671, 601-1135, 803-858, 871-928, 954-989, 968-1048, 580-600, 996-1032

(69) 2¼ miles west of Haileybury, cut on the Fleming Corner road at the first brooklet

3 feet leached clay
3 feet stiff, brown clay, 115 varves. Thicknesses not good
5½ feet rather stiff, brown clay, varves 635-785. Fine
Disturbed zone. Depth to substratum unknown

Series measured: 635-785

(70) 2¾ miles west-northwest of Haileybury, bluff on Dickson creek, 300 yards north of the junction of the brooks

Several feet contorted and covered
4½ feet coarse silt with thin winter layers, 59 varves. Not connected with the normal curve
Till

(71) 2 miles north-northwest of Haileybury, bluff on the north side of Dickson creek just west of the railway track

4 feet disturbed clay
12 feet rather stiff, brown clay, varves 580-742. Fine
Clay somewhat disturbed. Depth to substratum unknown

Series measured: 601-742 (580-600)

(72) 2½ miles southwest of New Liskeard, bluff on the south side of Wabi creek 500 yards west of the sharp bend towards the north

Surface of erosion
4 feet leached clay
4 inches granulary clay, drainage layer
2 inches varved clay
8 inches granulary clay, drainage
5 inches silt with thin winter layers. Varves up to an inch in thickness
1⅓ feet granulary clay with small balls of clay, drainage
8 inches granulary clay, distinctly limited upwards; another drainage
4 feet silt with thin winter layers
4 feet strongly contorted, stiff clay
1 foot slightly disturbed clay
5¾ feet rather stiff, brown clay, varves 1036-1195. Good
1⅓ feet thoroughly contorted clay
1⅓ feet pressed clay
3 feet covered to creek level. Depth to substratum unknown

Series measured: 1036-1195

(73) 500 yards east of locality 72, bluff on the west side of Wabi creek at its junction with the brook

Many feet covered and contorted clay
1 foot 2 inches stiff, brown clay, varves 1123-1182. Good
Fault. Varves 1106-1122 missing
2¾ feet stiff, brown clay, varves 1037-1105. Good
3 inches slidden zone, 2 varves missing
8 inches stiff, brown clay, varves 1018-1034
Fault
14 inches clay, varves 996-1022
3 feet covered to creek level. Depth to substratum unknown

Series measured: 1018-1034, 1037-1105, 1123-1182 (996-1022)

(74) 200 yards east of locality 73, slide on the east side of Wabi creek

2 feet leached silt
1¾ feet rather lean clay, 48 varves
3 feet lean clay, somewhat slidden. Varves up to ⅜ inch thick
7 feet fat, disturbed clay
2 feet clay. Varves so thin that they cannot be surely measured
2 feet fat, blue clay, varves 1186-1270. Good
2 feet pressed clay
15 feet covered to creek level
Far to bottom

Series measured: 1186-1270

(75) 2 miles southwest of New Liskeard, bluff on the north side of Wabi creek where its course is westerly, half-way between the two great bends

10 feet clay, partly disturbed. Not measured because of lack of time
4½ feet rather fat, brown clay, faulted at varves 1024 and 1063. Varves 990-1096
25 feet covered to creek level. Far to bottom

Series measured: 990-1096

(76) 1¼ miles south of New Liskeard, slide on the west side of the railway track, where the trolley line leaves the railway track

2 feet leached clay
2⅓ feet rather stiff, brown clay, varves 793-831. Good
Far to bottom

Series measured: (793-831)

(77) 15 miles north-northwest of New Liskeard, 4 miles southwest of Earlton Junction, 1 mile north of McCool, roadcut on the south side of Evanturel creek

Surface of sedimentation
6½ feet extremely fat clay, not measured
22½ feet rather stiff clay, varves 887-1270. Varves consist of light silt and brown winter layers. Fine
Many feet clay to creek level. Not measured on account of lack of time. Far to bottom

Series measured: 887-1270, 1035-1121, 1145-1185

(78) 2½ miles southwest of Earlton Junction, slide on the east side of Evanturel creek 100 yards north of the sharp bend towards the north

Many feet covered
10½ feet fairly stiff, distinctly varved clay, varves 608-733
Creek level. Depth to bottom unknown

Series measured: 608-733

(79) 400 yards north of locality 78, 500 yards south of the east-westerly road, slide on the east side of the creek

Many feet covered
3½ feet fairly fat clay, varves 671-711. Varves consist of light silt and brown winter layers. Fine
6 inches contorted clay, varves 653-670 missing
4 feet rather fat clay, varves 608-652. Fine
Disturbed zone at creek level. Depth to substratum unknown

Series measured: 608-652, 671-711

(80) 1¾ miles west of Earlton Junction, bluff on the east side of Evanturel creek 200 yards north of the east-westerly road

Several feet contorted clay
5 feet rather stiff, distinctly varved clay, varves 1061-1104

Series measured: 1061-1104

(81) 2¼ miles south of Englehart, bluff on the south side of Englehart river ¾ mile west of the bridge across the Evanturel, just where the Englehart turns towards the north

Plain of sedimentation
20 feet mostly covered, but probably varved silt all through. Not measured
10 feet grey silt and thin, dark winter layers, very lean, probably varves about 1670-1750
4¾ inches silt—2 varves or 16 varves
4 inches lean silt, 11 varves
3¾ feet chocolate-brown clay, exceedingly fat in the lower part but gradually becoming lean upwards, probably varves 1526 to about 1640
5 feet clay, very fat and chocolate-brown in the lower part but growing leaner and grey-brown in the upper part, and thus distinctly limited against the overlying clay, varves 1323 probably to 1525. About 50 varves in this zone are likely slidden away, though no disturbance was observed when the measuring was made
About 10 feet mostly covered, not measured
9½ feet silty clay, 79 varves. Winter layers and summer layers about equally thick. Silt often forms lenses, so that thicknesses not characteristic. Not connected with the normal curve
1 foot contorted clay
8½ feet silty clay, varves 980-1064. Some silt in lenses, so that thicknesses not always good
4 inches disturbed clay—2 varves

36 feet rather stiff, distinctly varved clay, varves 667-977. Varves consist of light silt and dark-brown, stiff clay; both layers of about equal thickness. Silt in some cases forms lenses, but on the whole the thicknesses of the varves are characteristic

River level. About 10 feet to bedrock—limestone—which outcrops in the river bed close by

Series measured: 1323-1400 (667-977, 980-1064, 1401 with gaps to about 1750)

(82) 2¼ miles south of Englehart, bluff on the west side of Evanturel creek 300 yards west of the bridge

Surface of sedimentation

6 feet leached silt

5⅓ feet light-yellow silt with thin winter layers, 94 varves (Plate IX)

