

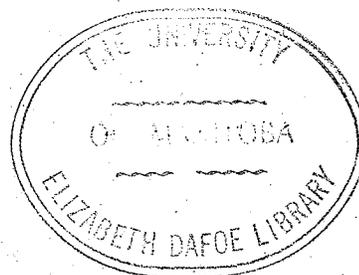
DIFFERENTIAL THERMAL ANALYSIS
OF THE SEDIMENTS OF THE LAKE AGASSIZ
BASIN IN METROPOLITAN WINNIPEG, MANITOBA

by
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A B S T R A C T

Samples of the Keewatin (Gray) till bed, the blue-gray and brown clay beds of Lake Agassiz I Unit, and the yellow silt, greenish brown clay and fine sand beds of Lake Agassiz II Unit were collected from 2 sample locations in the Metropolitan Winnipeg area. Differential thermal analysis under dynamic gases, nitrogen, oxygen and carbon dioxide, of nearly 60 samples followed up by X-ray powder diffraction photographs, X-ray fluorescence analysis, and standard chemical analysis of 9 selected samples showed a common mineralogy between all the beds except the fine sand bed of Lake Agassiz II Unit.

The dominant clay mineral component of the clay fraction was identified as an interstratified dioctahedral illite-montmorillonite. X-ray diffraction analysis indicated that the coarse clay fraction tended to be more dominantly illite and the fine clay fraction tended to be more dominantly montmorillonite, but differential thermal analysis showed both fractions to be similar interstratified illite-montmorillonite. Chemical formulae calculations showed that the over-all fraction had a slight dominance of montmorillonite over the illite and that the montmorillonite was near the montmorillonite end of the montmorillonite-beidellite series. A minor amount of a 7\AA phase either kaolinite and/or chlorite was identified from the X-ray powder photographs.

The non-clay mineral fraction was found to be composed of dolomite, calcite, quartz and feldspar with traces of hydrous iron oxides, possible traces of pyrite and occasional traces of gypsum.

In the gray till the non-clay mineral fraction, dominantly dolomite, was the main component with only 5 to 20% clay mineral fraction. In the blue-gray and brown clay beds the clay mineral fraction with dominantly calcium exchangeable ion, as determined from the thermograms, was the main component.

The non-clay mineral fraction fell steadily from approximately 40% at the base of the blue-gray clay bed to approximately 10% at the top of the brown clay bed. The greenish brown clay bed had a similar composition to the top of the brown clay bed except in this case the dominant exchangeable cation appeared to be magnesium. The basal part of the yellow silt bed was found to be composed of nearly equal proportions of clay and non-clay mineral fractions each present in thin pure layers. The fine sand bed was composed almost exclusively of non-clay minerals, dominantly dolomite with accessory quartz, feldspar, and traces of calcite.

Interpretation of the mineralogy of the beds showed that most of the material in the till originated from the Paleozoic carbonate rocks, with lesser contributions from the Precambrian rocks and minor contributions from the Cretaceous shales. The Lake Agassiz I beds received most of their material from the Cretaceous shales with contributions from the glacier steadily decreasing as it retreated northward. The yellow silt bed was interpreted as receiving its material as wind blown till from the northwest, with the fine clay layers originating from minor fluvial or intermittent lacustrine activity. The greenish brown clay bed had a similar mineralogy and origin as the brown clay. The fine sand bed appeared to have originated from a nearby dolomite outcrop probably Stony Mountain, through fluvial activity.

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CHAPTER I

INTRODUCTION

INTRODUCTION

During the academic year 1960 to 1961, G. A. Russell, P. Eng., Associate Professor of Geology, with the support of a National Research Council Grant, collected information on the subsurface geology of Metropolitan Winnipeg as related to construction and water supply. It was decided to study the mineralogy of the deposits as well as their structural aspects, and differential thermal analysis was proposed as a rapid method. Two sample sites were chosen, the main one at 194 Yale Avenue where the deposits from surface to bedrock were continuously sampled by shelby tubes, and the second at the Geology Building at the University of Manitoba where the deposits from surface to refusal were randomly sampled by auger.

The purpose of the present study was, therefore, to develop a rapid method of routine differential thermal analysis for the Lake Agassiz deposits. It was soon discovered that differential thermal analysis could not be used solely on its own so it was coupled with X-ray diffraction and used to establish the mineralogy, while X-ray fluorescence was used to give quantitative chemical data. The results of the three techniques were then combined to give information on the clay minerals.

Once this method was established, the mineralogical differences between the various beds were determined as well as the vertical mineralogical variations within the beds. Since the sampling was restricted to two sites no attempt could be made to interpret the horizontal variations. To conclude the study a general glacial history was interpreted from the mineralogy.

GLACIAL LAKE AGASSIZ

Lake Agassiz (Fig. 1) has been studied by a number of geologists since the early 1800s. The interested reader is referred to the bibliography in Elson (1961) which lists most of the authors who have studied Lake Agassiz. A few brief notes on the more important authors are given here to provide background information for the main part of the study.

Upham (1895) in the first major work on Lake Agassiz, proposed that meltwater from the retreating Laurentide Glacier (later named Labradorean) was trapped in the northward draining Red River Valley between the ice to the north and east, the divide to the south and the higher ground to the west. The lake thus formed drained to the south until the retreat of the ice allowed the lake to drain to the east and finally to the northeast into Hudson Bay. Upham estimated that the lake existed six to eight thousand years ago for a period of less than one thousand years.

Tyrrell (1896) discovered two glaciers were involved in the formation of Lake Agassiz: the Labradorean to the east

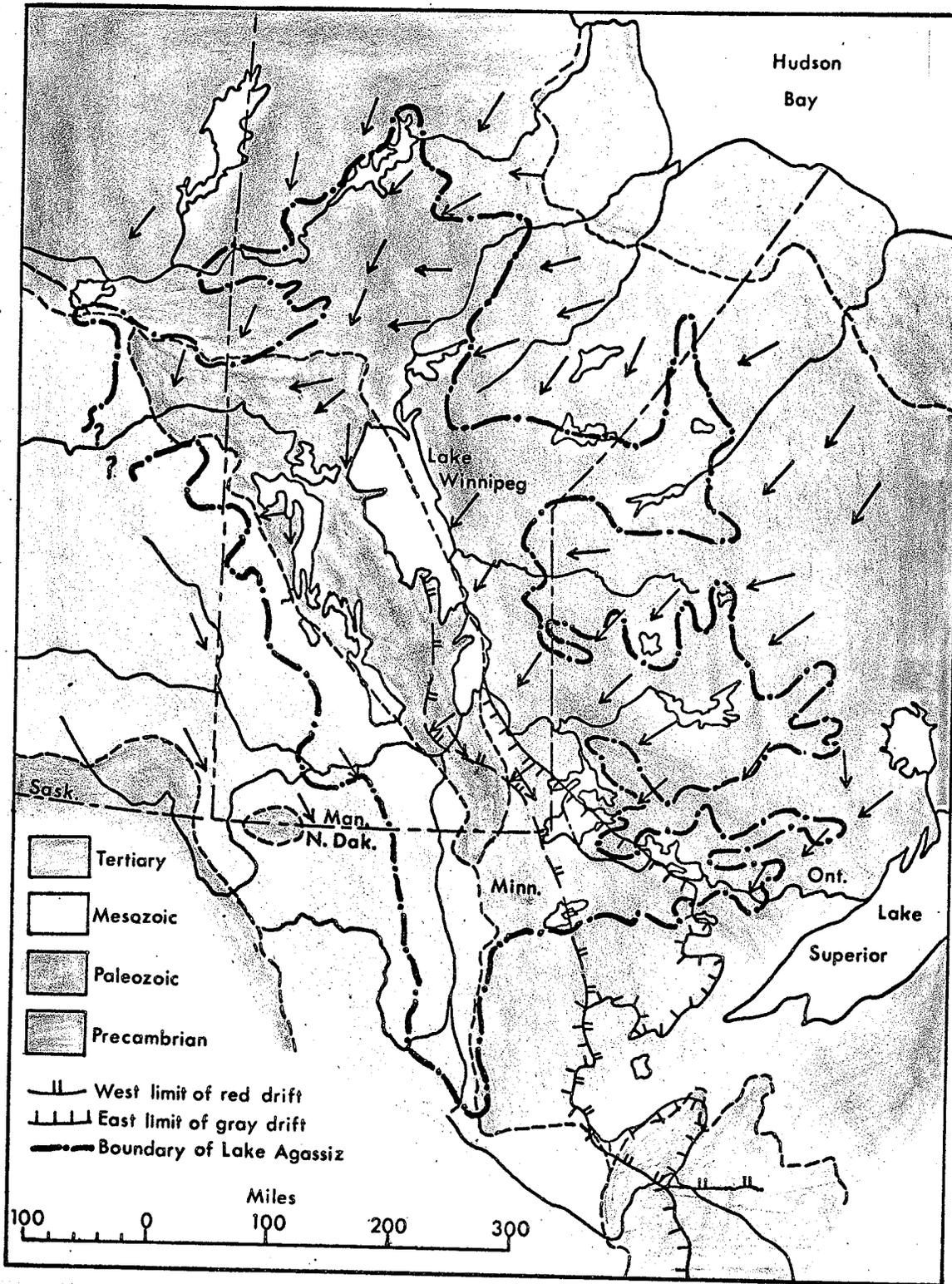


Figure 1. Boundary of Lake Agassiz, bedrock geology and direction of last ice movement. After Elson (1961).

and the Keewatin to the west. The Keewatin advanced first and was channelled southward by the Manitoba escarpment; it reached far into North Dakota, Minnesota and Iowa, then retreated back into Manitoba. It stopped retreating when it met the Labradorean glacier advancing from the east. The trapped meltwater formed Lake Agassiz which drained off to the south in this stage. Eventually the Keewatin glacier resumed its retreat, and drainage to the northeast developed when the two ice masses separated. About this time the Labradorean glacier also retreated and Lake Agassiz drained away.

Johnston (1915, 1916, 1917) generally agreed with Tyrrell but he was the first to suggest two separate lakes. He proposed that the advancing Keewatin glacier had trapped its meltwater between itself and high ground to the south and later advanced into the lake overriding the lake clays. Then the glacier retreated but the lake (Early Lake Agassiz) remained until the glacier retreated far enough north to allow northern drainage of the lake. Once the northern outlets opened, Early Lake Agassiz was largely drained, leaving the lake bottom to be weathered and eroded. From this point on Johnston largely agreed with Tyrrell. The advancing Labradorean glacier met the re-advancing Keewatin glacier, sealing off the northern drainage and once again the trapped meltwater formed a lake. Johnston called this lake the True Lake Agassiz. Once the glaciers joined, the Keewatin glacier remained almost stationary and did not advance as far as it had in Early Lake

Agassiz. Drainage was to the south as described by Upham (1895) and Tyrrell (1896) until a further retreat of the ice once again allowed northern drainage and Lake Agassiz rapidly drained again.

Elson (1961) has done the most recent work on Lake Agassiz by compiling an extensive review of all manner of investigations that have been carried out in the Lake Agassiz basin. These investigations include glacial studies, radio-carbon datings, grain size analysis, chemical analysis, mineralogical analysis, wave counts, and soil mechanics studies.

In addition to this, Elson has published two earlier papers, 1957a and 1957b, that deal with the history of Lake Agassiz. These were based on reviews of earlier work, radio-carbon datings and an aerial photograph study of the area north and west of Lake Superior. The aerial photograph study along the eastern part of the Lake Agassiz basin revealed the presence of previously unreported moraines, wave-cut terraces and several broad dry or nearly dry spillways which he interpreted as being eastern outlets of Lake Agassiz.

Basically, Elson (1957a, 1961) follows Johnston with, of course, some refinements and changes in terminology to conform with the present system of substages of the Wisconsin glacial stage. Advancing Wisconsin ice overrode an early proglacial lake. The northward retreat of this Mankato ice (Keewatin Glacier) resulted in the formation of Lake Agassiz I

in the manner proposed by Upham (1895). This lake initially discharged to the south but as the retreating ice opened eastern outlets, drainage to the east began. A minor readvance temporarily closed these outlets but the recommencement of the retreat reopened them and Lake Agassiz I drained. The ensuing dry interval, the Two Creeks interstadial period, allowed erosion of the lake floor and the development of fauna through the basin. Lake Agassiz II formed when advancing Valdres ice (Laurentide Glacier) blocked the eastern outlets and reinitiated the southern drainage. The eventual retreat of the Valdres ice allowed the discharge to the east again. A northern outlet into Hudson Bay developed as the retreating Laurentide ice separated from the residual ice in the Keewatin district. The final discharge, through the Nelson River system, commenced with the melting of the residual ice in this area and Lake Agassiz II drained off into Hudson Bay.

Through the use of radiocarbon datings carried out by the United States Geological Survey and Yale University, Elson was able to date certain phases of the Lake history quite accurately. The oldest radiocarbon dating, older than 36,000 years, was from the till sheet underlying the Lake Agassiz deposits at Bronson, Minnesota (samples W-102, Suess, 1954; W-468, Wright and Rubin, 1956; and Rubin and Alexander, 1958). A sample from Moorhead, Minnesota, from the base of the Lake Agassiz II unit was dated at $9,930 \pm 280$ years (sample W-388, Wright and Rubin, 1956 and Rubin and Alexander, 1958). Several

samples distributed through the Lake Agassiz I - II interval and the rising stage of Lake Agassiz II were taken in the Assiniboine Valley. These dates are $12,400 \pm 420$ years, and $11,230 \pm 480$ years (samples Y-165 and Y-166 respectively, Preston, Person and Deevey, 1955), and $10,550 \pm 200$ years, $9,110 \pm 110$ years, and $8,020 \pm 100$ years (samples Y-411, Y-415 and Y-416 respectively, Barendsen, Deevey and Gralenski, 1957). Elson (1961) gives the rounded dates of "...the dry interval between Lake Agassiz I and II occurred about 10,000 years ago and that Lake Agassiz II discharged southward from about 9,000 years ago or earlier to 8,000 years ago. It was drained perhaps 7,000 years ago (Rubin and Alexander, 1960)".

CHAPTER II

REVIEW OF THE LITERATURE ON THE LAKE AGASSIZ DEPOSITS

INTRODUCTION

This chapter contains a description of the Lake Agassiz deposits compiled from work done throughout the basin but it is intended essentially to describe the Winnipeg deposits. Reference to work done outside the Winnipeg area is made where it aids in understanding the Winnipeg deposits. The result is a fairly complete description of all the important work and it is applicable over much of the Lake Agassiz Basin. A general description of the deposits through the basin, along with maps and cross sections, by G. A. Russell, P. Eng., is given in Appendix A.

Extensive foundation investigations in and around the Winnipeg area have presented an excellent opportunity to study the Lake Agassiz deposits in detail. However, since most of these investigations are initiated for engineering purposes, detailed descriptions of the units, or peculiarities of the units, encountered in the test holes are rarely kept. The terminology used to describe the units varies from driller to driller and makes comparison difficult, if not impossible, in many cases. Most of the laboratory testing that has been done on these deposits has been solely for foundation purposes.

Thus the unfortunate situation exists, in which these deposits have been extensively drilled and much information gathered, but the opportunity for a comprehensive study has not been grasped.

DIVISION OF LAKE AGASSIZ DEPOSITS

Lake Agassiz deposits fall into several distinct units. The uppermost is generally called the silt unit and more recently Lake Agassiz II unit. It is composed of silty clays, silts and fine sands. Lying unconformably below this is a clay unit or Lake Agassiz I unit which can usually be divided into a varved dark gray or brown clay unit and a massive dark gray or blue-gray lower unit. Below these lies the gray Keewatin till sheet with associated glaciofluvial gravels and sands. In some areas (Fig. 1, p. 3) the Patrician or red till is found instead of the gray till. However, since it does not extend as far west as Winnipeg no samples of red till were recovered and it is not discussed in this report.

General descriptions of the Lake Agassiz deposits from various parts of the basin have been made by Upham (1895), Tyrrell (1892 and 1913), Tyrrell and Dowling (1900), Johnston (1915, 1917, 1921 and 1934), Leverett (1932) and Allison (1932), Dennis, Akin and Worts (1949), Rominger and Rutledge (1952), Ellis (1938), Ellis and Pratt (1953), and Elson (1961).

One of the earliest detailed descriptions of the silt unit and the upper part of the clay unit in the Winnipeg area

was made by Wallace and Maynard (1924). A brief outline of all the deposits from surface to and including the surface of the bedrock was given by Fosness (1926). Macdonald (1937) gave a fuller account of the complete series which was restated by Riddell (1950). The first authors approached their work from a geologic standpoint, whereas the last three authors were concerned with foundation problems.

BEDROCK

Bedrock in the Winnipeg area is Ordovician dolomite, dolomitic limestone, and some limestone. It lies between 15 to 20 feet below ground level in the north and west part of the city and 40 to 60 feet in the southern part (Johnston, 1934). The bedrock surface has a relief of about 50 feet (Johnston, 1934), and this is considerably more irregular than the surface. Vertical fissures running in irregular lines are often encountered in the bedrock surface. These are at least 45 feet deep, 10 to 12 feet wide at the top and are filled with unglaciated shattered limestone, fine sand and some white clay (Fosness, 1926). Johnston (1934) reports that ground water from these fissures can be obtained to a depth of 150 feet from surface indicating that this fracture system extends as deep as 90 to 130 feet from the bedrock surface.

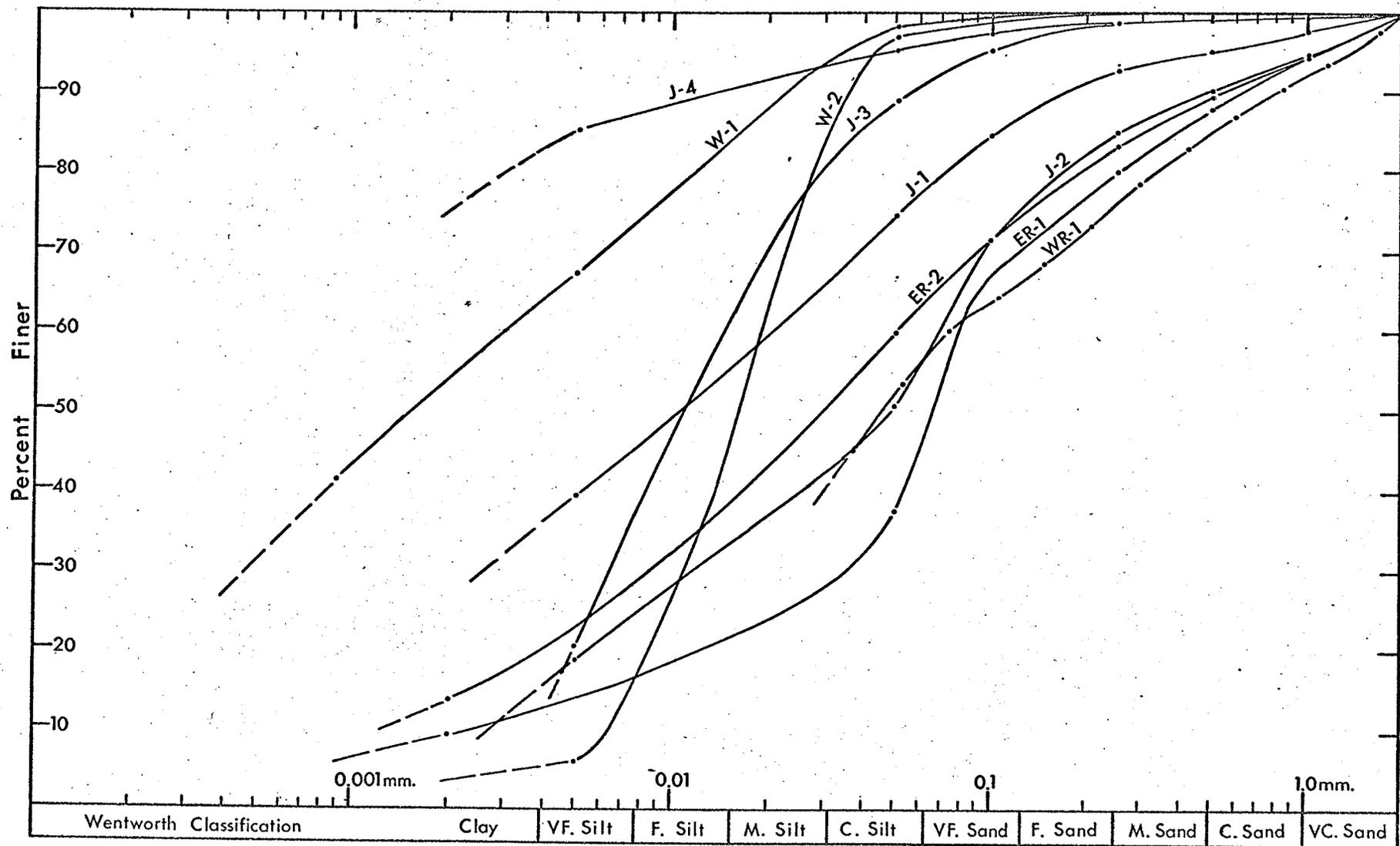
GRAY TILL

Gray till is generally found lying on the bedrock although in many locations sands and gravels have been encountered

between the till and bedrock (see below glaciofluvial deposits). The till is the Keewatin gray drift as described by many authors of which Johnston (1915), Kruger (1937), Ehrlich and Rice (1955) and Elson (1961) are the most important.

The gray drift develops a buff colour when it is weathered but it is gray where it is protected and unoxidized (Johnston, 1915). In the Winnipeg area it ranges in thickness from 1 to 20 feet (Johnston, 1934). Generally it is well graded ranging from sand to clay sizes. Elson (1961) has compared the grain size analysis (Fig. 2) of Johnston (1915) and Ehrlich and Rice (1955) and states that "...the median grain sizes range from .01 to .07 mm. The effective grain size is generally well within the clay-size range, which indicates impermeability to water" (p. 59). These samples from widely separated areas show a marked similarity. Samples J-1, J-2, ER-1 and ER-2, (Fig. 2) all pure tills, lie in a fairly close group. Sample J-3 falls part way between these tills and the clays, J-4 and W-1, because it is not a pure till but a mixture of reworked till and blue-gray clay.

Boulders and pebbles are present but not as commonly as might be expected (Johnston, 1915). They are angular, sub-angular or subrounded, and dolomite and limestone are by far the dominant rock types in the Winnipeg area. Precambrian boulders and pebbles are not rare. Where the till overlies



Johnston (1915).
Rainy River, Ont.

J-1 Unweathered till.
J-2 Weathered till.
J-3 Glacio-lacustrine clay.
(Grayish-white band).
J-4 Glacio-lacustrine clay.
(Bluish-gray band).

Wallace (1927).
Winnipeg, Man.
Ehrlich and Rice (1955).
Komarno and Isafold, Man.
Wicks and Russell.
St. James, Man.

W-1 Upper brown clay.
W-2 Yellow silt.
ER-1 (Gray) till.
ER-2 (Gray) till.
WR-1 Gray till.
(oxidized).

Figure 2. Grain size analyses of Lake Agassiz, and associated sediments.

carbonate rocks as it does in the Winnipeg area, the carbonate content can be as high as 65% equivalent CaCO_3 (Ehrlich and Rice, 1955, Table 1). This decreases to the south and south-east in North Dakota and Minnesota where new materials, mainly Cretaceous shales picked up in southern Manitoba, were added. Kruger (1937) gives an average value of 16% equivalent CaCO_3 for several gray drift samples approximately 50 miles west of Minneapolis. The decrease in carbonate content is obvious once the bedrock is no longer a carbonate rock.

The results of a light, and heavy mineral analysis carried out by Ehrlich and Rice (1955) on the fine sand fraction (0.10-0.25 mm.) of two till samples from the Interlake area north of Winnipeg, are given in Table 1. These results give some idea of the proportions of the various minerals present and hence the various rock types contributing to the till can be determined. The glacier originated in the Precambrian Shield area north of the sample location and the igneous and metamorphic rocks of this area are represented by feldspar, quartz and most of the heavy minerals. The influence of the underlying carbonate rocks is illustrated by dolomite and calcite. They make up 40 and 44% of the light minerals in the fine sand fraction, and the carbon dioxide content of the total sample represents 57 and 65% when calculated as calcite. The contribution of the Cretaceous shales is minor but not insignificant. The clay size fraction (less than 2 microns) of the two samples is approximately 13 and 9% of the total sample and most of this

Table 1

Mineral Composition Of the Fine Sand Fraction (0.1-0.25 mm.)
of Gray Till, Interlake Area, Manitoba (After Ehrlich & Rice, 1955)

	<u>Sample 6</u> <u>(Isafold)</u>	<u>Sample 7</u> <u>(Komarno)</u>
Total Carbonate (calculated as CaCO ₃) of Total Till Sample	57.4	65.0
Fine Sand Fraction as % of Total Till	11.50	12.62
% of Light Minerals in Fine Sand Fraction	96.06-99.73	
Light Mineral Composition		
Quartz	6.2	6.8
Feldspar	45.9	46.1
Calcite and Dolomite	44.4	39.7
Others	3.5	7.4
- TOTAL	100.0	100.0
% of Heavy Minerals in Fine Sand Fraction	3.94-0.27	
Heavy Mineral Composition		
Amphiboles	66.2	56.2
Pyroxenes	10.1	17.3
Garnet	4.1	5.0
Apatite	1.6	1.6
Iron Oxides	6.4	1.0
Others	11.6	17.9
TOTAL	100.0	100.0
Clay Fraction (< 2 microns) as % of Total Till	13.39	9.09

clay size material is illite and montmorillonite (see Table 2, p.34) from the Cretaceous shales.

For engineering purposes the till in the Winnipeg area has been divided into three beds: the hardpan, semi-hardpan and light coloured gray clay (Macdonald, 1937; Riddell, 1950). This classification is very useful for engineering purposes although it is not strictly necessary from a geological and mineralogical viewpoint.

The following description is based mainly on Macdonald's (1937) and Riddell's (1950) papers and the observation of the author.

The bottom bed commonly called the hardpan is generally five to ten feet thick and has a characteristic light gray colour. It is described by Macdonald (1937) and Riddell (1950) as poorly cemented till, when it is in a dry or slightly damp state. However, to call it well-compacted till is more accurate as it is not actually cemented and disintegrates readily when it is brought in contact with water. Glacial boulders and cobbles are common in this material.

The hardpan grades into what is called the semi-hardpan which is the same material except that it contains somewhat fewer boulders and cobbles, and is not as well compacted as the hardpan. One to two feet is the general range in thickness of this material.

Approximately five feet of light coloured gray clay (Macdonald, 1937 and Riddell, 1950) overlies the semi-hardpan and hardpan. Although this material does contain more clay material and fewer boulders and cobbles than the beds below it, it still is till and should be called till rather than clay. Unlike the other two beds which are relatively dry and well compacted, this bed is usually saturated. When samples of this material are brought to the surface they often flow under their own weight.

J. Mishtak (pers. comm.) of the Water Control and Conservation Board, Province of Manitoba, proposed shortly before his unfortunate death, the two terms, cemented glacial till and non-cemented glacial till. In the present report geological terms rather than engineering terms will be used. Hardpan and semi-hardpan are called well consolidated till and the light coloured gray clay is called poorly consolidated till.

GLACIOFLUVIAL DEPOSITS

Glaciofluvial deposits of sand and gravel with varying small amounts of silt and clay are found associated with the till. According to Dennis, Akin and Worts, (1949,, p.25), "The composition of the coarse constituents is essentially the same as those of the till, with the important exception that the percentage of shale pebbles is much smaller in the better-sorted deposits."

In the Winnipeg area very little is known about these

deposits, except that they are the main aquifers in the surface deposits. Elson (1961) describes these deposits as "...thin sand and gravel bodies" lying "...between till and the overlying lake clay" (p. 62) and he suggests that they have a considerable east-west lateral extent. Johnston (1934) describes the main aquifer in the Winnipeg area as a gravel bed between the till and bedrock lying north-west of the city. However, water is not supplied exclusively from this layer for much of it comes from fissures and fractures in the bedrock.

Dennis, Akin and Worts (1949) have made the most complete description of these deposits to date, but their work was done in Cass and Clay counties (Fargo-Moorhead) in North Dakota and Minnesota, 240 miles south of Winnipeg. They found that the glaciofluvial deposits occurred throughout the till unit rather than being restricted to lying above or below the till as stated by other authors. They describe beds varying in thickness from 1 foot up to the full thickness of the till sheet which is 180 feet thick in those areas. The degree of sorting also varies so materials, ranging from well sorted gravels to unsorted till, are found. The shape of the beds are elongate lenses trending north to north-east and passing laterally into till.

The origin (Dennis, Akin and Worts, 1949) suggested for these beds was deposition by streams emerging from crevasses in the face of the ice sheet, mainly during periods of retreat. Their second suggested origin was sorting by wave and current

action in shallow waters. Elson (1961) was not satisfied with either of these explanations but he states that no satisfactory mechanism is known to him. The present author (F.J.W.) suggests that the first origin, proposed by Dennis, Akin and Worts, is the most logical and satisfactory explanation for these deposits.

LAKE AGASSIZ I UNIT

Overlying the gray till is the Lake Agassiz I Unit (Elson 1961), sometimes called the clay unit (Dennis, Akin and Worts, 1949). In the Winnipeg area and throughout most of the Lake Agassiz Basin this unit can be divided by its colour and structure into two beds: an upper, the brown clay bed; and a lower, the blue-gray clay bed. In the Winnipeg area it is generally 20 to 40 feet thick but in the more central portions of the basin it is 60 to 90 feet thick, and on the western edge just east of the Manitoba escarpment it is 50 to 110 feet thick.

Blue-gray Clay

The blue-gray clay is generally described as blue clay, bluish-gray clay, or gray or dark gray clay. Whatever term is used to describe the clay, a definite blue cast, which becomes more pronounced with depth, is generally present. In the Winnipeg area the thickness is 20 to 30 feet and the clay itself is softer and siltier (Baracos and Bozozuk, 1958) than the overlying brown clay. Isolated, small calcareous silt lenses and limestone pebbles with a lesser number of igneous rock pebbles are common, particularly in the lower sections of the bed. Except for these silt lenses and pebbles, the blue-

gray clay bed is massive.

The origin of the pebbles and the silt lenses is generally thought to be ice rafting of till. The silt pockets formed when frozen particles of till hit and embedded themselves in the soft sedimentation surface. Thus when they thawed, they maintained their original shape and separate identity. The thawed material that was lost from the silt bodies as they settled and from those that thawed and disintegrated completely before reaching the bottom, settled with the clay material making the clay matrix enclosing the silt lenses slightly calcareous. This is borne out by the fact that the clay matrix is calcareous but it becomes less calcareous towards the top of the bed similarly to the decrease in the silt lense content.

Riddell (1950) states that the clay content of both this bed and the overlying brown clay "...is over 80% with colloidal clay nearly 50%". Johnston (1915) gives a grain size analysis (J-4, Fig. 2, p. 12, this report) of bluish-gray clay that shows 83% of the material to be clay size in the Wentworth scale, or 74% clay size in the U.S. Dept. of Agriculture scale.

Elson (1961) suggests two origins for this blue-gray bed by (a) deposition in deep water near the edge of the retreating glacier, or (b) deposition in shallow water away from the glacier where wave action could reach the sediments and prevent varves from forming.

Brown Clay

The brown clay lies conformably on the blue-gray clay.

It is generally called the brown or chocolate brown clay due to its characteristic milk-chocolate brown colour. However some researchers, Wallace and Maynard (1924), have described the upper portion of this bed as dark gray to yellow-gray, and Riddell (1950) described it as dark gray to chocolate. In the Winnipeg area the thickness (20 to 30 feet) of this unit is approximately equal to the blue clay under it (Baracos and Bozozuk, 1958).

The distinctive feature of this bed is its thin, very fine varves. Wallace and Maynard (1924) counted 49 varves in one inch of sample, from the top of the bed, sampled from an excavation on Rupert Street in Winnipeg. The average thickness of these varves is $1/40$ to $1/50$ of an inch. Variations from these limits are few but one sandy varve is $1/10$ of an inch thick. Elson (1961) states that the varves are not this thin in other locations. This bed generally contains less calcareous silt than the blue-gray clay bed. The only grain size analysis available (Wallace, 1927, W-1, Fig. 2, p. 12 this report) is misleading because it was taken from the top of the brown clay and has more silt than the blue-gray clay sample J-4 (Fig. 2). This analysis is not representative of the entire brown clay bed but is presented here because it is the only analysis available. The median grain size of the sample is 0.0018 mm. (Elson, 1961) and 67% of the material is clay size in the Wentworth classification or 54% is clay size in the U. S. Dept. of Agriculture classification.

Baracos and Bozozuk (1958) give a composition of the brown clay as "...approximately 30% montmorillonite and the remainder practically all illite". Elson (1961) suggests that this refers to some fine fraction of the clay, rather than the total sample. Riddell's (1950) statement of a clay content of "...over 80% with colloidal clay nearly 50%" refers to this bed as well as the blue clay as already mentioned in the blue clay section.

Elson (1961) states that the deposition of this bed took place in deep water far from the glacier front, thus allowing the formation of the fine varves.

Colours of the Lake Agassiz I Unit

The division of the Agassiz I unit into the blue-gray clay and brown clay beds may be somewhat misleading as the colour very likely has little to do with the deposition of the sediments. As is described in a later chapter, there is no mineralogical reason to divide the clays into two beds. The division should therefore be made on a textural or structural basis. However, the fact that the colours correspond fairly closely to, or perhaps are controlled to some degree by, the major structural features of the two beds -- the massiveness of the blue-gray clay and the varves of the brown clay -- makes it convenient to refer to the two beds by colour.

Blue and brown or yellow colours of clay, shales and other sedimentary rocks have long been attributed to the oxida-

tion state of the iron (Pettijohn, 1957, p. 347) in the rocks, the ferrous iron yielding the blue colour and the ferric iron yielding the yellow or brown. It has also been found that the colour often is a function of depth, or more precisely the water table, the yellow always being the surface colour overlying the blue.

Locally, Shepherd (1931) makes the following statement in a report on the Tyndall limestone. "The mottled limestone is divided into two broad classes by virtue of its colouration. That stone, which has a predominance of ferric iron over ferrous iron in the dolomitized areas, is known as the "Buff" variety. The stone having the larger amount of ferrous iron in the dolomite is known as the "Blue" variety. This difference in colour is not due to any inherent difference in the limestone. It depends entirely upon the degree of oxidation occasioned by the percolating ground water. It is not uncommon to find a band of buff stone in a bed of blue where a fracture or parting in the rock has afforded access for oxygen-rich ground water".

It has already been noted that Johnston (1915) has attributed the buff colour of the weathered bluish gray till to the oxidation of the iron content. Upham (1910) has also described this feature in the till at Bird's Hill, Manitoba. "The bluish gray colour of the till, extending to the surface, is due to its relatively unoxidized and unleached condition, in contrast with the yellowish gray till of the esker, where

its height above the adjoining land has permitted infiltration of the water of rains, favoring the change of its iron element from the protoxide combinations to the rust-colouring sesquioxide."

In describing the "buff to yellow to gray" silt unit (Lake Agassiz II Unit) near Fargo, Dennis, Akins and Worts (1949), state, "The yellow to buff colour of the deposits is believed to be the product of weathering wherein the iron compounds in the deposit have been oxidized." (p. 18).

The most recent local description of this phenomenon is given by Bannatyne (1963) in Cretaceous bentonite beds near the villages of Miami and Thornhill, Manitoba. Beds of yellow bentonite grade into blue-gray bentonite where the overburden is 30 feet thick or more, at the Miami location, and at the Thornhill location some of the bentonites are blue where the overburden is very thick.

At this stage with the terms brown and blue-gray clay in wide usage it would be senseless to suggest a change but the limitations of this terminology should be clearly understood. The brown clay is lightest at the top of the bed and darkens and becomes grayer with depth until it passes into the blue-gray clay. Likewise the blue-gray clay increases in darkness and blueness with depth. Local variations can be found but overall the relationship holds according to the references quoted above.

The question of why the two colours in general seem to coincide with the two structures is as yet unanswered. Possibly the varved clays with their preferred orientation tend to aid oxidation of the iron whereas the massive clays inhibit the oxidation. Whether or not this oxidation of the upper part of the Lake Agassiz I unit is essentially a feature of the period of erosion and weathering between the two lakes and is now more or less dormant -- or whether it is still an active process associated with the present erosion surface is another unanswered question. Certainly the oxidation of the Lake Agassiz II unit is associated with the present surface but no one has attempted to determine to what depths this process may be effective.

It seems probable that the upper portions of the Lake Agassiz I unit would oxidize during the dry interval between the two lakes. However, evidence in North Dakota (see next section) would appear to contradict this.

North Dakota and Minnesota

Separate notes on the Lake Agassiz I deposits in North Dakota and Minnesota are included here because of the imaginative work that has been done on them. The till unit and the Lake Agassiz II unit of North Dakota and Minnesota can be correlated with the Winnipeg deposits but the Lake Agassiz I unit presents some correlation problems. It should be noted also that similar differences may exist in beds west of Winnipeg but information on them is not very extensive.

Dennis, Akin and Worts (1949) describe the Lake Agassiz I beds in the Fargo-Moorhead area as "...predominantly a dark-gray to blue-gray clay with occasional ice-rafted boulders, cobbles, pebbles, and sand" (p. 20). They make no mention of any varves and presumably they are not there. On the other hand Laird (1944), described an exposure of 16 feet of laminated gray clay located approximately 10 miles west of Grand Forks, 70 miles north of Fargo. "Laminations are paper thin alternating between thicker dark-gray clay layers and thinner light-gray clay layers" (p. 21). This bed lies immediately below the Lake Agassiz II unit and although the till was not exposed at the site the thickness of the clays could be limited to not greater than 20 feet.

Rominger and Rutledge (1952) described Lake Agassiz I clays at three locations, Grand Forks, and Fargo, North Dakota, and Crookston, Minnesota, and were able to successfully subdivide and correlate the three sections through the imaginative use of soil mechanics and geology. Liquid and plastic limits, plasticity index, natural water content, relative water content, and preconsolidation stress (Appendix B) were the soil mechanic properties determined. Using these properties and the lithology they were able to divide the clay unit into four distinct beds. Uppermost is an 11 to 24 foot "...blue, brown or dark green clay, mostly nonstratified", unconformable with the overlying silt unit and the underlying clay bed due to the presence of two drying surfaces. The upper one is probably the dry interval

between the two lakes and the lower one a minor drying surface. Below this lies 3 to 22 feet of blue, nonstratified clay, 18 to 25 feet of dark-blue or black, nonstratified clay with a very high liquid limit and natural water content, and finally 5 to 35 feet of blue, nonstratified clay with silt lenses, calcareous concretions and isolated pebbles. These divisions were impossible to make with geology alone and the old drying surfaces were only discovered through the consolidation tests.

In comparison to the Winnipeg area, brown clay is rarely present and with the exception of Laird's (1944) location, varves are rare if present at all. Whether this means that the brown clay bed was not deposited in this area, or that it is generally gray rather than brown, is not known. Unfortunately the possibility exists, that the differences are not real but only differences in terminology. This idea appears to be supported by the report of Freers and Carlson (1963) for they refer to the Lake Agassiz II silt unit and the Lake Agassiz I clay unit in North Dakota as yellowish-brown.

The imaginative and successful combination of soil mechanics and geology carried out by Rominger and Rutledge (1952) points to the solution of this problem. Similar work should be done in the Winnipeg area, and indeed throughout the entire lake basin, so that drying surfaces could be identified, and correlations could be made without having to depend on such poorly definable characteristics as colour.

LAKE AGASSIZ II UNIT

The silt unit of Lake Agassiz II lies disconformably "...on the undulating drying surface of the clay unit" (Elson, 1961, p. 67). In the Winnipeg area the silt unit is 8 to 15 feet thick, and has a greater range of particles than the clay unit, from clay to sand with occasional isolated gravel lenses. The colour varies from yellow to yellow-brown to brown and it has been credited to the ferric iron compounds (Dennis, Akins and Worts, 1949), as discussed previously. A few authors have described it as gray or dark green. Varves are well developed in some beds but poorly developed or non-existent in others. Cross-bedding is found in the basal bed. Fresh water shells (Johnston, 1915, 1921) and wood dated at 9,900 years (Rubin and Alexander, 1958) are found at the base of this unit.

The Lake Agassiz II unit is divided into a basal yellow silt bed, a middle varved silty clay bed, and an upper fine sand-sandy clay bed, based on the reports of Wallace and Maynard (1924), Macdonald (1937), and Riddell (1950). This division is convenient but not absolute and could be improved by more detailed work. Only the middle varved silty clay bed is a strictly lacustrine deposit associated with Lake Agassiz II. The underlying yellow silt is considered to be an eolian deposit (Ellis and Pratt, 1953) from the Lake Agassiz I - II interval and the overlying fine sand - sandy clay is a fluvial or fluvio-lacustrine deposit associated with the predecessor of the Red River and possibly the late stages Lake Agassiz II. It is convenient

however to refer to these three beds under the single heading of Lake Agassiz II deposits.

Yellow Silt (Loess)

The basal bed is the most distinctive and problematic of the Lake Agassiz II unit. It is a pale to bright yellow, highly pervious calcareous silt, varying from half a foot to 4 feet in thickness, lying from 8 to 12 feet below the surface (Wallace and Maynard, 1924). Macdonald (1937) and Riddell (1950) call the silt a silty clay and Wallace and Maynard (1924) call it a sandy clay; however, the grain size analysis (W-2, Fig. 2, p. 12) indicates that 91% of the material lies in the silt sizes (Wentworth Scale). As this bed is continuous over a large area with little variation it should generally be called the yellow silt bed with modifiers such as sandy or clayey used only when justified by the presence of appreciable quantities of these sizes of material. Some cross-bedding with a northward dip was found in one location by Wallace and Maynard (1924). A bearing strength of 0.5 tons per square foot compared to 2.0 tons per square foot for the clay makes this bed a major problem for shallow foundations (Figures quoted are bearing strength values permitted by the Winnipeg Building Code).

Varved Silty Clay (Greenish-brown Clay)

Above the silt bed is a dark gray (Wallace and Maynard, 1924) or brown (Macdonald, 1937 and Riddell, 1950) clay bed, two to four feet thick. Varves are generally present but they are not developed to as high a degree as in the Lake Agas-

siz I unit. The bed as a whole, is also siltier than the upper clays of the clay unit, and it develops a cubic parting (nuggety structure) when dried. (Wallace and Maynard, 1924).

Fine Sand-Sandy Clay Bed

An irregularly bedded fine sand two to four feet thick, described by Wallace and Maynard (1924) as being composed of sand with varying amounts of clay, lies two to four and one-half feet below the surface. Macdonald (1937) and Riddell (1950) call this bed a silty clay and Wallace and Maynard (1924) give a range in composition from sand to sandy clay. Judging from their previous use of the term 'sand', Wallace and Maynard mean a very fine sand if not a silt.

Soil Profile

Finally the soil profile completes the section. It is two to four and one-half feet thick and is generally ignored by geologists and engineers as their interests lie deeper. Macdonald (1937) and Riddell (1950) describe the topsoil simply as "roughly two feet of loam with clay subsoil, very sticky and plastic when wet". Ehrlich, Poyser, Pratt and Ellis (1953) give excellent descriptions of the different soil profiles developed in the Winnipeg area. For further details the reader is referred to their report, and to the reports of the Manitoba Department of Agriculture for soil profiles developed on Lake Agassiz sediments in other parts of the province.

Origin of the Lake Agassiz II Unit

The origin of the silt unit has been a subject of much dispute. Initially Upham (1895) believed that the silt unit was exclusively a flood-plain deposit developed after Lake Agassiz had disappeared. Johnston (1915, 1916, 1921) realized that there were two stages to Lake Agassiz and he identified the entire silt unit as lacustrine and fluviolacustrine deposits associated with the second Lake Agassiz. In either case all the sedimentation was done in the lake, the lacustrine deposits developing from wave erosion of the shores and bottom of the shallow portions of the lake, the fluviolacustrine deposits developing as sub-aqueous valley fills and deltas, and both types of deposits being intermingled and mixed by wave action.

Wallace and Maynard (1924) accepted the theory of two stages but proposed that the yellow silt bed at the base of the silt unit and the sand bed near the top were deposited by rapid, northward-flowing rivers that periodically flooded, forming the observed deposits. The river that formed the yellow silt bed existed between the two lakes, and the river that formed the upper sandy bed was a proposed early predecessor of the present Red River. This left the laminated clays between the two beds representing Lake Agassiz II.

Dennis, Akin and Worts (1949) supported this theory, attributing the laminated clay beds to deposition by a shallow transgressing lake and acknowledging the possibility of the silt

being flood-plain deposits.

Both groups of authors mention the presence of thin beds of gravel and sand associated with the yellow silt bed, and both groups present this as further evidence of stream deposition of the yellow silt.

The presence of a drying surface, and thus positive proof of a dry interval, was eventually proven to exist by Rominger and Rutledge (1952) with the measurement in the clay of the preconsolidation stress (Appendix B) caused by the drying during the Lake Agassiz I - II interval.

The only other interpretation that has been put forward to explain the formation of the yellow silt bed was made by Ellis and Pratt (1953). "In the northern portion of the basin, calcareous straw-colored fine sandy clay has been carried from the high-lime till of the Interlake area, and deposited as a tongue-like intrusion over the lake clay in a section extending from the Stonewall - Selkirk districts, through the site of the City of Winnipeg and as far south as St. Norbert and La Salle" (p. 15). A similar deposit between Emerson and St. Jean, originated from the Walhalla delta, is also described in the same paper.

When the wide extent of the yellow silt bed in the Winnipeg area is considered, as well as the occurrence of similar deposits in other parts of the province (Johnston, 1921, and Ellis and Pratt, 1953), the deposition by wind is certainly

the most logical explanation. The wide expanse of this almost structureless bed of very uniform composition and grain size is not the sort of deposit one would expect to find as a flood-plain deposit. It is, however, what one would expect as a wind deposit. The presence of some small streams is indicated by the thin gravel beds, but these have deposited the gravel and not the silt bed. It is not unlikely that there was a river flowing north at this time probably down the present Red River channel but it would be similar to the present river and would not deposit a very pure yellow silt bed. It appears, therefore, that the yellow silt bed is a loess and not a flood-plain deposit.

On the contrary, the upper highly variable sandy clay bed is the sort of deposit one would expect to find in a flood-plain. So the northward flowing predecessor of the Red River, associated with the final stage of Lake Agassiz appears to be the best explanation for the origin of the upper sand-sandy clay bed. Some of this material may also have been deposited by wind in a similar manner to the basal bed.

The varved clays lying between the two silty beds are of course the deposits of Lake Agassiz II.

CLAY MINERALOGY OF PRAIRIE TILLS AND CLAYS

A review of the clay mineralogy of the prairie tills and clays is helpful in understanding the origin of the clay minerals in the Agassiz sediments. Very little clay mineralogy

has been published in Manitoba on any of the variety of clay bearing rocks, tills or soils. In Saskatchewan and Alberta the clay mineralogy has been somewhat more widely determined in a number of bedrock and overburden occurrences. Much of the work has been done by soil scientists investigating changes due to soil development. While these changes are not of direct interest to the geologist, the mineralogy of the parent material (C-horizon - Appendix C) is of great interest. Thus in the soils reports mentioned in the following review the pedogenic changes are ignored and only the clay mineralogy of the parent material is mentioned.

Table 2 is a summary of the results obtained by many researchers in the Prairie Provinces. The materials studied were for the most part from tills, but one loess, some glacial-fluvial sands, and several lacustrine clays were also examined. The similarity of the clay mineralogy is striking. In almost all cases montmorillonite is the dominant mineral with major to minor amounts of illite. Minor amounts of kaolinite are generally present and chlorite is often present in minor or trace amounts. Minor amounts of quartz and feldspars have also been identified in some soils. In those studies in which the clay has been fractionated montmorillonite generally dominates the finer fractions almost to the exclusion of other minerals. The exception to these trends occurs in Alberta where the clay fractions generally have more illite up to the point where illite is the dominant mineral in a few soils (Pawluk, 1960, 1961).

Table 2 Published Data on the Clay Mineralogy of Prairie Surface Deposits

References	Ehrlich, Rice and Ellis (1955)			Warder and Dion (1952)	Christiansen (1959)	Christiansen (1960)	Greer and Christiansen (1963)	
Province	Manitoba			Saskatchewan	Saskatchewan	Saskatchewan	Saskatchewan	
Area	General			General	Swift Current	Qu'Appelle	Wynyard	
Type of Deposit	Tills			Varied	Tills, Lacustrine, Loess	Tills	Tills and Lacustrine	
No. of Samples	10			8	9	?	25	5
Size Studied (Microns)	2-0.2	0.2-0.1	<0.1	<1.0	<2.0	<2.0	<2.0	
Montmorillonite	nil-20 ¹	10-40	15-70	55 ¹	55-90 ¹	dominant	dominant	minor
Illite	5-25	5-25	nil-15	45	5-30	minor	minor	major
Mica	nil-25	nil-20	nil-15	-	-	-	-	-
Mixed-Layer	nil-10	nil-20	nil-20	-	-	-	-	-
Kaolinite	nil-10	nil-10	nil	trace	5-20	minor	minor	major
Chlorite	nil-15	nil-15	nil-10	-	-	-	-	-
Quartz	1-30	nil-5	nil	-	-	-	-	-
Feldspar	1-25	5-15	5-15	-	-	-	-	-

¹ Figures opposite minerals are %.

Table 2 (cont'd)

Reference	St. Arnaud and Mortland (1963)		Rice, Forman and Patry (1959) ²			Rice, Forman and Patry (1959) ²			Pawluk (1960)
Province	Saskatchewan		Saskatchewan			Alberta			Alberta
Area	Touchwood Hills		Sceptre, Regina, Melford			Three Hills, Nampa			Fort McMurray
Type of Deposit	Tills		Lacustrine Clays			Lacustrine Clays			Glaciofluvial Sands
No. of Sample	4		3			2			2
Size Studied (Microns)	2.0-0.2	<0.2	2.0-0.2	0.2-0.1	<0.1	2.0-0.2	0.2-0.1	<0.1	<2.0
Montmorillonite	major	74-80	dominant -major	dominant	dominant	major -minor	dominant	dominant -major	major
Illite	major	19-22	major -minor	minor	trace	dominant -major	major -minor	nil - dominant	dominant
Mica	-	-	included with illite			included with illite			-
Mixed-layer	(possible illite-mont.) ³		-	-	-	-	-	-	major (illite-mont.)
Kaolinite	minor	trace	minor	minor	tr-nil	minor	minor	tr-minor	minor
Chlorite	minor	trace	minor-tr	nil	nil	tr-minor	nil	nil	trace
Quartz	minor	nil	minor	nil	nil	major -minor	nil	nil	minor
Feldspar	minor	nil	trace	nil	nil	trace	tr-nil	nil	-

²Same Report

³Mont. is used as an abbreviation for montmorillonite.

Table 2 (concluded)

Reference	Pawluk (1961) ₂		Pawluk (1961) ₂		Kodama and Brydon (1965)
Province	Alberta		Alberta		Manitoba, Saskatchewan and Alberta
Area	General		Alcan		General
Type of Deposit	Tills		Till		Tills, lacustrine
No. of Samples	4		1		7
Size Studied (Microns)	2.0-0.2	<0.2	2.0-0.2	<0.2	0.2-0.04
Montmorillonite	minor-major	dominant	trace	trace	reported as mixed-layer
Illite	major	tr-minor	major	minor-major	10Å mica-mont. phase
Mica	-	-	-	-	1 - 7
Mixed-layer	trace (illite-chlorite-mont.)	nil	minor (illite-mont.)	minor-major	89 - 97 (mont.-mica) ⁴
Kaolinite	minor	trace	minor	minor	2 - 4
Chlorite	trace	nil	trace	nil	nil
Quartz	minor	trace	minor	trace	trace
Feldspars	-	-	-	-	-

⁴ Three values of illite and montmorillonite in the interstratified material were obtained with three analytical methods. X-ray diffraction peak heights 50-65% mont., Fourier transforms from X-ray diffraction intensities 55% mont., and chemical analysis 76-82% mont. The balance is mica-like mineral (illite, F.J.W.) in all cases.

SOURCE AND NATURE OF THE CLAY MINERALS

Table 3 is a tabulation of the clay mineral analysis of Cretaceous shales in the prairie provinces and the northwestern states. Comparison of this table with Table 2 reveals a marked similarity between the clay mineralogy of the shales and the glacial and glacio-fluvial deposits. Many authors have noted this similarity. A few, Warder and Dion (1952) and Pawluk (1961), have studied soils on glacial deposits that can be related directly to the Cretaceous shales. Others, Christiansen (1959, 1960), Greer and Christiansen (1963), Forman and Rice (1959) and Rice, Forman and Patry (1959), have analyzed both glacial and shale deposits, and all have reached the conclusion that the Cretaceous shales are the source of the clay minerals in the glacial and glaciofluvial deposits. This conclusion is not a surprising one for one would expect glacial deposits to reflect the mineralogy of the bed rock the glaciers passed over. What is significant about this similarity of mineralogy is that it indicates that very little diagenitic change has occurred, (Kodama and Brydon, 1965), during the transportation and deposition of the clay minerals from the shales to the tills, loesses and lake clays.

In the X-ray diffraction results of most researchers the 18\AA peak from glycerol-saturated samples has been identified as montmorillonite and the 10\AA peak as illite. Both have been reported as separate minerals in most reports. A few researchers have reported mixed-layer types as well, tables 2 and 3. St.

Table 3 Published Data on the Clay Mineralogy of Prairie Cretaceous Shale and Associated Rocks

References	Christiansen (1959)		Christiansen (1960)	Proctor (1959)	Greer and Christiansen (1963)
Province	Saskatchewan		Saskatchewan	Saskatchewan	Saskatchewan
Area	Swift Current		Qu'Appelle	Southwest	Wynyard
Formation	Bearpaw	Whitemud	Marine Shales, Bearpaw & Belly River	Blairmore Vanguard Shale	Bearpaw & Marine Shales
Rock Types	Shale	Sand-clay	Marine Shales	Shale & Siltstone	Shale
Size Studies (Microns)	<2.0	<2.0	<2.0	?	<2.0
Montmorillonite	80-85 ¹	-	dominant	major	dominant
Illite	15-20	-	minor	dominant-major	minor
Mica	-	-	-	-	-
Mixed-layer	-	-	-	major-minor (mont.-illite) ²	-
Kaolinite	-	100	minor	minor	minor
Chlorite	-	-	-	minor-nil	-
Quartz	-	-	-	-	-
Feldspars	-	-	-	-	-

¹ Figures opposite minerals are %.

² Mont. is used as an abbreviation of montmorillonite

Table 3 (concluded)

References	Forman and Rice (1959)		Byrne and Farvolden (1959)	Lerbekmo (1963)	Tourtelot (1964)
Province	Saskatchewan		Alberta	Alberta	N.S., S.D., Neb., Mont.,
Area	South of Saskatoon		Southern Alberta	Southern Foothills	and Wyoming, U.S.A.
Formation	Bearpaw		Bearpaw	Belly River	Pierre Shale
Rock Types	Shale		Shale	Mudstone & Shale	Shale
Size Studied (Microns)	2.0-0.2, <0.2		<2.0	<2.0	<2.0 (?)
Montmorillonite	dominant	dominant	dominant	5-65 ³	10-20
Illite	major-minor	tr.-nil	major	30-80 ³	10-20
Mica	-	-	-	-	-
Mixed-layer	-	-	-	15-25 (mont.-illite)	30-50 (mont.-illite)
Kaolinite	minor	nil	-	5-20	-5
Chlorite	trace	nil	minor	nil	-5
Quartz	tr.-nil	tr.-nil	-	-	-
Feldspar	tr.-nil	tr.-nil	-	-	-

³Lerbekmo also found samples of both montmorillonite and illite, but only the polymineralic samples are included in the table.

Arnaud and Mortland (1963), table 2, have suggested that the illite and montmorillonite are interlayered. Kodama and Brydon (1965), table 2, demonstrated this by means of Fourier transforms. They identify the 18\AA phase as a dioctahedral, randomly interstratified mixture of montmorillonite and mica (illite, F.J.W.). This is the main clay mineral assemblage but a small amount of randomly interstratified mica (illite, F.J.W.)-montmorillonite is also present as the 10\AA phase. Their work is the first to show that the montmorillonite and mica-like (illite) layers are actually interstratified.

The composition of the montmorillonite member of the interlayered unit has been suggested by St. Arnaud and Mortland (1963) to be intermediate between beidellite and nontronite. Warder and Dion (1952) and Kodama and Brydon (1965) have calculated the composition of the montmorillonite to be intermediate between montmorillonite and beidellite.

Trace or minor amounts of kaolinite and chlorite have been reported by most researchers (Table 2). However, a few, Warder and Dion (1952) and Kodama and Brydon (1965) have found only kaolinite and no chlorite. When present, chlorite is generally in lesser amounts than kaolinite which itself is a minor to trace component. Pawluk (1960, 1961) has reported increased amounts of chlorite in the B-horizon (Appendix C) of some Alberta soils and suggests that in this position of the soil profile at least, it is authigenic. He does not state whether the small

amount of chlorite in the parent material (C-horizon) is authigenic or not.

St. Arnaud and Mortland (1963), on the contrary, found that in some Saskatchewan soils chlorite was less abundant in the B-horizon. In fact, although it was present in the coarse fraction of the C and A-horizons, it was present only in the fine fraction of the B-horizon. These results were interpreted by St. Arnaud and Mortland to mean that the chlorite was being weathered to finer sizes in the B-horizon.

Chlorite thus appears to be more highly variable than the other clay minerals mainly because it is more easily affected by weathering. The weathering can either form chlorite as Pawluk found or destroy the chlorite as St. Arnaud and Mortland found. This variability with weathering explains why it is generally present in small, lesser amounts than the kaolinite and in some cases is not present at all.



CHAPTER III

DESCRIPTION OF TEST HOLES

BEDROCK

Bedrock was reached in the Yale Avenue test hole at 45 feet 6 inches, and a little over 45 feet of diamond drill core was recovered. It was logged by Dr. H. R. McCabe of the Manitoba Mines Branch, and identified as the Upper Dolomite Unit of the Red River Formation, of Ordovician age (Table 4). A study of the bedrock is beyond the scope of this report but it is of interest to note, with respect to ground water movement in the bedrock, the sand filled fissure at 65 feet (Johnston, 1934).

No bedrock was recovered from the Geology Building test hole.

GRAY TILL

The till bed (Table 4) encountered in the Yale Avenue test hole is nine feet, three inches thick and is medium gray except for the lower foot and a half which is pale yellowish gray. The natural water content, although undetermined, is high and when extruded from the sampling tubes the samples have very low mechanical strength. The same samples, when dried, have a considerable strength. The till bed itself is fairly

uniform and essentially massive. It is composed of fine and very fine grained highly calcareous material with rounded to subangular pebbles occurring singly or in rough layers located randomly throughout the bed. The largest pebble recovered was naturally just slightly smaller than the diameter of the sampling tube (2 inches). Larger pebbles are undoubtedly present but they are not common. Pebbles are dominantly limestone or dolomite with minor igneous types. Several thin sinuous gray clay seams are present mainly in the upper half of the bed, but one is situated one foot above the bottom contact. In a few instances these occur close to, or are associated with, pebble layers, and probably represent disintegrated shale pebbles.

The lower contact could not be studied for refusal was reached at 44 feet 11 inches, one and a half feet above bedrock, but samples were recovered continuously across the upper contact. The till passes sharply into blue-gray clay. There is a minor amount of interfingering of blue-gray clay lenses in the top one inch of the till bed, and three inches below the contact a three inch wide band of till with 35% blue-gray clay and yellow silt pockets is present, but aside from these minor features the contact is sharp and easily determined.

Only one sample of till was recovered from the Geology Building test hole (Table 5). The till in this location appears to be less than five feet thick and is composed of

Table 4

Log of Yale Avenue Test Hole

<u>Depth of Bed</u>	<u>Sample Number</u>	<u>Sample Depth</u>	<u>Description</u>
0' - 4'11"			<p><u>Topsoil and Fill</u></p> <p>Dominantly black organic clay with roots and organic smell. Contains three bands of massive very fine grained yellow sand at 1'1" to 1'5", 2'1" to 2'3" and 4'0" to 4'1". Occasional silt lenses throughout and scattered gypsum lenses at 2'9". 58% sample recovery.</p>
			<p><u>Lake Agassiz II Unit</u></p>
4'11"-5'7"	Y-5 Y-5/6	5'0" - 5'6" 5'6" - 5'7"	Fine to very fine grained yellow sand, slightly calcareous with very fine, crude bedding. Upper contact is sharp, lower contact slightly interlayered with greenish gray clay. 100% sample recovery.
5'7" - 7'9"	Y-5/7 Y-6/9 Y-7	5'7" - 5'9" 6'9" - 7'0" 7'0" - 7'6"	Greenish brown clay, poorly developed varves, irregular dark to medium greenish brown and yellowish brown bands. Each band made up of several varves. Trace of silt lenses. 42% sample recovered.
7'9" - 8'4"	Y-8/1 Y-8/3	8'1" - 8'3" 8'3" - 8'4"	Yellow silt (loess), calcareous, banded with thin light brown waxy clay seams. Clay seams dominant in lower 2 inches; silt bands dominant above. Delicate micro-structure exists with silt and clay bands. (see text). Upper contact not recovered, lower contact sharp with brown clay. 43% sample recovered.

Table 4 (Cont'd)

<u>Depth of Bed</u>	<u>Sample Number</u>	<u>Sample Depth</u>	<u>Description</u>
			<u>Lake Agassiz I Unit</u>
8'4" -23'0"	Y-8/6	8'6" - 9'0"	Brown clay, well developed varves, slightly calcareous, colour deepens from light and medium brown clay at top to dark brown at bottom. Blue-brown bands at 19'7", 20'7" and 22'6". Sinuous banding due to local colour differences, generally parallel to varves, though a coarser feature. Gypsum filled fractures at 10'2" and 12'3"; gypsum lenses scattered through all but lower 7'. Yellow silt lenses make up 2-3% of material in lower 7' but decrease to traces in upper portion of bed. Varves highly distorted at 21'9" due to ice rafted boulder? Upper contact sharp, lower contact has minor interbedding with blue clay. 91% sample recovered.
	Y-10	10'0" - 10'6"	
	Y-11/2	11'2" - 11'6"	
	Y-12	12'0" - 12'6"	
	Y-13	13'0" - 13'6"	
	Y-14	14'0" - 14'6"	
	Y-15	15'0" - 15'6"	
	Y-16	16'0" - 16'6"	
	Y-17	17'0" - 17'6"	
	Y-18	18'0" - 18'6"	
	Y-19	19'0" - 19'6"	
	Y-20/1	20'1" - 20'4"	
	Y-21/6	21'6" - 22'0"	
	Y-22	22'0" - 22'6"	
23'0" -36'2"	Y-23	23'0" - 23'6"	Blue-gray clay, massive calcareous, colour at upper contact slightly brownish, deepens with depth to dark blue-gray. Faint sinuous banding due to local colour differences parallel to bedding. Yellow silt lenses, 15-20% at base decreases to 2-3% at top. Local concentration up to 35% at 35'6". Silt lenses generally below 1" in diameter. Pebbles, mainly limestone scattered throughout bed. Minor interbedding at both upper and lower contacts. 95% sample recovered.
	Y-24	24'0" - 24'6"	
	Y-25	25'0" - 25'6"	
	Y-26/4	26'4" - 26'6"	
	Y-27	27'0" - 27'6"	
	Y-28/1	28'1" - 28'6"	
	Y-29	29'0" - 29'6"	
	Y-30	30'0" - 30'6"	
	Y-31	31'0" - 31'6"	
	Y-32	32'0" - 32'6"	
	Y-33	33'0" - 33'6"	
	Y-34	34'0" - 34'6"	
	Y-35	35'0" - 35'6"	
	Y-36	36'0" - 36'2"	

Table 4 (cont'd)

<u>Depth of Bed</u>	<u>Sample Number</u>	<u>Sample Depth</u>	<u>Description</u>
			<u>Keewatin (Gray) Till</u>
36'2" - 44'11"	Y-36/2	36'2" - 36'6"	Gray till, massive, very high moisture content, highly calcareous. Mainly fine grained material with isolated pebbles and bands of pebbles. Pebbles mainly dolomite or limestone with some igneous material. Very thin sinuous gray clay seam occurs parallel to bedding mainly in the upper half of the bed, but one series occurs at 44'6". Upper contact has minor interbedding with blue-gray clay, lower contact sharp. 72% sample recovered.
	Y-36/6	36'6" - 36'9"	
	Y-36/9	36'9" - 36'11"	
	Y-37/1	37'1" - 37'6"	
	Y-38	38'0" - 38'6"	
	Y-39/6	39'6" - 40'6"	
	Y-41/1	41'1" - 41'6"	
	Y-42	42'0" - 42'6"	
	Y-43/1	43'1" - 43'6"	
	Y-44	44'0" - 44'6"	
	Y-44/6	44'6" - 44'11"	
44'11" *			Auger refusal.
44'11" - 45'6"			Light gray pebbly sandy clay.
			Bedrock 45'6"
			<u>Red River formation (upper dolomite unit)</u>
45'6" - 56'4"			Dolomite, pale yellow to yellowish buff, very finely saccharoidal to finely crystalline granular, massive, badly broken in upper part. Trace vuggy porosity.
56'4" - 57'6"			Argillaceous dolomite; medium grayish red, finely speckled light gray; badly broken, some shaly partings.
57'6" - 58'4"			Dolomite, light gray, massive, dense with patches reddish gray mottled as above.
58'4" - 58'7"			Argillaceous dolomite, medium reddish gray mottled as above, very shaly in part grading to shale.

Table 4 (cont'd)

<u>Depth of Bed</u>	<u>Sample Number</u>	<u>Sample Depth</u>	<u>Description</u>
58'7"	-65'0"		Dolomite, very light gray to white, hard, massive, sublithographic. Bottom 1' shows good pin point to fine vuggy porosity (fossil solution?).
65'0"	-65'6"		Loose sand, medium to coarse - cave?
65'6"	-67'4"		Dolomite, as above, very light gray to grayish buff, very fine grained dense to sublithographic, hard, compact, massive.
67'4"	-69'0"		Dolomite, light brownish to reddish gray mottled, probably slightly argillaceous.
69'0"	-85'0"		Dolomite, light gray to yellowish buff, dense to sublithographic as above, few bands pin point porosity. Lost water at 83.4' where dolomite becomes platy or slaty, possibly slightly argillaceous.
85'0"	-94'2"		Dolomite, variegated purplish and reddish gray to yellowish -- some light gray interbands, probably slightly argillaceous: 1" shaly band at approximately 94'.
94'2"	-94'10"		Dolomite, light grayish buff, almost lithographic, fair pin point fine vuggy porosity. 95% sample recovered.

* 44'11" to 94'10" logged by Dr. H. R. McCabe,
Manitoba Mines Branch.

similar medium gray calcareous material as already described. This sample and all those with the exception of the last foot and a half, recovered from the Yale Avenue test-hole, are unoxidized till samples as indicated by their gray colour. The lower foot and a half is a pale yellowish gray which indicates some oxidation but not total oxidation.

Over six feet of massive yellow buff till is naturally exposed on the north side of the Assiniboine River just west of the St. James Bridge. This till is the same as that found in the Yale Avenue and Geology Building test holes, except that it is highly oxidized to a yellow clay because of its exposure to the air. Also boulders up to two feet across were observed because the limitations of a sampling tube were not imposed on this site. These boulders make up only a small proportion of the material for most of it is very fine grained. Curve WR-1, Fig. 2, p. 12, shows the grain size distribution of the less than 2.0 mm. fraction of a sample taken down the face of the outcrop. Its similarity to the samples of Johnston (1915) and Ehrlich and Rice (1955) is readily noted (J-4, ER-1 and ER-2, Fig. 2).

LAKE AGASSIZ I UNIT

Blue-Gray Clay

Approximately thirteen feet of blue-gray clay was encountered in the Yale Avenue test hole (Table 4). At the upper contact, the top twelve to eighteen inches of the bed has a definite brownish cast but this gives way to a bluish

Table 5

Log of Geology Building Test Hole

<u>Depth of Bed</u>	<u>Sample Number</u>	<u>Sample Depth</u>	<u>Description</u>
			<u>Lake Agassiz II Unit</u>
0' - 10'	G-5	5' - 10'	Brown clay, medium to light brown, slightly greenish cast, calcareous. Scattered silt lenses mixed with yellow silt bed.
			<u>Lake Agassiz I Unit</u>
10' - 25'	G-10	10' - 15'	Brown clay, medium to dark chocolate brown, colour darkens with depth, slightly calcareous, traces of varves, traces of gypsum in G-20.
	G-15	15' - 20'	
	G-20	20' - 25'	
25' - 50'	G-25	25' - 30'	Blue-gray clay, dark bluish-brown at top to very dark blue-gray at bottom, massive, calcareous. Silt lenses present, most abundant towards the base. Limestone and dolomite pebbles scattered throughout the bed, most abundant towards the base.
	G-30	30' - 35'	
	G-35	35' - 40'	
	G-40	40' - 45'	
	G-45	45' - 50'	
50' - 55'	G-50	50' - 55'	<u>Keewatin (Gray) Till</u> Gray till, medium to light gray, very high moisture content, highly calcareous. Limestone and dolomite pebbles present but not abundant.
55'			<u>Auger Refusal</u>

cast which increases in intensity through the bed until at the base of the bed the colour is a very definite dark blue-gray. The natural water content of the clay was not determined but it is quite high. The structure of the blue-gray clay bed can best be described as massive, although there are faint suggestions of what may be varves in a few places in the upper portion of the bed. Local variations in colour over fractions of inches, from medium to dark bluish-gray, yield a faint somewhat sinuous banding, more or less parallel to the bedding, in parts of the bed. Small yellow silt pockets are present throughout the bed. In the lower portion the silt pockets make up 10 to 15% of the material present, except for local variations over short intervals where they make up 35% of the material; up through the bed the silt pocket content drops continuously to 5% or less at the top. The silt pockets themselves are generally subangular to rounded, equidimensional bodies with their contact with the blue-gray clay sharply defined. Most of the silt pockets are smaller than $\frac{1}{4}$ inch in diameter but some are as large as an inch in diameter.

Plus sixty mesh material is present throughout the bed but the amount steadily decreases towards the upper contact. This material is mainly dolomite and limestone with some igneous rock types. Isolated rounded to subangular pebbles of similar composition and of diameters of up to one inch are randomly distributed through the bed. The source of this material -- silt pockets, plus sixty mesh material, and the isolated pebbles --

is thought to be from ice rafting and the continual decrease of these components upwards through the bed suggests that the ice margin was progressively retreating farther and farther (northward) from the site of deposition.

The upper contact of the blue-gray clay with the overlying brown clay is fairly sharp, although the bluish cast of the lower bed becomes less pronounced and actually becomes brownish within one foot of the contact. Narrow bands of brown clay are present in the blue clay for approximately six inches below the contact. The few inches of the exact contact was not recovered and therefore cannot be described. One and one-half inches of blue-gray clay, at a slight angle to the horizontal, is present in the brown clay about six inches above the contact. Also about two to three and one-half feet above the contact two bands, four inches and seven inches wide, of varved blue-gray clay occur within the varved brown clay. These two bands of blue-gray clay are definitely varved whereas the lower blue-gray clay is massive. This serves to emphasize that the terms brown clay and blue-gray clay are only convenient, well-known terms and that the division of these two beds should be made on a structural basis, that is, the presence or absence of well developed varves.

The blue-gray clay bed encountered in the Geology Building test hole, Table 5, is approximately twenty-five feet thick. Five samples of very moist, highly plastic, massive,

blue-gray clay were recovered in a disturbed state. These samples well illustrate that colour is a function of depth for the colour varies from a brownish blue-gray clay just below the upper contact to a bluish gray and finally to a dark blue-gray in the lower ten feet. Silt pockets and some pebbles are present in all samples. The plus 60 mesh fraction increases with depth as it did in the Yale Avenue test hole.

Brown Clay

The brown clay bed in the Yale Avenue test hole is just slightly less than fourteen feet thick, Table 4. As in the lower beds the colour is not uniform throughout the bed. The general trend is to vary from dominantly a medium to dark brown clay with some bluish clay in the lower portion of the bed to a medium brown and light brown clay in the upper portion of the bed. The general trend is thus a deepening of the brown with depth. There are local exceptions to the trend.

Thin varves are present almost without exception through the bed. In some places in the lower portion of the bed the varves are slightly less well developed than those in the upper portion. In most cases varves appear to be horizontal but there is some distortion from the sampling. At approximately 23 feet considerable distortion is present in the form of a drag fold-like structure and must be due either to slumping or more probably to the results of the impact of a fairly large boulder dropped by ice rafting.

Banding, due to lighter and darker colours, is also a prominent feature. It generally follows the varves except that it is a much coarser feature than the fine varving and numerous varves are included within a single colour band. There is evidence that compositional differences exist between the bands, (Wallace, 1927). The lighter brown bands are siltier than the darker brown or blue-gray bands, but in some instances, mainly in the dark coloured bands, the colour banding seems to wander somewhat irregularly across varve planes. Two different factors appear to control the colour of the clay, 1) the depth that produces a darkening of colour through the browns to blue, and 2) the original differences in composition that produces the local differences in colour.

Small calcareous silt pockets are present in the brown clay bed but they are considerably less abundant than in the blue-gray clay bed. Near the lower contact the silt pockets make up approximately 3% of the material but this drops off to only traces near the top of the bed. The presence of pebbles and the amount of plus 60 mesh material decrease in a similar fashion and disappear part way through the bed. In several isolated locations traces of limonite occur in some of the silt pockets. Small scattered gypsum pockets occur in the upper half of the bed in much the same manner as the silt pockets. In the top four and one half feet several vertical or near vertical fissures are filled with very fine white gypsum. (These fissures are probably shrinkage cracks developed in the Lake Agassiz I - II dry interval.) The gypsum occurring in the

fissures is secondary, but the origin of the gypsum pockets is debatable. Since they generally occur only in the top half of the brown clay bed a secondary origin seems the most probable.

The lower contact has already been described in the previous section. The upper contact with the Lake Agassiz II sediments is sharp, abrupt and no inter-bedding occurs.

Three disturbed samples of the brown clay bed were recovered from the Geology Building test hole, Table 5. Although many of the details were lost due to the disturbed nature of these samples, the obvious features could still be studied. The medium brown colour became darker with depth, continuing the trend of the blue-gray clay bed. Remnants of varves and banding were visible as were scattered, very small, silt pockets.

LAKE AGASSIZ II UNIT

Yellow Silt (Loess)

The yellow silt bed in the Yale Avenue test hole is about one foot thick but only four inches of the lower part of the bed, including the lower contact, was recovered during sampling. The lower contact with the brown clay is sharp and no inter-bedding occurs. The material recovered consisted of alternating bands of medium brown clay and a slightly brownish yellow silt. Either band never exceeds more than one-half inch in thickness. The silt bands are dominant in the upper two inches and the clay bands are dominant in the lower two inches.

Within each band very complex microstructure exists. The silt bands are composed of very fine beds or layers of very fine pure calcareous silt with the occasional thin clay layer. Even on this fine scale there is no mixing of the clay or silt. The clay bands generally appear to be more massive but there is some suggestion of layers in places. Very fine cross bedding is present in the silt bands, but the whole structure is complicated and somewhat hidden by an intricate series of tiny normal faults that produce a miniature graben and horst structure. There is no way of determining if this is a primary structure or whether it was caused by the shelby tube sampling.

All these micro-features described can only be seen with a hand lens or binocular microscope.

The pure silt bands are soft and crumble easily when dried but the alternate clay bands have some dry strength and tend to support the whole structure.

The silt portion of the bed is somewhat calcareous. The clay proportion has a peculiar, almost waxy lustre, and appears to be very pure clay with a low organic content. Secondary limonite stains occur associated with either worm bore holes or root passages.

No sample of this bed was recovered from the Geology Building test hole.

Greenish Brown Clay (Varved Silty Clay)

Slightly over nine inches of sample were recovered from the greenish brown clay bed in the Yale Avenue test hole, Table 4, but the bed itself is at least two feet thick. The bed can be distinguished from all others in the series because it has a distinctive greenish brown colour. It is very finely varved and finely banded. The banding is finer than, and more closely related to the varves than in the brown clay bed. Dark bands, greenish brown and greenish gray, are the dominant ones, but thin yellowish brown bands (made up of several varves) are also present.

The grain size is all in the fine silt and clay range. Pebbles and gypsum pockets are not present in this bed but a few small silt pockets occur in the lower three inches of the bed. The carbonate content is very low although a small amount is present.

The lower contact was not recovered during sampling, but the upper contact was recovered. The greenish brown clay passes sharply into the overlying fine sand bed with tongues and lenses of the clay occurring in the lower inch or two of the fine sand bed.

The name varved silty clay of Macdonald (1937) and Riddell (1950) does not apply too well to this bed in this location for although it is varved, it has a very low silt content and is almost a pure clay, therefore, it is referred to as

the greenish brown clay in this report.

The greenish brown clay sample recovered from the Geology Building test hole was contaminated with material from the underlying yellow silt bed.

Fine Sand - Sandy Clay

The entire thickness of this bed was recovered from the Yale Avenue test hole, Table 4, but none of it was recovered from the Geology Building test hole. In the Yale Avenue test hole the bed is composed of eight and one half inches of a pale buff yellow fine to very fine sand. The sand is composed largely, but not totally, of carbonates. It is finely but crudely bedded and crumbles easily when it dries out.

Except for the contacts which are finely interbedded with the underlying and overlying beds, the sand is very pure and contains very little clay or other impurities.

SOIL PROFILE

No information regarding the soil profile was recovered from the Geology Building test hole, but in the Yale Avenue test hole, Table 4, a thickness of five feet of soil and other materials overlies the fine sand bed. This five feet consists of several layers of black or dark gray topsoil alternating with layers of massive fine buff yellow sand. Sample recovery was poor in this section and because this material appears to be a mixture of buried soil profiles and fill it was not studied.

CHAPTER IV

DIFFERENTIAL THERMAL ANALYSIS WITH DYNAMIC GASES

INTRODUCTION

The DTA¹ for this study was carried out with a model DTA-13M DTA unit produced by Robert L. Stone Company of Austin, Texas. The equipment heated at the rate of approximately 10°C per minute and the sample holder was constructed to allow gases to flow through both the sample cavities. Similar equipment has been described by Stone (1952), although the author's (F.J.W.) equipment had no pressure chamber assembly.

The use of dynamic gas atmospheres during DTA (Stone, 1951) causes changes of varying degrees in the thermograms of some of the minerals encountered in this study. This chapter describes briefly the thermodynamic reasons for these changes, and presents the standard curves of the minerals encountered under both static and various dynamic gas conditions. The quantitative methods used for the major minerals are also explained.

THERMODYNAMICS OF STATIC VS. DYNAMIC GAS TECHNIQUES

The theoretical considerations of DTA have been the

¹The standard abbreviation DTA for "differential thermal analysis" is used throughout this text.

subject of numerous papers, but this complex subject is beyond the scope of this paper. For those who wish to study the theory, two papers -- Sewell and Honeyborne (1957), and Smothers and Chiang (1958) -- are worth studying. Both papers review, correct and incorporate earlier work as they develop their own mathematical approximations of the factors that influence the DTA curves. A somewhat similar treatment is given by Stone (1951) who also demonstrates the effect of dynamic gases on the DTA curves. This latter topic is the point of interest for this chapter.

Most of the reactions encountered in a DTA study of clay and associated minerals are either phase-inversions or first-order reactions. Quartz inversion is a good example of the former. Alpha quartz passes into beta quartz during heating at 573°C with the absorption of heat, and beta quartz passes into alpha quartz during cooling at 573°C with a loss of heat. Since one or the other variety of quartz is the only product formed and heat is the only variable, pressure being constant, it is apparent that the reaction will not be affected by the dynamic or static condition of the atmosphere. This is in fact the case, in thermograms repeated in both types of atmosphere the reaction always occurs at 573°C , the same temperature.

First-order reactions, such as the dehydroxylation of clay minerals or the decomposition of the carbonate minerals,

can be described by van't Hoff law. For purposes of this discussion it is expressed in a form of the Clausius - Clapeyron equation (Stone, 1951):

$$\ln p = - \frac{\Delta H}{2RT} + C$$

where

- p = the vapor pressure of the gas given off
- ΔH = the heat of reaction
- R = a constant
- T = the temperature (in degrees absolute)
- C = the mean specific heat of the sample.

This means that for a vapor producing reaction at a temperature T vapor will be produced until a vapor pressure of p is established. During DTA T is constantly being increased so more and more vapor is released in an attempt to establish equilibrium, until no more vapor is available and the reaction is completed. Thus when the dehydroxylation temperature of a clay mineral is reached during heating the liberation of water vapor is very gradual at first but it rapidly increases as the temperature continues to increase. Stone (1951) describes what happens in the sample cavity during the reaction under static conditions. The initial atmosphere will be air trapped in the sample when it was packed in the sample holder, but as the water vapor is released it will gradually displace the air until the atmosphere is totally water vapor at one atmosphere pressure. The remaining water vapor will then be released into a constant atmosphere of water vapor at one atmosphere pressure. Under dynamic con-

ditions with a dry gas such as nitrogen or carbon dioxide flowing through the sample the atmosphere will be continually changing throughout the run. Thus as the initial reaction temperature is reached and the water vapor begins to be driven off, the dynamic gas will carry it out of the sample. This will keep the partial pressure of the water vapor almost negligible throughout the reaction so that the escaping water vapor will not be opposed by a one atmosphere water vapor pressure as it is under static conditions. The result is that the reaction progresses at a greater rate and the peak, and generally the whole curve, is shifted to a slightly lower temperature than attained under static conditions. The kaolinite curves (Fig. 3) illustrate this feature.

The same situation will develop in the decomposition of a carbonate mineral, such as calcite, in a dynamic nitrogen atmosphere, but a different situation develops in a dynamic carbon dioxide atmosphere. Here the partial pressure of the carbon dioxide will be near one atmosphere throughout so that the decomposition peak will be raised to a slightly higher temperature. This feature is described more fully in the section on dolomite and calcite.

KAOLINITE

The peak temperature for the dehydroxylation of well-ordered kaolinite is generally given in the literature, Grim (1953), Holdridge and Vaughin (1957), between 500 and 600°C.

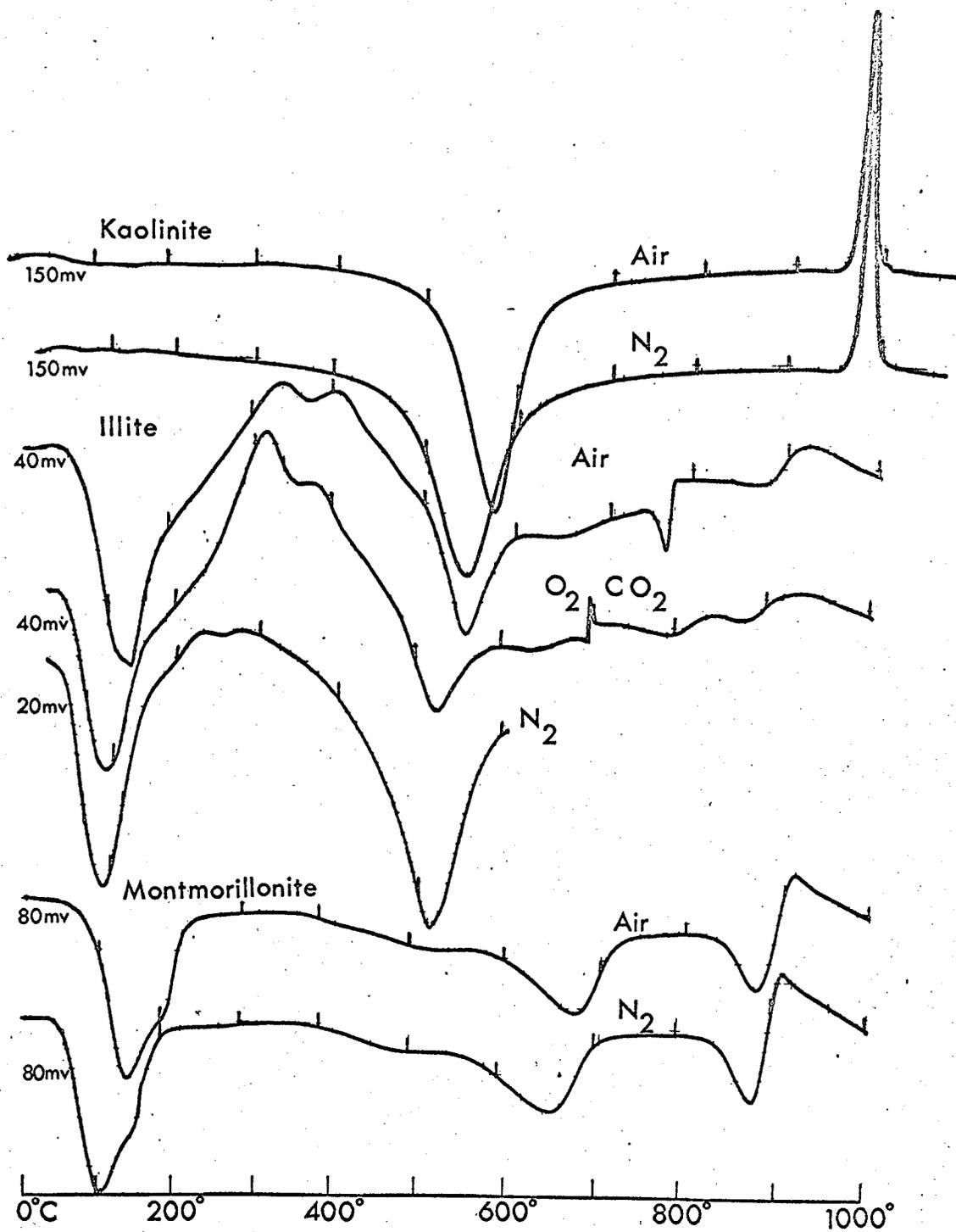


Figure 3. DTA curves of kaolinite, illite, and montmorillonite showing the effect of various dynamic gases. Note the different scale settings.

Under static conditions in the author's equipment (first curve, Fig. 3) the peak temperature occurs at 575°C and this is lowered further to 545°C (second curve, Fig. 3) under dynamic conditions in nitrogen.

The sharp exothermic peak due to the formation of γ -Al₂O₃ or mullite or both (Grim, 1953 and Vaughin, 1957) is stated in the literature as occurring between 980°C to just below 1000°C. In both the static and dynamic thermograms (Fig. 3) the peaks occur at 980°C and are equivalent to the values given in the literature.

ILLITE

Descriptions of typical illites under static conditions (Grim, 1953, Taboadella and Ferrandis, 1957) state that the loss of interlayer water occurs between 50°C and 250°C with the peak at approximately 130°C. This peak is generally of moderate size but it is variable with the relative humidity. The second peak is endothermic also and represents the loss of hydroxyl. The peak temperature is approximately 550°C but the peak extends through the range 450°C to 650°C. A third small endothermic peak in the 850° to 900°C region represents the final loss of hydroxyl and is generally followed immediately by an exothermic peak in the 900° to 1000°C range.

A great deal of controversy exists over what actually causes the final endothermic-exothermic pair, and what minerals are formed during the reaction. It does appear that for diocta-

hedral illites the substitution of magnesia and ferric iron for alumina in the octahedral layer, (Taboadella and Ferrandis, 1957), is responsible for the endothermic-exothermic pair, because illites with low substitution of alumina do not appear to have this reaction.

The third curve, Fig. 3, is the thermogram of an illitic material (it also contains some abnormal montmorillonite, see below Interlayered Illite-Montmorillonite) analyzed under static conditions in the author's equipment. The fourth and fifth curves, Fig. 3, are the same material analyzed under dynamic conditions, first in oxygen to 700°C and completed in carbon dioxide to 1000°C, and second in nitrogen to 600°C. Under static conditions the interlayer water peak extends from 45°C to 250°C with the peak temperature at 130°C; this is altered by the dynamic atmospheres to a range of 20° to 225°C with the peak at 90°C.

The beginning of the dehydroxylation peak, under static conditions, is obscured by the oxidation of the organic matter, but the peak temperature appears to occur at 545°C and the peak ends at 600°C. Under dynamic conditions in nitrogen the entire peak is recorded. It extends from 300°C to 600°C with the peak temperature occurring at 520°C.

The sharp endothermic peak at 770°C under static conditions is due to the simultaneous dissociation of calcium and magnesium carbonate in minor dolomite impurities. Under dynamic

carbon dioxide this is altered to a broad endothermic peak in the 790°C region due to magnesium carbonate dissociation and an endothermic peak at about 920°C due to calcium carbonate dissociation. This latter peak is obscured by the simultaneous exothermic reaction of the illite.

The endothermic-exothermic pair is similar under both static and dynamic carbon dioxide atmospheres. It extends from approximately 840° to 970°C, with the endothermic peak occurring at 875°C and the exothermic peak at 925°C. The endothermic peak is intensified slightly by the dynamic atmosphere but the exothermic peak is decreased due to the simultaneous endothermic peak of the dissociation of the calcium carbonate from the dolomite.

MONTMORILLONITE GROUP

The size and position of the interlayer water peak of montmorillonite minerals is greatly dependent on the exchange cation or cations present and the relative humidity at which the sample was stabilized prior to DTA. This aspect is discussed in the following section. In this section it will be sufficient to state that in the literature the peak is described as occurring between 50° and 300°C under static conditions.

The main dehydroxylation peak of the dioctahedral montmorillonite minerals, the ones commonly found in soils, is characteristic of each mineral in the series (Greene-Kelly, 1957). Montmorillonite, the well known member of the group,

has its main endothermic peak occurring at approximately 700°C and a minor endothermic peak occurring at approximately 900°C followed immediately by an exothermic peak. Beidellite has three endothermic dehydroxylation peaks with an exothermic following the final endotherm. The main dehydroxylation peak occurs at approximately 550°C and the two lesser peaks occur at approximately 700° and 850°C. Nontronite has its main peak occurring in the 550°C range with a second peak occurring near 900°C followed by an exothermic peak. The final endothermic-exothermic pair is highly variable and does not always occur for every nontronite.

Greene-Kelly (1957) has suggested that the high and low temperature dehydroxylation ranges for the montmorillonite minerals are related directly to the structural substitutions in the octahedral and tetrahedral layers. For montmorillonite, with most of its substitution being magnesia and ferric iron for alumina in the octahedral layer, the high-temperature endotherm occurs, but for beidellite and nontronite, with the greater part of the substitution being alumina for silica in the tetrahedral layers, the lower temperature occurs. Nontronite also has ferric iron substituting for alumina in the octahedral layer as does montmorillonite, but as this substitution is almost total replacement of alumina by iron, the significant substitution becomes the tetrahedral substitution of alumina for silica. Thus nontronite behaves like beidellite rather than montmorillonite.

There is considerable disagreement over this interpretation and other possibilities are discussed farther on in this chapter under the heading of mixed layeral illite-montmorillonite.

A dynamic atmosphere lowers the peak temperatures of all the montmorillonite minerals but only a single example is included in this text. The sixth and seventh curves, Fig. 3, are thermograms of a calcium montmorillonite from Laudshut, Germany.¹ The sixth curve, carried out in air, has the interlayer water peak occurring between 60° and 250°C with the peak temperature at 145°C. The main dehydroxylation peak occurs at 675°C with a minor peak at 525°C and a final significant endothermic peak at 880°C with the subsequent exothermic peak following at 910°C. The dynamic nitrogen atmosphere lowers the interlayer water peak temperature to 105°C and the range to be between 30° and 220°C. The shoulder at 155°C on the dynamic atmosphere thermogram and at 205°C on the static atmosphere thermogram is due to the calcium exchange ion. The main dehydroxylation peak is lowered to 660°C and the minor peak to 500°C, but the endothermic-exothermic pair remain unchanged although the endothermic peak is sharpened somewhat in the dynamic nitrogen.

Effect of Exchange Cations on Montmorillonite Thermograms

The expandable nature of the crystal structure of the

¹Sample courtesy of R. E. Grim, University of Illinois

montmorillonite minerals allows them to hold a large amount of interlayer or sorbed water as well as a variety of exchangeable cations. Because of this water a large endothermic peak develops on thermograms of these minerals between 30° and 300°C. The size and shape of this peak and its peak temperature are dependent to a minor extent on the actual mineral composition and to a very large extent on the exchangeable cations present and the relative humidity (R.H.) with which the clay was in equilibrium prior to DTA. As previously noted, the peak temperature is also dependent on the atmosphere of the sample during analysis, whether it be static or dynamic. However, the peaks, due to the various cations, all have the same relationship to each other within each atmosphere system.

Hendricks, Nelson and Alexander (1940) performed extensive experiments on the effects of various cations on the interlayer water peaks of montmorillonite thermograms. They found that for the divalent cations, magnesium and calcium, (Fig. 4), a double peak developed at about 5% R.H. and that a triple peak developed at 10% R.H. for magnesium and at 25% R.H. for calcium. For the monovalent cations, sodium and potassium, (Fig. 5), they found a single peak at low relative humidities and a double peak at 25% R.H. for sodium and at 40% R.H. for potassium.

For the divalent cations, Hendricks, Nelson and Alexander suggest that the highest temperature peak represents the

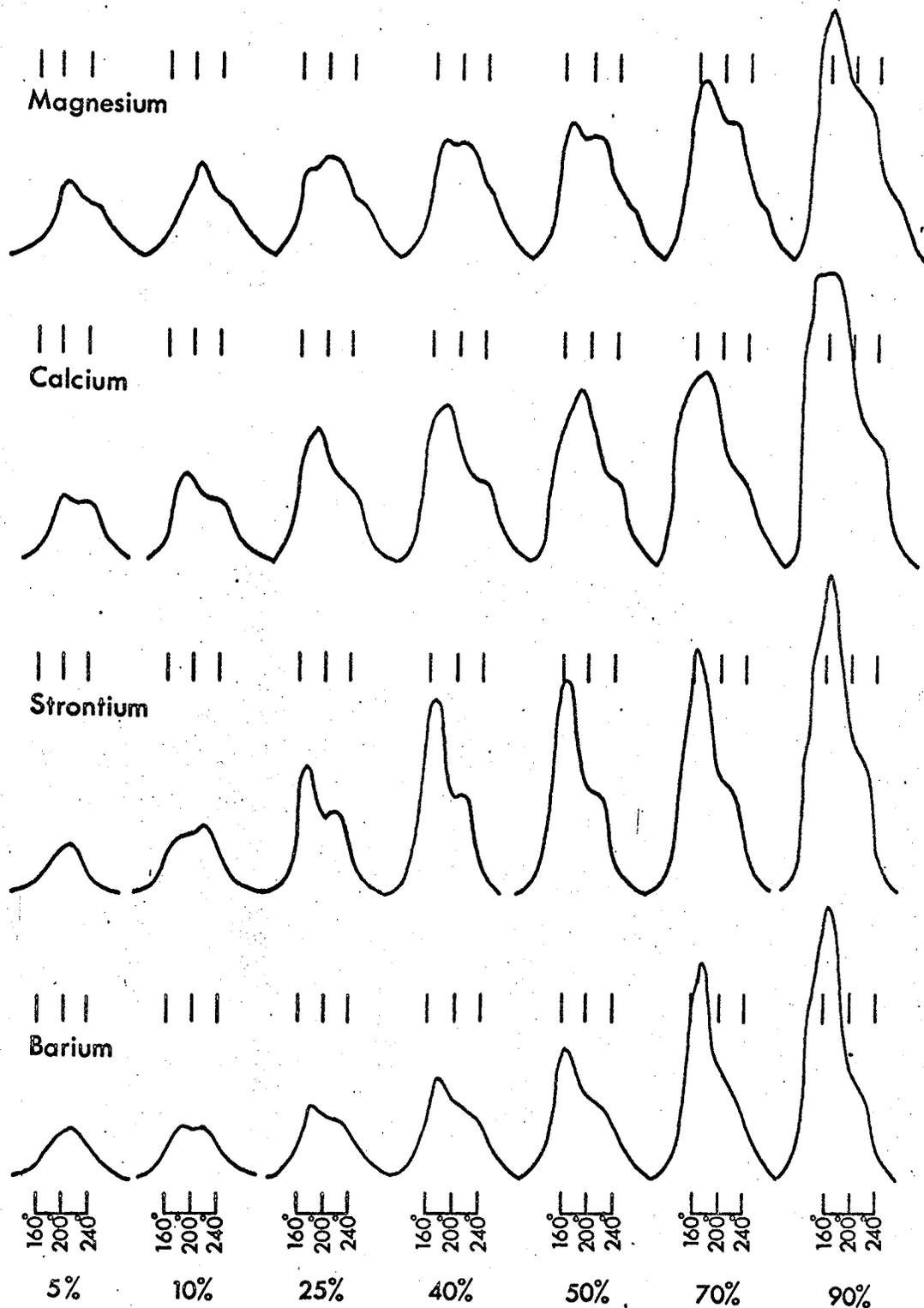


Figure 4. DTA curves, in static air, of Hendricks, Nelson, and Alexander (1940) for Mississippi montmorillonite saturated with various divalent cations at different relative humidities. Note that the curves are pointing up rather than down the page.