6 feet clay-silt, partly disturbed and mostly covered

3 feet contorted and faulted stiff clay

11 feet exceedingly fat—no silt—chocolate-brown clay, varves 1175-1389

66 feet clay-silt, varves 594-1174. One-half of the varves consists of light silt and one-half of fat, brown clay. Lamination exceedingly sharp. The silt chiefly occurs in lenses or is entirely squeezed out, so that thicknesses in many cases not good. Varve 594 bottom varve

4 feet till

Limestone

Series measured: 1175-1389 (594-1174, 500-626, 602-646, 700-722, and 94 varves at the top)

(83) 200 yards east of locality 82, bluff on the east side of the creek. 200 yards south of the bridge

Surface of erosion

1 foot leached clay

31 feet distinctly varved clay, varves 668-960. One-half of the varves consists of light silt, one-half of very fine, brown clay. Varves often wavy, so that thicknesses not reliable and in many cases not good

3 inches disturbed clay—2 varves

6½ feet clay-silt, varves 625-665

3 inches contorted clay, 1 varve missing

2⅔ feet silt-clay, varves 607-623

1½ feet contorted zone—2 varves

2 feet 2 inches silt-clay, varves 592-604. Varve 592 is bottom varve

2 feet till

Limestone

Series measured: (592-604, 607-623, 625-665, 668-960)

(84) 7½ miles north-northwest of Englehart, 2 miles west of Krugerdorf, bluff on the east side of Aidie creek 100 yards north of the east-westerly road

1½ feet leached, very thinly varved, brown clay

6 inches leached clay. Average thickness of varves ½ inch

3⅔ feet clay-silt, varves 1112-1164. Varves consist of light-yellow silt and brown winter layers. Good

20 feet clay-silt, mostly covered, not measured

6⅓ feet silt-clay, 47 varves. Not connected with the normal curve

6 inches contorted zone

10 feet sandy, distinctly varved silt, 20 varves. The lowermost varve 4 feet thick. Not connected with the normal curve

Creek level. Probably not far to substratum

Series measured: (1112-1164)

Localities North of the Height of Land

(Figure 37; Plates VII, VIII; cf. pages 127, 128)

(85) 7 miles southeast of Matheson, $2\frac{1}{4}$ miles southeast of Vimy Ridge, bluff on the tributary from the southwest to Black river, on the west side 200 yards south of the mouth

5 feet covered and leached
1 foot stiff, yellow to yellow-brown clay, varves 1528-1536
$1\frac{1}{3}$ feet very fat, yellow clay, varves 1490-1527
3 feet contorted clay—sand and very fat, yellow clay kneaded together
5 feet covered
4 feet contorted, sandy silt-clay, with scattered pebbles. The zone may indicate grounding of an iceberg
27 feet silt-clay, varves 1205-1416. One-half of the varves consists of light-grey silt, and one-half of dark grey-brown, fat clay. Fine
1 inch disturbed zone
7 inches silt-clay, 2 varves
5 inches contorted zone
7 feet silt, 17 varves
3 feet covered to creek level
Far to substratum, though not many varves. 50 yards to the south: lowermost, many feet sand with varves more than 3 feet thick and above a disturbed series of varved silt, all of which comes underneath the series measured

Series measured: 1205-1400, 1229-1292, 1490-1527 (1528-1536)

(86) $5\frac{1}{4}$ miles southeast of Matheson, $\frac{3}{4}$ mile southeast of Vimy Ridge, road-cut on the west side of the railway, at the brook

3 feet leached clay
$2\frac{1}{2}$ feet rather stiff distinctly varved clay, varves 1326-1368
3 feet faulted clay
Far to substratum

Series measured: 1326-1368 (1310-1350)

(87) 6 miles southeast of Matheson, $1\frac{1}{2}$ miles southeast of Vimy Ridge, bluff on the east side of Black river 300 yards north of the fall

Many feet covered
7 feet clay-silt, varves 1222-1291. One-half of the varve consists of light silt, and one-half of fat, brown clay
12 feet covered to river level. Far to substratum

Series measured: 1222-1291

(88) $6\frac{1}{2}$ miles east-southeast of Matheson, bluff on the south side of Pike river, 700 yards west of the bridge, at the sharp bend of the river towards the west

Plain of sedimentation
4 feet leached and contorted clay
$4\frac{1}{2}$ feet rather stiff clay, varves 1319-1392. One-third of the varve consists of grey-white silt, and two-thirds of fat brown clay. Good
5 feet repeatedly faulted clay. Partly not connected with the normal curve
About 40 feet covered to river level

Series measured: 1319-1392 (1295-1334)

(89) Bluff 300 yards east of locality 88

4 feet leached and disturbed clay
$5\frac{1}{2}$ feet clay, varves 1260-1324. Good, although faulted
Some 50 feet covered to river level