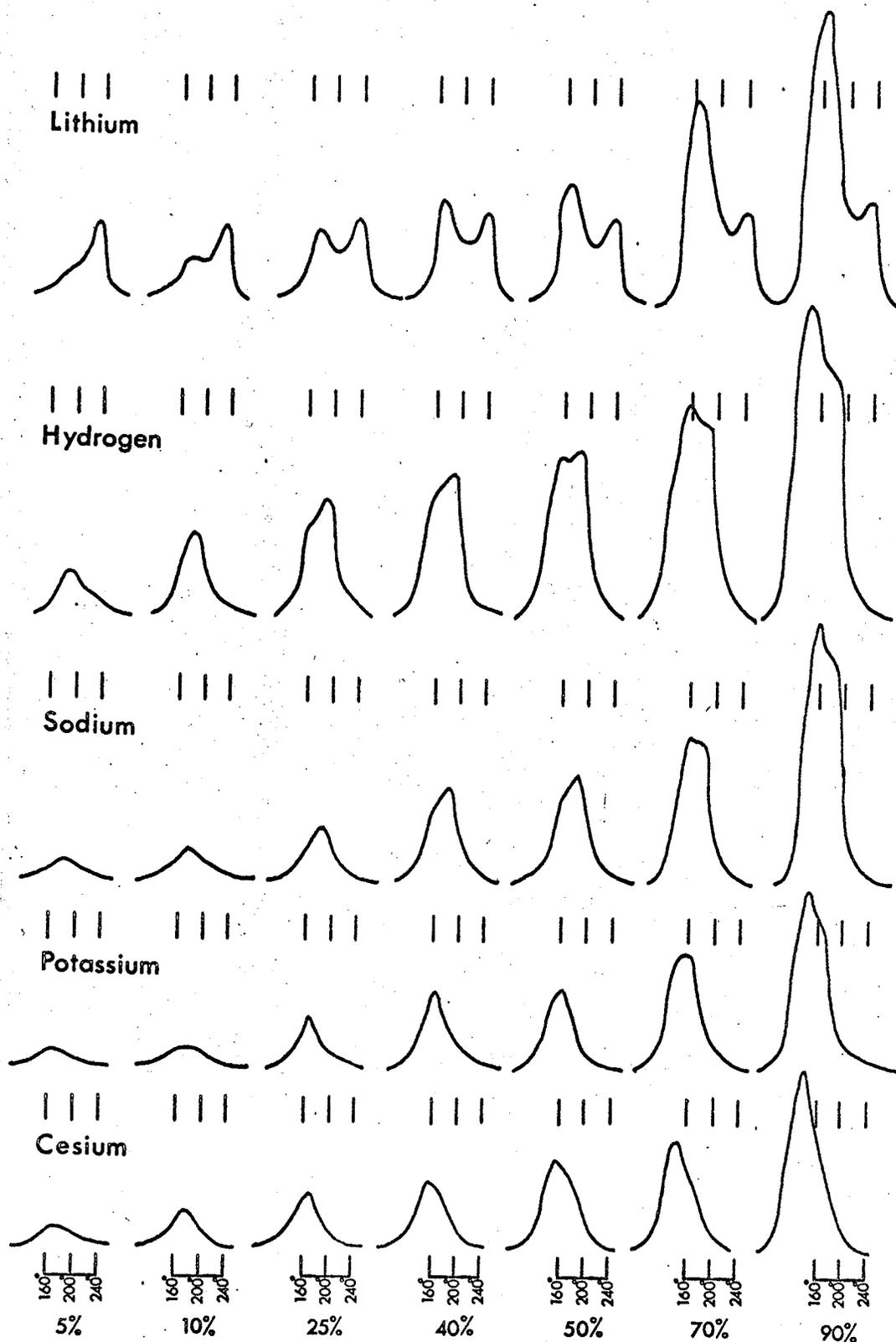


Figure 5. DTA curves, in static air, of Hendricks, Nelson, and Alexander (1940) for Mississippi montmorillonite saturated with various monovalent cations at different relative humidities. Note that the curves are pointing up rather than down the page.

water held by the exchange cation and that the second peak represents water completing the formation of a single water layer over the remaining (cation free) clay surface. The third peak appearing at higher relative humidities represents the formation of a second layer of water. Hendricks, Nelson and Alexander suggested, with some reservation, that the monovalent cations, other than lithium, do not hydrate and that only a single layer of water forms in the interlayer position of the clay particles. Mackenzie (1950) suggested that in fact the sodium does hydrate but that virtually all the water is orientated by the sodium ion in a single layer, hence only one peak was observed. The second peak formed at high relative humidities in this case is due to water associated with the clay surface in non-sodium areas and does not represent a double layer.

The author's research has not been directed towards this aspect but the opportunity arose in which it was possible to study the effects of four exchangeable cations, sodium, potassium, magnesium and calcium on four very similar, although not identical, Lake Agassiz clays. These samples were obtained from the Civil Engineering Department where they had undergone consolidation tests using chloride salt solutions as immersing liquids instead of distilled water. DTA was carried out on each sample after it had been stored at 45% relative humidity for several days. The results are shown in Fig. 6. The monovalent ions, sodium and potassium, have double peaks, the main one occurring at approximately 90°C for sodium and 75°C for

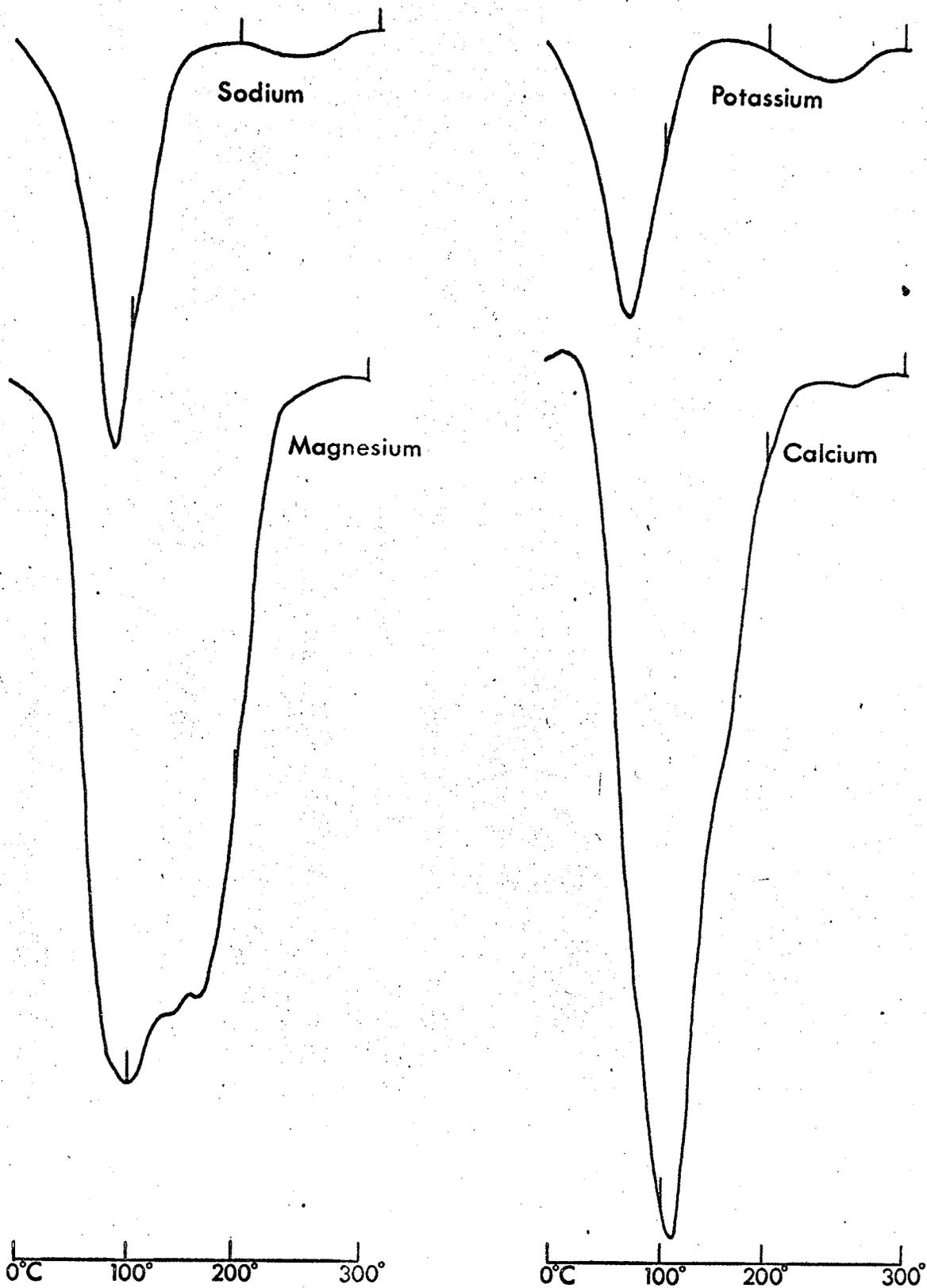


Figure 6. DTA curves, in dynamic nitrogen, of Lake Agassiz II clay saturated with various monovalent and divalent cations at 45 percent relative humidity.

potassium and the minor one as a very slight shoulder at approximately 105°C for sodium and 90°C for potassium. The divalent ions magnesium and calcium yield triple peaks as expected. For the magnesium saturated sample three distinct peaks occur at 100°, 145° and 170°C. For the calcium saturated sample the peaks were not as distinct but three are present as a slight shoulder at 85°C, a peak at 105°C and a moderate shoulder at 160°C.

These results (Fig. 6) compare favorably with those of Hendricks, Nelson and Alexander (Figs. 4 and 5) and also serve to demonstrate the effect of the dynamic nitrogen atmosphere on lowering the peak temperatures approximately 60°C.

INTERLAYERED ILLITE-MONTMORILLONITE

Differential thermal analysis alone cannot be used to identify either random or regularly interlayered sequences of illite and montmorillonite. Artificial mixtures of the two minerals do not necessarily and in fact rarely yield the same results as naturally occurring interlayered mixtures, so no standard reference curves can be produced.

Cole (1955) and Cole and Hosking (1957) state that the main endothermic peaks for interlayered minerals generally fall within two temperature ranges: 525° to 625°C, and 650° to 725°C. (These temperatures refer to static conditions and are 25° to 30°C lower in dynamic gas conditions, F.J.W.) The authors further note that normal illites fall into the lower

temperature range and normal montmorillonites fall into the upper temperature range. However some illites yield peaks in the upper temperature range and similarly some montmorillonites yield peaks in the lower temperature range. These are called abnormal illite and abnormal montmorillonite respectively. Categorizing interlayered illite-montmorillonites by the position of the main dehydroxylation peak, Cole (1955) divided these minerals into two series, normal illites passing into abnormal montmorillonite (low temperature endotherm), and normal montmorillonites passing into abnormal illites (high temperature endotherm).

Nagelschmidt (1944) and Jackson et al. (1952) proposed the following weathering series: mica \rightleftharpoons illite \rightleftharpoons intermediate \rightleftharpoons montmorillonites, and Cole and Hosking (1957) suggested that the two series, illite to abnormal montmorillonite and montmorillonite to abnormal illite are both parts of this series. At the same time Cole and Hosking supported the work of Kerr, Kulp and Hamilton (1949) and proposed that the normal and abnormal structures were due to differences in hydroxyl bonding, rather than substitution in the octahedral layer as discussed by Greene-Kelly (1957) (see previous section on montmorillonite). Actually hydroxyl bonding and cation substitution in the octahedral layer are two ways of saying almost the same thing for the two are intimately related.

Cole and Hosking (1957) make the following statement

concerning the proposal of Kerr, Kulp and Hamilton (1949):

"...since hydroxyl groups in montmorillonite are attached only to certain cation positions a non-statistical distribution of replaceable ions in the octahedral sheet could produce a high temperature reaction in a highly-substituted montmorillonite while a slightly-substituted montmorillonite might show a low-temperature reaction. Such an order-disorder phenomenon is believed by the present authors (Cole and Hosking), to be the cause of different dehydroxylation temperatures for illite and montmorillonite minerals." (p. 262).

Cole and Hosking (1957) conclude from their work on the two illite-montmorillonite series that one can "...predict from differential thermal analysis the direction in which the transformation is taking place". They also make the reservation that more work must be done on this idea before it can be used with certainty.

Applying this theory to the present thermograms on Fig. 3, the calcium-montmorillonite (curve 7) with the main peak at 660°C and the minor at 500°C would be placed in the montmorillonite-abnormal illite series and the Lake Agassiz II clay (curves 4 and 5) with the main peak at 520°C and the minor at 640°C would be placed in the illite to abnormal montmorillonite series.

QUANTITATIVE DETERMINATION OF THE CLAY MINERALS

No quantitative determinations of the clay minerals

were attempted directly from the thermograms. The difficulty, if not impossibility, of obtaining pure clay mineral samples that were identical to each of the samples studied was the main problem. Semi-quantitative estimations of amounts and compositions of the clay minerals were attempted by the combined use of DTA, X-ray diffraction, X-ray fluorescence and some standard chemical analysis. This is dealt with in Chapter VI.

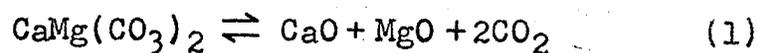
DOLOMITE AND CALCITE

Calcite and dolomite give the strongest endothermic peaks of the minerals encountered in this study. Because the carbonates produce such strong, easy-to-detect peaks and because of their abundance in rocks, the carbonate mineral group has been widely studied by thermal analysis. The results obtained have varied according to the experimental conditions, but a standardization and a realization of the effect of the variables involved have made it possible to understand and reconcile seemingly divergent results.

Dolomite yields two large endothermic peaks on its thermogram. Haul and Heystek (1952) in reviewing work of other researchers found the average peak temperatures to be $810 \pm 15^{\circ}\text{C}$ and $940 \pm 25^{\circ}\text{C}$, the lower temperature peak being due to the dissociation of magnesium carbonate and the higher temperature peak being due to the dissociation of calcium carbonate. These results are from pure samples run in air. When dolomite is run under various partial pressures of carbon dioxide different

results are obtained. The partial pressure of carbon dioxide in contact with the sample controls the peak temperature of the high-temperature peak but it has no effect on the peak temperature of the low-temperature peak. The effects of carbon dioxide atmospheres, both static and dynamic, on the peaks of the carbonates have been studied by Rowland and Lewis (1951), Haul and Heystek (1952) and Stone (1952, 1954). The latter used dynamic carbon dioxide and pressures greater than one atmosphere as well as pressures less than one atmosphere. The Haul and Heystek (1952) results are shown in Fig. 7.

These results could not be duplicated with the author's equipment as no variable pressure chamber was available; however, a similar effect is demonstrated by running a dolomite first in dynamic nitrogen (no CO₂ pressure), then static air (less than 1 atmosphere CO₂ pressure), and finally in dynamic carbon dioxide (slightly in excess of 1 atmosphere CO₂ pressure). (Fig. 8). The high-temperature peak rises from 880° to 935° to 960°C respectively with the increasing partial pressure of carbon dioxide. From their results (Fig. 7) Haul and Heystek concluded that at very low carbon dioxide pressures the single endothermic peak represented the total breakdown of the dolomite and was represented by the equation:



Increase in carbon dioxide pressure brings on the development of a second peak appearing at higher and higher temperature with increasing carbon dioxide pressure. This peak is due to the

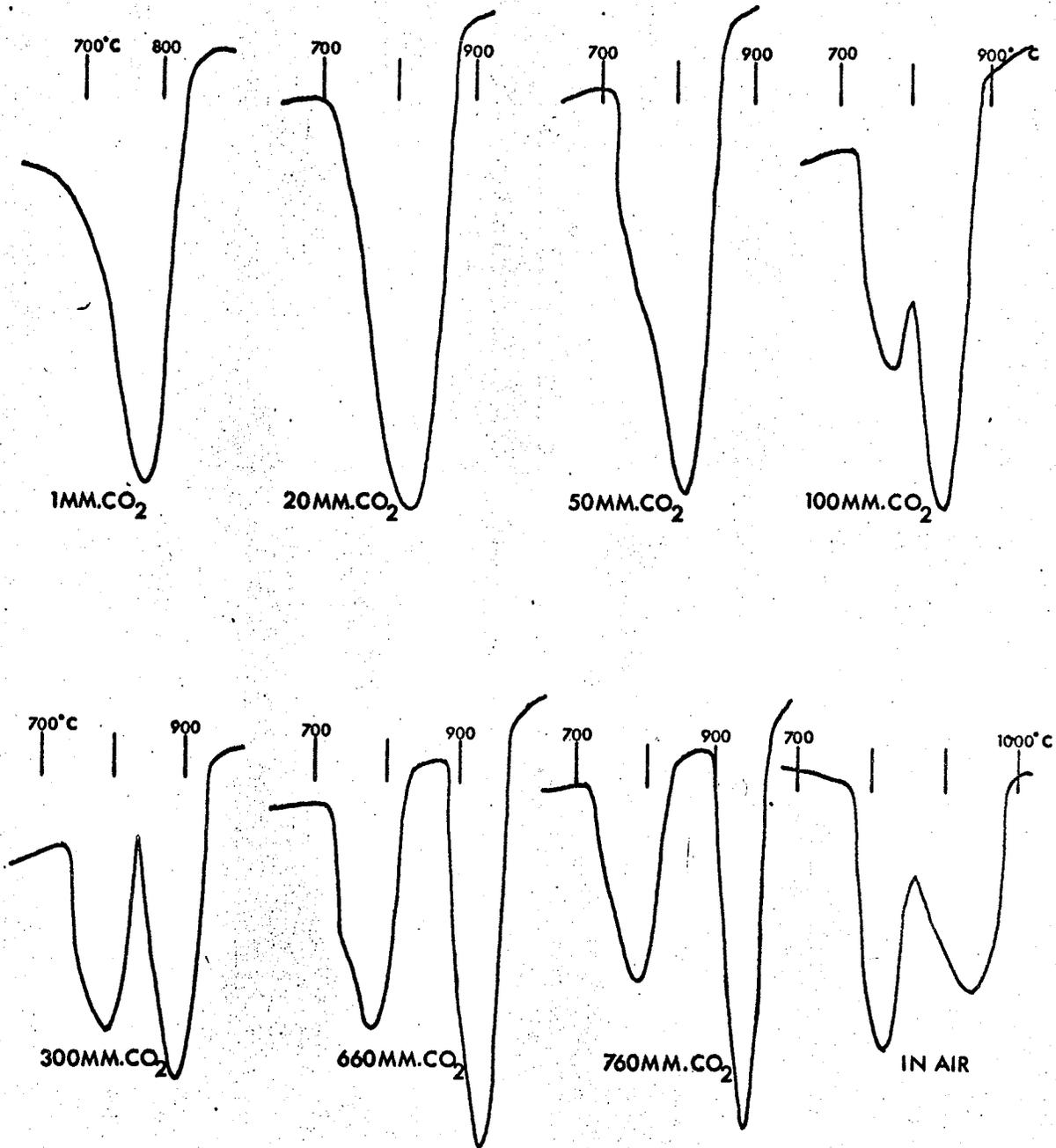


Figure 7. DTA curves of dolomite at different static carbon dioxide pressures. After Haul and Heystek (1952).

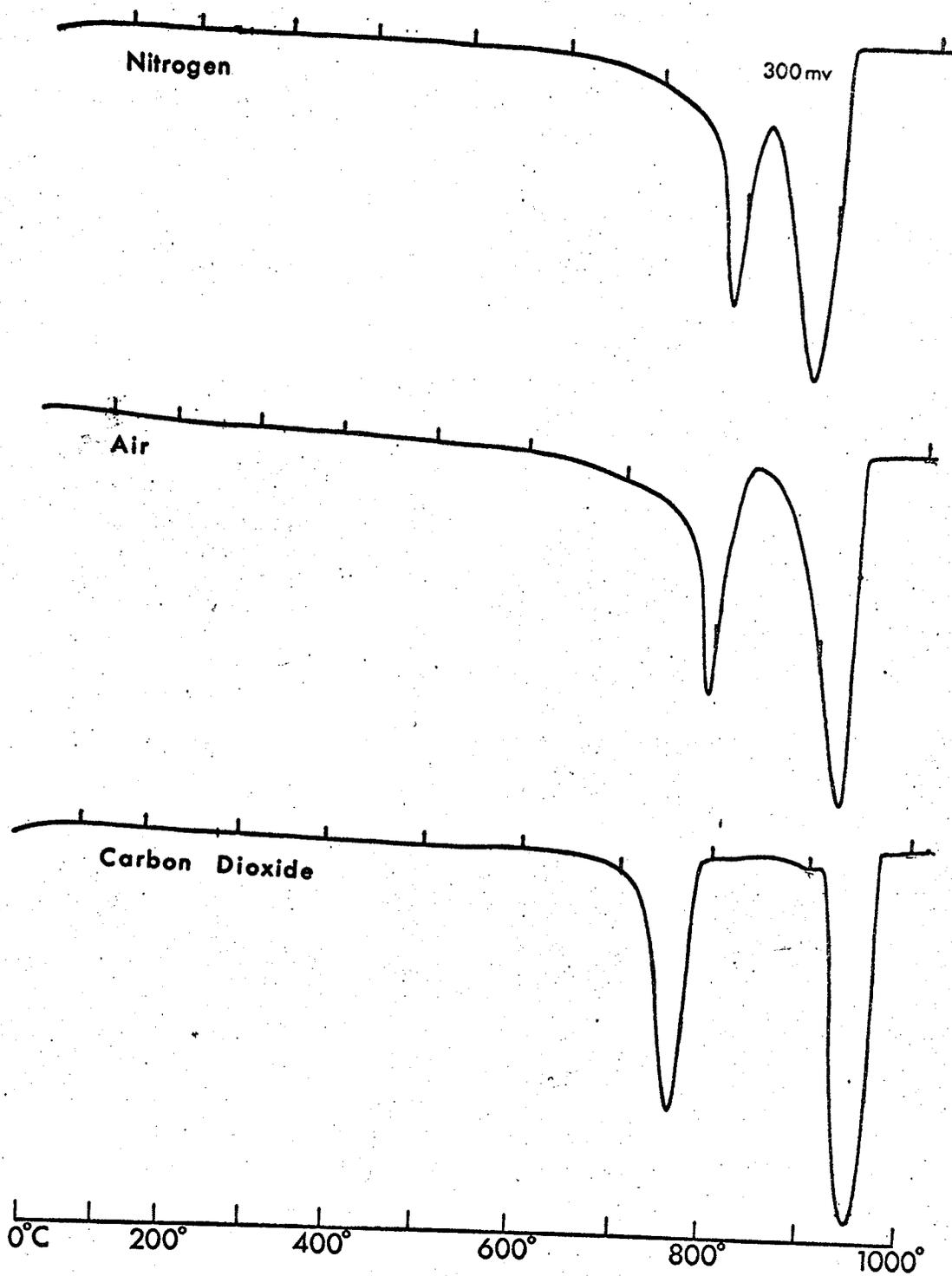


Figure 8. DTA curves of dolomite (Upper Dolomite Unit, Red River Formation, Manitoba) in dynamic nitrogen, static air, and dynamic carbon dioxide atmospheres.

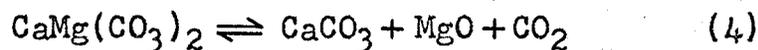
reaction:



and its peak temperature is related to the partial pressure of carbon dioxide by simple thermodynamic principles. On the other hand the low-temperature peak does not increase in peak temperature so that it obviously is not governed by the same principles and thus can not be represented by the simple equation:



but instead by the more complex equation:



Wilsdorf and Haul (1951) have demonstrated the validity of this assumption by X-ray diffraction studies. These authors found that the products of the low temperature reactions (600° and 640°C at 100 mm. CO₂) are fine grains of magnesium oxide and crystallites of calcium carbonate orientated in the directions as the original dolomite lattice. These orientated crystallites show considerable lattice distortion and their abundance is dependent on the rate and maximum temperature of heating. Samples heated to the higher temperature (800°C at 650 mm. CO₂) show considerably less orientation and more disorientated calcium carbonate, but the orientated crystallites never completely disappear.

Haul and Heystek (1952) thus explained the apparent immobility of the low-temperature peak as being due to the energy requirements necessary for diffusion, within the lattice, of magnesium ions to points of local concentration where they

form crystallites of magnesium oxide, simultaneously converting the surrounding dolomite lattice to calcium carbonate crystallites. It is because of these energy requirements, necessary to cause this diffusion which initiates reaction (4), that the reaction is restricted to the narrow temperature range at approximately 800°C.

In actual practice it has been found that the low-temperature dolomite peak can be totally shifted to a lower temperature (Fig. 9) by a process of rigorous and prolonged grinding of up to 400 hours (Bradley, Burst and Graf, 1953). In this case the energy required to cause the lattice diffusion necessary for equation (4) to take place is supplied by the grinding and thus the temperature of the dissociation of the magnesium carbonate is lowered to the temperature of ordinary magnesite.

The thermogram of the mineral calcite has a single endothermic peak generally between 930° and 1010°C, depending on experimental conditions (Webb and Heystek, 1951). The temperature of this peak, like the high-temperature peak of dolomite, is related directly to the partial pressure of the carbon dioxide in contact with the sample. Under similar conditions however, this peak occurs at a slightly higher temperature than the dissociation of calcium carbonate from dolomite. The reason for this difference in peak temperature appears to be the result of the difference in crystallite size, not grain

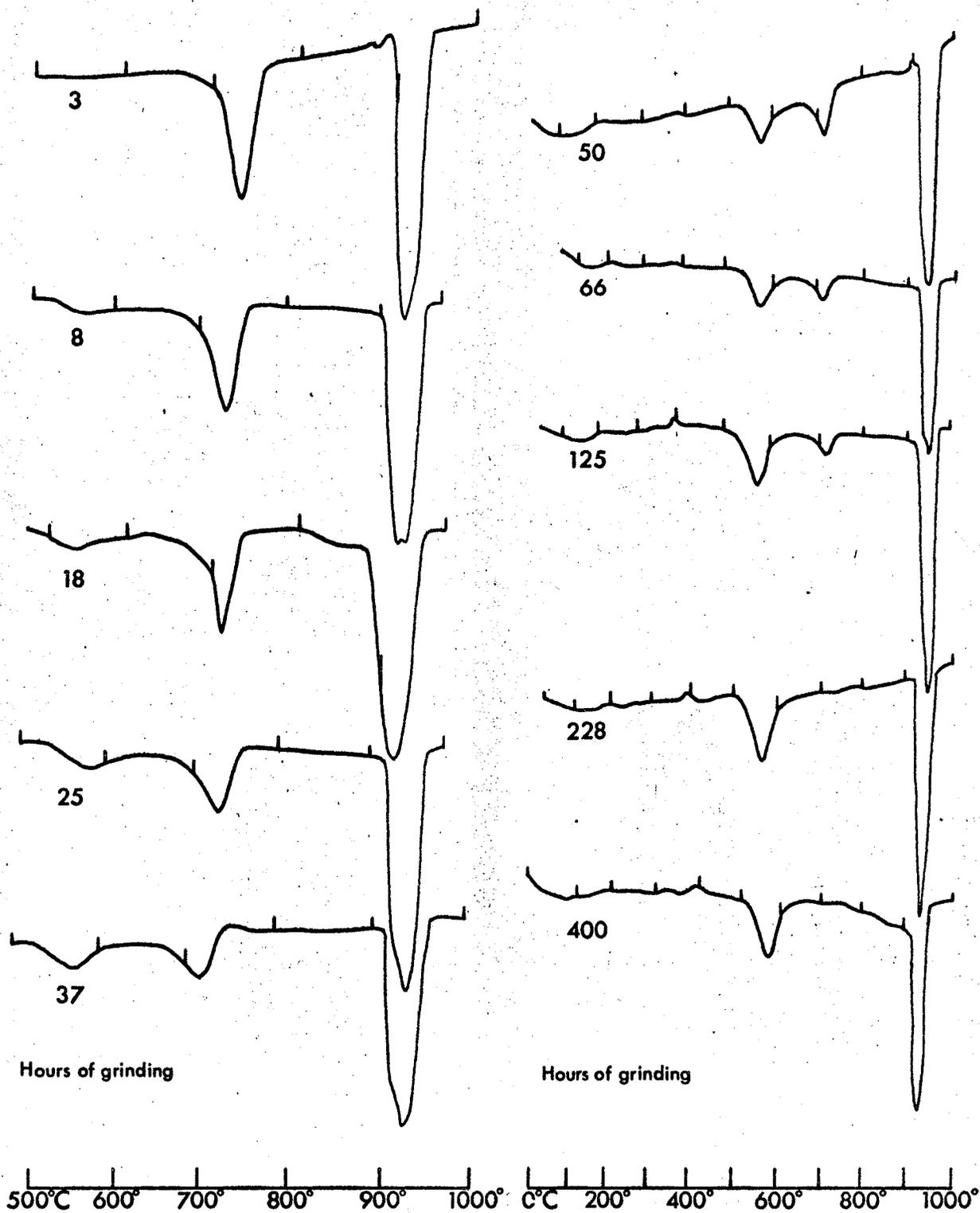


Figure 9. DTA curves of U.S.B.S. Dolomite No. 88 in static carbon dioxide, showing the effects of prolonged grinding. After Bradley, Burst, and Graf (1953).

size, between pure calcite and the calcium carbonate phase formed from dolomite. Rowland (1955) has demonstrated this difference in peak temperature with a series of curves (Fig. 10) of a pure calcite with an addition of small amounts of dolomite. In all cases the calcite peak at approximately 990°C was preceded by a small shoulder at 925°C due to the dissociation of the calcium carbonate from the dolomite. This effect can be seen in reverse in the curve Y-5/6, Fig. 27 (p.115) where a small amount of calcite is present in a dolomitic material. The calcium carbonate dissociation of the dolomite occurs as a large endotherm at about 920 to 925°C and the dissociation of the pure calcite occurs as a small endotherm at 945°C at the base of the dolomite peak.

Kulp, Kent and Kerr (1951) Graf (1952) and Webb and Heystek (1957) have investigated the variation of the peaks with the grain size of the material. In all cases some variation did occur in the very fine grain sizes. However for the Lake Agassiz sediments used in the present study all the material has the same origin and thus much the same grain size distribution so that the thermograms are all comparable to each other.

The effect of soluble salt impurities, mainly the chlorides of the alkaline metals, on the peaks of dolomite has been established by Berg (1943, 1945) and Graf (1952). These salts affect only the low-temperature peak of dolomite, but if

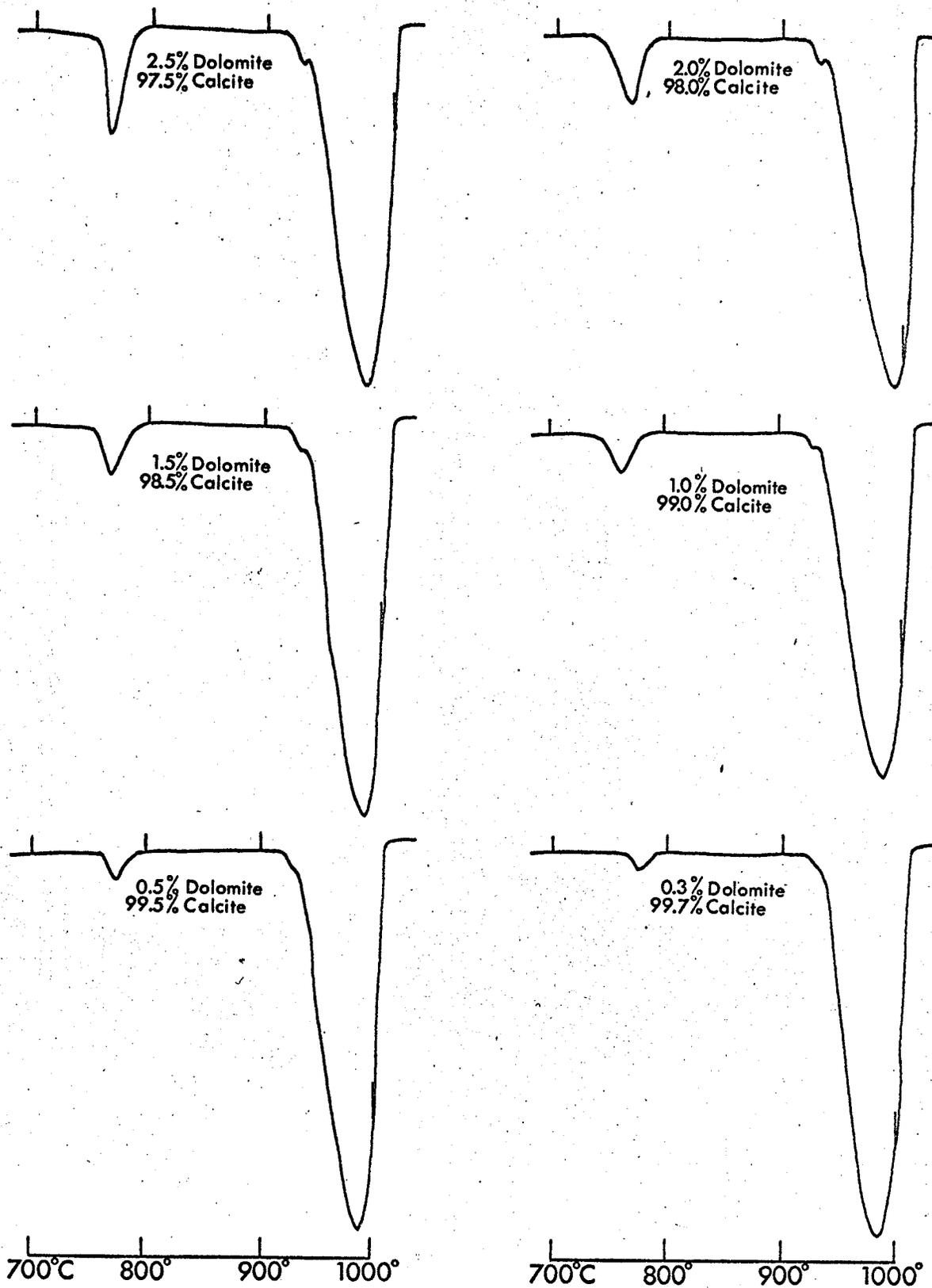


Figure 10. DTA curves of small percentages of dolomite in calcite. After Rowland (1955).

present even in trace amounts they cause the low temperature peak to start at a lower temperature and much more gradually than for normal pure dolomite. As the soluble salt content increases (Fig. 11) a small shoulder may develop in the 700° to 800°C region prior to the normal endothermic peak, and in extreme cases of high salt content the normal peak can be partially or totally shifted to a peak temperature of 750°C.

Graf (1952) experimented with an addition of 15% illite in dolomite to determine whether the illite caused any interference in the low-temperature dolomite peak. During the final dissociation of illite in the 900° to 1000°C region the central octahedral layer forms spinels and the outer tetrahedral layers combine with the interlayer cations, mainly potassium, to form an amorphous alkaline silicate phase. It was this alkaline glass that Graf thought might alter the low-temperature dolomite peak. He concluded that the alkaline glass did not effect the low-temperature dolomite peak because the glass did not form until after the low-temperature peak was completed. He failed to note, probably because he was using a static air atmosphere, what was later stated by Rowland (1955), that the high-temperature peak of the dolomite in a carbon dioxide atmosphere can be partially or completely eliminated by the presence of "...a sericite-like mica". In this case the carbon dioxide atmosphere elevates the high-temperature peak up into the region of the alkaline silicate glass so that a reaction does take place. This effect can be noted

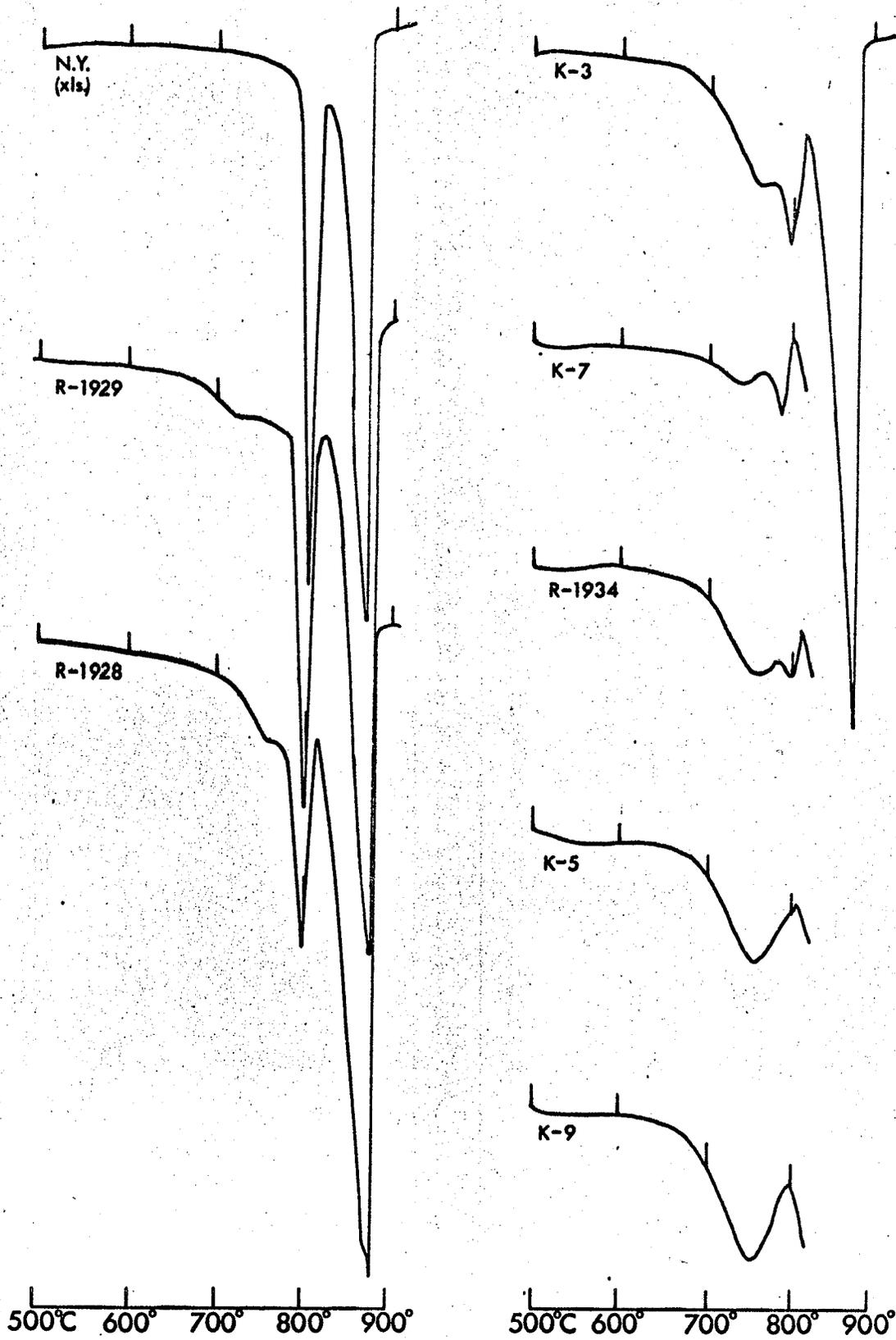


Figure 11. DTA curves of dolomite showing the effect of increasing amounts of soluble salts. After Graf (1952).

particularly in curves Y-14 and Y-15, Fig. 24, (p.112) where a definite low-temperature dolomite peak is present but where there is no trace of the high-temperature dolomite peak.

QUANTITATIVE DETERMINATIONS OF DOLOMITE AND CALCITE

The initial thought in the present study was that a simple measurement of the total area under the low-temperature and high-temperature peaks of the dolomite could be used for an estimation of the dolomite content. The actual calcite content was thought to be very small as suggested by the small peak at 950°C, at the base of the high-temperature dolomite peak on the curves in Fig. 16, (p.104). A peak of this order of magnitude was thought to be insignificant in the estimations. This simple solution was soon discarded when results such as the curves in Figs. 22 to 25 were recorded. In these, although the total carbonate content is small, a definite low-temperature dolomite peak develops but no peak or only a slight trace of the high-temperature peak develops. The shape of the endothermic-exothermic peak in this region is broadly "S" shaped, but irregularities and modifications from the normal shape are common in these curves. This indicates that a reaction is taking place between the alkaline glass being formed from the clay minerals, and the calcium carbonate, both the calcite proper and the calcium carbonate from the dolomite. The end result is a complete or nearly complete obliteration of the carbonate peaks, and a modification of the clay minerals exothermic peaks as described by Rowland (1955) and discussed above.

As small amounts of illite-montmorillonite are present even in the high carbonate till samples, only the low-temperature dolomite peak is unaffected by the alkaline silicate glass, and can be used for estimations of the carbonate content. Even these low-temperature dolomite peaks show the signs of a slight alkaline salt abnormality by the gradual early commencement of the peak at approximately 630°C in the till samples, (Figs. 16 and 17, pp. 104, 105). This effect is slight and uniform for all the samples so it does not prevent the peak from being used for quantitative estimations.

Fig. 12 shows the plot of the peak area of the low-temperature dolomite peak versus the chemically determined total carbon dioxide content of several samples of Lake Agassiz materials. The values used for plotting are given in Table 6. All points, with one exception, fall within 1% of the best curve. The single point that does not plot well is 3.5% off the best curve. This sample has a different origin than the other samples and X-ray work (Chapter VI) shows that it contains very little calcite. The other samples all have substantial amounts of calcite with the dolomite (Table 6); the calcite content increases uniformly, relative to the dolomite, with decreasing carbon dioxide content. This curve can, therefore, be used to estimate the total carbon dioxide content of any given sample of similar origin (till or lacustrine) to the samples used to plot the curve. Careful measurement of the dolomite peaks can yield estimations of the total carbon dioxide content within 1%.

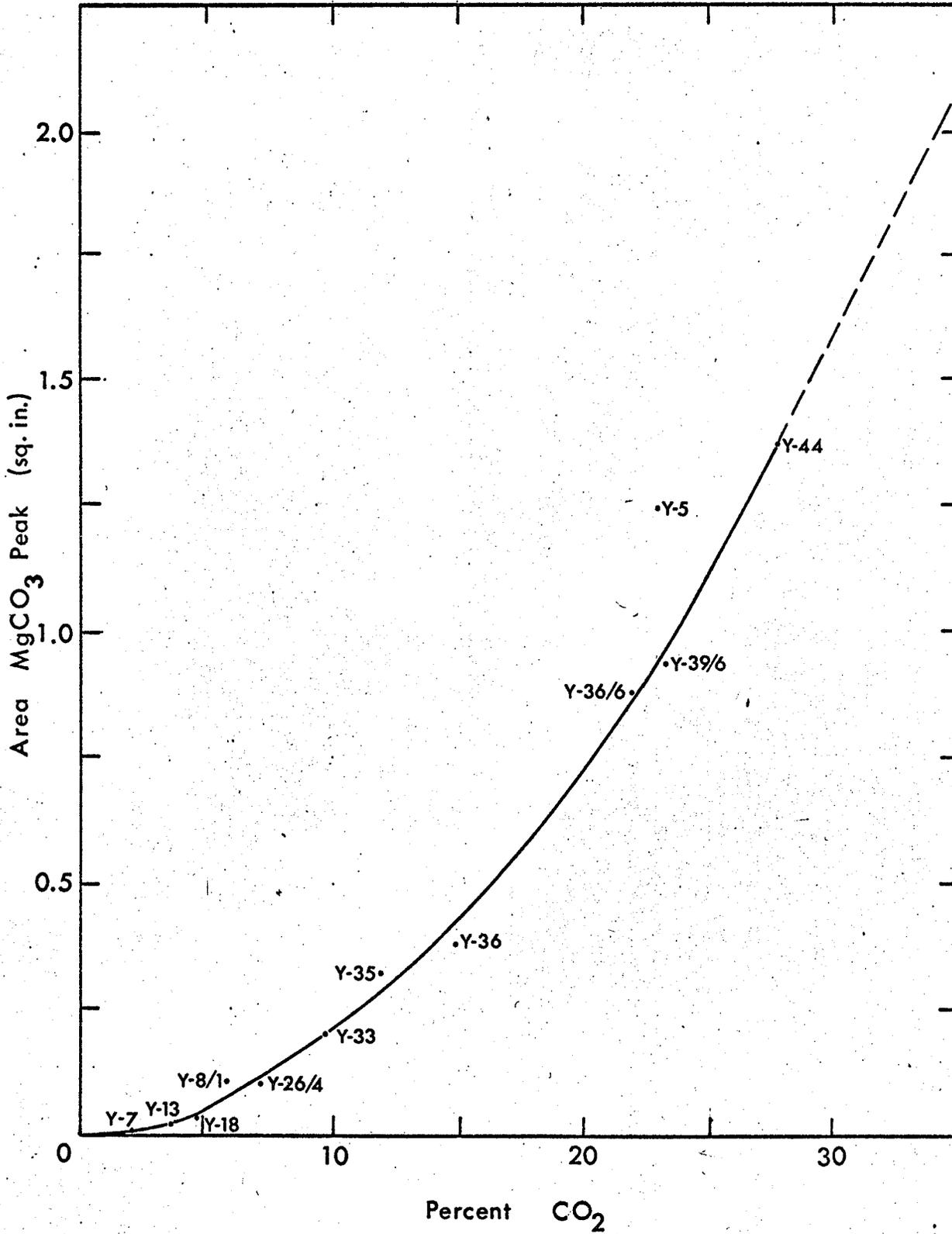


Figure 12. Calibration curve for carbon dioxide.