Series measured: 1260-1324, 1289-1324

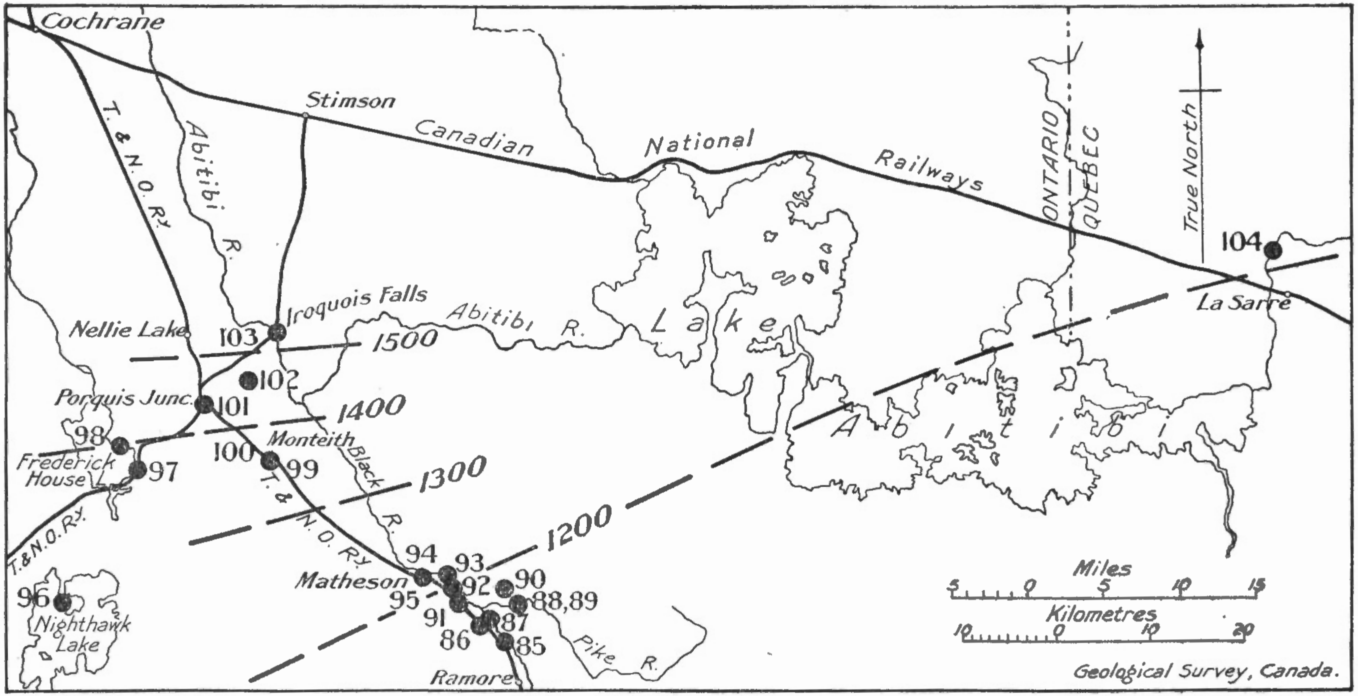

Figure 37. Localities examined in northern Ontario and position of ice edge every 100 years (localities 85-104).

(90) 5 miles east-southeast of Matheson, 3½ miles east of Belleek, ditch on the north side of the road near the brooklet

6 inches sandy loam
6 inches very fat clay
1 foot sandy clay-silt
1 foot stiff, thinly varved clay } Not measured
1½ feet very stiff clay. Varve limits difficult to distinguish. Probably varves 1711 to about 1817
1½ inches sand
8 inches slidden clay
9 inches extremely fat clay, varves 1679-1696
Fault
9 inches fat clay—varves 1681-1695
4 inches disturbed zone; no varve missing
Depth to substratum unknown

Series measured: (1662-1680, 1679-1696, 1681-1695, 1711 to about 1817)

(91) 2¾ miles southeast of Matheson, 1 mile south-southeast of Belleek, bluff on the west side of Black river at the river bend just east of the railway

The following series is compiled from two measurements near each other

3 feet leached clay
2 feet very fat, red-brown clay, 41 varves. Not connected with the normal curve
3½ feet sand and very fat clay kneaded together, probably by grounding iceberg
10¾ feet very fat, grey-brown clay, varves 1528-1709. Varves about 1 inch thick. Good
1½ feet very fat, brown, thinly varved clay, varves 1486-1527
Disturbed zone. 6 feet above river level, partly covered. Far to substratum

Series measured: 1486-1709, 1493-1522, 1573-1596

(92) 2¼ miles east-southeast of Matheson, slide on the east side of Black river, directly opposite Belleek

3 feet leached clay
3 inches exceedingly fat, red-brown clay, 6 varves
3 inches clayey silt. One or several varves
4½ inches stiff, red clay, many varves
1½ inches silt
1¾ inches clay, several varves
1⅝ inches silt and clay layer, 1 drainage varve, varve 1911
4 inches fairly fat clay, varves 1894-1910. Varve limits difficult to distinguish
11 feet chocolate-brown clay, mostly exceedingly fat, but some zones only moderately fat, varves 1681-1893. Varve 1711 drainage varve
15 feet covered to river level

Series measured: 1681-1911, 1681-1708

(93) 2 miles east of Matheson, slide on the east side of Black river, 400 yards south of the river bend towards the west

3 feet leached clay. One foot below the top 2 drainage varves, the lower one of which is the thickest
8½ feet clay, varves 1273-1426. One-half of the varve consists of light silt, one-half of brown clay. Fine
Many feet to till, mostly covered. About 60 feet to river level

Series measured: 1273-1400, 1341-1384 (1401-1426)

(94) ½ mile east of Matheson, bluff on the west side of Black river

Many feet covered and disturbed
4½ feet silt-clay, probably varves 1229-1246
Slidden and covered. 5 feet to river level
Probably not far to bottom

Series measured: (1229?-1246?)

(95) 400 yards south of Matheson, excavation for a house on the north side of the kettle lake

2 feet leached clay
6½ feet silt-clay, different varves slightly disturbed, perhaps varves 1212-1247. Varve 1212 more than 2 feet thick and may be bottom varve
Large boulders; between them sand

Series measured: (1212?-1247?)

(96) 16½ miles southwest of Porquis Junction, Nighthawk lake, bluff 300 yards east of the most westerly point of the peninsula 2½ miles south of the outlet—Frederick House river—or ¾ mile northwest of Peninsula gold mine

Plain of sedimentation
2 feet leached clay
1½ feet silt-clay, varves 1528-1553. Somewhat more than one-half of the varve consists of light silt, the rest of brown clay. Good
2 feet 2 inches very fat—hardly any silt—brown clay, varves 1431-1527
5⅓ feet clay-silt, varves 1353-1430. One-third to one-half of the varve consists of light-yellow silt, the rest of fine, brown clay
6 feet sandy silt, contorted, 20 to 25 varves
20 feet fine sand
Lake level. Probably not far to substratum

Series measured: 1353-1553, 1353-1396, 1404-1413, 1423-1491, 1512-1545 (1480-1511)

(97) 6 miles southwest of Porquis Junction, railway cut ½ mile south of McIntosh Springs

1 foot leached clay
3½ feet silt-clay—white silt and brown clay—varves 1389-1414
Some 20 feet sand and quicksilt down to the level of the track. Probably only a few varves. Depth to substratum unknown

Series measured: (1389-1414)

(98) 6 miles west-southwest of Porquis Junction, 1½ miles northwest of McIntosh Springs, slide on the abandoned shore of Frederick House lake whose south half was drained through diverting of Frederick House river in 1909 (Knight, etc., 1919, page 42)

5 feet disturbed and leached clay
3⅔ feet silt-clay, varves 1528-1585. One-half to two-thirds of the varve consists of light silt, the rest of fat clay
7 feet clay, varves 1409-1527. Varves up to 3 inches thick in the lower part, but rapidly decreasing upwards to become ¼ to ⅛ inch in thickness. The thick varves consist of light-yellow silt and grey-brown clay, the thin varves, almost exclusively of stiff clay. Good
10 feet above the ancient lake level. Probably about 10 to 15 varves to bottom

Series measured: 1412-1585, 1437-1467 (1409-1411)

(99) 6½ miles southeast of Porquis Junction, 600 yards south-southeast of Monteith, gravel pit east of Driftwood river

2 feet weathered clay
8 inches clay-silt, varves 1528-1541
7 inches stiff, thinly varved clay, varves 1494-1527. Varve 1494 is bottom varve
Gravel, exposed 5 feet, probably deep

Series measured: 1494-1541, 1505-1537

(100) 6 miles southeast of Porquis Junction, railway cut ½ mile northwest of Monteith on the east side of the track