Table 6

Data Relating CO₂ Content to MgCO₃ Peak Area
For Calibration Curve

<u>Sample Number</u>	<u>MgCO₃ Peak Area in Sq. In.¹</u>	<u>Wt.% CO₂²</u>	<u>Ratio³</u> <u>% Calcite vs % Dolomite</u>	
Y-7	0.006	2.01	37	63
Y-13	0.013	3.53	36	64
Y-18	0.027	4.67	37	63
Y-8/1	0.109	5.78	33	67
Y-26/4	0.101	7.05	30	70
Y-33	0.203	9.71	26	74
Y-35	0.333	11.86		
Y-36	0.384	14.86		
Y-36/6	0.880	21.94		
Y-5	1.240	22.87	<5	>95
Y-39/6	0.940	23.21	21	79
Y-44	1.380	27.63	17	83

¹Measured from thermograms

²Determined by standard chemical analysis

³Estimated from X-ray powder photographs

QUARTZ

Quartz and the various silica minerals have been subjected to extensive DTA research. This work is reviewed thoroughly by Grimshaw and Roberts (1957a). The inversion from alpha to beta quartz, or the reverse, takes place very rapidly at $573^{\circ}\pm 1^{\circ}\text{C}$, (Fig. 13). Variations from 573°C are very small, usually less than 1°C , except in rare cases where impurities are present. The grain size of granular samples has no effect on the size of the peak or the peak temperature, but the procedure for sample preparation is very important. Grinding quartz produces a coating of non-quartz material on the particles and low values are obtained if an estimation of the total quartz content is made. The value obtained is real however, in that it represents the actual quartz content of the sample after the grinding, but not the actual quartz content of the original material.

According to Grimshaw and Roberts (1957a), if care is taken in the sample preparation, accuracies of 1% can be achieved in quartz content estimations. They present a graph showing a linear relationship between the peak height of the quartz inversion and the total quartz content of a sample.

To establish a quartz calibration curve for the Agassiz sediments a series of standard mixtures of a very pure quartz sand (Table 7) and a very low quartz, Agassiz clay (Y-6/9, Figs 13 and 27) were run and rerun in dynamic nitrogen.

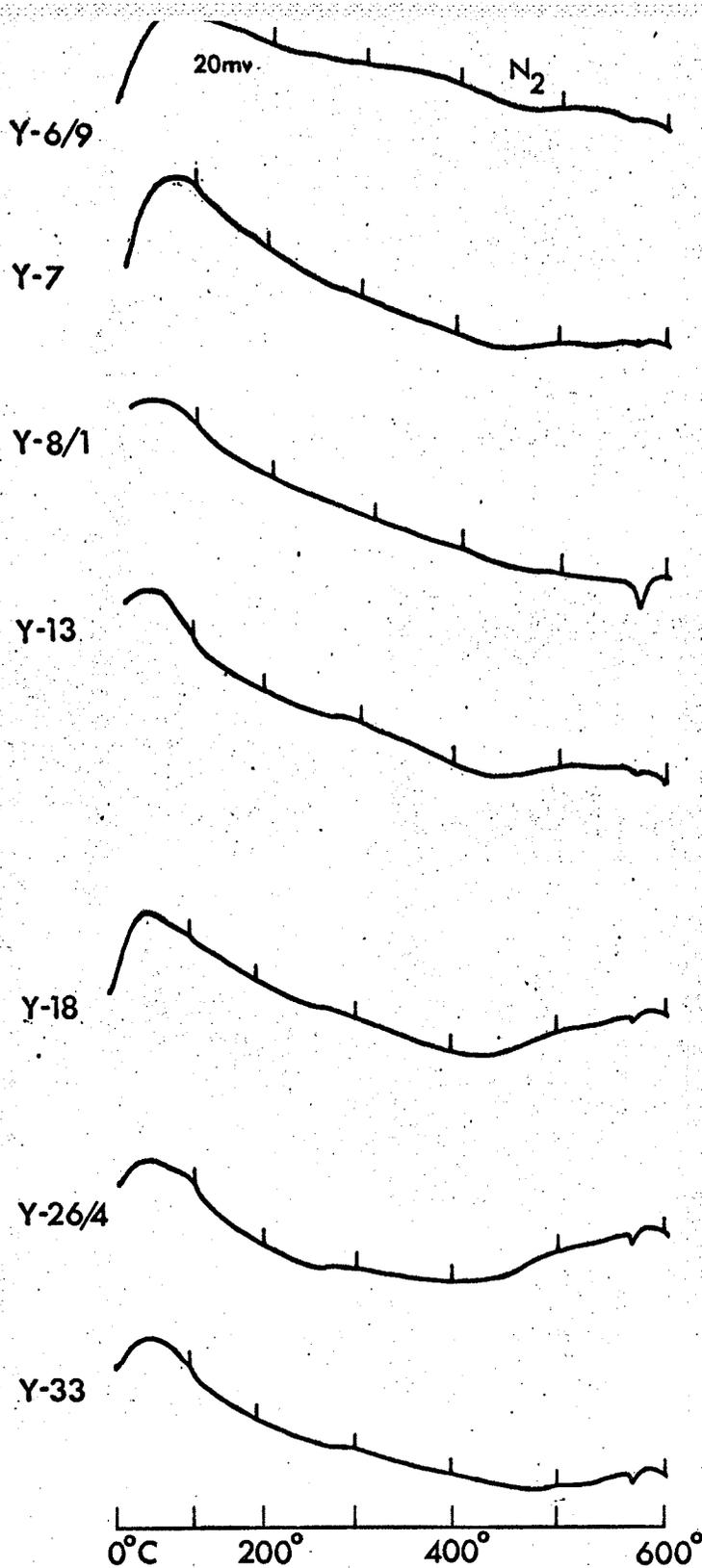


Figure 13. DTA curves of several Yale Avenue samples rerun for precise quartz determinations. Compare with Figures 18, 21, 23, 24, 26, and 27.

Table 7

Chemical Analysis
Of Selkirk Silica Sand¹

	Wt %
SiO ₂	99.588
Al ₂ O ₃	0.219
TiO ₂	0.0052
Fe ₂ O ₃	0.020
MgO	0.037
CaO	0.018
Ignition Loss	0.106
Total	99.993

¹National Testing Laboratories Limited,
Winnipeg, Manitoba

It was discovered that the clay peak partially masked the quartz peak in samples with moderate to low quartz contents, so all calibration samples were run twice before the quartz peak was measured.

The samples used in Chapter VI for careful quantitative chemical work were also run twice. The difference in results can be seen by comparing the curves on Fig. 13 with the appropriate curves on Figs. 18, 21, 23, 24, 26 and 27. In spite of this difference the quartz content for the rest of the samples was estimated from the original DTA curve. This resulted in an error, 2 - 3% low, in the low quartz samples but it still allowed the variations between samples to be determined and it saved considerable time.

Since it was impossible to get an entirely quartz-free Agassiz clay sample the quartz content of the clay Y-6/9, used in the standard mixtures, had to be estimated. This was done by assuming a linear relationship between the area of the quartz peak and the quartz content, and a straight line was drawn from the 100% quartz point to the origin, dashed line Fig. 14. Plotting the area of the quartz peak for Y-6/9 on this line yielded a quartz content of 3%. All the quartz contents were readjusted, Table 8, and plotted against the measured peak area, solid line, Fig. 14. As Piloyan (1965) has shown, this does not yield an exact linear relationship, but makes little change in the estimation of 3% quartz in the Y-6/9

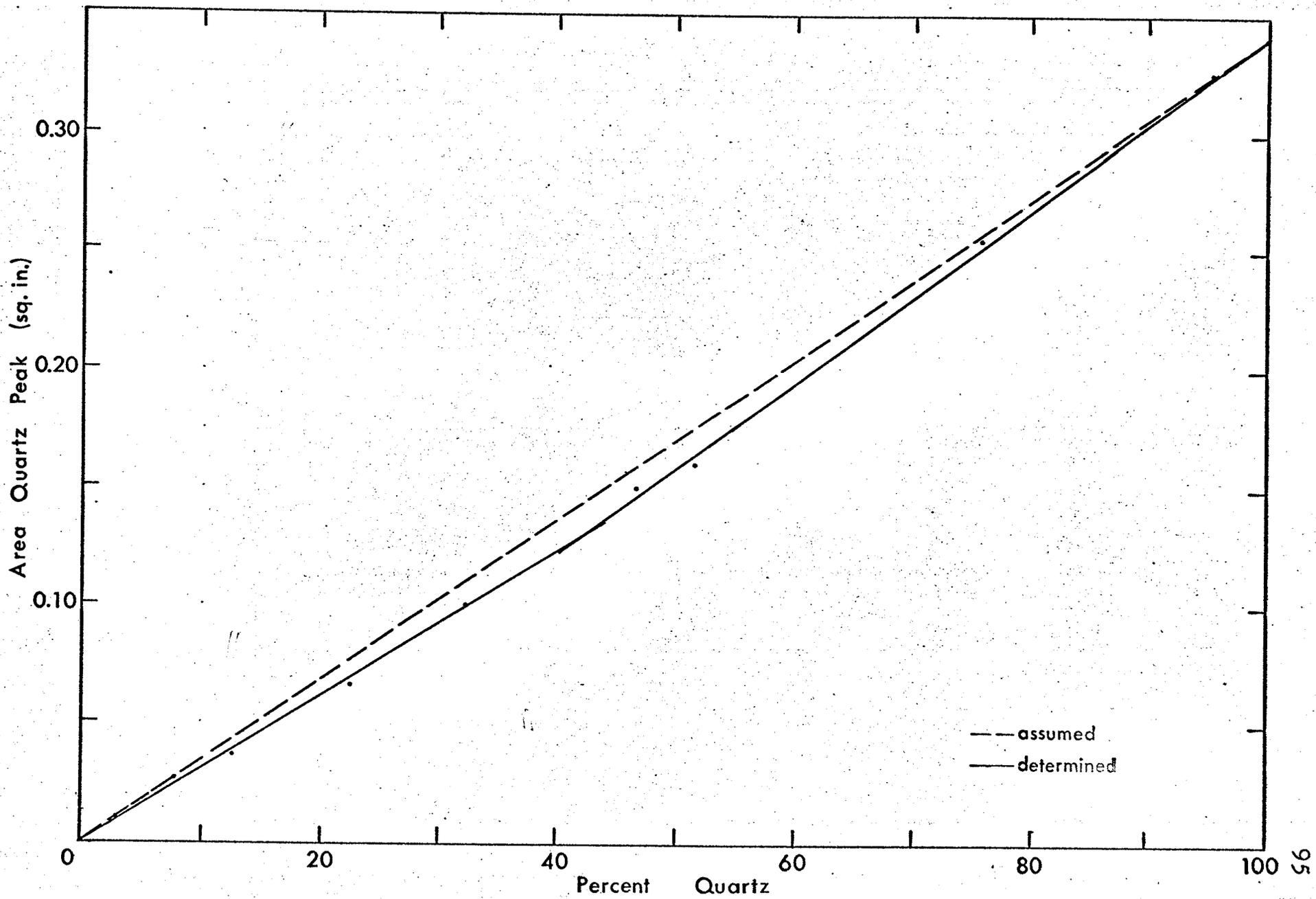


Figure 14. Calibration curve for quartz.

Table 8Data for Quartz Calibration Curve

Wt of Quartz	Area of Peak in Sq. Inches
100.0	0.340
95.15 ¹	0.326
75.75	0.255
51.50	0.170
46.65	0.150
32.10	0.095
22.40	0.066
12.70	0.037
7.85	0.026
3.00 ²	0.010

¹ total quartz = (wt. of quartz added) + (wt. of quartz in clay)

$$\text{wt. of quartz in clay} = \left(\frac{\% \text{ Clay} \times 3\%}{100} \right)$$

² estimate % of quartz in the original clay sample

sample.

All points lie within 1% of the line so an error of at least 1% can be expected for very careful work where the quartz peak measurements are determined after a second run. For routine work where the area is measured from the original DTA curve the results will be 2 to 3% low for samples with low clay contents (Tables 14 and 15, pp.146 and 150).

GYPSUM

Gypsum is one of the minor minerals found in this study but it does give a large, sharp endothermic peak and thus small quantities of it can be detected. McLaughlin (1957) states that under static conditions in air gypsum produces a large endothermic peak at 105°C corresponding to the loss of $1\frac{1}{2}$ molecules of water per formula unit, and a second smaller endothermic peak at 150°C corresponding to the loss of the remaining $\frac{1}{2}$ molecule of water. An exothermic peak generally occurs at 380°C (Kaufman and Dilling, 1950) when the soluble anhydrite changes to insoluble anhydrite.

The first curve, Fig. 15, is a thermogram of 50% gypsum - 50% alumina in a static atmosphere of air. The first endothermic peak occurs at 150°C, the second at 190°C, and the exothermic peak (very minor) occurs at 360°C. In a dynamic nitrogen atmosphere the two endothermic peaks change to a peak at 150°C with a shoulder at 170°C while the exothermic peak remains unchanged although it is stronger. The 1% gypsum sample

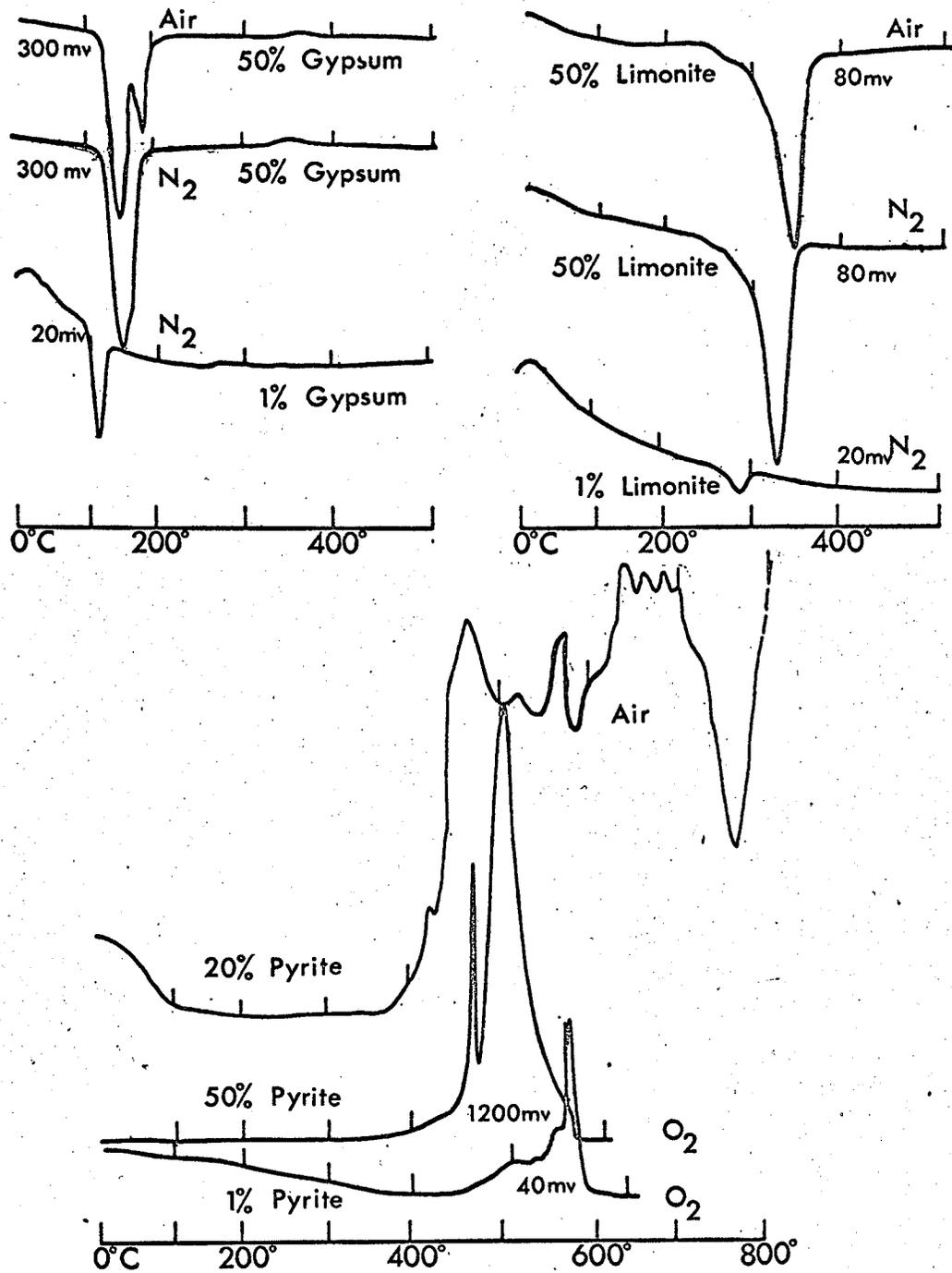


Figure 15. DTA curves of various amounts of gypsum, limonite, and pyrite in static air and dynamic nitrogen or oxygen.

gives a single peak at 115°C. The restriction of the gypsum endothermic peaks by the nitrogen atmosphere, although not particularly useful for the study of gypsum is very helpful for clay studies for it allows the more important details of the interlayer water peak to be recorded while still indicating the presence of gypsum.

Gypsum is present in many of the Agassiz samples studied but the peak, in most cases, is smaller than the 1% standard curve (Fig. 15) so no calibration curve was composed for this mineral. The strong, sharp peaks would lend themselves well to semi-quantitative estimations if appreciable quantities of the mineral did occur in a sample.

HYDROUS IRON OXIDES (LIMONITE)

Hydrous iron oxides exist in several forms, all giving endothermic peaks in the 200° to 400°C region (MacKenzie, 1957a). The endothermic peak at 250°C on the nitrogen thermograms of the Agassiz silts and clays is identified as being due to hydrous iron oxides or more commonly limonite. Curve Y-8/1, Fig. 26, p.114, shows this peak at its greatest development in the yellow silt. A sample of limonite from the University of Manitoba Geology Department Museum diluted to 50% with alumina run in air and in dynamic nitrogen (Fig. 15) yields endothermic peaks at 345° and 325°C respectively. The same limonite diluted to 1% with alumina gives a small peak at 290 C in dynamic nitrogen, (Fig. 15). These results

are slightly higher than those obtained from the Agassiz silts and clays; however the difference could be due to any one of several factors, such as a greater degree of crystallinity, a large crystallite size or a slight difference in composition of the museum sample over the material occurring in the Agassiz sediments.

PYRITE

The slight exothermic shoulder in the 400° to 500°C region of most of the Agassiz clay thermograms in dynamic oxygen is interpreted as being due to very minute traces of pyrite. Large quantities of pyrite in an oxidizing atmosphere give erratic differential thermal analysis results. McLaughlin (1957) describes a series of irregular exothermic peaks occurring between 355° and 700°C (20% pyrite curve, Fig. 15). The reason given for the extreme irregularity of the peaks is that the rapid oxidation of the pyrite depletes the oxygen supply so that oxidation is forced to stop until more oxygen is available to the pyrite. This produces an exothermic peak during oxidation periods and endothermic troughs during periods of oxygen depletion. In static air this produces a series of peaks and troughs; in dynamic oxygen where more oxygen is available only two peaks were produced: at 470° and 505°C (Fig. 15).

Small quantities of pyrite, in static air, produce a single exothermic peak at 445°C (McLaughlin, 1957). The thermogram of 1% pyrite in alumina in dynamic oxygen yields a

series of exothermic peaks between 450° and 560°C . These peak temperatures are higher than that cited by McLaughlin and the peak found in the Agassiz clays. However, the difference could be due to the difference between the fresh well crystallized museum sample used for the standard curve, and the fine grained, poorly crystallized pyrite found in the clay. Thus although it is not possible to positively attribute the slight shoulder at 450°C to minute traces of pyrite it appears to be very likely that it is due to pyrite; however, the possibility that it is due to organic matter can not be dismissed.

ORGANIC MATTER

Investigations into the carbonaceous material found in clay soils and rocks have not been extensive. Grimshaw and Roberts (1957b) give tabulations of the characteristic peaks of a series of carbonaceous material ranging from sawdust to graphite. The type of material found in clay soils is somewhat akin to peat or sawdust used in their tabulations. Two exothermic peaks characterize this organic material, one at 345° to 380°C and the second at 435° to 450°C . The first peak as described by Grimshaw and Roberts is due to the release of volatile matter from the organic compounds while the second peak is due to the burning of fixed carbon.

For young organic matter the first peak is dominant, but as the geologic age of the material increases it becomes converted from fresh organic matter towards the coals and the second peak, the fixed carbon peak, becomes dominant. Thus

the ratio of the two peaks could be used to determine the rank of the organic matter and hence be used in a crude way for age estimations.

The curves of illite contaminated with some organic matter (Fig. 3, p. 62) illustrate the effect of various dynamic gas atmospheres on the standard static atmosphere oxidation curve. The two peaks occur at 340°C and 405°C in air under static conditions, and at 315° and 375°C in a flow of oxygen. The use of nitrogen completely suppresses the oxidation.

CHAPTER V

QUALITATIVE MINERALOGY

INTRODUCTION

Mineral identification was made primarily from the thermograms, but X-ray powder photographs were also used to verify the DTA results and to identify thermally inert minerals. The samples were prepared by the methods described in Appendix D. The thermograms were run according to the procedure described in Appendix E. In all cases in the figures in this chapter the endothermic reactions point down the page and the exothermic point up the page.

DIFFERENTIAL THERMAL ANALYSIS

Gray Till

The minerals in the till identified by DTA are dolomite, calcite, quartz, normal illite-abnormal montmorillonite, and gypsum. Figs. 16 and 17 show the thermograms of the samples recovered from the Yale Avenue test hole. Curve G-50 (Fig. 28) is the thermogram of the single till sample recovered from the Geology Building test hole, and Curve WR-1 (Fig. 28) is the thermogram of a sample taken from an outcrop of buff till exposed on the north side of the Assiniboine River just west of the St. James Bridge.

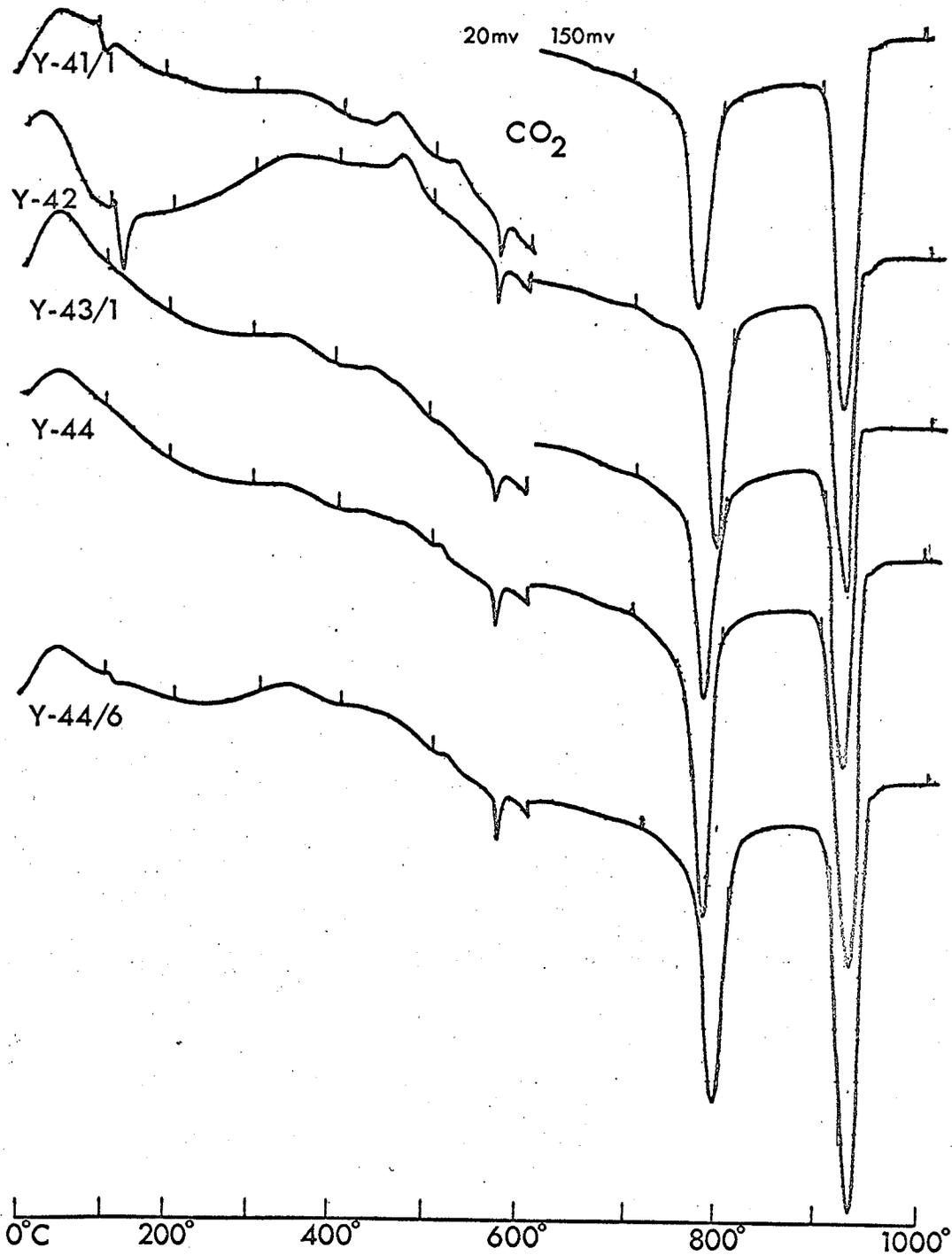


Figure 16. DTA curves, in dynamic gases, of Gray Till samples, Y-41/1, Y-42, Y-43/1, Y-44, and Y-44/6 from the Yale Avenue test hole.

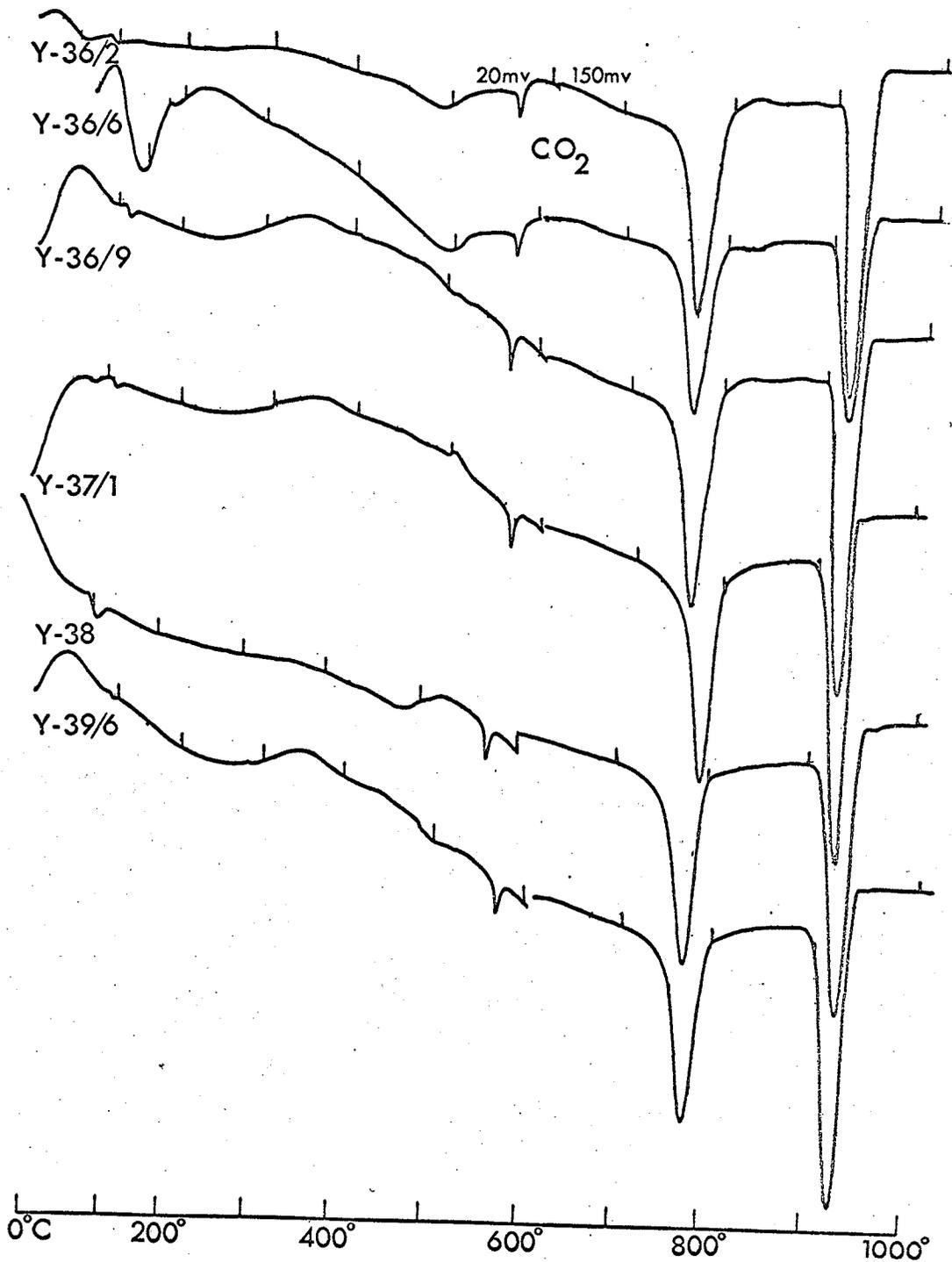


Figure 17. DTA curves, in dynamic gases, of Gray Till samples Y-36/2, Y-36/6, Y-36/9, Y-37/1, Y-38, and Y-39/6 from the Yale Avenue test hole.

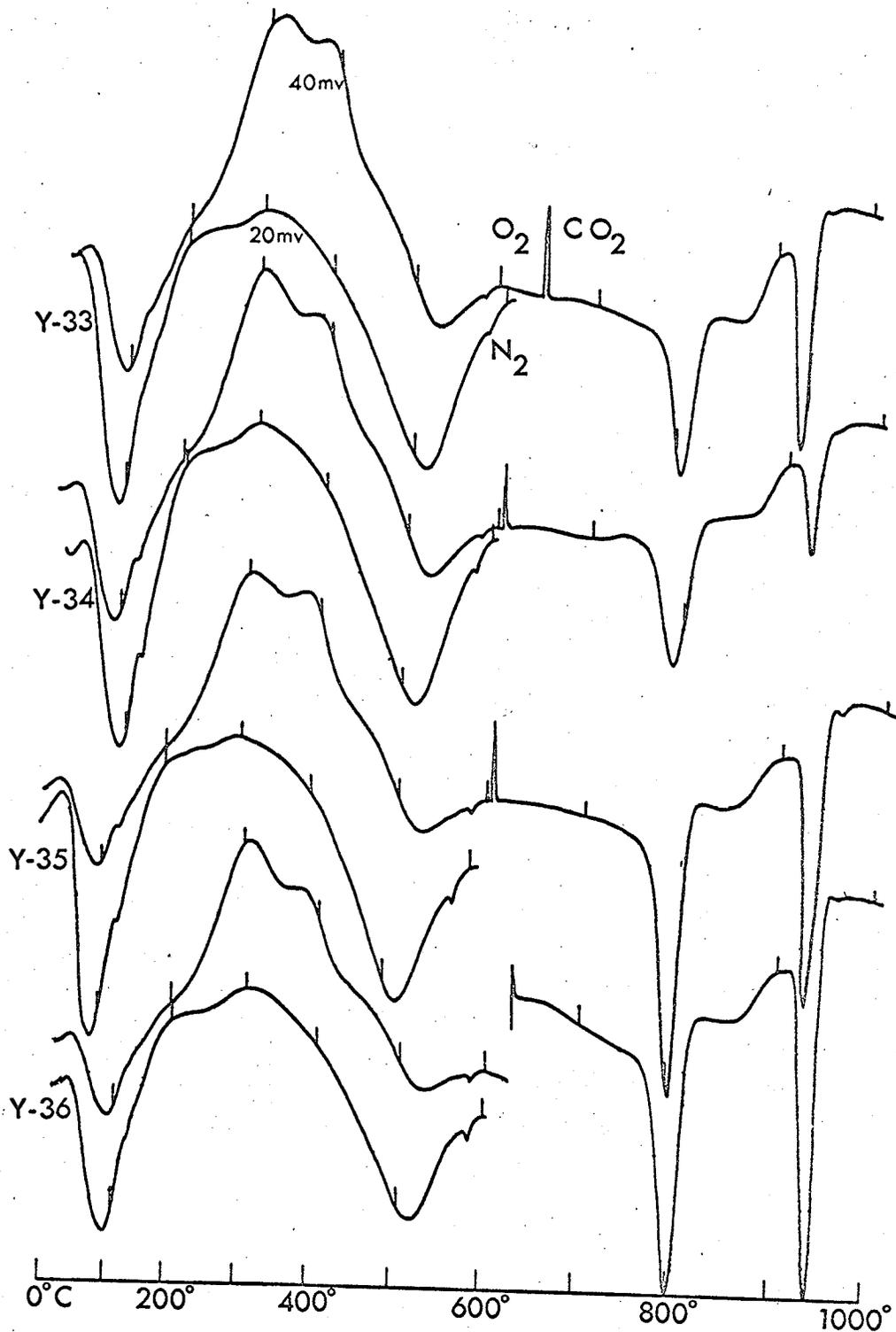


Figure 18. DTA curves, in dynamic gases, of Blue-gray Clay samples Y-33, Y-34, Y-35, and Y-36 from the Yale Avenue test hole.

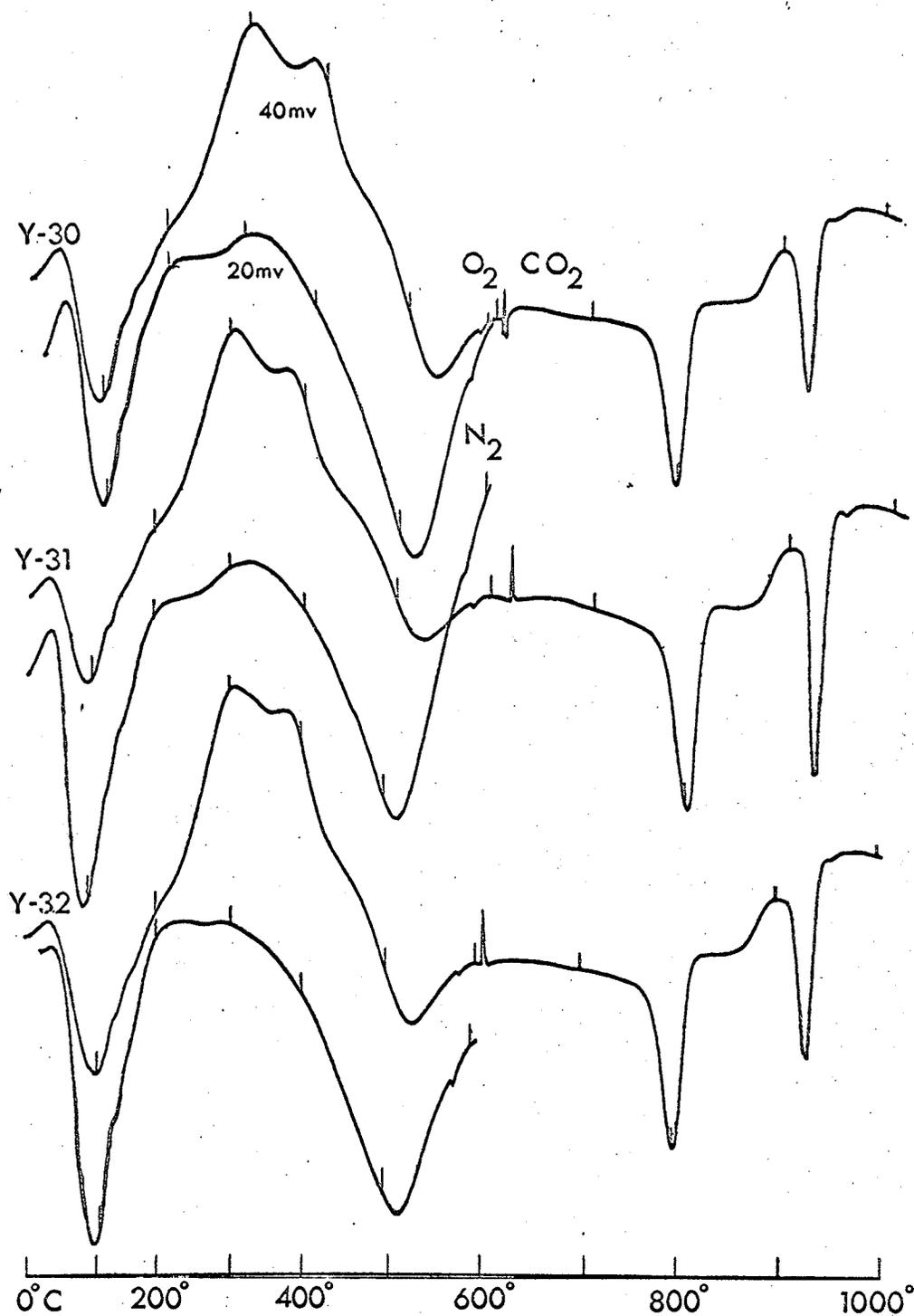


Figure 19. DTA curves, in dynamic gases, of Blue-gray Clay samples Y-30, Y-31, and Y-32 from the Yale Avenue test hole.

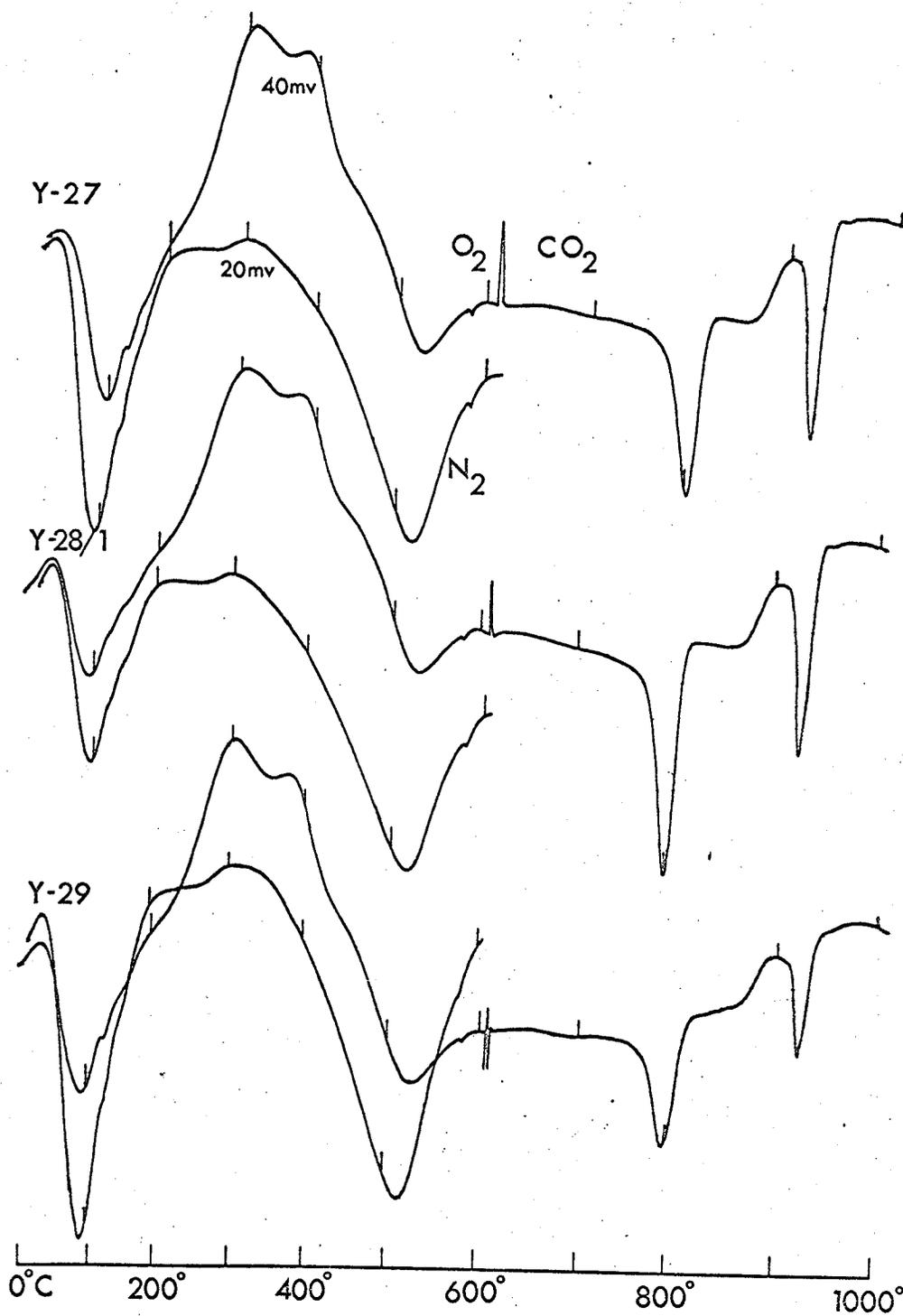


Figure 20. DTA curves, in dynamic gases, of Blue-gray Clay samples Y-27, Y-28/1, and Y-29 from the Yale Avenue test hole.

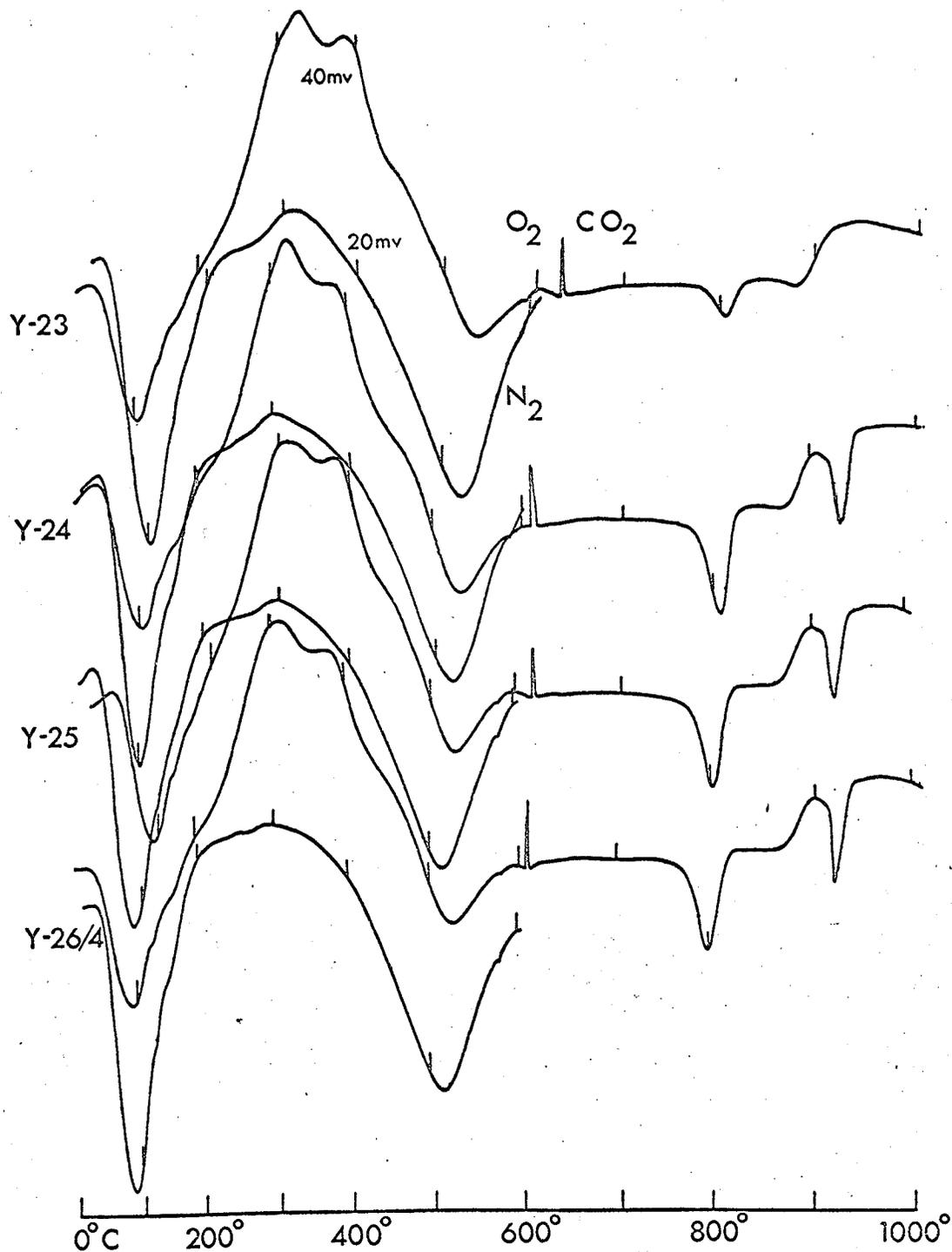


Figure 21. DTA curves, in dynamic gases, of Blue-gray Clay samples Y-23, Y-24, Y-25, and Y-26/4 from the Yale Avenue test hole.

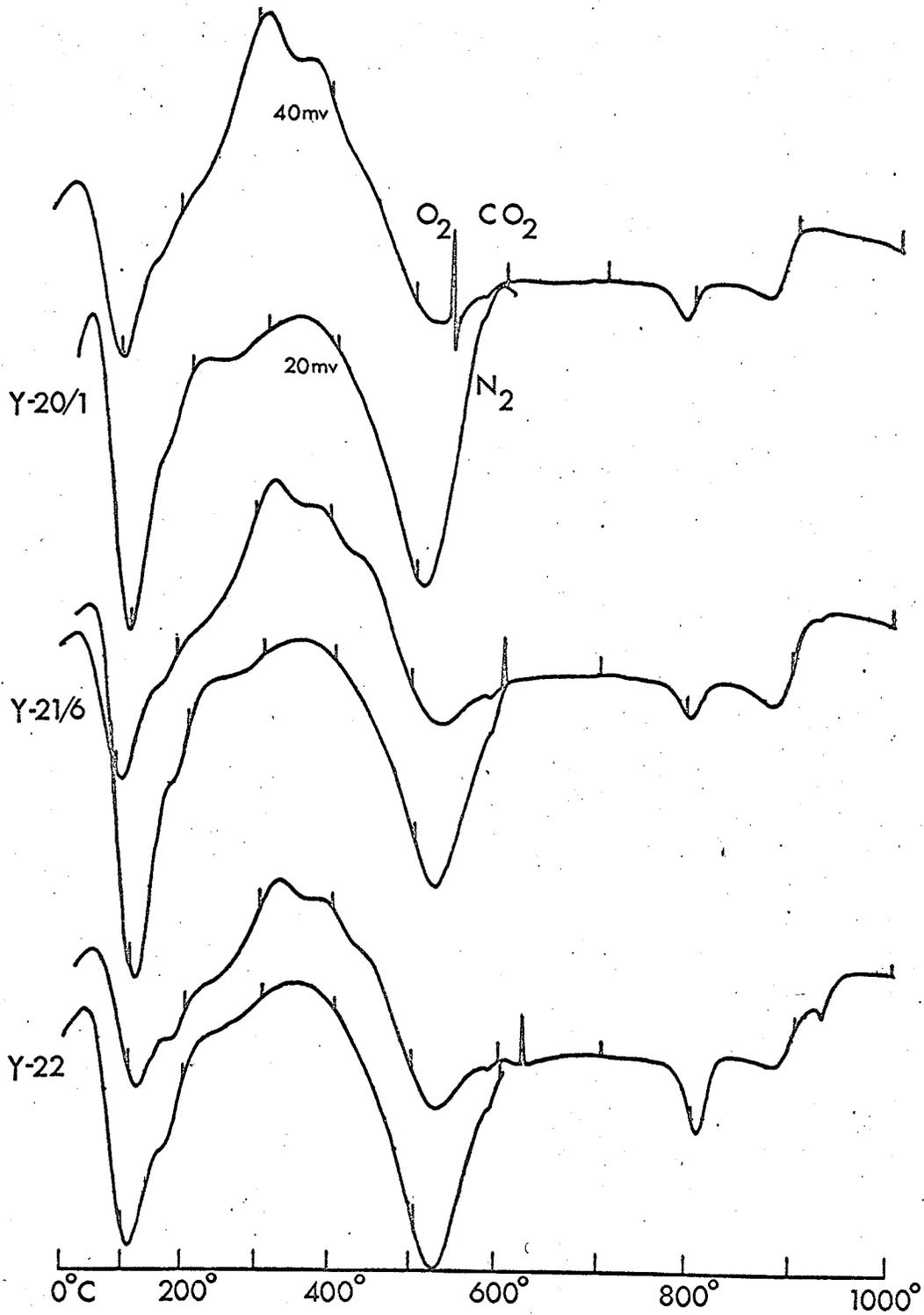


Figure 22. DTA curves, in dynamic gases, of Brown Clay samples Y-20/1, Y-21/6, and Y-22 from the Yale Avenue test hole.

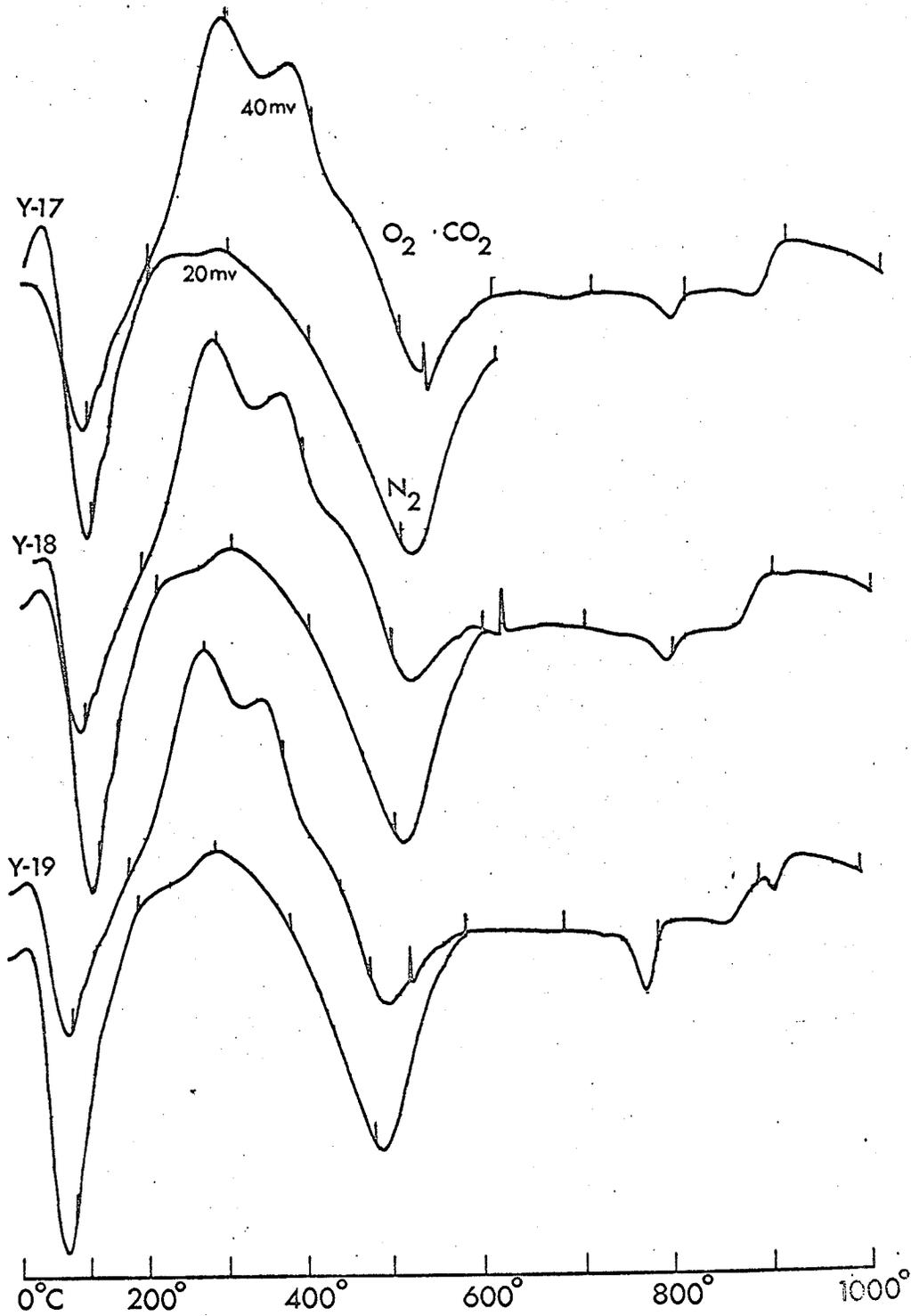


Figure 23. DTA curves, in dynamic gases, of Brown Clay samples Y-17, Y-18, and Y-19 from the Yale Avenue test hole.

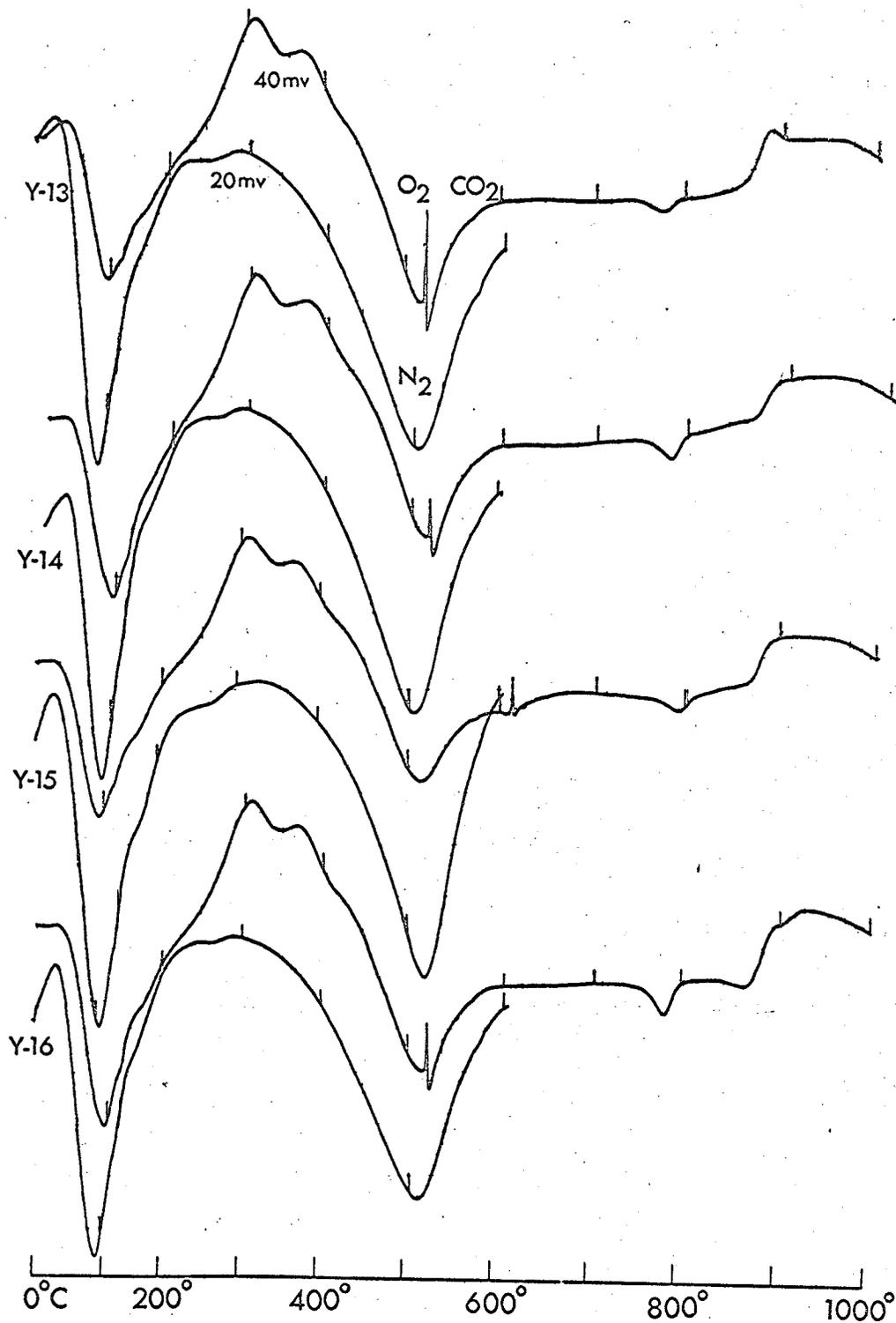


Figure 24. DTA curves, in dynamic gases, of Brown Clay samples Y-13, Y-14, Y-15, and Y-16 from the Yale Avenue test hole.

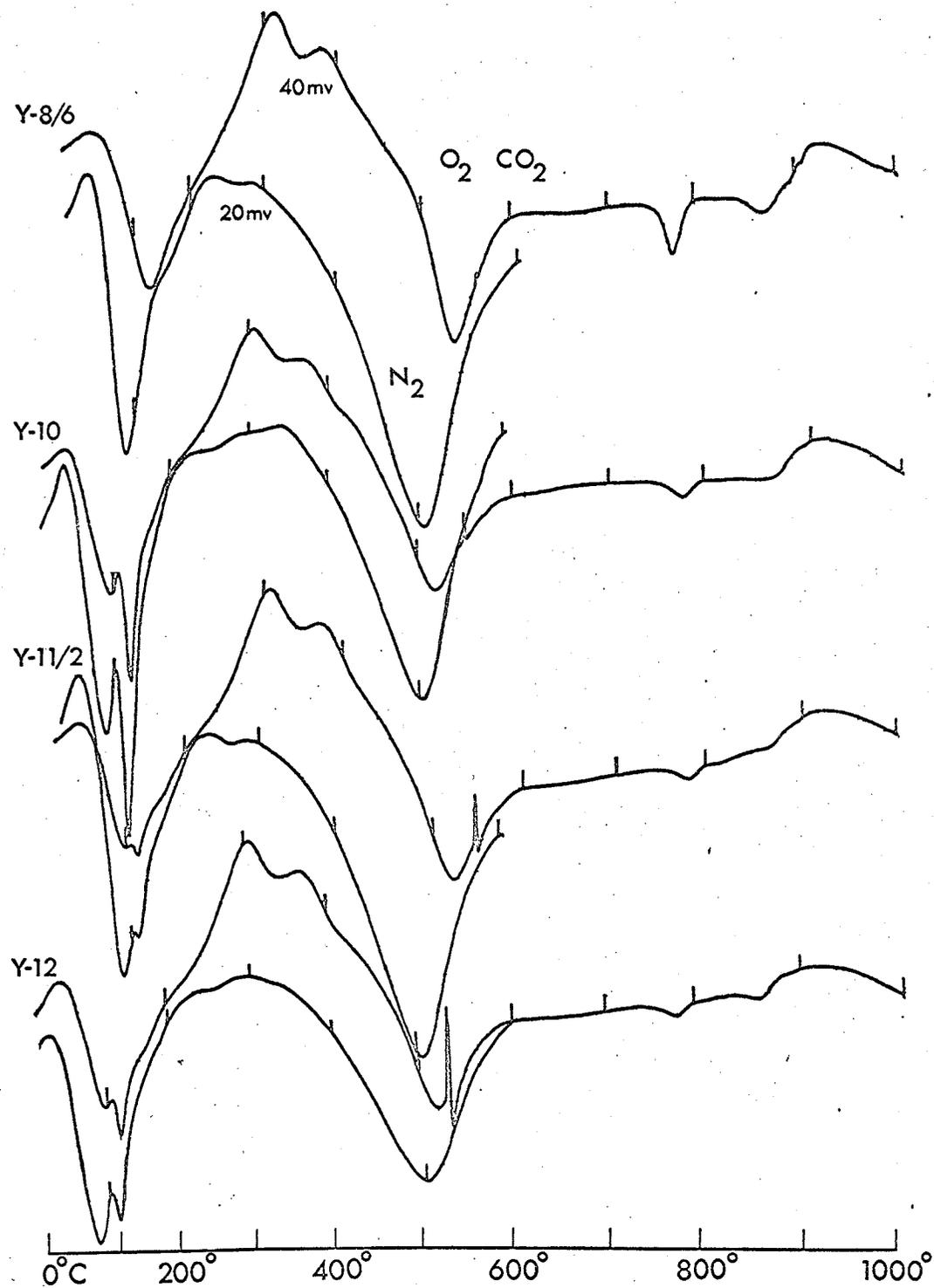


Figure 25. DTA curves, in dynamic gases, of Brown Clay samples Y-8/6, Y-10, Y-11/2, and Y-12 from the Yale Avenue test hole.

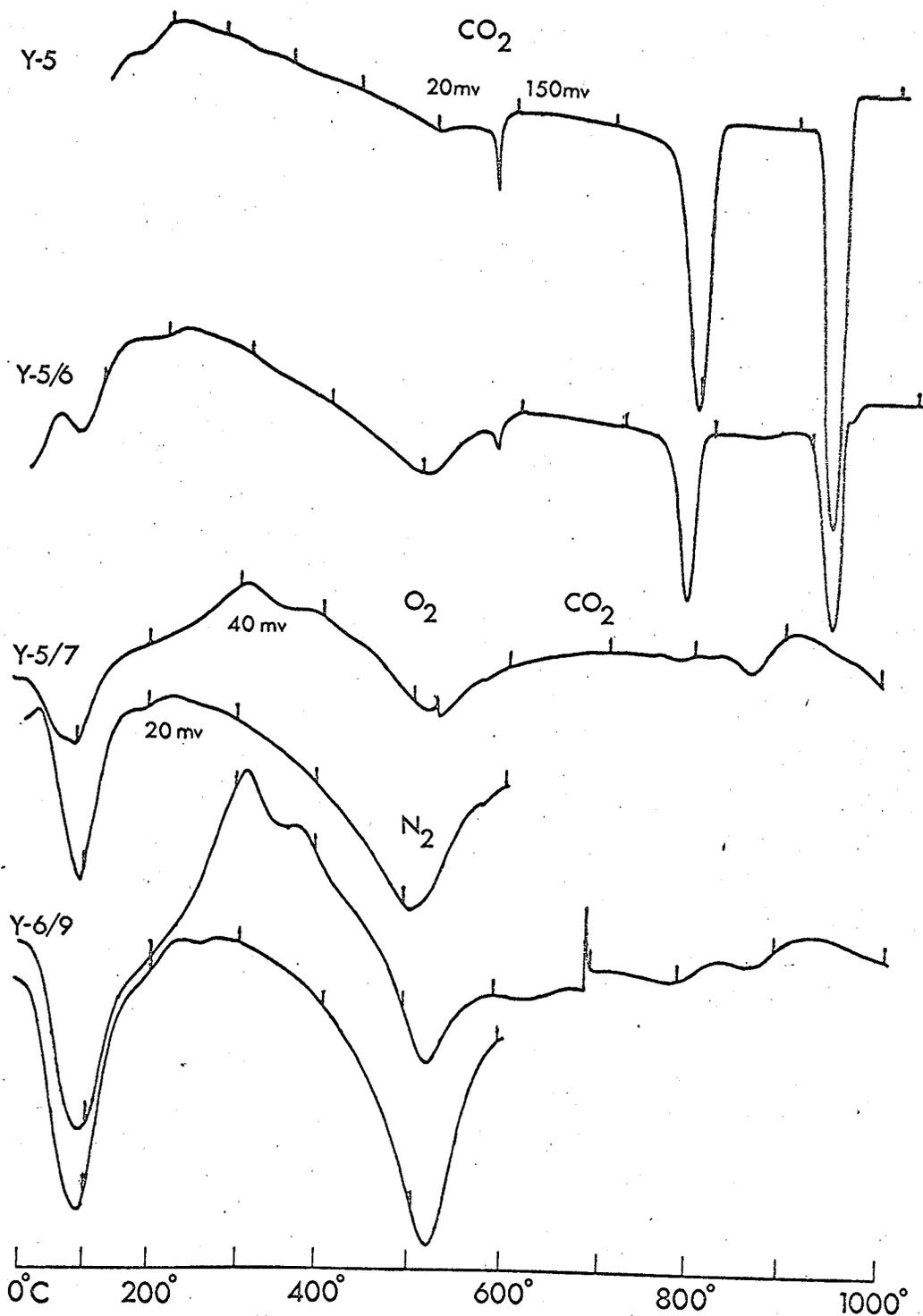


Figure 27. DTA curves, in dynamic gases, of Fine Sand samples Y-5, and Y-5/6, and Greenish-brown Clay samples Y-5/7, and Y-6/9 from the Yale Avenue test hole.

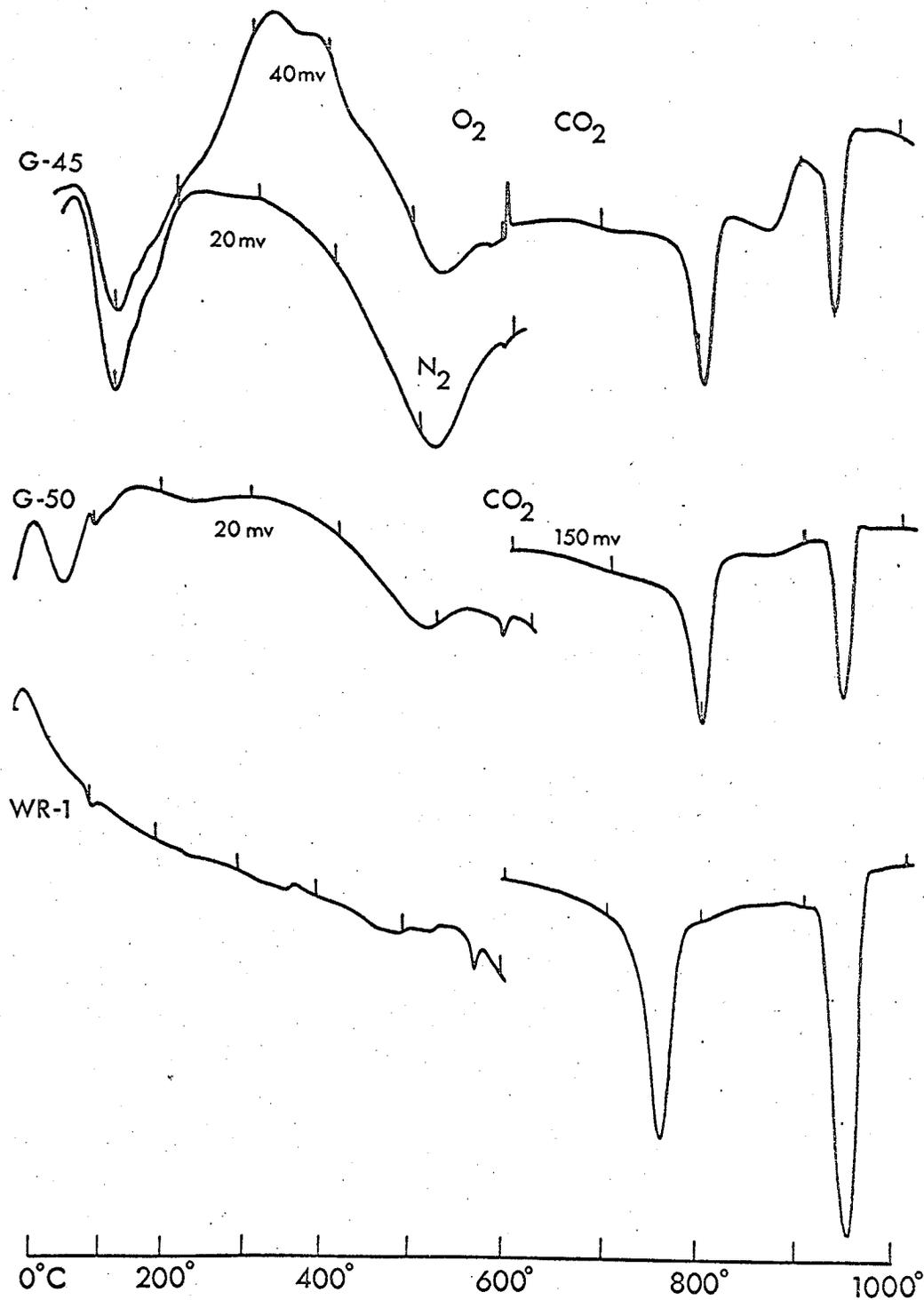


Figure 28. DTA curves, in dynamic gases, of Blue-gray Clay sample G-45 and Gray Till sample G-50 from the Geology Building test hole, and Gray Till sample WR-1 from the St. James Bridge outcrop.

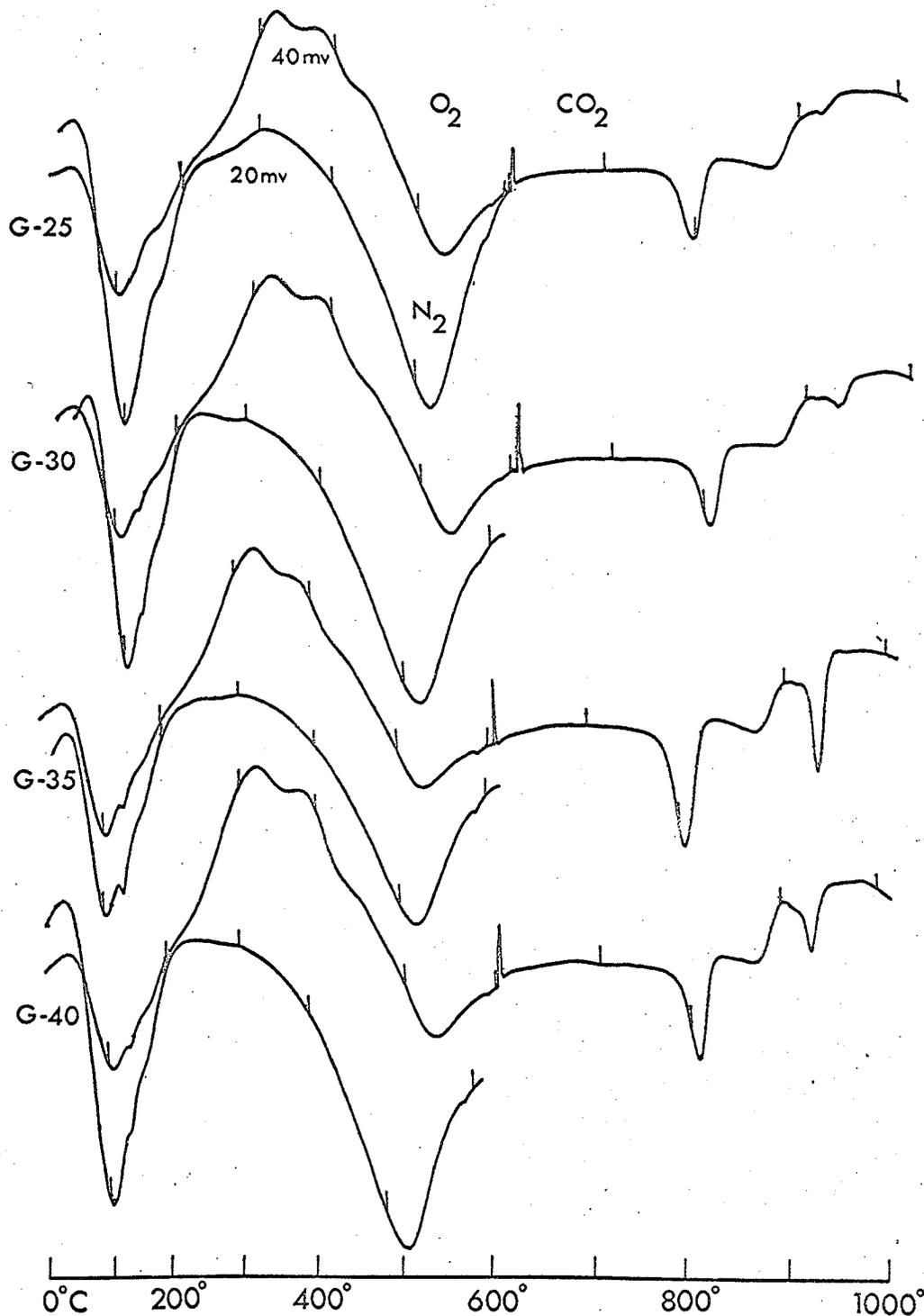


Figure 29. DTA curves, in dynamic gases, of Blue-gray Clay samples G-25, G-30, G-35, and G-40 from the Geology Building test hole.

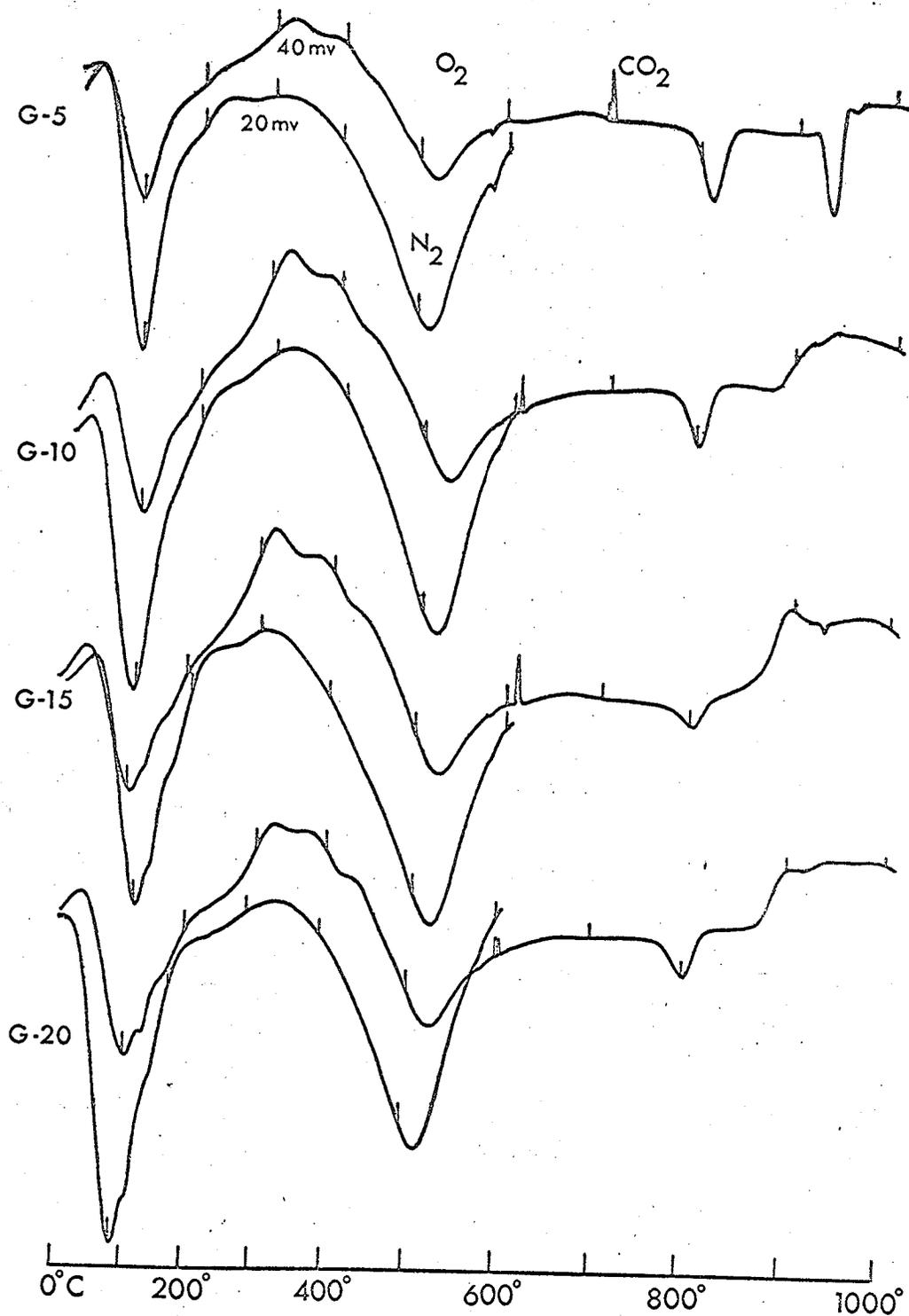


Figure 30. DTA curves, in dynamic gases, of Greenish-brown Clay sample G-5, and Brown Clay samples G-10, G-15, and G-20 from the Geology Building test hole.

Dolomite, the dominant mineral in these samples, is identified by the large endothermic peaks at approximately 780° and 925°C in all thermograms. Calcite is identified by the small endothermic peak at 945°C in some of the thermograms. The small size of this peak does not indicate that only minute quantities of calcite are present, for the high temperature dolomite peak and in particular the calcite peak are inhibited by the products (alkaline silicate glass) of the normal illite-abnormal montmorillonite exothermic reaction. (p. 85).

Quartz, the second most abundant mineral, is identified by its α to β inversion at 573°C.

Normal illite-abnormal montmorillonite, the main minor mineral, is identified by the small endothermic peak slightly below 500°C on some of the curves. The minor endothermic effects at about 90°C and 850°C on those thermograms with the lowest carbonate content, and hence a higher clay content, (Y-36/2 and Y-36/6, Fig. 17 and G-50, Fig. 28) are also due to the clay mineral assemblage.

Traces of gypsum (less than 1%) are indicated on most thermograms by the small, sharp endothermic peak at 115° to 120°C.

No indication of what caused the yellow colour in the bottom one and one-half feet of the Yale Avenue till or the St. James Bridge outcrop could be gained from the thermograms.

Lake Agassiz I Unit

Blue-gray and Brown Clay

No major mineralogical differences exist in the thermograms of the blue-gray clay and the brown clay so they can be described under a single heading. Figs. 18 to 25 show the thermograms of these two clay beds from the Yale Avenue test hole. Figs. 28, 29 and 30 show the thermograms of the samples from the Geology Building test hole. Normal illite-abnormal montmorillonite, montmorillonite, dolomite, calcite, quartz, gypsum, limonite and possibly pyrite are identified from the DTA curves of the blue-gray and the brown clay beds.

Normal illite-abnormal montmorillonite is the dominant mineral assemblage in these beds. It is identified by the large endothermic peak at approximately 500°C to 520°C, and by the large interlayer water endotherm at 100°C which varies widely with variations in relative humidity. The shoulder at 160°C on the interlayer water peak indicates that calcium is the main exchange cation on the montmorillonite. The endothermic-exothermic reaction pair in the 850°C to 950°C region is also due to the normal illite-abnormal montmorillonite. These peaks are present on most curves though they are distorted by the carbonate peaks.

On some of the thermograms on which the gas change was made well below 600 C° (the brown clay) suggestion of a broad endothermic peak in the 600° to 700°C region indicates the presence of a small quantity of normal montmorillonite.

This is best seen on curves Y-20/1 (Fig. 22), Y-17 (Fig. 23) and Y-8/6 (Fig. 25).

The effect of the alkaline silicate glass, from the clay minerals, in obliterating the calcium carbonate peaks (925° and 950°C) is well illustrated by the Yale Avenue series of thermograms. In the samples with a moderate carbonate content below about 24 feet (Figs. 18 to 21) the effect is not readily noticed, but it is still appreciable, particularly on the calcite peak.

In the thermograms of the sample Y-23 (Fig. 21), the calcium carbonate peaks, both the high-temperature dolomite and the calcite, are entirely obliterated while a small but very definite low-temperature dolomite peak remains at 800°C. Above this point, in the brown clay bed (Figs. 22 to 25), the effect is present throughout resulting in either entire obliteration of both the calcium carbonate peaks or only a slight suggestion of either of the peaks.

The quartz peak at 573°C, although strong through the lower portion of the beds (Figs. 18 to 22, 28 and 29), parallels the trend of the carbonates and gradually weakens until only the slightest trace of it is left in the thermograms of samples at the top of the brown clay bed (Figs. 23 to 25, and 30).

Traces of gypsum can be identified in almost all the

thermograms by the slight shoulder or very small peak at 115°C on the interlayer water peak. Near the top of the brown clay bed where gypsum-filled fissures are encountered, some of the samples (curves Y-10, Y-11/2 and Y-12, Fig. 25), have large very definite gypsum peaks. Despite their prominence the gypsum peaks represent less than 1% gypsum in samples Y-11/2 and Y-12 and no more than 2% in sample Y-10 based on the standard curves in Fig. 15, p. 98.

A small amount of limonite is present in all the samples. It is identified by a broad, shallow endothermic peak in the 250°C region.

The possible presence of a small amount of pyrite is suggested by the slight shoulder in the 450°C region of the oxygen thermograms. The main oxidation peaks at approximately 300° and 385°C on the oxygen thermograms are due to the organic matter in the clays. The characteristic form with the low-temperature peak dominant over the high-temperature peak indicates a low rank of hydrocarbons such as one would expect to find in recent sediments.

Lake Agassiz II Unit

Yellow Silt (Loess)

The minerals normal illite-abnormal montmorillonite, quartz, dolomite, calcite and hydrous iron oxides were identified from the thermograms (Y-8/1 and Y-8/3, Fig. 26) of the yellow silt bed. Minor amounts of kaolinite are suggested by

the broadening of the 500°C peak on the nitrogen curve and the double peaks 510° and 550°C on the oxygen curve. With this exception the clay minerals are similar to those in the blue-gray and brown clay beds although they are less abundant. The fact that there is no shoulder at 160°C on the interlayer water peak indicates that these clay minerals have appreciably less calcium exchange ions than those in the clays below. The slight peak at 190 C (Y-8/1, Fig. 26) probably indicates magnesium exchange ions. Both the low- and high-temperature dolomite peaks as well as a small calcite peak are present on curve Y-8/1 (Fig. 26) but only the low-temperature dolomite peak is present on the second thermogram Y-8/3 (Fig. 26). The second sample has a larger clay minerals peak, at 500°C. These differences in clay content reflect differences in the clay content already noted in the hand specimens. Quartz is indicated by its sharp peak at 573°C on both thermograms. The small but definite peaks, instead of the more usual broad shallow peaks or suggestion of peaks, at 250°C indicate that there is more hydrous iron oxide in these samples than in the other materials studied probably in excess of 2% based on the standard curves (Fig. 15, p. 98). No gypsum is indicated on the thermograms but a small amount of organic matter is indicated on the oxygen runs as well as the suggestion of pyrite.

The total magnitude of all the peaks is somewhat less than those recorded for other samples from other beds. This suggests that some other mineral that is thermally inert in the

range studied might also be present in the silt.

Greenish Brown Clay

The suite of minerals identified in the greenish brown clay bed is similar to, though not identical to, the suite of minerals identified in the brown and blue-gray beds.

Normal illite-abnormal montmorillonite is the main mineral assemblage with the main peak at slightly over 500°C . The interlayer water peak on the thermograms from the samples from both the Yale Avenue (curves Y-7, Y-6/9 and Y-5/7, Figs. 26 and 27) and Geology Building test holes (curve G-5, Fig. 30) has no shoulder at 160°C (present on all the interlayer water peaks of the Agassiz I clays). This means that, similarly to the clays in the underlying silt bed, calcium is not the main exchange ion as it is in the Agassiz I clays. The shoulder in the 190° to 200°C region suggests magnesium is the dominant exchange ion. A minor amount of normal montmorillonite is suggested by the slight endothermic peak in the 600° to 700°C region of most of these thermograms (Y-7 and Y-6/9, Figs. 26 and 27, and G-5, Fig. 30).

Dolomite is present in very small amounts in the Yale Avenue samples (Figs. 26 and 27) as indicated by the very small peak at approximately 780°C . There is no indication of any calcium carbonate peak. In contrast to these thermograms the thermogram of the single sample from the Geology Building test hole (curve G-5, Fig. 30) has an appreciable amount of dolomite,

and the high-temperature dolomite peak as well as a small calcite peak are both present although they are modified. This, in part, is due to a higher silt content of this bed but also to contamination, during sampling, from the underlying yellow silt bed. Similarly quartz is easily identified in the Geology Building thermogram but only traces of it are indicated, by minute endothermic bulges at 573°C , in the Yale Avenue samples.

There is no trace of gypsum in any of the samples, but very small amounts of hydrous iron oxides ($200^{\circ} - 300^{\circ}\text{C}$ region) are indicated on most of the thermograms, particularly that of the Geology Building test hole sample.

On the oxygen atmosphere portion of the thermograms there is the usual suggestion of pyrite at approximately 450°C and the normal oxidation peaks at 310° and 385°C of the organic matter.

Fine Sand

The thermograms of the fine sand from the Yale Avenue test hole show that dolomite and quartz are the main minerals with a very minor amount of normal illite-abnormal montmorillonite and traces of hydrous iron-oxides.

The thermogram of the fine sand (number Y-5, Fig. 27) has only the single high-temperature dolomite peak with no indication of a separate calcite peak. Calcite is indicated by a small peak at the base of the high-temperature dolomite peak in the thermogram of the sample from the contact zone (number

Y-5/6, Fig. 27). This sample contains lenses of the underlying clay bed and thus the dolomite and quartz peaks are less intense whereas the normal illite-abnormal montmorillonite peak is more pronounced. The interlayer water peak of the sample from the contact zone (Y-5/6, Fig. 27) exhibits no calcium cation shoulder at 160°C but a possible magnesium cation shoulder at 200°C, and thus is similar to the clays in the other two Agassiz II units. The clay content of the pure fine sand (curve Y-5, Fig. 27) above the contact zone is very low. No gypsum is indicated in either thermogram but the slight endothermic bulge at 250°C on both thermograms indicates small amounts of hydrous iron oxide.

X-RAY DIFFRACTION POWDER PHOTOGRAPHS

Introduction

X-ray powder photographs of selected samples from the Yale Avenue test hole were used to verify the identifications made from the DTA thermograms. Photographs of very thin collodion rollings (preparation described in Appendix D) were taken with Philips large (114.63 mm. diameter) cameras with the special clay mineral collimators and beam traps (Philips part number PW 1025) using 5 hour exposures of nickel-filtered copper radiation. Quartz, dolomite, calcite and feldspars were the only non-clay minerals identified. The clay minerals identified were illite, montmorillonite and either or both kaolinite and chlorite. The trace amounts of gypsum and pyrite found on the thermograms were not discovered on the powder photographs.

Some minor fogging developed on some photographs indicating the presence of minor amounts of iron, but it did not seriously interfere with the identification of the other minerals. The inconvenience of the iron fluorescence was well compensated for by the time saved in being able to use copper radiation as opposed to iron radiation. Also it was helpful to know that the iron was present.

Criteria For Mineral Identification

Despite the fact that each powder photograph contained from thirty to fifty reflections, it was possible to find, after eliminating coinciding reflections, one or two strong reflections with which to identify each mineral. Even the pattern of coinciding reflections, when coupled with relative intensities, was useful to identify some minerals and in the final analysis no reflection was left unaccounted for. See Brown (1961) for X-ray diffraction data for clay and associated minerals.

Quartz was identified on all powder photographs by its two strongest reflections at 4.26 and 3.34 Å. In addition to these almost all the quartz reflections could be identified on the photographs of the tills and loess. On the photographs of the clays only the ten strongest reflections could be identified.

Dolomite was identified mainly by its strong reflection at 2.89 Å but on most photographs at least four of the stronger lines could be identified and on some photographs of

the dolomite samples up to twenty reflections were identified.

Calcite is present in lesser amounts than dolomite in these samples, and consequently only the strongest reflection, at 3.04 \AA , could be identified. In till samples, because of their higher calcite content, at least three other reflections could be picked out to confirm the identification.

The strong reflection at 3.20 \AA and the medium reflection at 4.04 \AA were used to identify the feldspars. In samples with low feldspar contents these were the only lines identifiable. In others the 3.68 and 6.4 \AA reflections could also be identified. The 4.04 and 3.68 \AA reflections coincide closely with two very weak dolomite reflections but these are too weak to affect the feldspar reflections. The 4.04 \AA reflection might have coincided with the 3.95 \AA reflection of montmorillonite so only the 3.20 \AA feldspar reflection, the strongest, was left totally unquestioned. The 3.68 \AA reflection would seem to indicate that the feldspars are mainly plagioclase, and the general position of the lines further suggests a possible oligoclase composition. However, the diffraction patterns are too poorly developed to consider this observation to be anything more than a suggestion.

There is a possibility that the $2.98 - 3.04 \text{ \AA}$ and $2.84 - 2.89 \text{ \AA}$ feldspar reflections coincided respectively with the 3.04 \AA reflection of calcite and the 2.89 \AA reflection of dolomite. Fortunately the feldspar diffraction pattern was

generally so weak that its effect on the carbonate reflections could be considered to be negligible.

The clay minerals montmorillonite, illite and kaolinite or chlorite produced, in combination, a complex series of reflections in the 4.48 - 4.50, 3.54 - 3.57, 2.55 - 2.59, 2.37 - 2.39 and 1.49 - 1.50 Å regions. In addition to this series several individual peaks were used to identify individual minerals. A broad diffuse reflection in the 15.2 Å region indicated montmorillonite and a sharper reflection at 10 Å and a weaker, broader reflection at 5.0 Å indicated illite. The reflection at 1.50 Å was important for it indicated that the illite and montmorillonite are dioctahedral. In all cases a broadening of this reflection indicates that the two minerals are both contributing to the effect. A reflection at 7.1 Å indicated the presence of kaolinite or chlorite. A faint 14 Å reflection of chlorite was present on some photographs but it was not sufficiently resolved from the broad diffuse 15.2 Å montmorillonite reflection to be positively identified. Several other possible kaolinite or chlorite reflections could be identified on many photographs but they coincide with illite and montmorillonite reflections so that their help in identification was slight. No heat treatments were performed to distinguish between kaolinite and chlorite so no conclusion can be reached other than that one or the other or both minerals are present; judging from previous work (tables 2 and 3, pp. 34 and 38) kaolinite is the more probable. Most researchers found either minor kaolinite with no chlorite or minor kaolinite

with smaller amounts of chlorite. Some researchers, (Pawluk, 1960, 1961), found chlorite to be an alteration product of pedogenic processes and not a primary mineral in the soils.

The intensity of the 7 Å reflection in most cases bore the same relationship to the 15 and 10 Å reflections of montmorillonite and illite and indicated that the kaolinite and/or chlorite content is in a fixed ratio to the other two clay minerals.

Till

Dolomite, quartz, calcite and feldspars, in order of decreasing abundance, are the minerals identified by X-ray diffraction of the two samples, Y-39/6 and Y-44 (Table 9) from the Yale Avenue test hole. No reflections from clay minerals were recorded although the DTA revealed the presence of minor amounts of normal illite-abnormal montmorillonite.

Lake Agassiz I Unit

Blue-gray and Brown Clay

Diocahedral illite and diocahedral montmorillonite are the main clay minerals identified by X-ray diffraction, in the Lake Agassiz I Unit (samples Y-13, Y-18, Y-26/4 and Y-33, Table 9). Minor amounts of kaolinite or chlorite, dolomite, calcite, quartz and plagioclase were also identified. All the non-clay minerals increased in abundance with depth, but the intensities of all the clay mineral reflections bear the same

Table 9

Qualitative Results Of X-ray Powder Diffraction Analyses

Unit	Sample	Dolomite	Calcite	Quartz	Feldspar	Illite ²	Montmorillonite ²	Kaolinite or Chlorite
Lake Agassiz II Unit	Y-5	S ¹	VW	S	W	Nil	Nil	Nil
	Y-7	M	W	M	W	W	W	W
	Y-8/1	M	W	S	W	W	W	W
Lake Agassiz I Unit	Y-13	M	W	M	VW	W	W	W
	Y-18	M	W	M	VW	W	W	W
	Y-26/4	M	W	M	VW	W	W	W
	Y-33	M	W	M	W	VW	W	W
Till	Y-39/6	S	M	S	W	Nil	Nil	Nil
	Y-44	S	M	S	W	Nil	Nil	Nil

¹ The letters S, M, W, and VW represent Strong, Medium, Weak, and Very Weak and refer to the relative strength of the pattern as it appears on the film. Due to the difference in diffracting power of the various minerals, particularly the clay and the non-clay minerals, the strength of the pattern does not directly reflect the amount of each mineral present. Thus the clay mineral patterns are weaker than the non-clay mineral patterns even when the bulk of the sample is made up of clay minerals.

² Dioctahedral.

relationship to each other indicating that although the clay content decreases with depth the relative amounts of the different clay minerals remain the same.

Other than the slight changes in the amounts of the minor minerals, no differences were noted between the blue-gray and brown clay beds.

Lake Agassiz II Unit

Yellow Silt (Loess)

No single mineral appears to dominate the X-ray photograph of sample Y-8/1 (Table 9) from the yellow silt bed; dioctahedral illite and dioctahedral montmorillonite, dolomite, calcite, quartz and feldspars are present in moderate amounts. Minor amounts of kaolinite or chlorite are also present. The intensity of the 7 Å reflection, assumed to be mainly kaolinite, increases slightly over the intensity of the montmorillonite (15 Å) but not the illite (10 Å) reflections. This indicates that there is less montmorillonite and more kaolinite in the loess than in the underlying clay beds. This appears to confirm a similar indication on the thermograms.

Greenish Brown Clay

Dioctahedral illite and montmorillonite are the dominant minerals in this clay with minor amounts of quartz, dolomite, calcite and feldspars and traces of kaolinite or chlorite (sample Y-7, Table 9). The kaolinite or chlorite 7 Å reflection bears approximately the same relative intensity to the

montmorillonite reflection and indicates the kaolinite or chlorite is present in relatively the same proportion as it is in the lower beds.

Fine Sand

Dolomite is the dominant mineral of the fine sand (sample Y-5, Table 9) with major amounts of quartz and to a lesser extent feldspar, and minor amounts of calcite. No reflections from clay minerals were noted.

NATURE OF THE ILLITE-MONTMORILLONITE INTERSTRATIFICATION

Random interstratification of the illite and montmorillonite is not definitely proven by this study. Both minerals are present; this is verified by the X-ray diffraction, and the type of DTA curve is that which is given by what Cole (1955) calls normal illite-abnormal montmorillonite, that is randomly interstratified illite and montmorillonite. Unsupported DTA evidence is not enough to prove that the illite and montmorillonite are definitely interstratified, though it strongly suggests that they are. Recently Kodama and Brydon (1965) have shown by extensive X-ray work including one-dimensional Fourier analysis, that the 0.2 to 0.04 micron fraction of Lake Agassiz II clay is composed mainly of randomly interstratified montmorillonite-mica (illite), with the montmorillonite layers making up 50 to 65%. This has the same DTA curve as the Agassiz clay samples used in this author's study. In the clay fraction as a whole most researchers (Tables 2 and 3, pp. 34 and 38) have found both illite and montmorillonite, with illite dominating

the coarse fractions and montmorillonite dominating the fine fraction. This was found to be the case in Lake Agassiz II clays studied by X-ray diffraction in Manitoba (Beke and Wicks, 1962, and Smith, 1965). The coarse fraction (2-1 micron) was found to be dominantly illite, a 10\AA phase, and the fine (<1 micron) fraction was found to be dominantly montmorillonite, a 15\AA phase untreated or an 18\AA phase glycerol saturated. Fig. 31, from Beke and Wicks (1962), shows that the illite (10\AA) in the coarse fraction and the montmorillonite (15\AA) in the fine fraction, are very closely related. Both have endothermic peaks at approximately 500°C . The main difference is that the fine fraction, the dominantly montmorillonite material, has a much larger interlayer water peak (hence its larger C-spacing), and a slight peak at 700°C suggesting a minor amount of normal montmorillonite. Also the final endothermic-exothermic pair is barely discernible at 900°C for the coarse fraction but it is stronger and at a lower temperature 830°C for the fine fraction. This difference is due to the difference in interlayer cations, slight compositional differences, and possibly the difference in particle size.