1½ feet very fat, red-brown, varved clay
⅝ inch silt-sand, drainage varve
⅝ inch clay, a few varves
$\frac{3}{16}$ inch silt-sand, drainage varve

1 foot leached and faulted clay. Varves exceedingly thin in the upper part
1½ feet clay-silt, varves 1546-1583
Fault, varves 1532-1545 missing
1½ inches clay-silt, varves 1528-1531
6 inches stiff, thinly varved clay, varves 1493-1527. Varve 1493 is bottom varve
10 feet coarse sand down to the rails. Depth to bottom unknown

Series measured: 1493-1531, 1493-1531, 1546-1583, 1546-1572

(101) ½ mile south of Porquis Junction, ditch on the east side of the railway track (Plate IX)

3 feet weathered clay
1½ feet fat, brown clay, 15 varves. The lowest varves $\frac{3}{16}$ inch thick, the upper varves up to 3¼ inches thick. Not connected with the normal curve
1½ inches clayey sand
¾ inch clayey silt
¾ inch sand and clay
1¾ inches clay and sand
} Varves not distinguishable
9 inches very fat, blue, thinly varved clay, 39 varves. Not connected with the normal curve

(102) 3¼ miles east-northeast of Porquis Junction, and 1¾ miles south of Onagon, bluff on the south side of the brook

50-60 feet sediments, covered
7¾ feet silt-clay—grey-white silt and grey-brown clay—varves 1554-1646. Fine. 3 feet below brook level. Depth to substratum unknown

Series measured: (1554-1646)

(103) ½ mile north of Iroquois Falls, cut south of the railway bridge across Abitibi river (Plate I a)

(A) 100 yards south of the bridge

10 feet stiff, contorted clay
1½ inches clayey silt, disturbed
2½ inches clay-silt, varves 1899 (?)–1904 (?)
3⅜ feet exceedingly fat, dark-brown, thinly varved clay, varves 1644-1898 (?). Varves 1770-1840 contain minute white concretions, so that the clay is spotted
5 feet rather fat clay, varves 1554-1643. One-half of the varve consists of light silt and one-half of dark grey-brown clay. Fine
Covered. Depth to substratum unknown, but probably not great

Series measured: 1554-1710, 1554-1710, 1560-1591, 1569-1591, 1569-1591, 1611-1634, 1651-1710 (1711-1904, 1711-1870, 1711-1730)

(B) 50 yards south of the bridge (Plate IX)

Surface of erosion
5 feet disturbed and leached clay
4½ feet silt, distinctly varved, 28 varves from bottom, not connected with the normal curve
1 inch gravel
Bedrock

(104) 3 miles north-northwest of La Sarre, slide on the west side of La Sarre river at the southernmost rapid

3 feet weathered clay
4½ feet exceedingly fat, chocolate-brown clay, varves 1528-1577. This zone differs from the underlying one only by greater thicknesses of the varves
12 feet clay, varves 1234-1527. In the lowermost part the varves consist of a thin layer of light-yellow silt and a thick layer of grey-brown, fat clay. The clay becomes finer upwards, and from about varve 1360 is exceedingly fat and chocolate-brown in colour. Good
5 inches quicksilt, varves 1231-1233
13 inches silt-clay, varves 1219-1230
½ foot coarse silt, varves 1216-1218
2 feet sand, varves 1210-1215. Varve 1210 is bottom varve
5-6 feet till
Bedrock

Series measured: 1210-1550, 1483-1502, 1525-1541 (1551-1577)

CHAPTER XII

NORMAL CURVE

METHOD OF CONSTRUCTION

The curves on Plates IV to IX give as accurately as the material permits the average thickness and number of the annual clay layers formed during parts of the ice retreat from Bracebridge to Iroquois Falls, Ontario. They have been constructed in the following manner. All individual curves were first matched and corrected for number of varves. If, for example, out of three measurements two agreed, but one had one varve less or more than the others, the exact location of the mistake was determined and the curve corrected by dividing one varve in two or uniting two varves in one, so that this curve agreed with the two others. Then the curves or such parts of them as included undisturbed varves of normal variation and thickness were selected for constructing the normal curve, and those curves were discarded that showed great difference in thickness from the majority or poor agreement in the shape of the curve. The normal curve was constructed from the selected individual curves by calculating the average thickness of each single varve. When measurements of the same series were at hand from distant or differently situated parts of the same lake basin or from different regions, separate normal curves were worked out for each region. Varves abnormal at all localities have been marked by broken lines.

The curves were drawn so as to show the actual thickness of the varves. The distance between the vertical lines was 5 mm. In the plates the curves have been reduced one-half.

The varves in the different groups have been independently dated, as they have not been connected. The large-sized figures at the arrows indicate the number of the measurements used in constructing the curves, and the small-sized figures give the respective localities. The figures in brackets mark profiles used for control of the number of the varves, but not otherwise. The arrows show to what part of the curve the figures refer.

RELIABILITY AND SIGNIFICANCE OF THE VARVE CURVES

In the following a valuation of the curves, Plates IV to IX, based upon the character of the material, the agreement between the curves covering the same period, etc., is made. The causes of the change of the character of the clay are discussed and individual varves of special significance are described.

The individual curves on Plate IX and those from localities 1 to 8 are briefly treated under the description of the localities, pages, 95-98, etc.

Bracebridge, Localities 9-16, *Plate IV*

(1-7) Thicknesses not reliable, since the varves were deposited too close to the ice edge.

(8-26) Fairly good, though the variations are somewhat too great, because of the nearness of the ice and because of the position of the localities in the course of water current coming from the melting ice.

11, 12, 24, and 26 contain sand or coarse silt.

(27-200) Rather good, though the variations are partly exaggerated and parts of series B are not reliable. Profiles 12 and 13 agree very well, only that the varves at the latter are thicker. Profiles 9 and 11 agree well in some parts, not so well in other parts.

34, at locality 13, contains sand and is three times as thick as at locality 12.

43, at locality 12, consists of $\frac{7}{8}$ inch sand and $2\frac{1}{4}$ inches fairly fat clay. It records a drainage which hardly affected locality 12 where the varve is $2\frac{3}{8}$ inches thick.

52, 79, and 107, series A, are drainage varves. Varve 52, at locality 13, is $5\frac{1}{2}$ inches thick. Varves 79 and 107 are normal at locality 10 which, together with the fact that drainage varves at locality 10 are not recorded at locality 12, shows that the valleys were fed by different streams.

82, 98, 130, and others, though very thick in series A are also thick in the other valley and, therefore, perhaps record years of unusual ice melting.

159 and 163, series A, may be abnormally thick.

107, 125, and 166, series B, are too thick. Varves 125 and 166 peculiarly enough are thin at locality 112.

64, 78, 181, and 196, series C, most probably record drainages, the corresponding varves being thin in Muskoka River valley.