Kodama and Brydon (1965) have demonstrated that in the fine clay fraction the 15\AA phase is randomly interstratified montmorillonite-illite. They also suggest that the minor 10\AA phase they noted is randomly interstratified illite-montmorillonite. From this evidence and the evidence of the similarity of DTA curves of the coarse and fine clays it can be

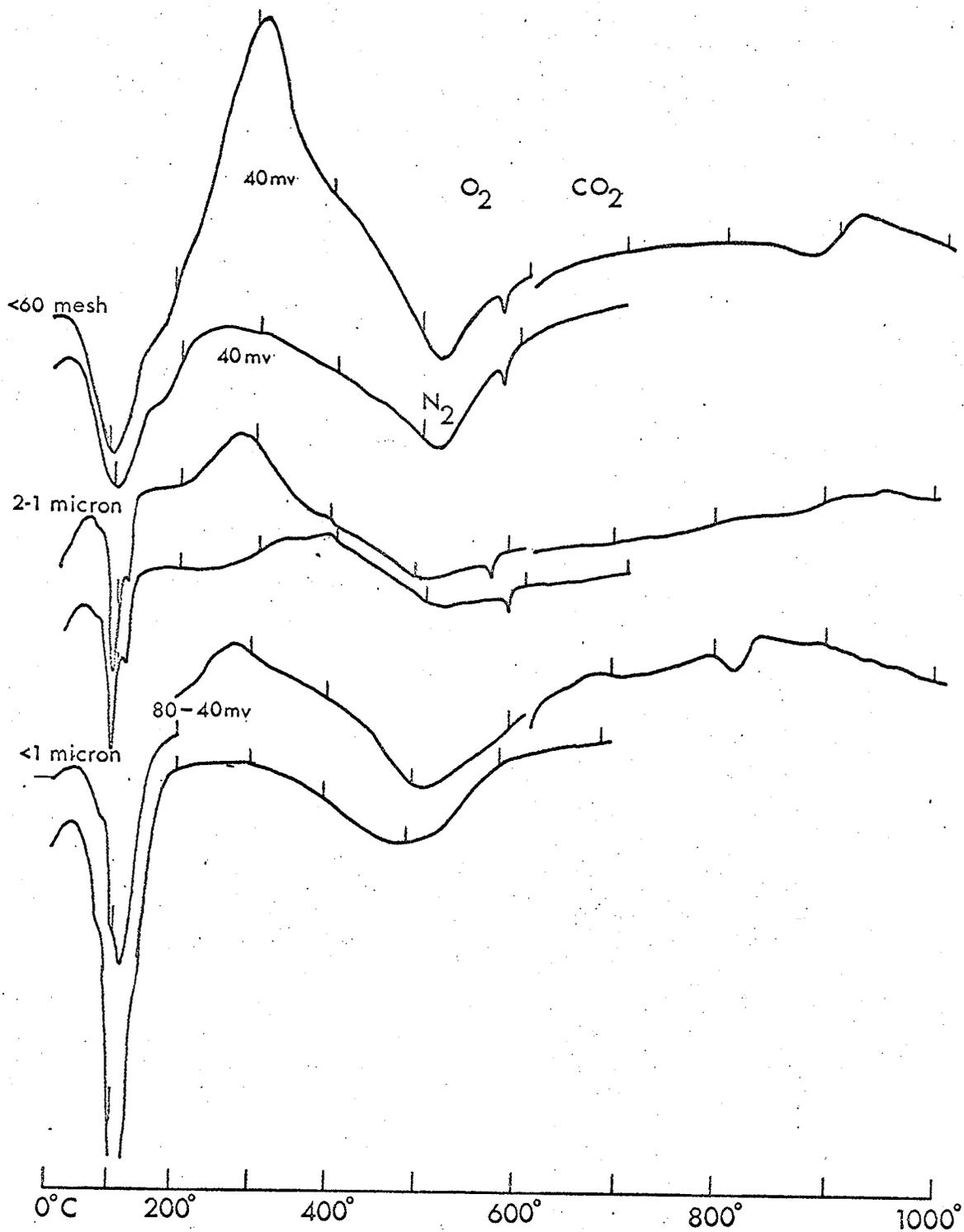


Figure 31. DTA curves, in dynamic gases, of the -60 mesh, 2-1 micron, and -1 micron fractions of a Lake Agassiz II clay. After Beke and Wicks (1962).

concluded that the entire clay fraction with the exception of minor kaolinite, quartz and feldspar, is interstratified illite-montmorillonite. The coarse fraction is dominantly an illite-montmorillonite (10Å phase) and the fine fraction is dominantly a montmorillonite-illite (15Å phase) but both mineral assemblages are very closely and intimately related to each other.

CHAPTER VI

QUANTITATIVE MINERALOGY

INTRODUCTION

Quantitative determinations of the carbon dioxide and quartz contents were made from all the thermograms of both the Yale Avenue and Geology Building test holes, using the methods described in Chapter IV. The presence or absence of minor quantities of gypsum was also noted. All the quartz determinations of the clay samples are a consistent 2 to 3% low because of the masking effect of the illite-montmorillonite peak on the quartz peak. These results are presented in spite of the error because they show the variation in amount even if the accuracy is poor. For more accurate determinations a sample was heated twice, once to eliminate the illite-montmorillonite peak and a second time to record the quartz peak for measurement.

Nine samples were chosen from the Yale Avenue test hole for extensive quantitative chemical and mineralogical analysis. The samples are listed in Table 10. One sample was taken from each of the Lake Agassiz II beds and two samples were taken from both the Lake Agassiz I beds and the till bed. X-ray photographs were taken of each of these samples and used to determine the relative amounts of calcite and dolomite by

Table 10

Description of Yale Avenue Samples
for Chemical Analysis

Sample Number	Bed	Unit
Y-5	Fine Sand - Sandy Clay	Lake Agassiz II
Y-7	Greenish Brown Clay	Lake Agassiz II
Y-8/1	Yellow Silt	Lake Agassiz II
Y-13	Brown Clay	Lake Agassiz I
Y-18	Brown Clay	Lake Agassiz I
Y-26/4	Blue-gray Clay	Lake Agassiz I
Y-33	Blue-gray Clay	Lake Agassiz I
Y-39/6	Gray Till	Keewatin Till
Y-44	Gray Till	Keewatin Till

the method of Tennent and Berger (1957). The proportion of calcite to dolomite was then applied to the total carbon-dioxide content and the weight percent of the various oxides for both minerals, with respect to the total sample, was thus determined (Table 11). Totaling the weight percent of the oxides gives the weight percent of calcite and dolomite in Table 12.

X-ray fluorescence analysis of the nine samples was carried out by Mr. K. Ramlal of the Geology Department, University of Manitoba, using an Applied Research Laboratories Incorporated, Vacuum X-ray Spectrometer. This unit consists of a Model 25,000 Vacuum X-ray Quantameter, a Model 24,500 X-ray console and a Model 12,350 X-ray power supply. The precision of the instrument is 0.1% of the amount present for heavy elements and 0.3% of the amount present for light elements. The samples were prepared as described in Appendix C and analysed for silica, alumina, total iron as ferric iron oxide, magnesia, lime, potash, titanium and manganese. Soda and phosphorus were also determined by Mr. Ramlal, the former by flame photometry, the latter colorimetrically.

The author (F.J.W.) determined water, carbon dioxide, and loss on ignition, on samples dried at 110°C for 24 hours. The method of Shapiro and Brannock (1956) was used for the water determination and the direct method of Kolthoff and Sandell (1952) was used for the carbon dioxide. The loss on ignition was determined as the loss of weight after heating to

Table 11

Weight Percent of CaO and MgODue to Calcite and Dolomite in Yale Avenue Samples

Sample Number	Carbon Dioxide Wt. %	Ratio		CaO (Calcite) Wt. %	CaO (Dolomite) Wt. %	Total CaO Wt. %	MgO (Dolomite) Wt. %
		Calcite vs Dolomite Wt. %	Wt. %				
Y-5	22.87	<5.0	>95.0	1.46	13.84	15.30	9.95
Y-7	2.01	37.0	63.0	0.95	0.81	1.76	0.58
Y-8/1	5.78	33.0	67.0	2.43	2.47	4.90	1.77
Y-13	3.53	36.0	64.0	1.62	1.44	3.06	1.03
Y-18	4.67	37.0	63.0	2.20	1.88	4.08	1.35
Y-26/4	7.05	30.0	70.0	2.70	3.14	5.84	2.26
Y-33	9.71	26.0	74.0	3.22	4.58	7.80	3.29
Y-39/6	23.04	21.0	79.0	6.17	11.60	17.77	8.34
Y-44	27.63	17.0	83.0	5.99	14.61	20.61	10.51

1000°C for 30 minutes. It is reported not as total weight loss but as the balance after the carbon dioxide and water determinations are subtracted. This balance represents mainly the organic content of the samples.

YALE AVENUE TEST HOLE

Gray Till

The combined analysis of two samples of the till, one from the main body of the till (Y-39/6) and the other from the lower few feet (Y-44), are given in Table 12. The higher carbonate content of the lower portion of the till is illustrated by these results. Both samples have approximately 11% calcite but the lower sample has 48% dolomite compared to 38% for the upper sample.

The mineralogy of the till sheet is uniform and only the lower 2 to 3 feet show any significant variation. Because dolomite, calcite and quartz are dominant minerals of the till Table 13, showing the carbon dioxide and quartz content of the till sheet, illustrates the uniformity very precisely. In order to emphasize the uniformity and minimize local variations, an average of each three adjacent values was calculated for both the carbon dioxide and quartz (Table 13). This shows that the till varies little from these average values of 23 to 24% carbon dioxide and 14 to 15% quartz, except for the lower 2 to 3 feet where the carbon dioxide rises to 27% and quartz to almost 19%. A second slight variation occurs in the 37 to 39

Table 12

Combined Quantitative, Semi-Quantitative
and Estimated Mineral Analysis of
Selected Yale Avenue Samples

Sample	Wt.% Dolomite ±3.0	Wt.% Calcite ±3.0	Wt.% Quartz ±1.0	Wt.% Feldspar ¹ estimated	Wt.% Illite- ² Montmor- illonite ±10.0	Dominant Exchange Ion	Gypsum	Hydrous Iron Oxides	Possible Pyrite ?	Wt.% Organic Matter ±0.1
Y-5	45.5	2.6	29.2	20	<5	?	Nil	Tr (?)	Nil	0.49
Y-7	2.7	1.7	1.8	2	90	Mg	Nil	Tr (?)	Nil	2.03
Y-8/1	8.1	4.3	16.3	16	50 ³	Mg	Nil	Minor (2-3%)	Tr (?)	2.59
Y-13	4.7	2.9	2.7	3	85	Ca	Tr	Tr	Tr	1.90
Y-18	6.2	3.9	3.3	3	80	Ca	Tr	Tr	Tr	1.85
Y-26/4	10.3	4.8	5.3	5	75	Ca	Tr	Tr	Tr	1.81
Y-33	15.1	5.7	5.6	6	65	Ca	Tr (?)	Tr	Tr	1.08
Y-39/6	38.2	11.0	13.9	14	20	?	Tr (?)	Nil	Nil	0.50
Y-44	48.1	10.7	17.1	17	5	?	Tr (?)	Nil	Nil	0.40

¹ Estimated as being equal to the quartz content, the exception of Sample Y-5.

² Figure arrived at by difference, and rounded off to nearest 5%, includes minor kaolinite and/or chlorite.

³ Amount of kaolinite greater than in other samples.

Table 13

Carbon Dioxide, Quartz and Gypsum Contents¹
of Yale Avenue Gray Till

Sample Number	Carbon Dioxide Wt. %	Carbon Dioxide Average ²	Quartz Wt. %	Quartz Average ²	Gypsum
Y-36/2	26.4	23.8 ³	14.3	13.7 ³	Tr.
Y-36/6	21.1	24.4	13.0	14.1	Tr.
Y-36/9	25.8	23.7	14.9	14.9	Tr.
Y-37/1	24.1	24.4	16.7	15.1	Tr.
Y-38	23.2	23.4	13.6	15.3	Tr.
Y-39/6	23.0	23.3	15.5	14.2	Tr.
Y-41/1	23.8	23.4	13.6	14.5	Tr.
Y-42	23.5	24.2	14.4	14.5	<1
Y-43/1	25.3	25.5	15.5	16.2	Nil
Y-44	27.6	26.7	18.8	17.5	Nil
Y-44/6	27.1	27.4 ³	18.2	18.5 ³	Tr.

¹ These were determined from the thermogram.

² Average of three adjacent values.

³ Only two values were used to obtain this average.

foot depth where the quartz content rises slightly.

The feldspar content was not determined quantitatively although feldspars are definitely present (Table 9). Ehrlich and Rice (1955) found the feldspar content of the clay fraction (<2 microns) of the similar gray till to be equal to the quartz content. If the same ratios held for the Yale Avenue till, the feldspar content could be equal to the quartz content. This approximation was used to establish the figures in Table 12.

The illite-montmorillonite content is low throughout the till, 15 to 20% in the main body and 5 to 10% at the base, Table 12.

A comparison of the 500°C peak on the thermograms (Figs. 16 and 17, pp. 104 and 105) shows that the peak is smallest in the high carbonate zone at the base of the bed and slightly larger in the main body of the bed. Curve Y-36/6 (Fig. 17, p. 105) has the best developed 500°C peak because this sample comes from a band of blue clay in the till just below the contact.

Gypsum is fairly consistently present in uniform trace amounts throughout the bed. The only variation is a very minor increase in gypsum at the 42 foot depth.

Lake Agassiz I Unit

Blue-gray Clay

The results of the combined analysis of two samples (Y-26/4 and Y-33) of blue-gray clay are given in Table 12. The decrease of clay minerals and increase in non-clay minerals with depth is illustrated by these two samples, but the variation is more fully illustrated by the carbon dioxide and quartz content shown in Table 14. The total carbon dioxide content of the blue-gray clay decreases from 14% at the bottom of the bed to about 4% at the top of the bed. The lower half of Table 14, column 3, is an average of three adjacent samples and illustrates the trend in variation in approximately 1 foot intervals. The initial drop off of carbon dioxide content is rapid from 14.0 to 9.5% in the bottom two feet of the bed but the average remains fairly constant between 9.4 and 10.6% up to four feet below the top contact. In this upper zone the carbon dioxide content drops rapidly from approximately 10% just outside the zone to 4% at the top of the zone. The variations of the quartz content (Table 14) do not follow the variations of the carbon dioxide exactly. The quartz falls from about 5% at the bottom of the bed to about 3% in the lower few feet and then gradually rises again up to nearly 6% four feet from the top of the bed. Above this point it decreases sharply to about 3%. These values are presented only to illustrate the variations with depth. They are 2 to 3% below the actual values.

Carbon Dioxide, Quartz and Gypsum Contents¹
Of the Yale Avenue Brown and Blue-gray Clays

Sample Number	Bed	Carbon Dioxide Wt. %	Carbon Dioxide Average	Quartz ² Wt. %	Quartz Average	Gypsum
Y-8/6		4.6	3.9 ³	Tr.	-	Nil
Y-10		3.2	3.7	Tr.	-	<2
Y-11/2		3.2	3.2	Tr.	-	<1
Y-12		3.2	3.1	Tr.	-	<1
Y-13		3.0	3.2	Tr(2.7)	-	Tr.
Y-14		3.5	3.3	Tr.	-	Tr.
Y-15	Brown	3.3	3.5	Tr.	-	Tr.
Y-16	Clay	3.7	3.4	Tr.	-	Tr.
Y-17		3.3	3.5	Tr.	-	Tr.
Y-18		3.5	3.8	Tr(3.3)	-	Tr.
Y-19		4.7	3.8	Tr.	-	Nil
Y-20/1		3.3	3.9	2.6	-	Nil
Y-21/6		3.6	3.9	3.1	2.9	Nil
Y-22		4.9	4.3	3.0	3.1	Tr.
Y-23		4.4	5.3	3.3	2.8	Tr.
Y-24		6.5	5.8	2.2	3.1	Tr.
Y-25		6.6	6.5	3.9	3.0	Tr.
Y-26/4		6.4	7.8	2.9(5.3)	4.2	Tr.
Y-27		10.3	9.4	5.8	4.5	Tr.
Y-28/1		11.4	10.1	4.9	5.0	Tr.
Y-29	Blue-gray	8.6	10.0	4.4	4.3	Tr.
Y-30	Clay	10.1	10.0	3.7	3.8	Tr.
Y-31		11.2	10.6	3.4	3.6	Tr.
Y-32		10.6	10.5	3.7	3.3	Tr.
Y-33		9.6	9.9	2.9(5.6)	3.3	Nil
Y-34		9.5	10.7	3.4	3.7	Tr.
Y-35		12.9	12.1	4.7	4.3	Tr.
Y-36		14.0	13.5 ³	4.7	4.7 ³	Tr.

¹ These were determined from the thermograms.

² These values were determined from the original thermograms and are 2 to 3% low. The figures in brackets were determined from special quartz reruns.

³ Only two values were used to obtain this average.

Feldspar was definitely identified (Table 9) but it was not quantitatively determined. From the work of Ehrlich, Rice and Ellis (1955) on the clay fraction (less than 2 microns) of tills a feldspar content approximately equal to the quartz content can be expected.

Interstratified illite-montmorillonite is the dominant mineral assemblage and increases in amount from about 60% at the base to about 75% at the top of the bed, as the non-clay minerals decrease, (Table 12). The change is so slight that it does not show up in the thermograms of adjacent samples but an increase in the 500°C peak between the thermograms at the bottom and the top of the bed can be seen (Fig. 18, p. 106 and Fig. 21, p. 109). Calcium appears to be the dominant exchange ion throughout the series although other cations, most probably magnesium, are very likely present.

Gypsum (Table 14) occurs in trace amounts throughout most of the bed.

Brown Clay

The analysis of the two selected samples of the brown clay are given in Table 12. The trend noticed in the blue-gray clay bed continues uninterrupted through this bed as well. The upper half of Table 14 gives the actual values of carbon dioxide content and quartz measured, as well as the averages continued on from the blue-gray clay bed.

The maximum carbon dioxide content is 4.9% at the 22

foot depth, the bottom of the bed, and the minimum is 3.0% at the 13 foot depth, near the top of the bed. The decrease in carbon dioxide between these points is uniform and steady with little deviation from the average values. Above the 13 foot depth the carbon dioxide content is constant at about 3% except for the top foot which sharply rises to 4.9%. The quartz values shown in Table 14 are low for the same reasons described earlier but show the variation falling from what would be 5 to 6% (corrected) at the bottom to 2 to 3% at the top.

Feldspar was identified in the two samples X-rayed (Table 9) but was not estimated quantitatively. It is assumed to be approximately equal to the quartz content (see till and blue-gray clay sections).

Interstratified illite-montmorillonite is the dominant mineral assemblage throughout the bed. The clays make up 75-80% of the material at the blue-gray/brown clay contact and about 90% at the top of the bed. The prominence of the calcium exchange ion seems to decrease slightly in the upper part of the bed though no indication of any other cation appears to be emerging.

The occurrence of gypsum is more irregular in this bed (Table 14) than in the blue-gray clay bed. In most of the lower portion of the bed faint traces of gypsum can be detected but in the upper four feet where gypsum-filled cracks are found, the gypsum content rises to minor amounts, 1 to 2%, particularly

at the 10 foot level.

Lake Agassiz II

Yellow Silt (Loess)

The results of the combined analysis of a single loess sample (Y-8/1) are given in Table 12. A second sample (Y-8/3) was analysed by differential thermal analysis only. The DTA results of both samples are given in Table 15. The chief difference between the two samples is the increased clay content of the contact sample (Y-8/3) as revealed by the lower quartz and carbon dioxide contents and increased 510°C endothermic peak on the thermogram (Fig. 26, p. 114). This was noted in the hand specimen as an increased number of thin clay laminations.

The clay minerals make up 50% of Y-8/1, Table 12, and appear to be mainly illite-montmorillonite similar to the clay found in the other beds but there are indications of a greater proportion of kaolinite than in the other beds. There is no trace of calcium exchange ion and the shape of the peak suggests magnesium exchange cation.

Since the thickness of the bed is unknown and since only two samples were studied, little can be said about the variation of the mineralogy within the bed. However it appears that the lower contact has a higher clay content than the main part of the bed.

Table 15

Carbon Dioxide, Quartz and Gypsum Contents¹
of Yale Avenue Lake Agassiz II Beds

Sample Number	Bed	Carbon Dioxide Wt. %	Quartz ³ Wt. %	Gypsum
Y-5	Fine Sand	26.3 ²	31.9	Nil
Y-5/6		17.5 ²	12.7	Nil
Y-5/7	Greenish Brown Clay	Trace	4.7	Nil
Y-6/9		4.1	Tr (3.1)	Nil
Y-7		2.0	Tr (2.1)	Nil
Y-8/1	Yellow Silt	6.7	16.3	Nil
Y-8/3		4.6	6.9	Nil

¹ These were determined from the thermograms.

² 3-4% high.

³ The low quartz contents were determined from the original thermograms and are 2 to 3% low. The figures in brackets were determined from special quartz reruns.

Greenish Brown Clay

Table 12 gives the analysis of a sample (Y-7) from this bed. Further results of thermograms alone are given in Table 15. This sample has the lowest non-clay mineral fraction, about 10%, of any of the samples studied. Although in the hand specimen the varves are not as well developed as they are in the upper brown clay bed of Lake Agassiz I, mineralogically the two materials are very similar.

The clay minerals appear to be the same as those in the other clay beds, and make up about 90% of the material. Similar to the clay fraction of the yellow silt, calcium does not appear to be the dominant exchange ion. The slight shoulder at 190 C (Figs. 26 and 27, curves Y-7, Y-6/9 and Y-5/7, pp. 114 and 115) probably represents magnesium exchange ion. Varying amounts of both calcium and magnesium are very likely present.

Sampling is too infrequent to make a positive study of the mineralogical variations with depth. Judging from the samples recovered variations appear to be very slight.

Fine Sand

The mineralogical analysis of the fine sand (Y-5) is given in Table 12, and further differential thermal analysis data are given in Table 15. In the main body of the bed the clay mineral content is only a trace but it is slightly higher in the lower contact zone where some mixing of the beds occurs.

The ratio of quartz to feldspar, 1 to 1, as used for the other materials, does not appear to be valid for this bed as it yields a total of over 100%. This is to be expected for the origin of this material differs greatly from the other Agassiz sediments.

Similarly to the other Lake Agassiz II sediments too few samples were recovered to thoroughly study the mineralogical variations with depth.

GEOLOGY BUILDING TEST HOLE AND ST. JAMES BRIDGE TILL

DTA results only, Table 16, are available for the Geology Building test hole sample and the St. James Bridge outcrop sample. The sampling interval was too infrequent to yield a detailed study as carried out on the Yale Avenue material, particularly in the till and Lake Agassiz II unit. However, Table 16 shows that the trends discovered in the detailed study hold true for the Geology Building sample site.

The results of the analysis of the oxidized till from the St. James Bridge are also given in Table 16 and are similar to the results of the analysis of the till from both the test holes. Any changes in the till due to the weathering that produced its yellow colour do not show up in the general mineralogy.

Table 16

Carbon Dioxide, Quartz and Gypsum Contents¹
of Geology Building Samples and St. James Bridge Sample

Sample Number	Unit	Carbon Dioxide Wt.%	Quartz ² Wt.%	Gypsum
G-5	LAKE AGASSIZ II (Greenish Brown Clay)	5.6	4.0	Nil
G-10	LAKE AGASSIZ I (Brown Clay)	5.2	1.7	Nil
G-15		4.3	Tr.	Tr.
G-20		4.8	1.3	Tr.
G-25	LAKE AGASSIZ I (Blue-gray Clay)	5.8	2.1	Nil
G-30		5.6	1.5	Tr.
G-35		7.9	2.6	Tr.
G-40		7.2	1.6	Tr.
G-45		10.0	3.3	Tr.
G-50	Gray Till (unoxidized)	24.0	7.3	Tr.
WR-1	Gray Till (oxidized)	26.3	12.3	Tr.

¹ These were determined from the thermograms

² These values were determined from original thermograms and are 2 to 3% low.

COMPARISON OF THE CHEMICAL ANALYSIS RESULTS
TO THE LITERATURE

The results of the chemical analyses of the nine selected Yale Avenue Agassiz samples (Table 17) agree closely with the published results of similar sediments (Table 18). The analysis of the clay unit of Lake Agassiz II (Y-7, Table 17) is reasonably similar to Wallace and Maynard's (1924) analysis (column 2, Table 18) and Elson's (1961) average analysis (column 4, Table 18). The yellow silt analyses (Y-8/1, Table 17 and column 3, Table 18) are also very similar. The analyses of the Lake Agassiz I samples (Y-13, Y-18, Y-26/4 and Y-33, Table 17) do not compare quite as closely to the analysis of Wallace and Maynard (1924) (column 1, Table 18) of a similar clay but this is accounted for by the higher carbonate content of the Yale Avenue samples. Yale Avenue sample Y-13, the one with the lowest carbonate content, is close to the Wallace and Maynard result. The gray till samples (Y-39/6 and Y-44, Table 17) compare very closely to Ehrlich, Rice and Ellis' analysis (columns 7 and 8, Table 18).

DISTRIBUTION OF ELEMENTS AND MINERALOGY

A general examination of the chemical analyses given in Table 17 supports the mineralogical determinations. Although no single element is present exclusively in one mineral, the amounts of the various elements present indicate in a general way the amounts of the various minerals present.

Table 17

Chemical Analyses of Agassiz Sediments from Yale Avenue Test Hole

	Lake Agassiz II			Lake Agassiz I				Gray Till	
	Y-5	Y-7	Y-8/1	Y-13	Y-18	Y-26/4	Y-33	Y-39/6	Y-44
SiO ₂	40.55	55.09	60.57	52.55	52.89	51.38	48.85	34.57	31.47
Al ₂ O ₃	6.30	18.26	11.77	18.73	16.74	15.31	14.25	6.27	4.50
Fe ₂ O ₃ (total iron)	1.61	7.73	4.67	7.02	6.24	5.96	5.27	1.59	0.86
TiO ₂	0.20	0.81	0.61	0.73	0.63	0.62	0.57	0.10	0.02
MnO	0.04	0.06	0.15	0.15	0.11	0.07	0.05	0.02	0.01
MgO	9.43	2.69	2.03	2.91	3.46	4.26	4.65	8.22	8.45
CaO	14.73	2.15	5.07	3.90	4.57	6.32	8.26	22.50	22.70
Na ₂ O	0.92	0.54	0.90	0.50	0.58	0.60	0.61	0.82	0.86
K ₂ O	1.41	3.11	2.54	2.97	2.89	2.62	2.58	1.64	1.29
P ₂ O ₅	0.15	—	0.16	—	0.16	—	0.14	0.06	0.05
H ₂ O ^(m)	0.94	5.58	3.25	5.45	5.15	4.42	3.93	0.97	0.61
Ig. Loss ¹ CO ₂	22.87	2.01	5.78	3.53	4.67	7.05	9.71	23.04	27.63
Balance (Organic Carbon ?)	0.49	2.03	2.59	1.90	1.85	1.81	1.08	0.50	0.40
TOTAL	99.64	100.06	100.09	100.34	99.94	100.42	99.95	100.30	98.85

¹ H₂O, CO₂, Ignition Loss determined by the author; all other determinations by K. Ramlal, Geology Department, University of Manitoba.

Table 18
Published Chemical Analyses
Of Lake Agassiz and Associated Sediments

	1	2	3	4	5	6	7*	8*
SiO ₂	55.48	56.02	58.63	55.83	52.25	52.27	32.61	31.05
Al ₂ O ₃	20.37	20.53	9.12	15.77	21.30	25.13	7.12	4.33
Fe ₂ O ₃	5.54	5.35	8.42	6.36	9.97	6.08	0.97	0.98
TiO ₂	—	—	—	—	0.73	none	0.10	0.15
MnO	—	—	—	—	—	trace	0.07	0.07
MgO	3.29	3.81	3.62	3.84	3.52	1.38	11.53	11.94
CaO	4.95	2.75	6.41	5.73	—	0.93	18.84	19.62
Na ₂ O	0.82	0.80	1.60	2.13	—	0.24	0.90	0.50
K ₂ O	1.86	2.22	2.85	2.51	2.57	2.91	0.94	0.51
P ₂ O ₅	—	—	—	—	—	none	0.04	0.05
Ig. Loss	8.77	9.02	9.00	8.02	9.49	10.71	—	—
H ₂ O(+)	—	—	—	—	(5.48)	—	—	—
CO ₂	—	—	—	—	—	—	25.24	28.58
Organic Carbon	—	—	—	—	—	—	0.1	0.1
TOTAL	101.08	100.50	99.65	100.19	99.83	99.65	98.46	97.88

1. Wallace and Maynard (1924), Lake Agassiz I Clay, Winnipeg.
2. Wallace and Maynard (1924), Lake Agassiz II Gray Clay, Winnipeg.
3. Wallace and Maynard (1924), Lake Agassiz II Lower Sandy Clay (Yellow Silt), Winnipeg.
4. Elson (1961) Lake Agassiz II, Average of 8 Analyses, Manitoba, Ontario and Minnesota.
5. Warder and Dion (1952), Average Analyses of the Less Than 1 Micron Fraction of 6 Saskatchewan Soils.
6. Kodama and Brydon (1965), Lake Agassiz II Clay, 0.2 to 0.04 Micron Fraction, Winnipeg.
7. Ehrlich, Rice and Ellis (1955), Gray Till, Isafold, Interlake Area, Man.
8. Ehrlich, Rice and Ellis (1955), Gray Till, Komarno, Interlake, Area, Man.

* Recast by the author (F.J.W.) from the values of the oxides after ignition to include the carbon dioxide content.

The presence of the clay minerals is reflected directly in the water and alumina contents. These are both highest in the high clay samples (Y-7 and Y-13, Table 17) and decrease down through the Agassiz I clays as the amounts of clay minerals decrease (Y-18, Y-26/4 and Y-33, Table 17). They are lowest in the sand (Y-5), silt (Y-8/1) and tills (Y-39/6 and Y-44, Table 17), the low clay sediments.

The carbon dioxide, and to a lesser extent the magnesia and lime, indicate the carbonate content. These values are highest in the high carbonate samples, the tills (Y-39/6 and Y-44, Table 17) and sand (Y-5), and lowest in the clays (particularly Y-7 and Y-13, Table 17). The increasing carbonate content down through the Lake Agassiz I clays is illustrated by samples Y-13, Y-18, Y-26/4 and Y-33 (Table 17).

Because the clay minerals, quartz, and feldspar all contain silica, the silica content does not directly reflect any particular mineral. Generally it increases with the clay content but the yellow silt (Y-8/1) is an exception to this trend.

The feldspar content is reflected in a general way by the soda content. It is highest in the low clay mineral and high non-clay mineral sediments (Y-5, Y-8/1, Y-39/6 and Y-44, Table 17) the high clay mineral, low non-clay mineral sediments (Y-7, Y-13, Y-18, Y-26/4 and Y-33).

CHEMICAL FORMULAE OF THE CLAY MINERALS

Chemical formulae of the prairie clays have been calculated by Warder and Dion (1952) and Kodama and Brydon (1965). Both results are given in Tables 19 and 20. Warder and Dion's formula (Table 19) is an average from the less than 1 micron fraction of six Saskatchewan soils of various origins. The formula represents the over-all layer of an interlayered illite-montmorillonite. Kodama and Brydon's work was based on the results from the 0.2 to 0.4 micron fraction of seven prairie tills and lacustrine clays. Two over-all layer formulae for interlayered illite-montmorillonite are given. The first (Table 19) is an average formula for seven prairie soils. The second (Table 20) is the formula for a single sample of Lake Agassiz II clay from the Winnipeg area. The swelling layer and non-swelling layer formulae also presented in these tables will be discussed later.

In both the above cases as many of the impurities as possible, organic matter, carbonates, etc., were removed and the material carefully sized. (Warder and Dion, however, did not remove the free iron or free aluminum oxides.) This careful preparation of the material prior to analysis reduces the impurities to a minimum, making the result more meaningful. The effect of this type of preparation can be observed by comparing Warder and Dion's and Kodama and Brydon's analyses in columns 5 and 6, Table 18 to the analysis of bulk samples of similar material, columns 1 and 2 in the same table.

The most significant changes are the 3 to 4% decrease in silica and the 1 to 5% increase in the amount of alumina in the fractionated samples over the bulk samples.

No sample preparation of this nature was carried out in the present study but chemical formulae were calculated from the analysis of the bulk sample of -60 mesh material, after the appropriate ratios of oxides representing the non-clay minerals were subtracted. The procedure followed is given in Appendix F.

The usefulness of these calculated formulae even on carefully purified samples has been strongly doubted by Kelley (1945), and thus calculated formulae based on analyses of unsized, unpurified material must be subject to a great deal more doubt. With these limitations well understood, the calculations were carried out on the four Lake Agassiz I clay analyses and the one Lake Agassiz II clay analysis. It must be emphasized that these results are not to be accepted as the actual mineral formulae but rather as a general indication of the nature of the formulae, that can be compared to the more careful work of Warder and Dion, and Kodama and Brydon.

The formulae are calculated on the basis of ten oxygens and two hydroxyls per 1/2-unit cell. When the determined water contents were calculated on this basis they yielded more than two hydroxyls per 1/2-unit cell. The excess water was considered to be sorbed water and was calculated by

Table 19

Published Chemical Formulae of Prairie

Illite-Montmorillonites

Warder and Dion (1952) - Average of Six Samples

Over-all Layer

K* (B.E.C.)** (Al Fe Mg) (Si Al) O (OH)
 0.22 0.22 1.26 0.51 0.36 3.55 0.45 10 2

Kodama and Brydon (1965) - Average of Seven Samples

Over-all Layer

(K Na Ca) (Al Fe Mg) (Si Al) O (OH)
 0.21 0.04 0.10 1.61 0.20 0.25 3.62 0.38 9.77 2.31

Swelling Layer

0.79 x (Na Ca) (Al Fe Mg) (Si Al) O (OH)
 0.06 0.13 1.51 0.25 0.32 3.79 0.21 9.80 2.40

Non-swelling Layer

0.21 x K Al (Si Al) O (OH)
 2 3 10 2

* Fixed Potassium

** Basic Exchange Capacity

Table 20

Chemical Formulae for a Lacustrine Clay, Winnipeg
 (After Kodama and Brydon 1965)

<u>Over-all Layer</u>									
(K	Na	Ca	(Al	Fe	Mg	(Si	Al) O	(OH)
0.21	0.03	0.08	1.63	0.24	0.16	3.69	0.31	9.895	2.21
Based on a Muscovite Non-swelling Layer									
<u>Swelling Layer</u>									
0.79 x	(Na	Ca	(Al	Fe	Mg	(Si	Al) O	(OH)
	0.04	0.10	1.53	0.30	0.20	3.87	0.13	9.87	2.26
<u>Non-swelling Layer</u>									
0.21 x	KAl	(Si	Al)	O	(OH)				
	2	3	10	2					
Based on an Illite Non-swelling Layer									
<u>Swelling Layer</u>									
0.53 x	(Na	Ca	(Al	Fe	Mg	(Si	Al) O	(OH)
	0.06	0.13	1.69	0.20	0.11	3.80	0.20	9.90	2.20
<u>Non-swelling Layer (Schultz 1964)</u>									
0.47 x	K	(Al	Fe	Mg	(Si	Al	O	(OH)	
	0.06	1.55	0.25	0.20	3.35	0.55	10	2	

the method of Mackenzie (1957b), (Appendix F). This water can be considered to be present in the interlayer positions as H_3O^+ as suggested by Brown and Norrish (1952).

The total tetrahedral positions were assumed to be equal to four, and to be filled with silicon with any deficit made up by aluminium.

Both titanium and manganese were included in the octahedral positions although their presence could also be attributed to minor amounts of rutile and rhodochrosite respectively. In either case the error involved is too small to be significant. A further slight but insignificant error could be incurred by assuming all the iron to be in the ferric state. Magnesium was also assigned solely to the octahedral position although recent work by J. P. Smith (1965) shows a Lake Agassiz II clay, collected at the University, to contain nearly equal amounts of calcium and magnesium exchange ions, with minor sodium. Foster (1951) has also noted the presence of exchangeable magnesium.

Potassium, sodium and calcium were assigned to the interlayer position along with the excess water as mentioned above.

The formulae calculated are given in Table 21 and represent the over-all general formula for interlayered illite-montmorillonite. No correction was made for the minor kaolinite,

Table 21

Chemical Formulae For Yale Avenue Clay Samples

Sample Number	% Sorbed Water	Over-all Layer
Y-7	1.50	(K Na Ca)(Al Fe Mg Ti Mn)(Si Al) O (OH) 0.28 0.07 0.03 1.31 0.41 0.22 0.04 0.004 3.78 0.22 10 2
Y-13	1.60	(K Na Ca)(Al Fe Mg Ti Mn)(Si Al) O (OH) 0.28 0.07 0.07 1.34 0.39 0.21 0.04 0.009 3.70 0.30 10 2
Y-18	1.48	(K Na Ca)(Al Fe Mg Ti Mn)(Si Al) O (OH) 0.28 0.09 0.04 1.34 0.36 0.24 0.04 0.007 3.82 0.18 10 2
Y-26/4	1.03	(K Na Ca)(Al Fe Mg Ti Mn)(Si Al) O (OH) 0.28 0.10 0.04 1.32 0.37 0.25 0.04 0.005 3.82 0.18 10 2
Y-33	0.78	(K Na Ca)(Al Fe Mg Ti Mn)(Si Al) O (OH) 0.29 0.11 0.04 1.36 0.35 0.18 0.04 0.004 3.86 0.14 10 2

feldspar and free iron oxide contents in these samples and the effect of the increasing feldspar content can be seen by the slight increase in the sodium content through samples Y-13 to Y-33. The samples with the least impurities, Y-7 and Y-13, should therefore be the best results to study. Of the two, the results from Y-13 appear to give the closest match to the published formulae.

Compared to Warder and Dion's formulae (Table 19) the formula for Y-13 (Table 21) has appreciably higher silica and lower alumina, iron, and magnesium. Compared to Kodama and Brydon's average formula (Table 19) Y-13 has higher silica and iron, and appreciably less alumina. The magnesium is nearly the same. Compared to Kodama and Brydon's Agassiz clay formula (Table 20) the silica in Y-13 is almost identical, alumina is, however, considerably lower. Iron is higher in Y-13 but it should be for no free iron oxides were removed. Magnesium is also higher in Y-13.

Of the two sets of formulae, Warder and Dion's and Kodama and Brydon's, the latter's should be the more accurate because their sample preparation was more thorough. Warder and Dion assumed that no free iron oxides were present in their samples and made no analysis for them. Kodama and Brydon found free iron oxides in all their samples.

Examining the five calculated formulae (Table 21) more closely a definite trend is evident, at least in samples

Y-13, Y-18, Y-26/4, and Y-33. Sodium and silicon increase with depth while aluminium and iron decrease. These changes are very likely due to the increase in impurities unaccounted for in the corrections.

Further, more detailed work is necessary on this aspect of quantitative mineralogy and chemical formula, but this was not possible in the present study. Having progressed to this stage it was decided to carry the chemical formula calculations one stage farther on the basis of further assumptions. It is of course well understood that these further results are an approximation and must not be accepted literally.

The formula from sample Y-13, Table 21, was chosen for further calculations. This was one of the samples with the least amount of impurities, and the most reasonable calculated formulae. Beginning with the original analysis the ratio of oxides calculated for the carbonates were subtracted as before, and 2% Fe_2O_3 to compensate for the free iron oxides, based on Kodama and Brydon, (1965), was subtracted. The correction for free quartz was boosted from 2.7% to 5% and a further correction for 5% feldspar (assumed to be half potassium and half plagioclase feldspar, Schultz 1964) was made. These results were recast to 100% and then a correction for 3% kaolinite (based on Kodama and Brydon, 1965) was subtracted. (Appendix F). The resulting over-all layer formula, given in Table 22, has a silicon content close to that of Kodama and

Table 22

Estimated Chemical Formulae for Sample Y-13, Yale Avenue

% Sorbed Water	Chemical Formula
2.04	<p align="center"><u>Over-all Layer</u></p> $(K \quad Na \quad Ca) (Al \quad Fe \quad Mg \quad Ti \quad Mn) (Si \quad Al) O (OH)$ $0.28 \quad 0.05 \quad 0.06 \quad 1.39 \quad 0.32 \quad 0.24 \quad 0.05 \quad 0.01 \quad 3.71 \quad 0.29 \quad 10 \quad 2$
2.04	<p align="center">Based on a Muscovite Non-swelling Layer</p> <p align="center"><u>Swelling Layer</u></p> $0.72 \times (Na \quad Ca) (Al \quad Fe \quad Mg \quad Ti \quad Mn) (Si \quad Al) O (OH)$ $0.07 \quad 0.09 \quad 1.16 \quad 0.45 \quad 0.33 \quad 0.07 \quad 0.02 \quad 3.99 \quad 0.01 \quad 10 \quad 2$ <p align="center"><u>Non-swelling Layer</u></p> $0.28 \times KAl_2 (Si_3 Al) O_{10} (OH)_2$
1.85	<p align="center">Based on an Illite Non-swelling Layer</p> <p align="center"><u>Swelling Layer</u></p> $0.52 \times (Na \quad Ca) (Al \quad Fe \quad Mg \quad Ti \quad Mn) (Si \quad Al) O (OH)$ $0.10 \quad 0.12 \quad 1.25 \quad 0.40 \quad 0.28 \quad 0.05 \quad 0.02 \quad 3.93 \quad 0.05 \quad 10 \quad 2$ <p align="center"><u>Non-swelling Layer (Schultz 1964)</u></p> $0.48 \times K (Al \quad Fe \quad Mg \quad Ti) (Si \quad Al) O (OH)$ $0.06 \quad 1.50 \quad 0.25 \quad 0.20 \quad 0.05 \quad 3.45 \quad 0.55 \quad 10 \quad 2$

Brydon, Table 20, although the aluminium content is still low. The other cations are reasonably similar.

In order to calculate the composition of the swelling layer Kodama and Brydon assumed the composition of the non-swelling mica-like layer to be that of ideal muscovite. In their X-ray analysis they found the mica-like (illite) component to make up 35 to 50% of the sample and the montmorillonite component to make up the remaining 50 to 65%. In their chemical formula calculations, by assigning the potassium to an ideal muscovite composition they obtain average values of 20% mica-like layers and 80% montmorillonite (Table 20).

Following the procedure of Kodama and Brydon and assigning the potassium content to an ideal muscovite formula, sample Y-13 yielded the results given in Table 22, giving a ratio of 28% mica-like and 72% montmorillonite layers. These results are close to the 21% and 79% obtained by Kodama and Brydon (Table 20).

The use of an ideal muscovite formula for the non-swelling layer is a debatable point. Interlayering of illite and montmorillonite would produce an illite with a composition closer to that of montmorillonite rather than muscovite. Schultz (1964) gives an approximate formula for the illite of the Pierre Shales (the continuation of the upper part of the Manitoba Cretaceous shales in the United States) repro-

duced in Tables 20 and 22. Assigning the total potassium content of sample Y-13 to this formula yields the formula for the montmorillonite given in Table 22, and the proportion of 48% illite and 52% montmorillonite (Appendix F). If the same illite formula were used for the Kodama and Brydon data, instead of the ideal muscovite formula, a similar proportion of illite to montmorillonite results. Table 20 contains Kodama and Brydon's formula for a sample of Winnipeg clay calculated with the ideal muscovite formula and then recalculated with the illite formula of Schultz. The recalculation changes the estimated mica-like layers from 21% to 47% illite and the montmorillonite from 79% to 53%. These new results agree with the X-ray results of Kodama and Brydon and substantiate the use of Schultz' illite formula over the muscovite formula.

The comparison of the composition of the swelling layer for sample Y-13 (Table 22) to that of Kodama and Brydon's swelling layer for the Winnipeg clay (Table 20) shows that the former is very close to the montmorillonite end of the montmorillonite-beidellite series whereas the latter is more beidellitic although still a montmorillonite (see Ross and Hendricks (1945) and Foster (1954)). Caution must be exercised in using the formula of sample Y-13 because of the numerous assumptions used during the calculations.

One important result that can be accepted is the

proportion of 48% illite to 52% montmorillonite calculated from sample Y-13. This is almost identical with Kodama and Brydon's Winnipeg sample which in turn was very similar to the six other prairie clays in their study. The conclusion can thus be drawn that, because of the similarity of DTA and X-ray diffraction results in all the samples used in the present study, the ratio of illite to montmorillonite found for sample Y-13 applies to all the Lake Agassiz clays studied. That is, the ratio of illite to montmorillonite is very close to equality with a slight dominance of montmorillonite. The montmorillonite itself appears to be near the montmorillonite end of the montmorillonite-beidellite series. Also the assumed chemical formula of illite used by Schultz seems to be a practical one for these studies.

C H A P T E R V I I

S U M M A R Y A N D C O N C L U S I O N S

I N T R O D U C T I O N

Conclusions based on the results of this study, coupled with the results of similar studies, are drawn in two fields, 1) the DTA method and 2) the mineralogy of the deposits. Finally an interpretation of the glacial history is developed, based on the mineralogy.

D I F F E R E N T I A L T H E R M A L A N A L Y S I S

Concerning the DTA method it can be concluded:

1. The dynamic gas technique of DTA is superior and more useful than the static technique.
2. The dynamic DTA method is an excellent method of rapid routine analysis, with a minimum of sample preparation.
3. Excellent quantitative carbon dioxide contents can be determined to within 1% by a careful standardized procedure.
4. Quantitative quartz content determinations can be made to within 3% from ordinary thermograms and results within 1% can be obtained if the determination is made after a sample is heated twice. More chemical work is necessary to firmly establish this method and to further check the accuracy of the results.

5. Illite, montmorillonite, kaolinite, gypsum, limonite and pyrite can all be identified even in minor amounts, with the exception of minor kaolinite in illite-montmorillonite.
6. The dominant exchange cation on montmorillonite can generally be determined from the thermogram.
7. DTA can not be used solely on its own, but has to be coupled with other methods of analysis particularly X-ray diffraction and chemical analysis. However, DTA is a good method to rapidly study a large number of samples from which a few can be selected for more careful detailed work by other methods.

MINERALOGY

Gray Till

1. The same randomly interstratified illite-montmorillonite present in the clay beds occurs in amounts between 15 and 20% in the main body of the till bed and as low as 5% at the base of the till bed.
2. The main non-clay minerals are dolomite, calcite, quartz and feldspar, with traces of gypsum.
3. Dolomite makes up about 40% of the material in the main body of the till and nearly 50% at the base.
4. Quartz, feldspar, and calcite make up the balance of the non-clay minerals and are present in roughly equal amounts.

Clay Beds

1. The three clay beds, the greenish brown clay bed of Lake Agassiz II, and the brown and blue-gray clay beds of Lake Agassiz I, have a common mineralogy.
2. The beds are composed mainly of randomly interstratified dioctahedral illite-montmorillonite.
3. The coarse clay fractions tend to be more dominantly illite and the fine clay fractions tend to be more dominantly montmorillonite.
4. The over-all ratio of illite to montmorillonite is close to 50:50 with a slight dominance of montmorillonite.
5. Calcium is the dominant exchange cation through the Blue-gray and all but the top few feet of the Brown clay.
6. Magnesium is the dominant exchange cation of the Greenish brown clay.
7. A 7Å phase, probably kaolinite but possibly chlorite, is present in minor amounts in the clay fraction.
8. Non-clay minerals found with the clays are dolomite, calcite, quartz and feldspar, with traces of gypsum, limonite and possibly very minute traces of pyrite.
9. Of the non-clay minerals, dolomite is dominant and becomes more so with depth, whereas the others, calcite, quartz and feldspar are generally present in equal amounts.

10. In the Greenish brown clay the non-clay minerals make up approximately 10% of the material.
11. In the Brown clay the non-clay minerals make up approximately 25% of the material present at the base of the bed, and decrease continually to about 10% at the top.
12. In the Blue-gray clay the non-clay minerals make up approximately 40% of the material present at the base, and decrease continually to approximately 25% at the top of the bed.

Yellow Silt (Loess)

1. Fifty percent non-clay and 50% clay minerals make up the yellow silt material at the base of the lower silt bed. (It is thought to contain less clay above the lower contact).
2. Randomly interstratified dioctahedral illite-montmorillonite makes up most of the clay minerals.
3. Magnesium is the dominant exchange cation on the interstratified clay.
4. Kaolinite, though minor, appears to be more abundant in this bed than in any other.
5. Non-clay minerals are dolomite, calcite, quartz and feldspar, with minor limonite.
6. Quartz and feldspar are the dominant non-clay minerals.
7. More limonite, 2 to 3%, is present in this bed than any other.

Fine Sand

1. Illite-montmorillonite is present in a minor amount, less than 5%.
2. Dolomite is the main mineral and comprises 46% of the material.
3. The other minerals are quartz, 30%, feldspar approximately 20% and minor calcite, nearly 3%.

SIGNIFICANCE OF THE RESULTS

Mineralogy of the Bedrock of Manitoba

The bedrock of Manitoba falls into three main divisions: Precambrian, largely granites, granodiorites, quartz diorites, granitic gneisses and some volcanic and sedimentary rocks; Paleozoic, largely dolomites and some calcareous formations with minor argillaceous beds, and one sandstone formation; and Mesozoic, largely shales and calcareous shales with minor gypsum and limestone beds. The rock types in these three areas each have their own distinctive mineralogy, so that each area can be characterized by certain individual minerals. Some of these characteristic minerals do occur in more than one area but in such cases the second occurrence is minor in comparison with the main occurrence. It is possible, therefore, by studying the mineralogy of the Agassiz sediments to determine the proportion that each of the three source areas has contributed to any particular sediment.

The clay minerals found in the Agassiz beds can be

attributed ultimately to the Cretaceous shales. The possible contribution of argillaceous members of the Paleozoic rocks is too minor to be significant. Quartz and feldspars originate mainly from the igneous and metamorphic rocks of the Precambrian although a minor amount of the quartz will have originated from the Winnipeg Sandstone at the base of the Paleozoic. Dolomite and calcite have two possible source areas, the major source being the Paleozoic rocks and the minor source being the Cretaceous rocks. Despite the two possible source areas, the carbonate content of the Agassiz sediments can be attributed mainly to one source or the other because distinct differences exist between the two sources. In the Paleozoic rocks dolomite is the dominant mineral although calcite is also a major mineral. However in the Cretaceous rocks calcite is the dominant carbonate mineral and dolomite is very minor if present at all. Gypsum -- a trace mineral throughout the Agassiz sediments -- originates from the Jurassic Amaranth formation and from the Silurian (?) gypsum formation of the Interlake Group of formations.

Interpretation of the Mineralogy of the Yale Avenue Testhole

Gray Till

The dominance of dolomite, approximately 40%, indicates that the underlying Paleozoic dolomites were the major source of the till at this site. The importance of the immediately underlying bedrock in the mineralogical make-up of the till has

been pointed out by Ehrlich and Rice (1955) and the above results agree with their studies of Manitoba tills. Approximately 15% quartz and an assumed 15% feldspar reveals that the Precambrian rocks were the secondary source material for the till. The contribution of the Cretaceous shales to the till is minor for the clay content is low, 15 to 20%, throughout. It appears therefore that the shales had been largely removed from the general sample area prior to the last glaciation.

A slight increase in the dolomite, up to nearly 50%, and the quartz, up to approximately 17%, and a consequent decrease in the clay content in the lower two to three feet of the till to 5-10%, illustrate the influence of the immediately underlying bed rock and indicate that the Cretaceous shales made even less contribution to the lower portion of the till than to the main body of the till.

Lake Agassiz I Unit

Visual examination of the lake clays indicate that the ice-rafted pockets of yellow calcareous till decrease up through the beds. This till material is characterized by the non-clay minerals, dolomite, calcite, quartz and feldspar. When these minerals are found in the lake clays they can in general be attributed to having originated mainly from the till. The source of the clay minerals in the lake clays cannot be attributed to the till for the clay mineral content of the till is low and thus the contribution of clay minerals

from the till must necessarily be minor. The abundance of clay minerals in the lake clays must, therefore, be attributed to some action other than liberation from the melting glacier. The only other possible origin is transportation to the lake by runoff water passing over the exposed Cretaceous shales to the west and south of the lake.

Because the minerals of the lake clays have these two possible sources it is possible to learn something of the movement of the glacier by studying the distribution of the clay and non-clay minerals.

The immediate increase in clay minerals at the base of the Blue clay bed indicates that the lake regime immediately followed the glacial regime. The rapid decrease in non-clay minerals, as represented by the carbon dioxide content, (Table 14, p. 146) in the basal few feet of the blue-gray clay bed shows that the glacier was retreating fairly rapidly. During this period the non-clay material would be originating from ice-rafting as well as direct contribution from the melting glacier. This direct contribution would drop off as the glacier moved away from the site and ice-rafting would become the main source of the non-clay minerals. This change in source may account for the slower rate of decrease of the non-clay minerals up through the main body of the blue-gray clay bed as compared to the rate of decrease in the basal few feet. A change in the rate of retreat of the glacier is another possible explanation. If sampling could be carried out over a large

area some interpretation of the rate of glacier retreat could be made, based on the rate of decrease of the non-clay minerals. However, from a single site one can only say that the steady decrease in non-clay minerals up through the Lake Agassiz I beds marks the continued retreat of the glacier carrying the main source of the non-clay minerals farther and farther from this site of deposition.

The structureless nature of the blue-gray clay bed has been attributed by Elson (1961) as due to either 1) deposition in deep water near the edge of the glacier, or 2) deposition away from the glacier in water shallower than the wave base where wave action could prevent varves from forming. He has further stated, "The upper laminated clay was laid down in deep water when the ice margin was remote." (p.65). In view of the mineralogical evidence stated above the first hypothesis is unlikely except for the early stages, for the glacier appears to have continually retreated from the deposition site. The second hypothesis also appears unlikely for Lake Agassiz I was at its greatest depth in its early stages and gradually decreased in depth throughout its history, (Davies, Bannatyne, Barry and McCabe, 1963) so that the lake was shallower when the varved brown clay was deposited than it was when the massive blue-gray clay was deposited. Also although no varves are present in the blue-gray clay numerous silt pockets are present as sharply defined separate entities with no mixing at the contact between the silt and clay. If

wave action was strong enough to destroy varves it most certainly would have destroyed the silt pockets.

The transition between the massive blue-gray clay and the varved brown clay occurs relatively sharply over a few feet. The lack of any mineralogical difference between the blue-gray clay and the brown clay and the smooth uninterrupted decrease in the non-clay minerals (Table 14, p. 146) illustrates that the change that occurred in Lake Agassiz I at this time did not affect the material that was being carried into the lake, but only the structure of the material as it was deposited.

Grain size analysis of the light layers (summer) and the dark layers (winter) of varves¹ have been carried out by Wallace (1927), Rittenhouse (1934), Legget and Bartley (1953) and Eden (1955), on materials either deposited in Lake Agassiz or in associated glacial lakes. In all cases careful work has shown that the dark layers are finer grained than the light layers. The dark layers also have a higher plasticity and moisture content (Legget and Bartley, 1953, and Eden, 1955). Eden (1955) found that the light layers he studied were composed of quartz, carbonates, feldspar, a small amount of clay, and traces of organic matter; and the dark layers were composed of quartz, feldspar, more clay (probably montmorillonite) and slightly more organic matter than the light layers.

¹ A varve is not a single layer but is composed of two layers, a summer and a winter layer. Thus, a single varve represents a single year of sedimentation.

The classic explanation of varve formation is that during the spring and summer months silt and clay size material is carried into the lake by streams from the land and the glacier. The larger size particles, the silt, naturally settles faster than the clay size particles and forms the light coloured layer during the summer months. During the fall and winter months when the lake is frozen and no new material is introduced to the lake, the clay size material settles and forms the dark coloured layer. The repetition of this process year by year produces a sequence of varves.

Antevs (1951) has listed the conditions affecting varve formation: 1) restriction of appreciable mud supply to the warm months, 2) fluctuation of the mud influx and of the currents, 3) differential rate of settling of unequal grains and particles, 4) low temperature and accompanying high viscosity and density of the lake water, 5) semi-stratification or isothermy of the lake water and oscillations between these states, 6) height of fall of the particle, and 7) variations in the concentration of the electrolytes.

Burwash (1938) has suggested that the carbon dioxide content of the water is very important. He pointed out that cold water has a greater capacity for the solution of gases. Thus a cold lake could hold more carbon dioxide in solution than a warm lake. During the winter a lake would be more acidic and be able to hold carbonates in solution, however as

the lake warmed in the summer carbon dioxide would be released and calcium carbonate would be deposited. As an example of this Burwash mentioned the white partings often found in the summer layer. Eden (1955) supported this theory and suggests that this may be the reason no carbonate minerals were found in the winter layer of the varves he studied.

The deposition of massive blue-gray clay followed by varved brown clay in the same lake from the same material is an unanswered problem. Obviously one or several of the conditions listed above were not right for varve formation in the early stages of Lake Agassiz I. The most probable factors are the concentration of salts, depth of water, the thermal currents, or the proximity of the glacier and its effect on the climate. Concerning the effect of salts, Flint (1957) has stated, "The salts in sea water, acting as electrolytes, are believed to flocculate suspended sediment, causing it to settle as homogeneous masses of particles of various sizes and preventing the creation of mechanical rhythmites. Apparently, however, flocculation is not confined to sea water." (p. 295). He goes on to say that Eden (1955) found that the material within individual dark and light layers of a varve have no gradation of particle size and thus did not have graded bedding as suggested by Antevs (1951). He attributed this to flocculation of the particles during deposition, possibly due to variations in carbon dioxide concentrations as proposed by Burwash (1938). It seems possible in the case of the lower

Lake Agassiz I sediments, that carbon dioxide or some dissolved salt prevented the varves from forming by continuously flocculating the particles to produce the homogeneous bed instead of just flocculating material with individual varve layers as Eden found.

The effect of the depth of a lake as a factor in varve formation has not been considered in the discussions of Lake Agassiz. The depth of Lake Johnston, a varve forming glacial lake at Steep Rock, Ontario contemporaneous with Lake Agassiz, was estimated by Antevs (1951) as being 125 to 275 feet deep. Lake Louise, Alberta, where varves are forming at the present time is approximately 180 feet deep. Lake Agassiz I was initially 650 feet deep in the Winnipeg area when the massive blue-gray clay was deposited, and it gradually lowered and eventually drained. At some point during the lowering the varved brown clay was deposited. Lake Agassiz II had a maximum depth of approximately 350 feet in the Winnipeg area. Lake Agassiz II clays are varved as are the upper Lake Agassiz I clays, but the lower Lake Agassiz I clays are massive. It appears, therefore, that there may be a critical depth between 350 and 650 feet at which varves, for some reason, will not form.

The above possible explanations are only the more obvious possibilities, other factors or combination of factors could be involved. The problem requires much further detailed study.

The calcite-dolomite ratio (Table 11, p. 140) is also useful for interpretation. In the main body of the till, the ratio is 21% calcite to 79% dolomite. In the blue-gray clay 3 feet above the contact the ratio is 26% calcite to 74% dolomite. Seven feet above this, still in the blue-gray clay, the ratio is 30% calcite to 70% dolomite. In the brown clay where the decreasing carbonates content is more gradual the results are 36% and 37% calcite to 64% and 63% dolomite. This increase in calcite may indicate a minor change in the character of the source till, or an introduction of more calcite relative to the dolomite from Cretaceous shales, or the possible formation of calcite from solution as proposed by Burwash (1938).

Lake Agassiz I - II Interval Bed

The mineralogy of the two bulk samples from the loess bed is somewhat misleading. It shows a moderately high clay content of approximately 50% and moderate non-clay mineral content of calcite 4.3%, dolomite 8.1%, quartz 16.3% and feldspar assumed also to be 16% (Y-8/1, Table 12, p. 142). This is not representative of a loess; however, examination of the hand specimens offers an explanation. The sample is not a massive structureless silt deposit but is composed of fine laminations of clay between laminations of very pure wind-blown silt. The clay laminations are present in both the samples examined but are most abundant in the lower sample at the contact. Unfortunately the complete sample of this bed was not recovered so that its total thickness is unknown. Therefore

it is not known whether or not this finding applies to the complete bed or some unknown thickness of the lower portion.

The interpretation of the mineralogy and structure of the loess indicates that while the silt was blowing across the dry lake bottom some minor flooding, due to fluvial or intermittent lacustrine activity, was producing the thin pure clay lamination.

Lake Agassiz II Unit

The clay bed of Lake Agassiz II is mineralogically very similar to the upper part of the brown clay bed of Lake Agassiz I. Table 12, (p. 142) shows the similarity in calcite, dolomite, quartz and the clay minerals. Even the ratio of 37% calcite to 63% dolomite (Table 11, p. 140) is identical. This similarity is a result of nearly identical sedimentary conditions despite different glacier movements. During the final stages of Lake Agassiz I the glacier lay far to the north of the sample site so that contributions to the lake sediments from the till by ice-rafting were very minor and the bulk of the sediments were derived from the Cretaceous shales. Then the glacier retreated far enough to allow drainage into Hudson Bay and Lake Agassiz I to drain away. Lake Agassiz II formed when the Labradorean glacier joined the readvancing Keewatin glacier and blocked the drainage into Hudson Bay, trapping the meltwater. Thus although the glaciers were advancing rather than retreating, the source areas were still the same with the

Cretaceous shales providing the bulk of the material and the till providing very minor amounts.

The mineralogy of the fine sand bed found at the top of the Lake Agassiz II unit in the Yale Avenue test hole is distinctly different from the other sediments studied. The clay mineral content is nil and the calcite to dolomite ratio is 5% calcite to 95% dolomite. Both these features are in sharp contrast with all the other Agassiz sediments.

The relatively coarse grain size, lack of clay minerals and general occurrence suggest a fluvial or fluvial-lacustrine origin. The dominance of dolomite suggests a much more restricted source, probably Stony Mountain lying to the northwest, than the broader source of the other sediments.

FURTHER WORK

The limited success of this study suggests that with a thoroughly developed X-ray diffractometer method of analysis, the present system could be used for a systematic study of the Lake Agassiz sediments throughout the basin. The DTA method could be used as a control method to determine gross mineralogical variations through depth and lateral extent and also to select representative samples for more thorough X-ray diffraction and X-ray fluorescence analysis.

The success of Rominger and Rutledge's study in 1952 further suggests that an integrated geological, mineralogical, and engineering study would yield excellent results.

A P P E N D I X A

S U B S U R F A C E G E O L O G Y

O F

M E T R O P O L I T A N W I N N I P E G

by

George A. Russell, P. Eng.

I N T R O D U C T I O N

While the subsurface geology will be most clearly and most succinctly illustrated by the diagrams which accompany this appendix, it was thought that a brief general survey might be of value.

Most of the detail regarding the strata on which Winnipeg is founded and which underlie the area has been made available through the whole-hearted cooperation of many people, too numerous to list here. Exploratory boring companies, foundation specialists, consulting engineers and architects and municipal, provincial and federal agencies have literally 'opened their records' to the author's inspection. It must be pointed out that such cooperation represents a great deal of generosity because subsurface records in the field of foundations, much as in the field of water supply, represents valuable technical data acquired at great expense and the field is distinctly competitive.

GEOLOGICAL SETTING

Metropolitan Winnipeg (hereafter simply called Winnipeg) is located at the junction of the Red and Assiniboine Rivers on the broad, flat prairie which was previously the bottom of glacial Lake Agassiz. The prairie has an elevation of about 750 feet above sea level.

Going westward from Winnipeg the first topographical break is the Pembina Escarpment. The entire face of the escarpment is composed of Cretaceous shales which outcrop in various places where the rock projects through glacial till. The shales contain numerous beds and seams of bentonite and, on bedding planes, occurrences of gypsum are plentiful. At some places the shales contain sufficient hydrocarbons so that thin splinters may be ignited with a match.

Going eastward from Winnipeg the Agassiz clays thin out gradually and are replaced, as the mantle rock, by glacial till. The first outcrops of bedrock which are encountered to the east are Precambrian granite. Thus, the Agassiz clays cover the 'outcrops' of geological formations which include the lower part of the Cretaceous, the Jurassic (evaporite deposits with abundant gypsum) and all of the Paleozoic represented in this part of Manitoba. These sedimentary formations have an average strike of a few degrees west of north and dip to the west at low angles which, in the Winnipeg area, would not exceed five degrees.

Northward from Winnipeg the Agassiz clays are also gradually replaced as the surface mantle by glacial drift. Bedrock outcrops encountered 10 to 20 miles north of Winnipeg are of Paleozoic limestones.

Southward from Winnipeg the Agassiz clays continue for nearly 400 miles to the south end of former glacial Lake Agassiz. The clays, which have a maximum depth of about 55 feet in the Winnipeg area increase in thicknesses slightly to over a hundred feet in North Dakota. As would be expected, no significant topographic break occurs to the south.

The higher ground which borders the Agassiz basin on the north, west and east is the charging area for an artesian aquifer which underlies most of the valley. Numerous data indicate that the aquifer is, almost exclusively, a fractured zone at the surface of the underlying Paleozoic limestones. The water is impounded in this fractured zone by the highly impermeable, overlying glacial till. This glacial till forms the lowermost strata of the Pleistocene deposits in the Winnipeg area and is continuous with the glacial till which appears as the surface mantle to the north, west and east whenever the boundary of the glacial clays has been passed.

Static level varies from place to place and, in general, has been dropping for the past 30 years. One observation well in the centre of Winnipeg shows a level of 733 feet above sea level. Seven miles to the south, exploration for

bridge foundations showed a static level 17.5 feet above prairie level at an elevation of about 773 feet.

STRATIGRAPHY

Because the illustrations which follow have been plotted to scale, the present discussion will be entirely qualitative.

The uppermost layer is composed of weathered clays with varying amounts of organic material. Many test holes show superficial occurrences of foreign fill.

Below the organic layer is a zone of brown clay which has varying amounts of yellowish to buff silts. One bed of silt is quite extensive and lies at the approximate depth of thousands of residential foundations. It has been responsible for numerous failures. The brown clay in this zone often shows what is called nugget structure. This structure is the result of dehydration of the clay (with the formation of columnar jointing) either by evaporation or by freezing.

Below the zone of brown clay and silt is a layer of brown or chocolate clay which varies greatly in thickness but would average about 20 feet. Below this is a layer of blue or gray clay which is about 20 feet in thickness. The two are separated, in some places, by a transition zone composed of alternating layers of brown clay and blue clay.

The formation which occurs immediately below the blue

clay is called the 'putty layer' (by drillers) and non-cemented glacial till. This latter term was established by the late John Mishtak of The Water Control and Conservation Branch, Department of Agriculture, Province of Manitoba. It is composed essentially of rock flour formed by the comminution of limestone and dolomite. It may contain granite pebbles and is not continuous throughout the entire area.

Below the non-cemented glacial till is the last layer of the Pleistocene and has been designated as cemented glacial till by Mishtak. It contains abundant fragments of both Precambrian and Paleozoic rocks in a matrix of calcareous and argillaceous clay and silt.

The Pleistocene deposits are underlain by a zone of fractured and fissured limestone and dolomite which contain some red shaly sections. The surface is quite irregular having been channeled by pre-glacial streams and disrupted by ice action which has torn loose and tilted some large blocks of limestone. Isolated, pipelike conduits occur in the solid limestone bedrock giving rise, for example, to highly variable water conditions in a single foundation. In one well, a borehole television camera showed a 4-inch hole approximately in the centre of the 12-inch bore. Abundant water was available in this hole whereas holes only a short distance away were completely dry.

The irregularity of the bedrock surface has resulted

in a draping effect in the overlying glacial till. It is believed that this draping effect has created traps for gas, often under high pressure, which is encountered periodically in auger borings. A sample of gas from one hole which 'blew' continuously for about 12 hours was found to be 95% non-combustible by ordinary chemical means -- a mass spectrograph not being available. The gas has accounted for the death of one construction worker who was suffocated while belling out the lower portion of a caisson excavation.

The origin of the gas in the pockets is not certain. The author found records of gas samples from the Virden oil field in western Manitoba which had been mass spectrographed and proved to be 95.4% nitrogen. It is believed possible that this gas might have migrated up dip and been trapped under the impermeable glacial till and Agassiz clays. In this connection it is interesting to note that natural hydrocarbons have been found to occur within the area of Winnipeg in two ways. First, there have been reports of oily water in some wells east of the Red River. Second, the author, noting the similarity in colour between the brown or chocolate clays and oil well cores from Virden, carried out ether extractions on clays from a sufficient depth to obviate any suggestion of slop oils from motors, etc. When evaporated on a white china plate the ether extraction left a brown ring of hydrocarbon. Tests under ultraviolet light showed a strong fluorescence similar to that described for natural hydrocarbons.

ORIGIN OF THE DEPOSITS

Agassiz Clays

It is suggested that the material which forms the clays was eroded from the Cretaceous shales of the Pembina escarpment by wave action and flowed as heavy density underflows out into the valley. As a provenance, the Cretaceous shales are capable of supplying all components of the Agassiz clays including the active clay minerals, the calcium sulphate (gypsum) content of the clays and, perhaps, the hydrocarbons which occur in the clays, the latter being derived from those portions of the shale which, as has been reported previously in this paper, may be ignited by a match.

Silts

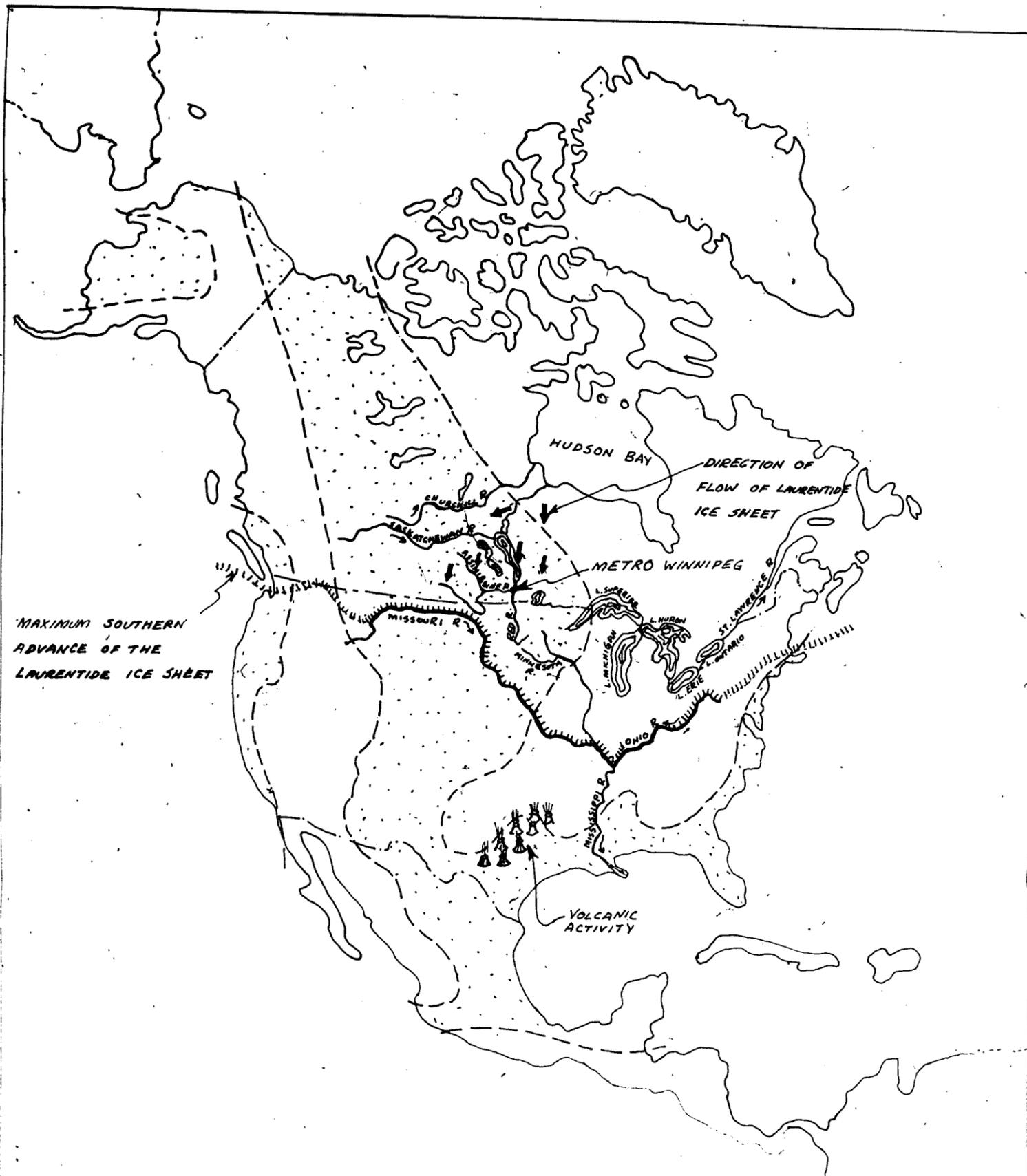
When Lake Agassiz was finally drained, the broad delta of the Assiniboine was exposed stretching from Brandon to Portage la Prairie. It is believed that denudation of this surface by wind action, with prevailing northwesterly winds, drifted fine silt eastward over the remnants of Lake Agassiz.

A second occurrence of silty material should be noted at this time. Numerous excavations reveal that both the Assiniboine and the Red rivers are operating in second channels, i.e., a first channel was cut and then for some reason filled in with river alluvium. The rivers are now cutting a second channel in this alluvium. The river alluvium extends outward

for a few hundred feet on each side of the rivers and has given rise to especially difficult foundation conditions. The alluvium, composed essentially of silt with abundant organic residue, contains practically no clay minerals which have been flushed out. As a result, the inherent binding strength of the clay minerals has been lost.

Glacial Deposits

The glacial deposits are a ground moraine deposited during the last retreat of the ice. In some places, test borings have shown thicker accumulations of drift under the city. It is believed that these are terminal moraines deposited by a longer halt of the ice front at the places where the moraines occur.

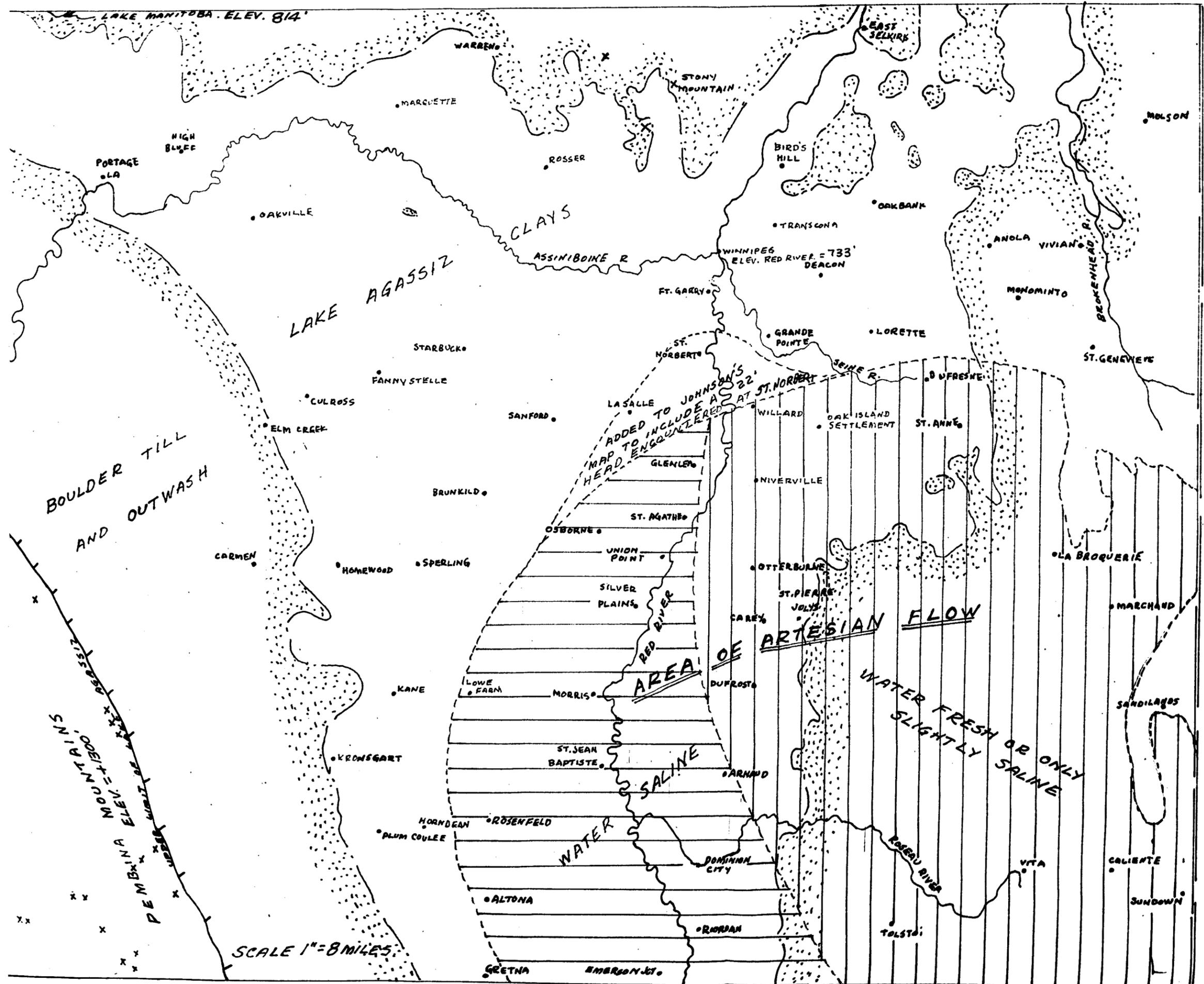


NOTE:-

The Cretaceous sea indicated represents the last epicontinental sea covering the area of Metropolitan Winnipeg. The area probably emerged as a vast, nearly flat, muddy plain characterized by numerous streams with a dendritic pattern similar to numerous tidal flats in existence today.

According to Upham, this broad flat plain was incised by a major, preglacial river system which may have been composed of the then northward flowing Missouri and Sheyenne Rivers which flowed into the preglacial Red River. This major valley cutting (whether the valley was cut to the harder, pre-Cretaceous sediments is not known at present) set the stage for the subsequent glaciation.

MAP SHOWING MAXIMUM EMBAYMENT OF NORTH AMERICA BY CRETACEOUS SEA.

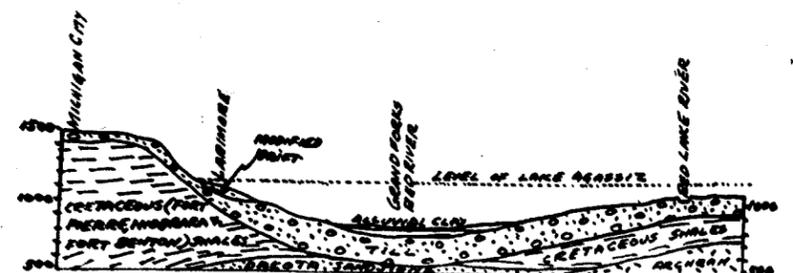




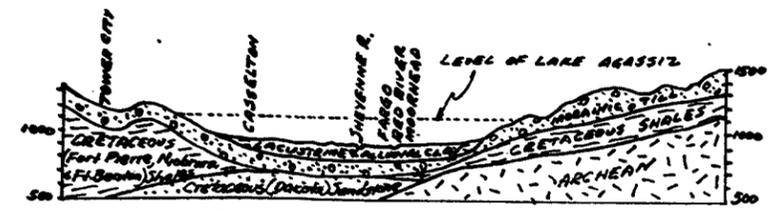
HORIZ. SCALE 1" = 15 MILES
SECTION AA (After Warren Upham)



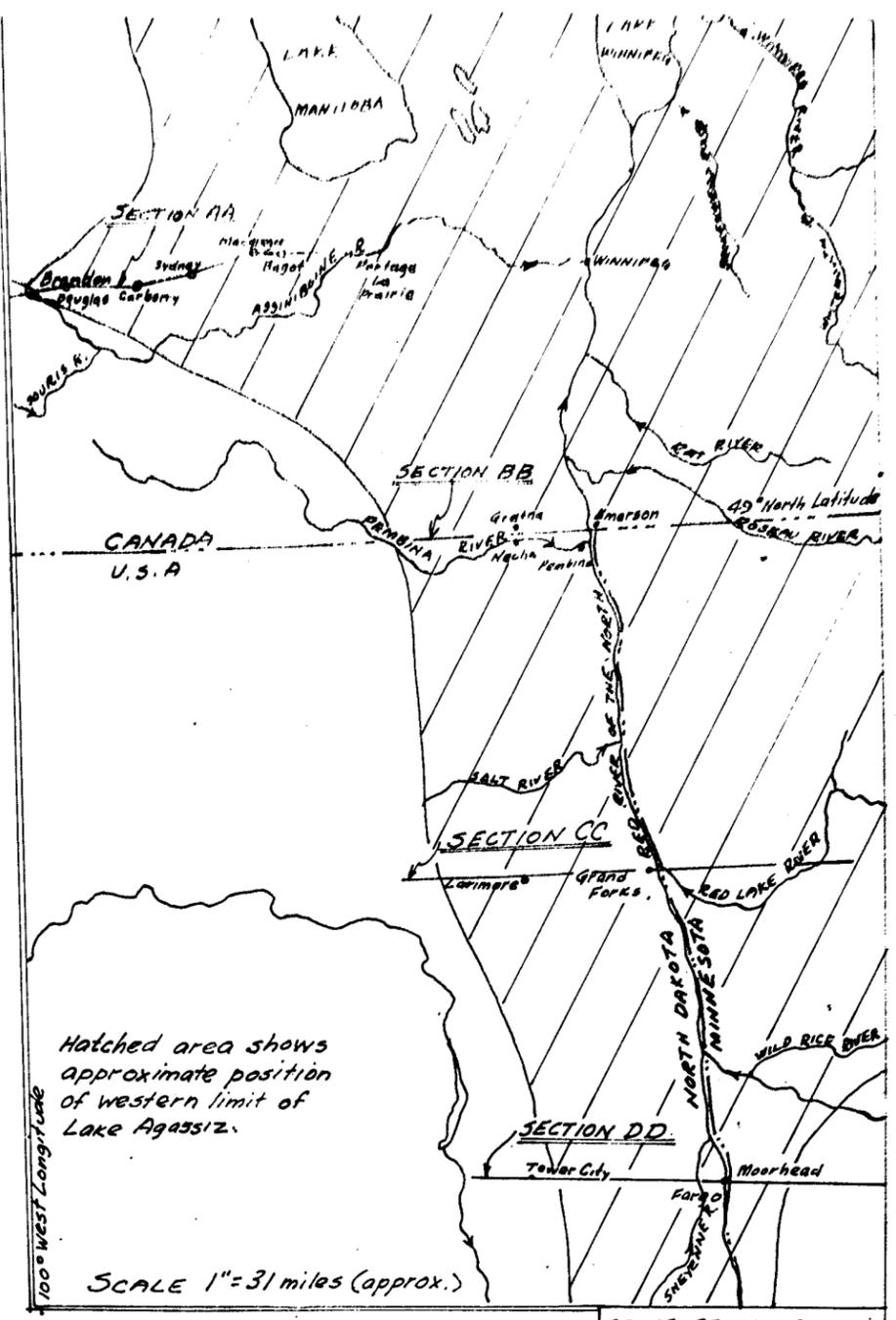
HORIZ. SCALE 1" = 20 MILES (After Warren Upham)
SECTION BB



HORIZ. SCALE 1" = 20 MILES (After Warren Upham)
SECTION CC



HORIZ. SCALE 1" = 20 MILES (After Warren Upham)
SECTION DD

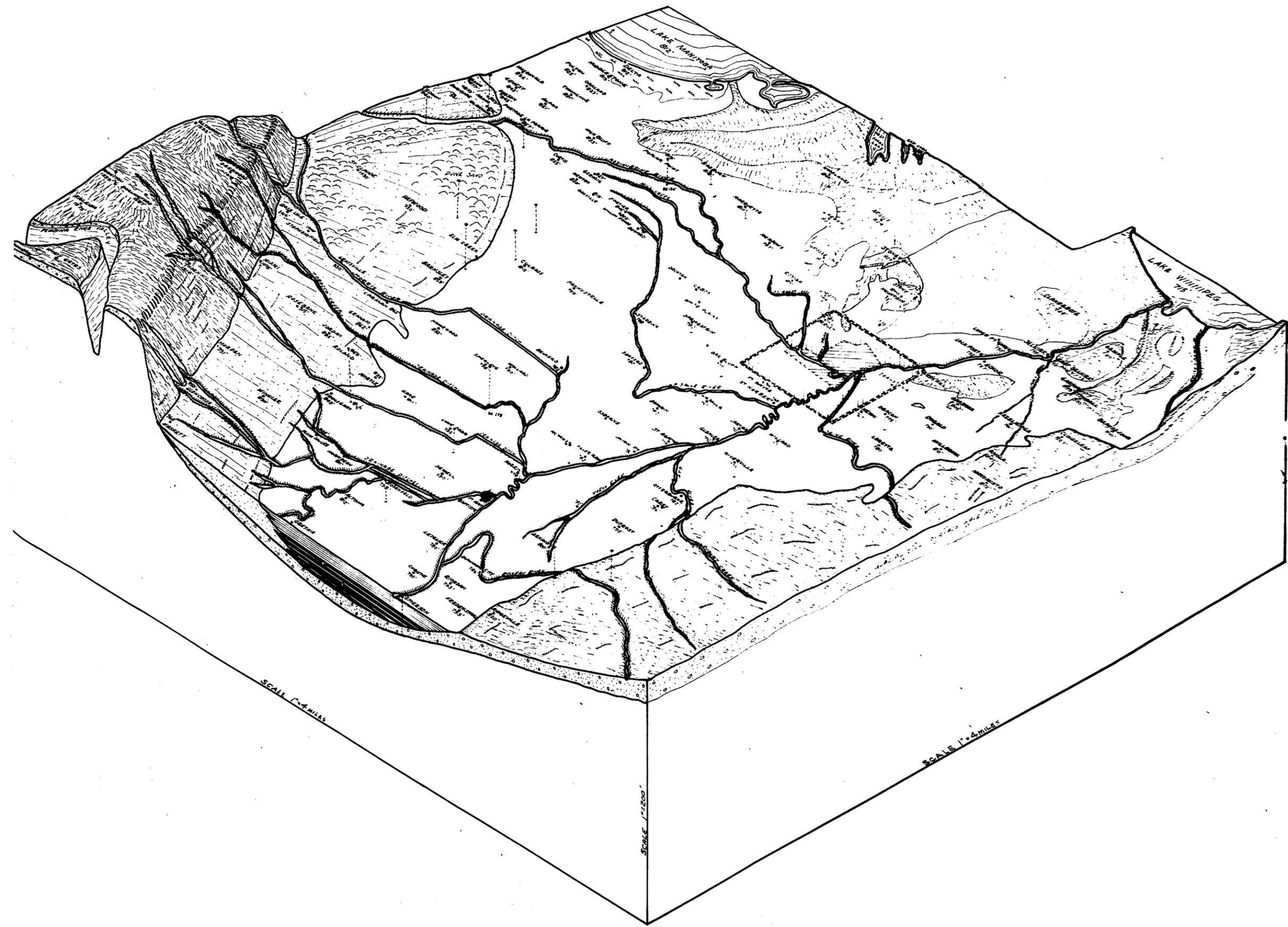


Hatched area shows approximate position of western limit of Lake Agassiz.

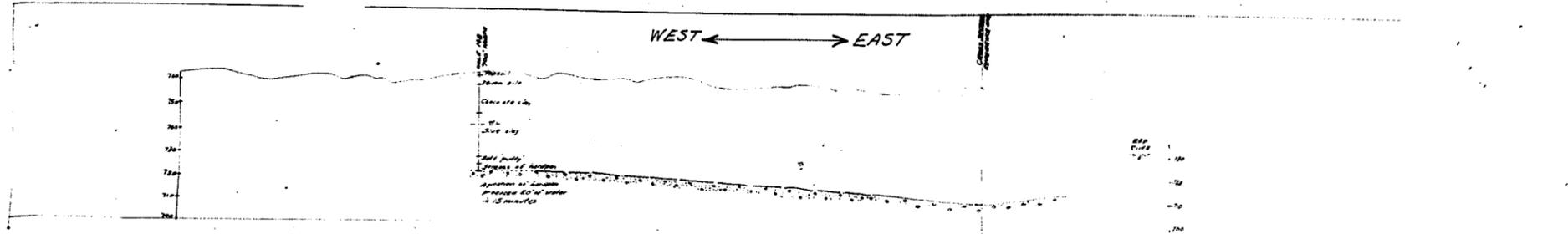
SCALE 1" = 31 miles (approx.)

Note especially Upham's interpretation of the pre-glacial surface as being composed almost entirely of Cretaceous shales. Upham further postulates that the shale terrain was dissected by a pre-glacial river system.

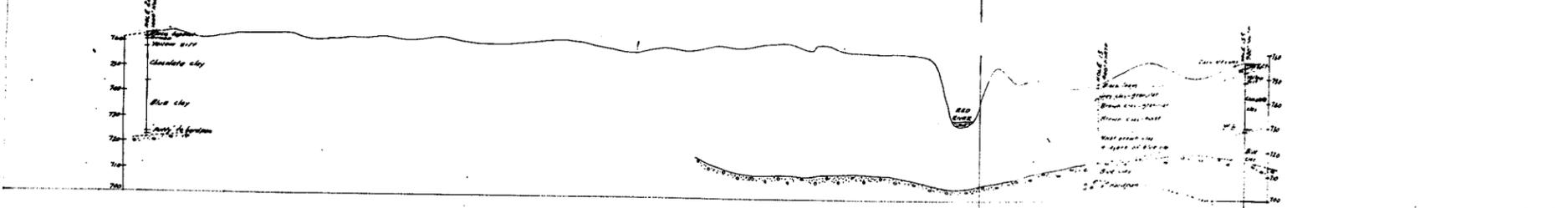
CROSS SECTIONS SHOWING W. UPHAM'S INTERPRETATION OF GEOLOGICAL SETTING OF GLACIAL LAKE AGASSIZ SCALES AS INDICATED



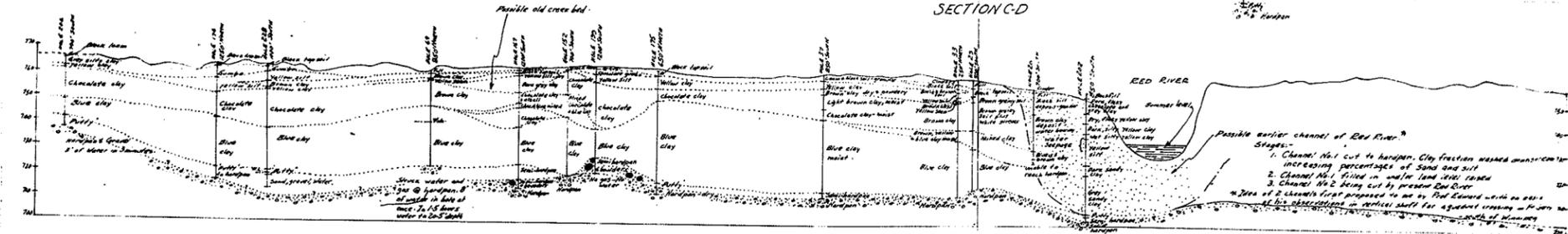
WEST ← → EAST



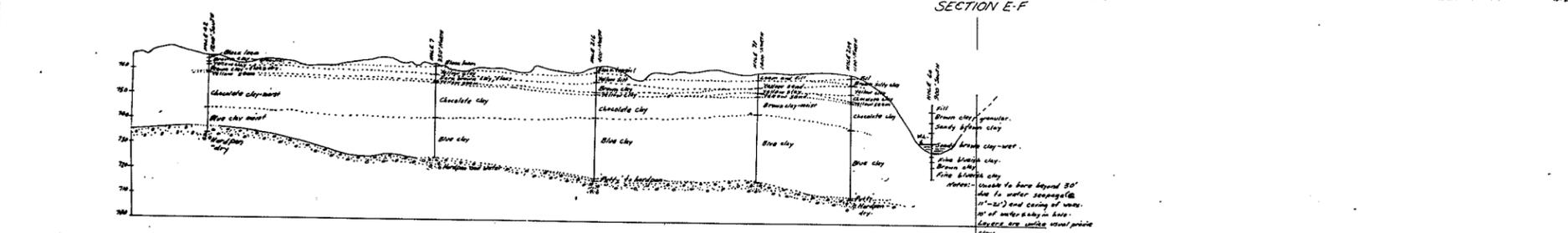
SECTION A-B



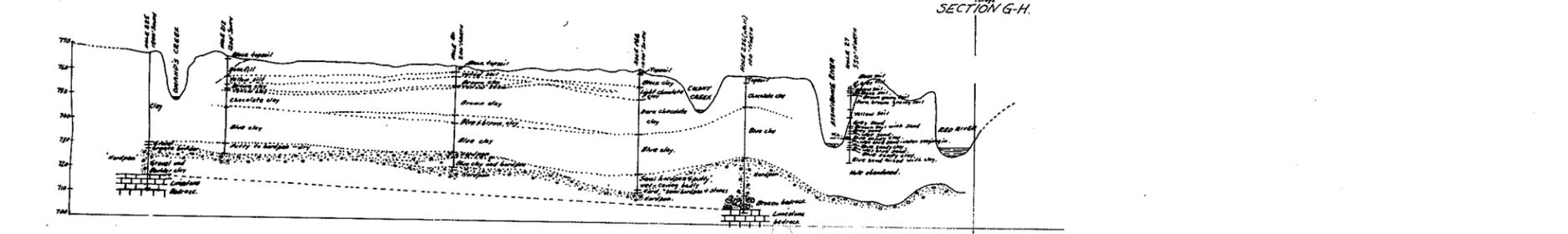
SECTION C-D



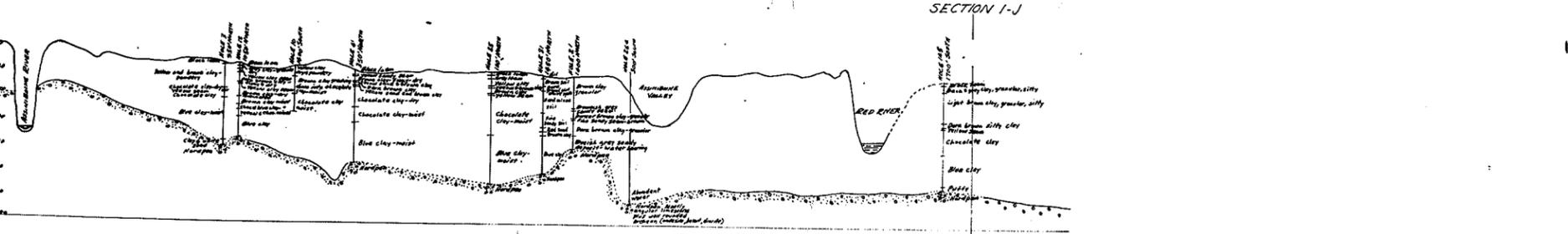
SECTION E-F



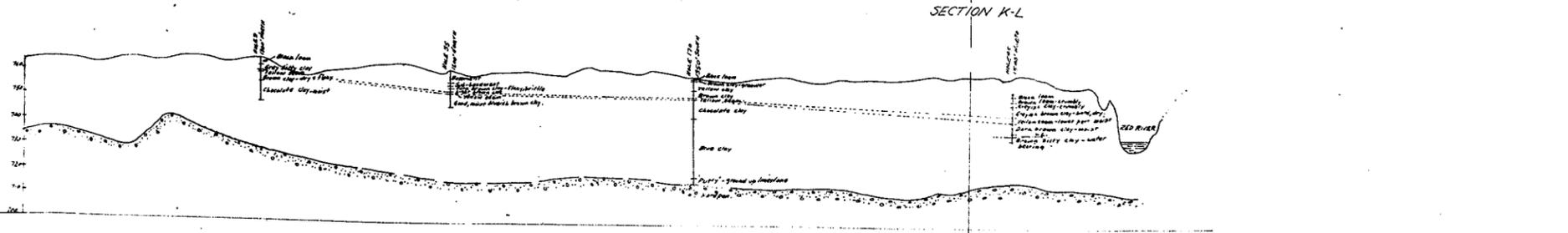
SECTION G-H



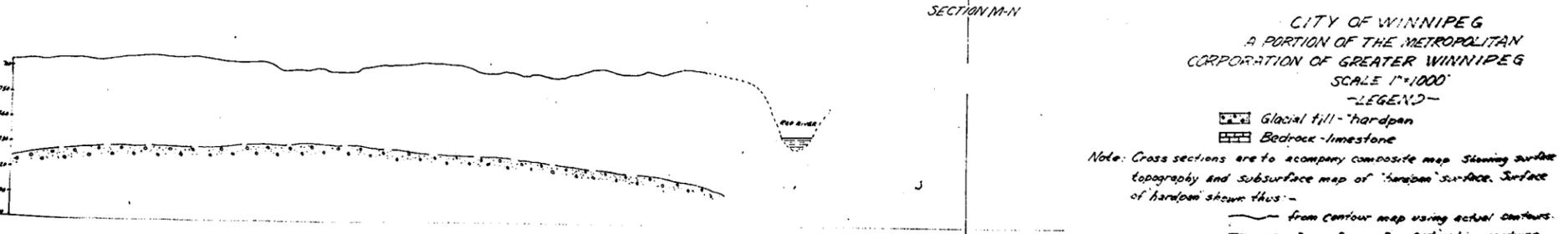
SECTION I-J



SECTION K-L



SECTION M-N



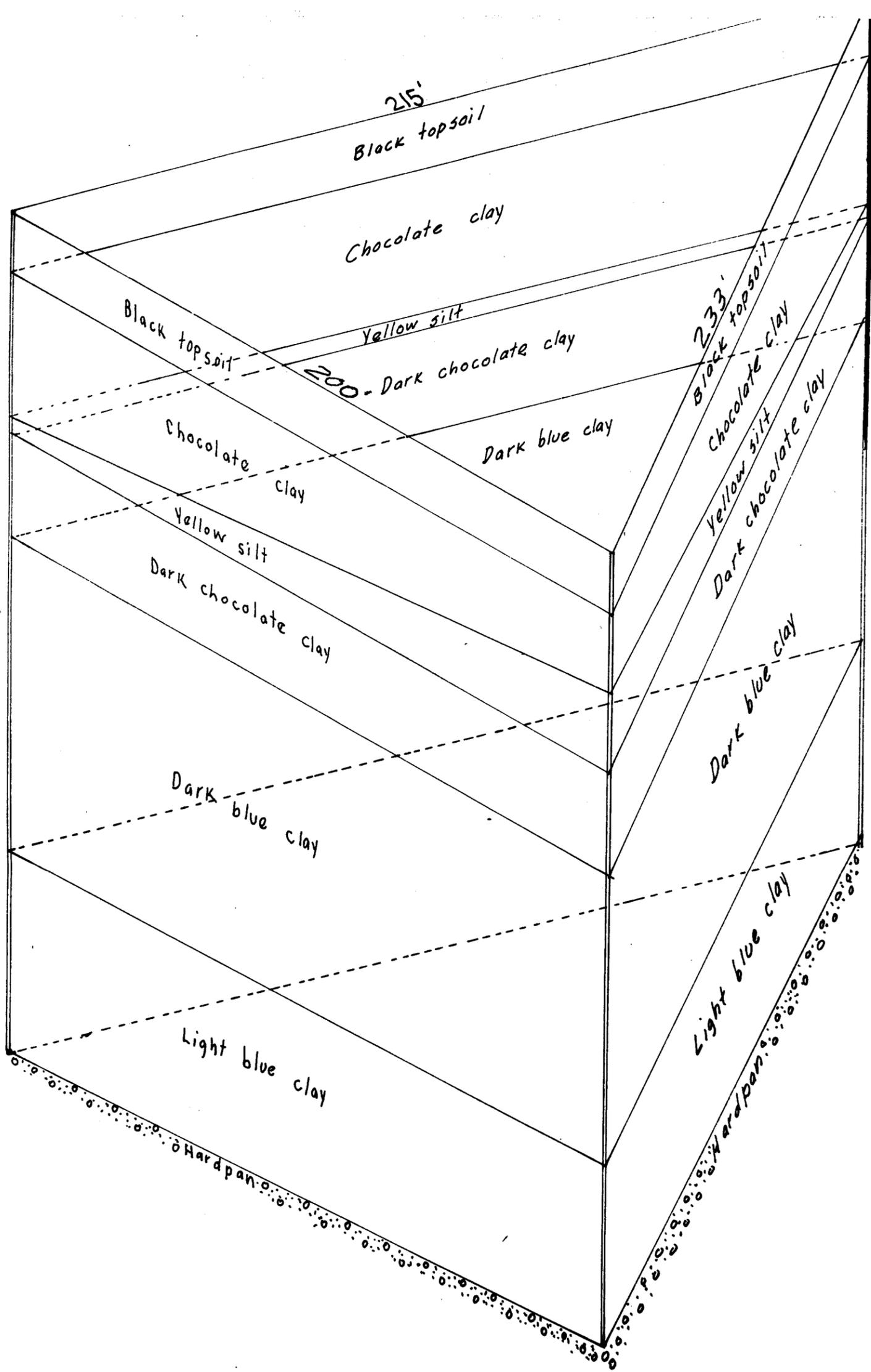
SECTION O-P

CITY OF WINNIPEG
A PORTION OF THE METROPOLITAN
CORPORATION OF GREATER WINNIPEG
SCALE 1"=1000'