(201-400) Rather good. The varves that are too thick have been indicated by broken lines. Most of these varves occur in curves A and B from the main Muskoka valley.

223 is thick at localities 9 and 11 only. The drainage must have occurred west of locality 12.

275, particularly in curve A, is remarkably thick, but since it is also thick at locality 10 it perhaps records a year of unusual melting.

(401-600) Not good. A number of varves indicated with broken lines are too thick. In some cases the varves are thick in curves A and B, in B and C, in some cases again only in one of the curves. The thick varves may be due to different causes as drainage, direction of current, and of wave-action, as the bays now had become shallow by sedimentation. At locality 16 the thin varves are somewhat thinner than at localities 13 and 14, but the thick varves are thicker. Several varves, particularly 411, 425, 437, 503, 534, and 542, contain sand and are considerably thicker than at the two other localities $1\frac{1}{2}$ miles farther down the valley.

508, 541, and 557 especially, record drainages at localities 10 and 11, but not at 12 and others farther north.

(601-800) A valuation cannot very well be made, since only one curve is at hand. Varves containing sand, and thus probably too thick, have been marked by dashing.

Huntsville, Localities 17 *and* 18, *Plate IV*

Since the two localities, only separated by an island, occurred in the same bay of the ancient Lake Algonquin, their clay varves, as could be expected, agree well. The character of the clay indicates that the sedimentation took place undisturbed, and the thicknesses, with few exceptions, may form a true record of the ice melting.

25, especially at locality 18, is very thick. It occurs, however, near the bottom.

33, at locality 17, had been overlooked in the measuring.

43 and 44, at locality 17, contain sand.

283, 295, and 356, at locality 18, and 356 at locality 17, contain coarse silt and perhaps are abnormally thick.

Burks Falls, Localities 19 and 20, Plate V

Only one series is at hand, but this may on the whole form a trustworthy record of the ice melting and the climatic conditions, judging from the character of the clay. A few varves contain silt or fine sand and have been marked by dashing, since they perhaps are abnormally thick.

225-252 are somewhat obscure and some varve limits may have been overlooked. In the zone next above the varve limits are too indistinct for a sure measurement. At the top of this zone is a disturbed layer $1\frac{1}{2}$ inches thick. These uncertain zones have arbitrarily been supposed to represent varves 253-310.

311-329 are somewhat uncertain, the varve limits being difficult to distinguish.

Powassan, Localities 21-25, Plates V and IX

PLATE V

(1-7) Deposited too close to the ice edge to be characteristic.

(8-103) Very good.

(104-162) Series A very good, series B fairly good. The lower varves in series B are rather thick, as being deposited close to the ice edge. The agreement between the two curves is fairly good.

104 and 105, at locality 23, are too thick.

112-115 do not agree in the two curves.

138 in curve B may be abnormally thick.

(163-200) Not very good. The separate measurements agree well in some parts, not well in other parts.

PLATE IX

(201-220) Not very good.

(221-227) Poor.

222 contains silt and is too thick.

(228-256) Very poor. Thicknesses valueless and even the number is uncertain.

(257-272) Fairly good.

263 contains sand and is abnormal.

(273-284) Not very good.

(285-301) Good.

(302-316) Varves consist of sandy silt and are too thick because of shallowing (cf. page 15).

Sudbury, Locality 30, Plate IX

(1-77) Only a few varves may have characteristic thicknesses.

1-19 were deposited close to the ice edge, and, therefore, are not reliable.

20-46 or 47 indicate a remarkable increase in the supply of material, which probably was due to drainage of a greater area into this part of Lake Algonquin.

53-55 are abnormally thick.

The varves below 66 consist largely of silt. Varve 66 begins a series of fat, rather thinly varved clay. Thus, the main current from the ice may have been diverted by some topographic obstacle to one side.

70 contains silt.

Espanola, Locality 33, Plate V

Locality 33, Espanola, probably became free from ice somewhat before the opening of the Trent Valley outlet (cf. page 60). The lower part of the profile was deposited, when the ice border was in the vicinity. The decreasing thicknesses of the varves indicate increasingly distant positions of the ice edge. The homogeneous clay bed may date from the time when the ice was a few tens of miles away and the water depth was about 500 feet. The gradual increase of the annual deposition, so that the varves became distinguishable and later fairly thick, may have been due to the approach of the shore as Lake Algonquin was emptied. Although the drainages during the years 286 and 288 evidently took place far away, as no coarse material was transported to the place, they appear to have been accompanied by increase of the area drained to this part of the lake. The circumstances of the drainage are not clear, but it was perhaps a drainage of a lake isolated by a barrier of drift as Lake Algonquin fell.

The 200 or so top varves may have been deposited in the Nipissing Great Lakes, whose shore-line in the region now stands at about 670 feet (Goldthwait, 1910, Figure 4, page 41). The altitude of the surface of the deposit is not exactly known but is probably about 650 to 660 feet, since the level of lake Huron is 581 feet, the waterfall at Espanola 62 feet, and the sandy top beds are somewhat above the river level above the falls.

It seems probable, therefore, that the homogeneous clay in the middle of the deposit represents a very long time and that no, or almost no, material was brought to the place for thousands of years. It has, however, not been thought advisable to date the varves according to this interpretation.

There is, however, one part which perhaps throws doubt upon the above interpretation. The uplift that put an end to the Kirkfield stage and caused the sea to fall to the very bottom of Ottawa valley (bed 4), should have affected Espanola region, but does not seem to be recorded in the profile by change in deposition, erosion, or by other features. This might mean that the district was uncovered first after the end of the Kirkfield stage.

(1-14) The great thicknesses of these varves may be due to the fact that they were deposited close to the ice edge.

(15-200) Judging from the character of the clay this series may be very good, though there is no other series that can serve as a check.

18 contains sand.

38 records a considerable drainage, and 87 a small drainage.

(201-340) Varves too thin to trustworthily record ice recession and climatic conditions. The varve limits, however, are fairly distinct, so that the number of the varves may be correct.

219, 267, 286, and 308 contain coarse silt and may be abnormally thick.

(341-400) Besides being entirely too thin, these varves are somewhat uncertain, since the varve limits are very difficult to distinguish.

Above varve 400 the varve limits become more and more obscure and finally become indistinguishable. Measuring and counting of the distinguishable varves and estimation of the varve number in the homogeneous zone give a total of between 700 and 800 varves. Because of this uncertainty in the number of the varves the measurable series at the top of the section has been arbitrarily begun with number 1201.

(1201-1281) Too thin to record the ice melting in a reliable manner.

1282, or possibly 1275, begins a series of thicker varves containing coarser material. The remarkable increase in deposition, as mentioned, most likely was due to regression of the shore of Lake Algonquin as the basin was tilted, so that locality 33 came to lie fairly near to the shore of the Nipissing Great Lakes which now came into existence. Deposition increased because of the nearness of the shore, because of concentration of the melt-water from a large area into a land river which may have had its mouth in the neighbourhood of the locality, and because this river picked up considerable quantities of material during its course. Moreover, as deposition must have taken place in basins traversed by the river, the thicknesses of the varves cannot be reliable records of the ice melting. Since, however, the amount of melt-water may largely have determined the amount of erosion and deposition, the thicknesses, although too great, may register the ice melting fairly accurately, particularly as the clay appears to have been deposited under quiet and favourable conditions.

1286 and 1288 though considerably thick do not contain more coarse material than the ordinary varves, but have a surplus of fine material. They may record large drainages which occurred so far away that only the finest material was transported to the locality. It is not known where the drainages took place (cf. page 19 and Plate I c).

(1401-1551) Similar to the zone immediately below varve 1401. The varves appear to be good, but it is not known how accurately they register the ice melting because of erosion and deposition by the land river (Plate I e).

1479 and 1515 may each consist of two varves.

Above 1551 and somewhat higher up are contorted zones of unknown numbers of varves. The lower disturbed zone is supposed to represent 3 varves, the upper one, 4 varves.

(1551-1570 and 1575-1600) Besides being influenced by river erosion and by deposition these varves initiate a general upward increase of the thicknesses evidently due to shallowing. The decreasing water depth probably was the result of deposition. The value of the varves evidently is little.

(1601-1800) The great thickness of the varves, the coarseness of the material, and the thin, to exceedingly thin, winter layers, indicate that a large part of the material was derived by wave erosion in shallow water. The winter layers in some zones are not only so thin that they are difficult to distinguish, but were locally eroded away during the spring following their deposition. Thus both the thicknesses and the numbers of the varves are uncertain.

1625, 1638, and 1663 might each be two varves.

1629 and 1630, 1635 and 1636, 1669 and 1670, 1671 and 1672, 1688 and 1689 might each be one varve.

Massey, Locality 34, Plate VI

(1-27) Mostly valueless as far as the thicknesses are concerned, for besides being deposited near the ice edge many of the varves record drainages.

(28-200) Rather good. A few varves contain silt and are perhaps abnormally thick; they have been marked by dashing.

Sault Ste. Marie, Locality 35, Plate VI

The whole series is more or less uncertain, both as to number and thicknesses of the varves. It shows, however, interesting drainages and changes in character and colour of the clay.

(1-42) Fairly good as far as the number of the varves is concerned. From varve 3 to 16 there is a remarkable decrease in the thicknesses and after a sudden increase a similar decrease from varve 17 to 26. These variations may have been caused by sudden increases and gradual decreases of the drainage area of the glacial river.

39 initiates a series of abnormally thick varves.

(43-57) Thicknesses valueless and number of varves somewhat uncertain. Varves 54, 55, and 57 record important drainages.

(58-99) A valuation is difficult. The fluctuations in thickness of the varves are surely very great, but the uniform character of the clay does not indicate drainages. The fluctuations appear to be periodic, judging from the gradual increases and decreases and the equal time-interval between the minima.

(100-120) Three drainages recorded by sand or coarse silt; the thicknesses of the other varves doubtful.

106 and 107 may be one varve.

(121-200) Not very good. Some varves contain silt and may have exaggerated thicknesses, and some varve limits are uncertain.

143 and 144, 156 and 157 may each be single varves.

170 may be two varves.

(201-323) Probably fairly good judging from the uniform character of the clay.

201 and 202 may be one varve.

303-305 were difficult to distinguish.

323 records a drainage.

Timiskaming, Localities 36-104, Plates VI, VII, VIII

(1-60) Based upon one measurement only which represents the lowest part of a section. The series, therefore, is not entirely reliable as regards the thicknesses of the varves.
35-38 and 57-58 are probably abnormally thick.

(61-200) Probably fairly good. The different measurements as a rule present similar fluctuations, but these are often of rather different magnitude because the transportation and deposition of material were influenced by the topography of the ancient lake. These discrepancies have been partly balanced in the normal curve which has a smoother course than the individual curves. It would have been better perhaps to have calculated two or three normal curves of the material rather than one. In spite of the flattening out, the curve shows some long and great fluctuations that may be due to variations in the ice melting. Most pronounced are minima from 39 to 56 and from 79 to 91 and maximum from 92 to 171.

(201-400) Series A, good. The different measurements mostly agree well, though the thicknesses of the varves, especially in the lower part of the curve, are somewhat influenced by the topography. The local character, however, may be largely eliminated through the great number of the measurements.
Series B. Also good.
The two series agree well.
235 and 295 in series B may be abnormally thick.
246 and 296, at 36 and other localities at low levels, contain coarse silt and perhaps are abnormally thick.

(401-505) Most probably good, the lamination being well developed.

(506-600) Both series are rather good, and agree quite well.
518, 542, 590, and 596 in series A, and 520, 523, and 553 in series B, may be too thick.

(601-663) Series A, good. The different measurements agree well, both as regards variations and absolute thicknesses of the varves, because the ice front was far away when the material was laid down. This has reference to locality 69 west of Haileybury, situated about 20 miles from 47, etc., at Béarn and on the opposite side of lake Timiskaming.
Series B, good.
Series C, very good. The different measurements agree excellently, though localities 78 and 79 lie 21 and 22 miles, respectively, north of locality 68.
Considering the fact that the series A, B, and C are based upon groups of measurements so far apart and essentially from opposite sides of the basin their correspondence is remarkable.

(664-800) Series A, very good. The different measurements show good to excellent correspondence, so that the normal curve calculated from a large amount of material may be an accurate record of the ice melting.

Series C, good, as proved by the correspondence between the different measurements and the consistency of the clay. The two series agree well at times, less well at other times. The differences are chiefly due to topographic conditions by which the main current was conducted different ways during different years.

723 and 738 in curve A are thick, in curve C, comparatively thin. It is noteworthy that though thin at locality 68 south of Haileybury they are thick at localities 65 and 69 at higher levels south-southwest and west of the town.

704, 760, 767, 771-773, 778, 784, and other varves, form small maxima in curve A, but not in curve C.

667, 749, and 798, on the other hand, form insignificant maxima in curve C, but not in A.

(801-1000) Series A, very good as computed from a great number of measurements that show excellent agreement with one another.

Series C, also good.

The correspondence between the curves after 830 is very good, for a long series even perfect, but before 830 not good.

803, 815, 820, and other varves, form small maxima in curve A, but not in curve C.

802, 805, 806, 816, and 819 on the other hand form insignificant maxima in curve C, but not in A. Diagram 65 follows curve C somewhat more closely than curve A.

(1001-1200) Series A, good to very good, though partly consisting of rather thin varves. The individual graphs agree very well.