LEGEND
Glacial till - hardpan
Bedrock - limestone

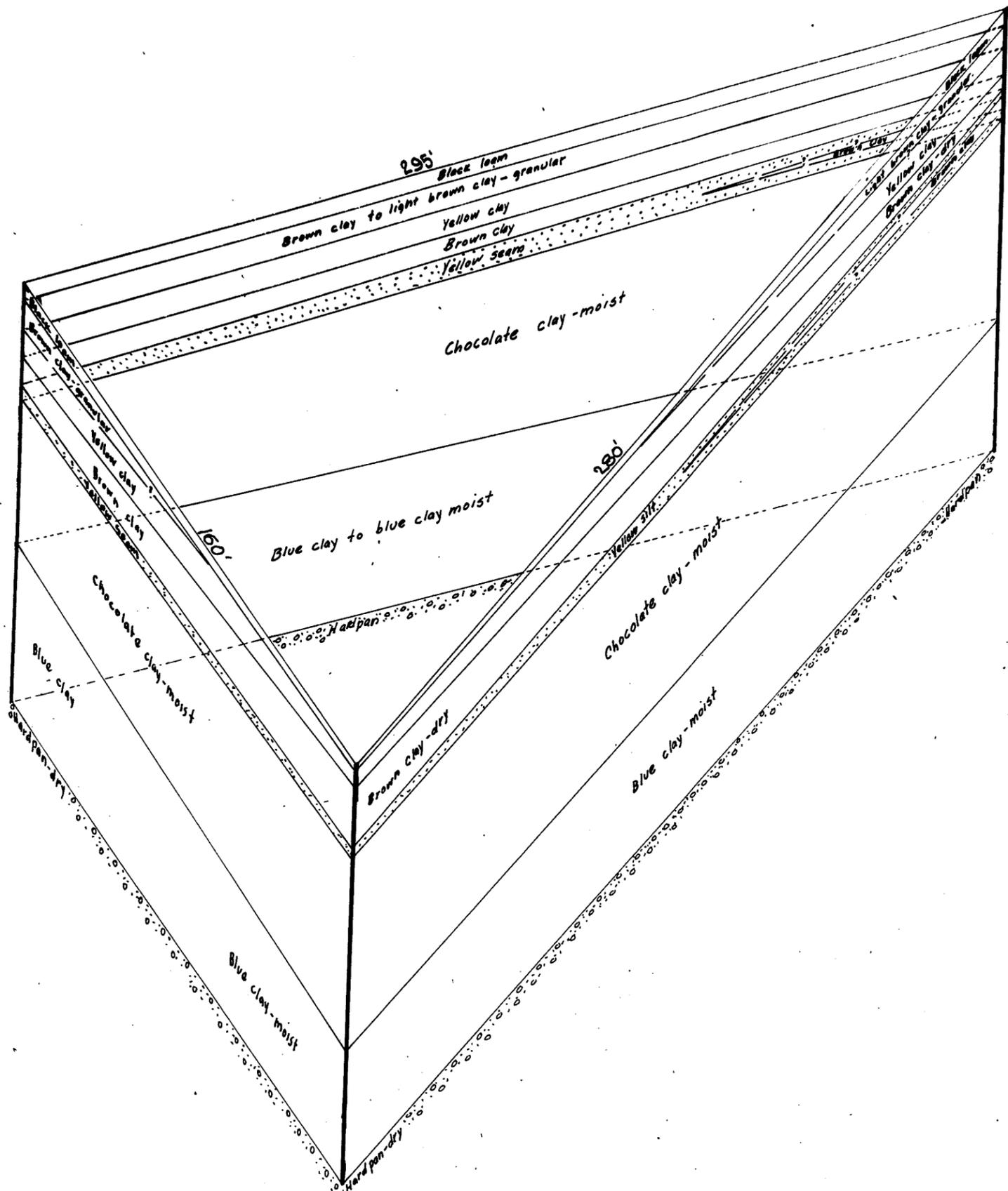
Note: Cross sections are to accompany composite map showing surface topography and subsurface map of hardpan surface. Surface of hardpan shown thus -

— from contour map using actual contours
- - - - - estimating contours
..... from borehole data

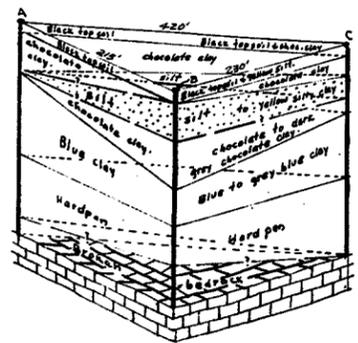
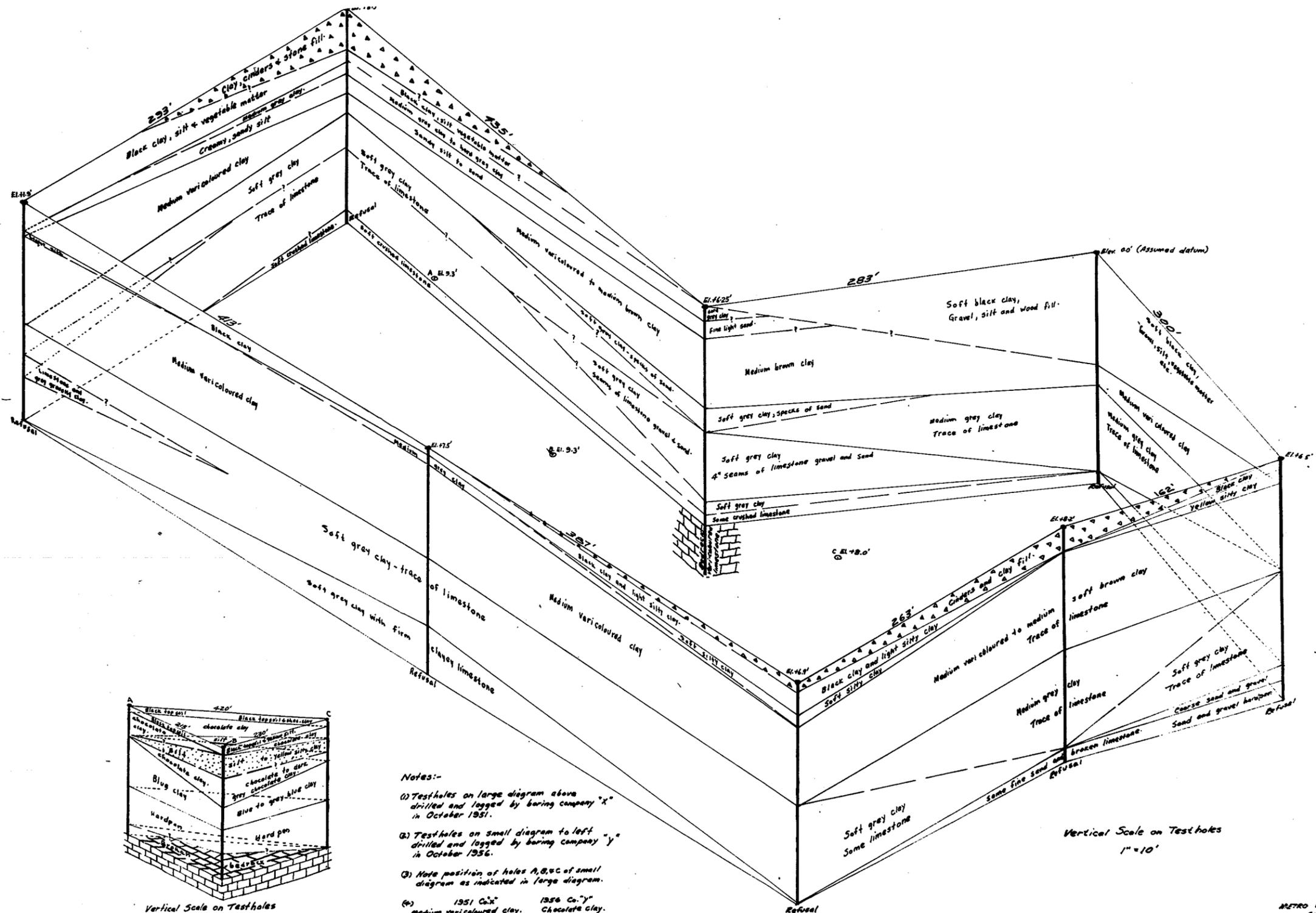


Vertical scale of test holes

1" = 5'



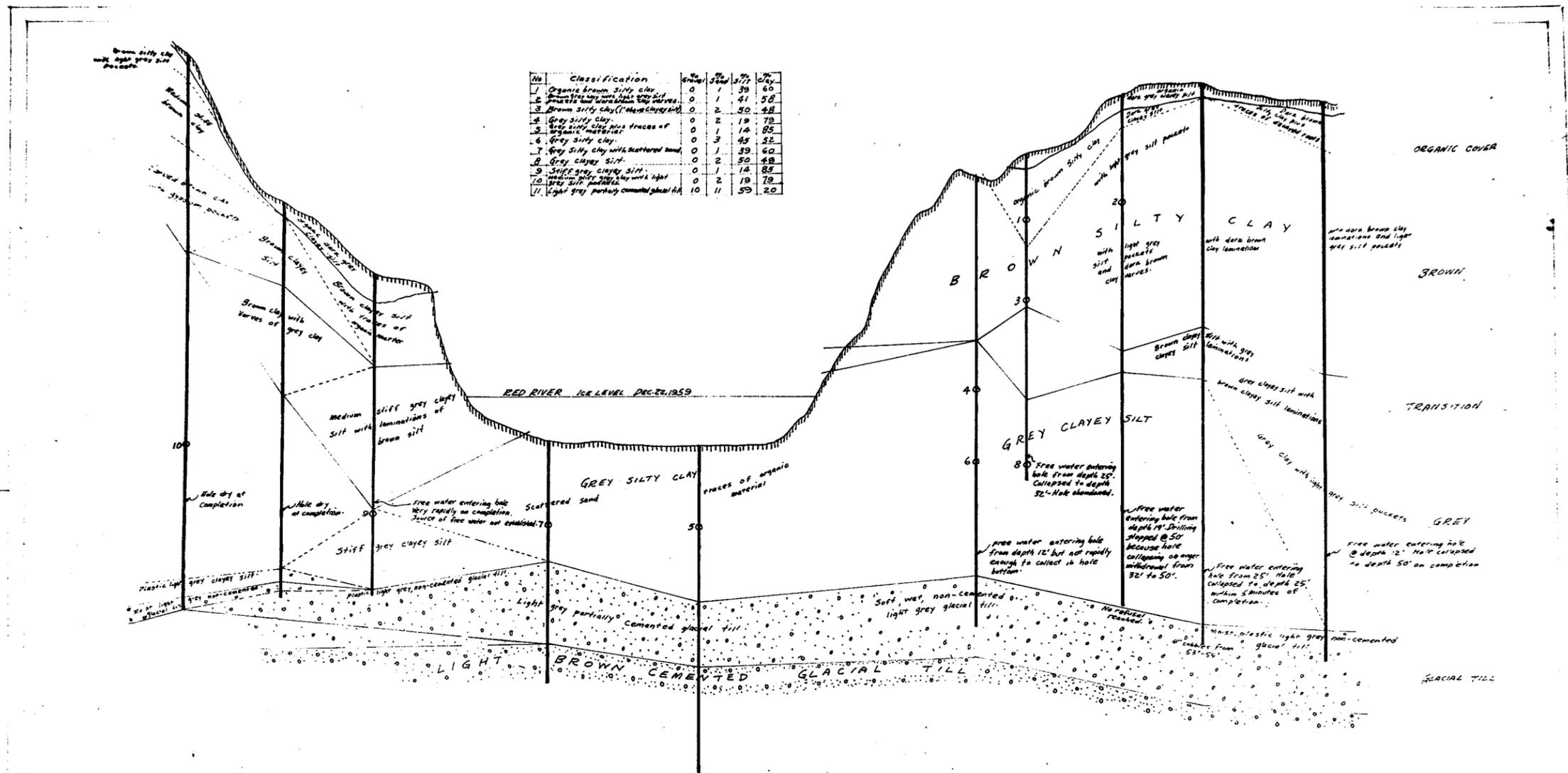
Vertical Scale of Test holes
 1" = 5'



- Notes:-
- (1) Testholes on large diagram above drilled and logged by boring company "x" in October 1951.
 - (2) Testholes on small diagram to left drilled and logged by boring company "y" in October 1956.
 - (3) Note position of holes A, B, C of small diagram as indicated in large diagram.
 - (4) 1951 Co. "x" Medium varicoloured clay. Medium to soft grey clay.
 - 1956 Co. "y" Chocolate clay. Blue to grey blue clay.

Vertical Scale on Testholes
1" = 10'

Vertical Scale on Testholes
1" = 20'



No	Classification	% Sand	% Silt	% Clay
1	Organic brown silty clay	0	1	39
2	Organic brown silty clay with traces of organic material	0	1	41
3	Brown silty clay (f. med. clayey)	0	2	50
4	Grey silty clay	0	2	19
5	Grey silty clay plus traces of organic material	0	1	14
6	Grey silty clay	0	3	45
7	Grey silty clay with scattered sand	0	1	39
8	Grey clayey silt	0	2	50
9	Stiff grey clayey silt	0	1	14
10	Medium grey clayey silt with light grey silt nodules	0	2	19
11	Light grey partially cemented glacial till	10	11	59

Geologic Column (Provisional)

- Organic cover.
- River worked alluvium.
- Prairie clays
 - Brown clay and silt
 - Transition zone - brown to grey clayey silt.
 - Grey clay.
- Glacial till.
 - Partially cemented (moist, plastic)
 - Light brown cemented (dry).

Notes:-

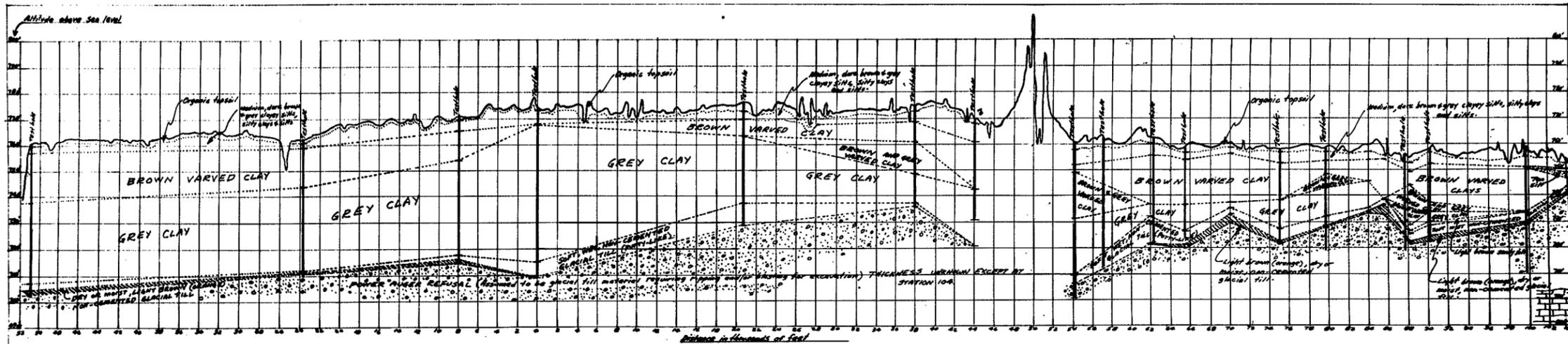
- Data modified from J. Mishak, Water Control And Conservation, Province of Manitoba
- Interpretation between test holes is highly conjectural because:-
 - (a) No dips or strikes of planar features (verves, beds etc) are available.

SCALES

HORIZONTAL 1"=50'

VERTICAL 1"=5'

METRO WINNIPEG
GEOLOGY
SITE STUDY No 5



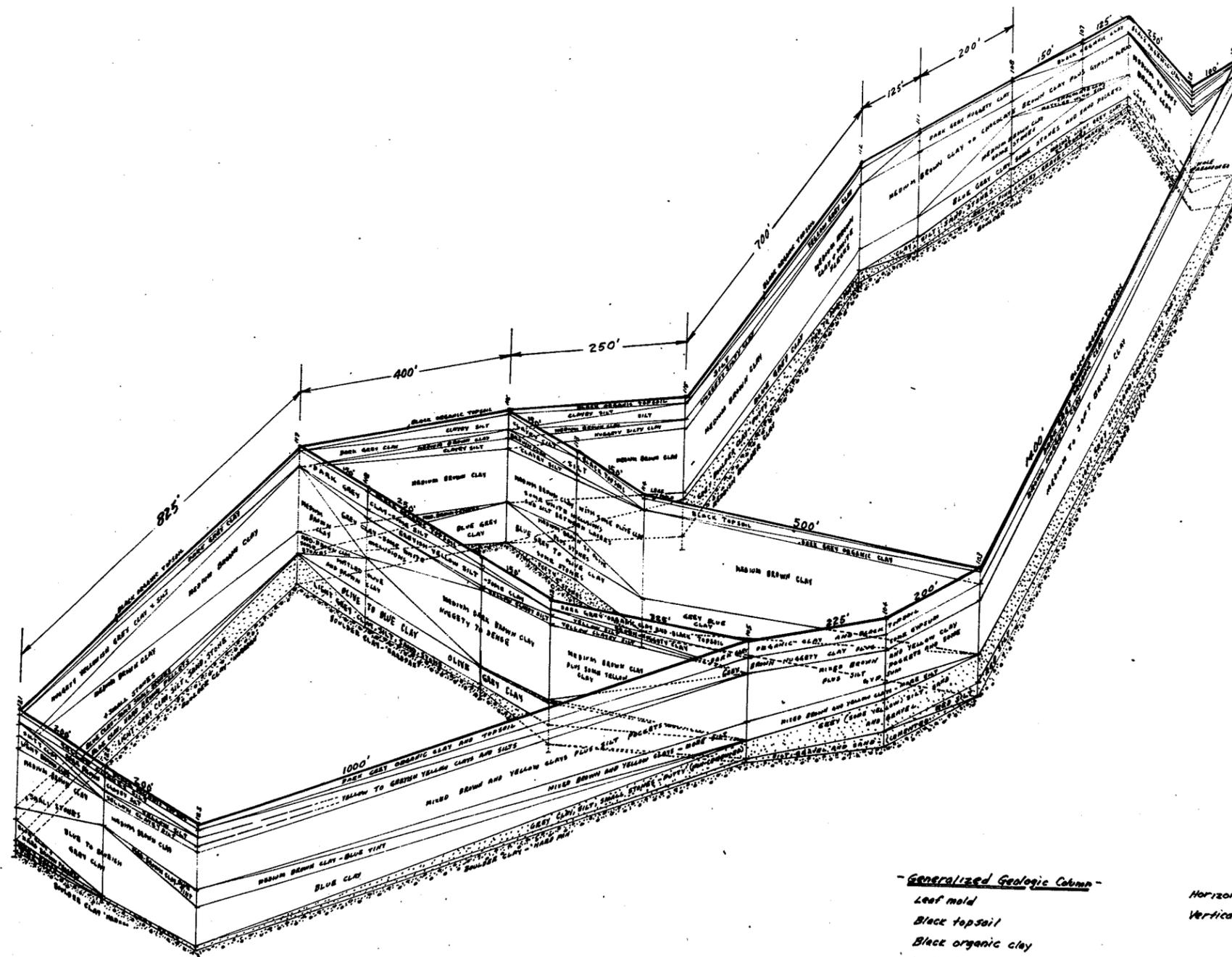
LEGEND

- Organic topsoil
- Medium, dark brown and gray clayey silts, silty clays and silts.
- Tan silt
- Brown varved clay
- Brown and gray varved clay
- Light brown sandy silt
- Brown gravel
- Soft, wet, non-cemented glacial till (Dirtly like)
- Dry or moist, light brown (orange) non-cemented glacial till
- Power auger refusal (Assumed to be glacial till material requiring ripping and/or blasting for excavation)
- Light brown (orange) dry or moist, non-cemented glacial till

Notes:-

- (1) Data modified from J. Nishtak, Water Control and Conservation Division, Province of Manitoba.
- (2) Cross section based on two separate periods of testhole drilling. Breaks in configuration of stratigraphic continuity, where testholes are not indicated, are second test period holes somewhat off section.

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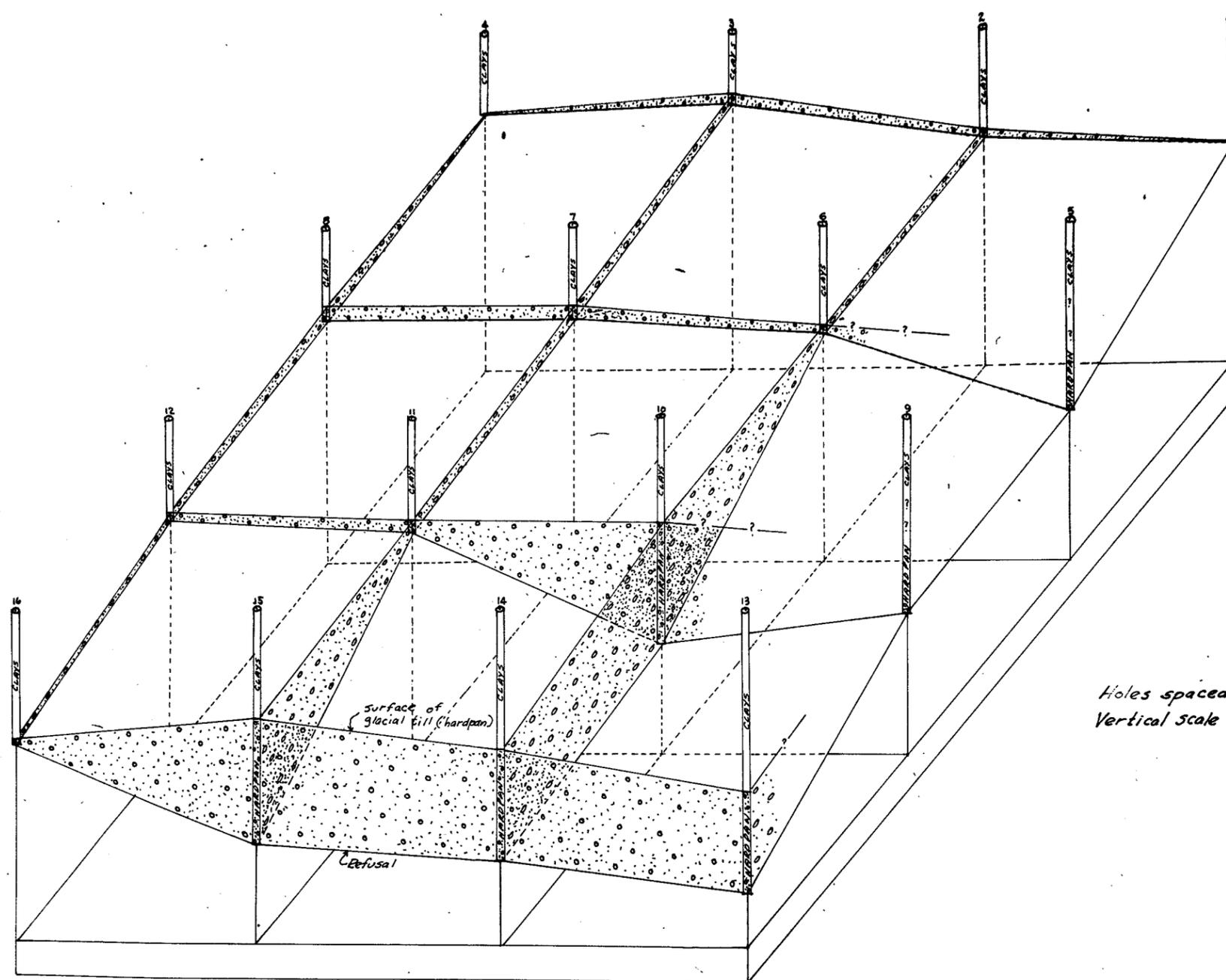


- Generalized Geologic Column -

- Leaf mold
- Black topsoil
- Black organic clay
- Yellow, tan or grey silts and silty clays
- Brown to chocolate brown clay - gypsum floors
- Mixed brown and grey to blueish grey clay
- Blueish grey to olive grey clay
- Glacial till - soft, non-cemented underlain by hard, cemented - grey, orange, red or pink coloured.

Horizontal distance between test holes as indicated
Vertical scale - 1"=10'

METRO WINNIPEG
GEOLOGY
SITE STUDY No. 7



Holes spaced on 150' centres.
Vertical scale - 1"=10'

Site study shows relationship between surface of glacial till (hardpan) and so called REFUSAL.

Some factors affecting position of refusal are:-

1. Kind of boring equipment employed, that is, auger or diamond drill and condition of equipment.
2. Size of auger bit.
3. Size of stones in glacial till.
4. Water content of glacial till.
5. Presence of erratic boulders.
6. Whether piles are to be cast-in-place or driven.
Agitation or vibration of glacial till may release free water making till give up its refusal.

Note:-

- (1) On one site, not the one illustrated here, three auger holes defined refusal at 28'. When piles were driven, shallowest refusal was encountered at 52'.
- (2) Even if refusal is defined at bedrock, known information cannot be extrapolated very far (for purposes of estimation) because of the irregular nature of the bedrock surface and the presence in it of cracks, fissures and probably preglacial drainage channels.

METRO WINNIPEG
GEOLOGY
SITE STUDY No. 8

A P P E N D I X B

S O I L M E C H A N I C S T E R M I N O L O G Y

The following definitions are included for the benefit of those unfamiliar with soil mechanics terminology. For descriptions of the standard procedures for performing these tests one should consult A.S.T.M. (1964).

Liquid limit is the moisture content expressed as a percentage by weight of the oven dried soil at which the soil will just begin to flow when jarred slightly. (Allen, 1942).

Plastic limit is the lowest moisture content expressed as a percentage by weight of the oven dried soil at which the soil can be rolled into threads 1/8 inch in diameter without breaking into pieces. Soils which cannot be rolled into threads at any moisture content are considered non-plastic. (Allen, 1942).

Plasticity index is the difference between the liquid limit and the plastic limit. It is the range of moisture content in which a soil is plastic. When the plastic limit is equal to or greater than the liquid limit, the plastic index is recorded as 0. (Allen, 1942).

Natural water content, expressed in percentage of dry weight, is the ratio of the loss in weight sustained by a sample in drying at 105 C to the weight of the dried sample. (Rominger and Rutledge, 1952, p. 163).

Relative water content (W_r), also called "water-plasticity ratio" or "liquidity index", is defined by the ratio of the difference between natural water content (W_n) and plastic limit (W_p) to the difference between liquid limit (W_l) and plastic limit (hence, to plastic index) (I_p):

$$\begin{aligned} W_r \text{ (in percent)} &= \frac{W_n - W_p}{W_l - W_p} \times 100 \\ &= \frac{W_n - W_p}{I_p} \times 100 \end{aligned}$$

Thus relative water content is a measure of the relation of natural water content to the liquid and plastic limits of the soil. (Rominger and Rutledge, 1952, p. 167).

Preconsolidation pressure is the pressure exerted on unconsolidated sediment by overlying material (sediments, water, ice, etc. F.J.W.) that resulted in compaction; the overburden may have been removed later by erosion (drainage, or melting and drainage F.J.W.). A.G.I. (1962)

A P P E N D I X C

S O I L P R O F I L E T E R M I N O L O G Y

The following definitions are included for the benefit of those unfamiliar with soil science terminology.

Soil profile. Succession of zones or horizons beginning at the surface that have been altered by normal soil-forming processes of which leaching and oxidation have been particularly important. (A.G.I., 1962)

A-horizon. Zone of eluviation. The uppermost zone in the soil profile, from which soluble salts and colloids have been leached, and in which organic matter has accumulated. (A.G.I., 1962).

B-horizon. Illuvial horizon. The lower soil zone which is enriched by the deposition or precipitation of material from the overlying zone or A-horizon. (A.G.I., 1962).

C-horizon. A layer of unconsolidated material, relatively little affected by the influence of organisms and presumed to be similar in chemical, physical and mineralogical composition to the material from which at least a portion of the overlying solum has developed. Any slight

alteration of the upper part of the C, such as reduction of calcium carbonate content in glacial till, unaccompanied by other changes, is designated as C₁. (A.G.I., 1962).

Solum. The upper part of the soil profile, above the parent material, in which the processes of soil formation are taking place. In mature soils this includes the A and B horizons. (A.G.I., 1962).

A P P E N D I X D

S A M P L E P R E P A R A T I O N

INITIAL SAMPLE PREPARATION

The shelby tube samples from the Yale Avenue test hole were split longitudinally just over half way through, with a fine wire, and then pulled apart. This produced a partially cut and partially broken face allowing a more thorough examination. One half of the sample was sealed in saran wrap and kept moist, while the second sample was allowed to air dry.

Both samples were used for observations. Slices six inches long and an eighth of an inch thick were peeled off the face of the moist sample at one foot intervals. This yielded 10 to 20 gms. air dried material. Contacts and other zones of special interest were sampled in greater detail. The slices were air dried and ground in a porcelain mortar and pestle to pass a 60 mesh screen. Natural 60 mesh material was not present in most samples but when it was encountered, in the lower blue-gray clay and till beds the air dried material was gently rubbed between two pieces of wood, instead of being ground in the mortar and pestle, so as not to crush actual 60 mesh material. The 60 mesh material was examined and discarded.

The random auger samples recovered from the Geology Building test hole were sampled by removing a representative sample large enough to produce approximately 10 gms. of air dried material from each of the auger cuttings.

DIFFERENTIAL THERMAL ANALYSIS PREPARATION

The -60 mesh material was split, with a micro-splitter, down to two 0.5 to 0.25 gm. samples. Both samples were kept at 45, $\pm 5\%$ relative humidity for at least 24 hours prior to analysis.

To load the sample holder the material was poured into the sample cavity and packed, initially by tapping the holder with a finger and finally with the static load of an 8-inch long 1/4 inch diameter glass rod. The holder was filled to the same point on each loading. On occasions material had to be added or discarded to keep the sample size constant. Testing revealed that sample weights are reproducible within 0.02 gms. if this method is followed carefully. This eliminated the necessity of using weighed samples for analysis.

X-RAY DIFFRACTION PREPARATION

For the X-ray powder photographs a very small amount of the -60 mesh material was finely ground in an agate mortar, mixed with collodion and rolled into a very fine thread approximately 0.06 mm. in diameter. The rolling was allowed to harden for a few hours and was then mounted and centered in a large, 114.63 mm., Philips Camera.

X-RAY FLUORESCENCE SPECTROMETER PREPARATION

A small 1 to 2 gm. portion of the -60 mesh material was ground to -200 mesh for X-ray fluorescence analysis. Pellets for analysis were made by mixing 0.50 gms. of sample, 0.50 gms. lanthanum oxide and 1.0 gms. lithium tetraborate, and fusing the mixture in a graphite crucible, at 1800°F for 20 minutes. The resulting bead was weighed and made up to 2.1 gms. with boric acid. The mixture was re-ground to approximately -300 mesh, spread evenly in the bottom of a pellet die, carefully topped with boric acid (to provide backing for the pellet) and subjected to thirty tons pressure for two minutes. The resulting pellet was used for analysis.

A P P E N D I X E

D I F F E R E N T I A L T H E R M A L A N A L Y S I S

D Y N A M I C G A S P R O C E D U R E

Two clay samples were used for differential thermal analysis. The first was heated to 600°C in dynamic nitrogen and the second was heated to 600°C in dynamic oxygen and continued to 1,000° in dynamic carbon dioxide. This procedure allowed the clay mineral peaks below 600°C to be recorded in dynamic nitrogen without interference from the oxidation of organic matter. Continuation of the run above 600°C in dynamic nitrogen resulted in a severe exothermic drift due to the break-down of the organic matter so 600°C was used as a cut-off for the dynamic nitrogen run. For the second sample oxygen was used to promote oxidation of the organic matter below 500°C so that the thermogram above this temperature could be recorded. The carbon dioxide was used to separate the calcium carbonate and magnesium carbonate peaks of dolomite. (see Chapter IV). The change-over from oxygen to carbon dioxide was initially performed at 600°C. The disturbance caused by this change-over affects a range of 30 to 40°C before equilibrium is re-established, so that peaks occurring between 600 and 640°C were not being recorded. Thus part way through the

program the change-over was lowered to 520°C, in the range already recorded in the dynamic nitrogen run. The range between 600° and 640° was then recorded and minor amounts of montmorillonite were detected in some samples.

Till samples, because of their low clay and low organic content were only run once in dynamic carbon dioxide to 1000°C. A few samples were run in nitrogen to 600°C followed by carbon dioxide to 1000°C but no difference was noted between the two gas procedures.

A recorder scale setting of 20 microvolts full scale deflection was used for the nitrogen runs to 600°C for the clay samples. The oxygen run was performed at 40 microvolts full scale deflection to 600°C and later 520°C, and continued at 40 microvolts with dynamic carbon dioxide portion if the carbonate content was low, or continued at 150 microvolts full scale deflection if the carbonate content was moderate or high. The till samples were run at 20 microvolts to 600°C and at 150 microvolts to 1,000°C.

The scale setting of 20 microvolts produced baseline drift problems in both the clay and particularly the till samples but the high sensitivity was necessary to record the quartz peak and to indicate the presence of minor amounts of clay minerals.

A heating rate of approximately 10°C per minute was used throughout.

A P P E N D I X F

C A L C U L A T I O N S

F O R

C H E M I C A L F O R M U L A E

The calculations used in deriving the various chemical formulae are outlined in Tables F-1 to F-4. The first step, Table F-1, was to subtract the various oxides representing the non-clay minerals (columns 3 and 4) and the P_2O_5 content and ignition loss (column 2). The result was recast to 100% (column 5). This step was not strictly necessary but was carried out so that the results could be compared to the analysis of other researchers. Next the mole proportion (column 6) was determined by dividing the percentage of each oxide (column 5) by its molecular weight. The cation proportion of each oxide (column 7) was arrived at by multiplying the mole proportion of each oxide (column 6) by the number of cations in the oxide. The anion proportion for each oxide (column 8) was determined in a similar manner by multiplying the mole proportion of each oxide by the number of oxygen atoms in the oxide. The total anion charge (bottom left, table F-1) was calculated by totaling the anion proportion for each oxide, subtracting the anion proportion for water, and multiplying by two. This

gave the total anion charge for the cations. To arrive at the total anion charge for the complete formula the cation proportion of hydrogen (representing the hydroxyl charge) was added to the above total. Finally the cation ratio (column 9) of each element per 1/2-formula unit was calculated by multiplying each cation proportion (column 7) by a factor "n". The "n" was calculated (bottom right, Table F-1) by dividing 22 by the total anion charge. These calculations yielded the formulae found in Table 21, p. 163).

In Table F-2 the corrections for non-clay minerals (columns 3, 4, 5 and 6) were made as described in Table F-1. These corrections were the oxides for the carbonates as determined plus an estimated 5% each of quartz and feldspar (half orthoclase, half plagioclase), and an estimated 2% of iron oxides. The balance was recast to 100% (column 7) and the oxides representing an estimated 3% kaolinite (column 8) were subtracted. The mole, cation and anion proportions plus the total charge and the cation ratio (columns 10, 11, 12 and 13) were determined from the balance of oxides (column 9) as outlined for Table F-1.

In Tables F-3 and F-4 the chemical formula for the swelling layer was calculated and based on an ideal muscovite non-swelling layer (Table F-3), and based on an illite (after Schultx, 1964) non-swelling layer (Table F-4). In both cases the oxides representing the non-swelling layer (column 3) were calculated by assuming all the potash to be in the non-swelling

layer. The percentage of illite or muscovite to montmorillonite was calculated at the bottom of each table.

In all cases the determined amount of water in the analysis was greater than the required 2 hydroxyls per 1/2-formula unit. This excess water was assumed to be present as H₂O (Brown and Norrish, 1952) or sorbed water (Mackenzie, 1957b) and was calculated as percentage of sorbed water by the method of Mackenzie (1957b) using the following formula:

$$x = n (O^{2-}) - 12$$

$$\frac{x}{n} \times 18 = \% \text{ excess moisture}$$

where

- x = the amount of water not removed at 110°C.
- n = the factor by which the ionic proportions must be multiplied to give the final formula. (see Table F-1)
- O²⁻ = the total ionic proportion of oxygen.

Table F-1

Calculation of Chemical Formula for Sample Y-13,
with Determined Non-Clay Mineral Corrections

1	2	3	4	5	6	7	8	9
Oxides	Analysis	Quartz (%)	Carbon- ates	Recast to 100%	Mole Prop.	Cation Prop.	Anion Prop.	Cation Ratio
SiO ₂	52.55	2.7		56.57	0.9419	0.9419	1.8838	3.704
Al ₂ O ₃	18.73			21.25	0.2085	0.4170	0.6255	1.640
Fe ₂ O ₃	7.02			7.97	0.0499	0.0998	0.1497	0.393
TiO ₂	0.73			0.83	0.0104	0.0104	0.0208	0.041
MnO	0.15			0.17	0.0024	0.0024	0.0024	0.009
MgO	2.91		1.03	2.13	0.0528	0.0528	0.0528	0.208
CaO	3.90		3.06	0.95	0.0169	0.0169	0.0169	0.067
Na ₂ O	0.50			0.56	0.0090	0.0180	0.0090	0.071
K ₂ O	2.97			3.37	0.0358	0.0716	0.0358	0.282
H ₂ O	5.45			6.18	0.3430	0.6860	(-0.3430)	2.698
CO ₂	3.53		3.53					
P ₂ O ₅	-							
Ig. Loss	1.90							
Total	100.34	2.7	7.62	100.00			2.4537	

Total Charge

Total oxygens less hydroxyls = 2.4537
 Oxygen charge = 4.9074
 Hydroxyl charge = .6860
 Total charge = 5.5934

Factor "n" for Cation Ratio

Total charge per $\frac{1}{2}$ unit cell = 22.00
 Total anion charge = 5.5934

$$n = \frac{22}{5.5934} = 3.9332$$

Table F-2
Calculation of Chemical Formula for Sample Y-13
with Estimated Non-clay Mineral Corrections

1	2	3	4	5	6	7	8	9	10	11	12	13
Oxides	Analysis	Quartz (%)	Feld- spar	Carbon- ates	Free Iron Oxides	Recast to 100%	Kaoli- nite	Balance	Mole Prop.	Cation Prop.	Anion Prop.	Cation Ratio
SiO ₂	52.55	5.0	3.15			56.33	1.38	54.95	0.9149	0.9149	1.8298	3.708
Al ₂ O ₃	18.73		1.10			22.37	1.20	21.17	0.2077	0.4154	0.6231	1.684
Fe ₂ O ₃	7.02				2.00	6.37		6.37	0.0399	0.0798	0.1197	0.323
TiO ₂	0.73					0.93		0.93	0.0116	0.0116	0.0232	0.047
MnO	0.15					0.19		0.19	0.0027	0.0027	0.0027	0.011
MgO	2.91			1.03		2.38		2.38	0.0590	0.0590	0.0590	0.239
CaO	3.90		0.15	3.06		0.88		0.88	0.0157	0.0157	0.0157	0.064
Na ₂ O	0.50		0.20			0.38		0.38	0.0061	0.0122	0.0061	0.049
K ₂ O	2.97		0.40			3.26		3.26	0.0346	0.0692	0.0346	0.281
H ₂ O	5.45					6.91	0.42	6.49	0.3602	0.7204	(-0.3602)	2.920
CO ₂	3.53			3.53								
P ₂ O ₅	-	(-)										
Ig. Loss	1.90	(1.90)										
Total	100.34	5.0	5.00	9.52	2.00	100.00	3.00	97.00			2.3537	

Total Charge

5.4278

Factor "n"

4.0532

Table F-3

Calculation of Chemical Formula for Swelling Layer of Sample Y-13,
Based on an Ideal Muscovite Non-Swelling Layer

1	2	3	4	5	6	7	8
Oxides	Balance Col. 9 of F-2	Muscovite	Balance Montmor- illonite	Mole Prop.	Cation Prop.	Anion Prop.	Cation Ratio
SiO ₂	54.95	12.47	42.48	0.7073	0.7073	1.4146	3.985
Al ₂ O ₃	21.17	10.59	10.58	0.1038	0.2076	0.3114	1.170
Fe ₂ O ₃	6.37	6.37	6.37	0.0399	0.0798	0.1197	0.450
TiO ₂	0.93		0.93	0.0116	0.0116	0.0232	0.065
MnO	0.19		0.19	0.0027	0.0027	0.0027	0.015
MgO	2.38		2.38	0.0590	0.0590	0.0590	0.332
CaO	0.88		0.88	0.0157	0.0157	0.0157	0.089
Na ₂ O	0.38		0.38	0.0061	0.0122	0.0061	0.069
K ₂ O	3.26	3.26	-	-	-		
H ₂ O	6.49	1.25	5.24	0.2909	0.5818	(-0.2909)	3.278
Total	97.00	27.57	69.43			1.6615	

Mica+Montmorillonite = 97.00%

Percent Mica = $27.57 \times \frac{100}{97} = 28.42\%$

Percent Montmorillonite = $69.43 \times \frac{100}{97} = 71.58\%$

Total Charge

3.9048

Factor "n"

5.6341

Table F-4

Calculation of Chemical Formula for Swelling Layer of Sample Y-13,
Based on an Illite Non-Swelling Layer

1	2	3	4	5	6	7	8
Oxides	Balance Col. 9 of F-2	Illite	Balance Montmor- illonite	Mole Prop.	Cation Prop.	Anion Prop.	Cation Ratio
SiO ₂	54.95	24.68	30.27	0.5040	0.5040	1.0080	3.934
Al ₂ O ₃	21.17	12.57	8.60	0.0844	0.1688	0.2532	1.318
Fe ₂ O ₃	6.37	2.33	4.04	0.0253	0.0506	0.0759	0.395
TiO ₂	0.93	0.47	0.46	0.0058	0.0058	0.0116	0.045
MnO	0.19		0.19	0.0027	0.0027	0.0027	0.021
MgO	2.38	0.93	1.45	0.0360	0.0360	0.0360	0.281
CaO	0.88		0.88	0.0157	0.0157	0.0157	0.123
Na ₂ O	0.38		0.38	0.0061	0.0122	0.0061	0.095
K ₂ O	3.26	3.26	-	-	-	-	-
H ₂ O	6.49	2.33	4.16	0.2309	0.4618	-0.2309	3.604
Total	97.00	46.57	50.43			1.1783	

Illite+montmorillonite = 97.00%

Percent Illite = $46.57 \times \frac{100}{97} = 48.01\%$

Percent Montmorillonite = $50.43 \times \frac{100}{97} = 51.99\%$

Total Charge

2.8184

Factor "n"

7.8058

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