Series C. Largely very good. The individual curves correspond remarkably well, in spite of the great distances between them. Localities 37 and 77 lie 40 miles apart.

The correspondence between the curves on the whole is very good.

1185 begins a long series of exceedingly fat clay without silt. The trapping of the silt north of the divide was perhaps due to a lowering of the water level.

(1200-1400) Curves A and C, years 1201-1270, are continuations of long series and can be connected with each other by means of the material presented, as can also curves B and C, years 1353-1400, and curves D and E. But these different groups of curves do not show sufficient agreement to permit connexions with one another. The connexions made are based upon material collected in 1924. A valuation of the curves and the different types will be discussed when the new material is published.

(1401-1411) Both series good.

(1412-1460) Series B and C are good, series A is fairly good.

The agreement between curves B and C is remarkable considering the fact that the localities lie 85 miles apart in a west-southwest east-northeast direction, that is in the direction of the ice border, and as a consequence were fed independently of each other.

(1461-1527) The varves are too thin to be characteristic, and the curves show only partial agreement.

(1528-1550) Rather good. The curves show some differences, but on the whole agree fairly well. As might be expected from the position of the localities, curves A and C correspond best. Which curve is correct in each case cannot always be said.

(1551-1553) Both series good.

(1554-1585) Series A and C agree fairly well, but series B could be connected with the other two only by means of material collected in 1924. When this material is worked up a more detailed valuation can be made.

(1586-1600) The curves show only partial agreement.

(1601-1710) Curve A is probably fairly good. Curve B, after 1650, has too thin varves to be characteristic. The correspondence is poor.

(1711-1800) Only one measurement is at hand, but this may record the ice melting rather well. The upward gradually decreasing thickness of the varves may largely be due to increasing distance from the ice border. Up to 1793 the clay is exceedingly fat; the varves consist almost exclusively of greasy chocolate-coloured clay. Varves 1793-1821 are moderately fat.

1711 consists of 4 inches quick silt and 10 inches very fat clay. It records a drainage rather far away.

1749 and 1760 contain relatively much silt and may be abnormally thick.

(1801-1911) A valuation is difficult since only one measurement is at hand. Varves 1801-1821 are moderately fat. The general increase in the thicknesses of varves 1822-1847 is remarkable. Since varve 1822 begins a series of exceedingly stiff clay this increased sedimentation may partly at least be due to more rapid flocculation of the fine particles. Partly it may be due to increased drainage area or increased ice melting. The subsequent decrease in varve thickness perhaps is due to decreased melting. About varve 1885 the clay begins gradually to become moderately lean.

1890 and 1899 contain relatively much silt and may be too thick.

1911 consists largely of silt and records a drainage.

CHAPTER XIII

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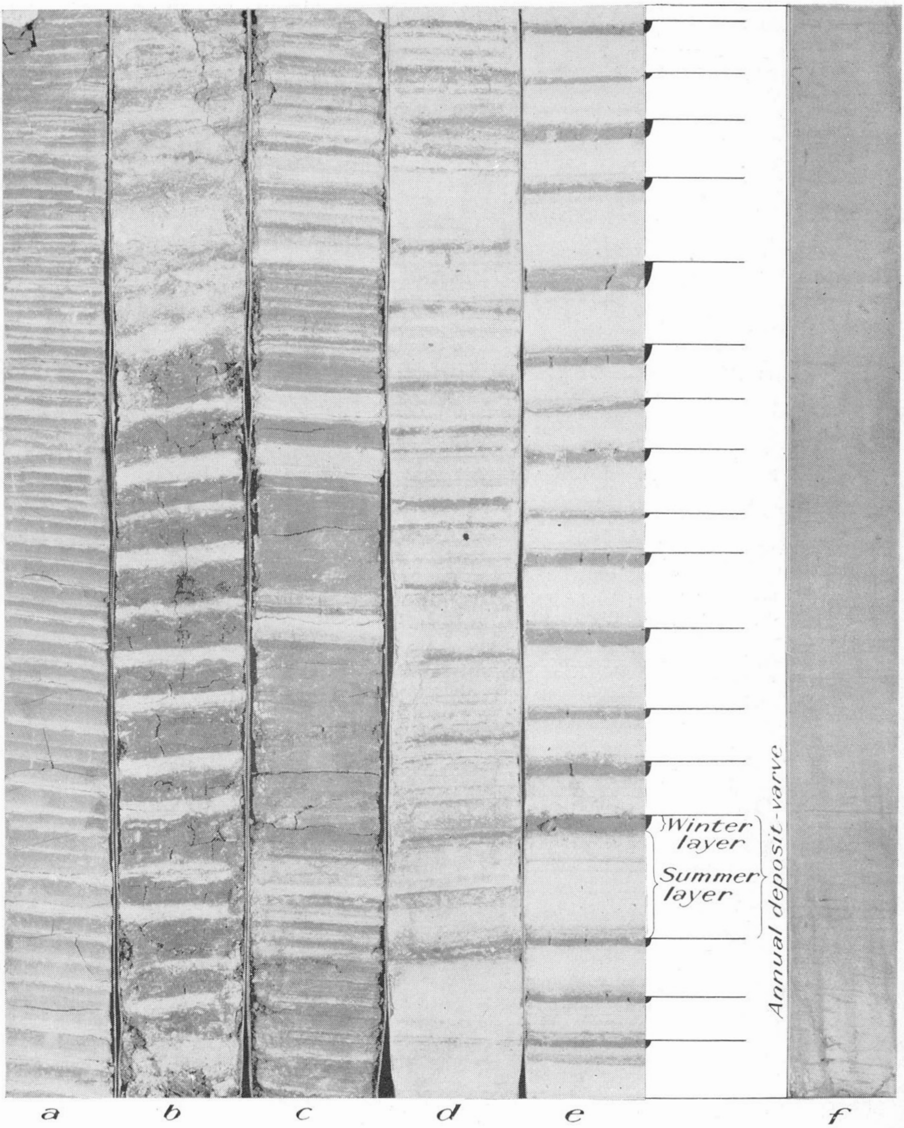


Clay samples. (Actual length of samples 1½ feet.)

a. Varved late-glacial clay from locality 103, Iroquois Falls, Ontario. A part shows symminct structure. (Page 14.)

b. Varved late-glacial clay from locality 29, Sturgeon Falls, Ontario. Lower limit of winter layers sharpest. Upper part of sample leached. (Page 13.)

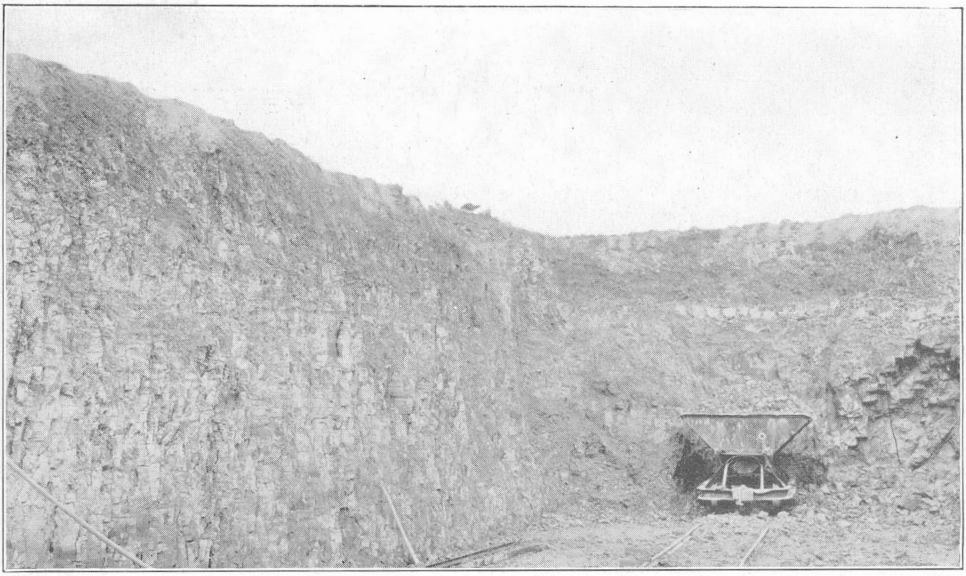
c. Varved late-glacial clay from lower part of section at locality 33, Espanola, Ontario. The two thick dark layers at and below the middle are drainage varves. (Pages 19 and 124.)

d. Varved late-glacial clay from locality 30, Sudbury, Ontario. Shows typical diatactic structure (page 13). Near the top there is an abrupt change from coarse-grained to fine-grained material (page 16).

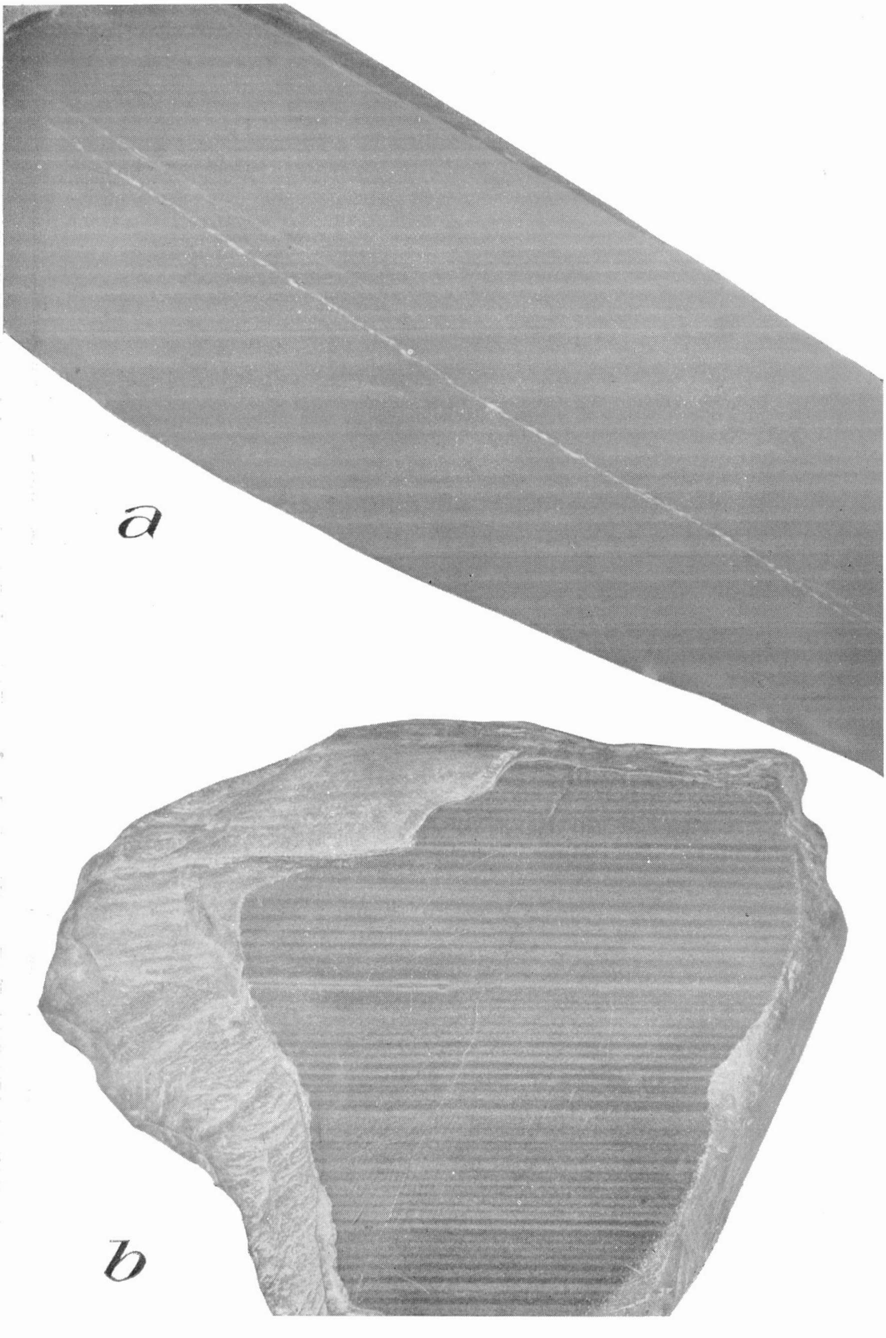
e. Varved late-glacial clay from upper part of section at Espanola, Ontario. Shows typical diatactic structure. Illustrates method of measuring the varves. (Page 12.)

f. Homogeneous post-glacial clay from La Sarre, Quebec. (Page 7.)

A. Varved late-glacial clay, Sandy Falls, 6 miles west-northwest of Timmins, Ontario. (From Rept. Ont. Bureau of Mines, vol. 21, pt. 1, 1912, p. 207.) (Page 4.)



B. Champlain (marine) clay, Renfrew, Ontario. Note the uneven fracture and homogeneous character in the lower part and faint lamination in the upper part, owing to deposition in brackish water far up the estuary. (Page 14.)



Varved glacial Upper Huronian (Precambrian) slate, Cobalt, Ontario. (Page 4.)

PLATES IV–IX

Normal curves showing the average thickness and number of the annual clay layers formed during parts of the ice retreat from Bracebridge to Iroquois Falls, Ontario, and individual curves (Plate IX) showing the thickness and number of the annual clay layers at isolated localities in Ontario and Quebec. (Pages 120-128.)