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BRAIDED RIVER DEPOSITS AND THEIR RELATIONSHIP TO THE PLEISTOCENE HISTORY OF THE CREDIT VALLEY, ONTARIO

By

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A Thesis

Submitted to the Department of Geology

in Partial Fulfilment of the Requirements

for the Degree

Master of Science

McMaster University

1970

MASTER OF SCIENCE (1970)

McMASTER UNIVERSITY
Hamilton, Ontario.

TITLE: Braided river deposits and their relationship to the Pleistocene history of the Credit Valley, Ontario.

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SUPERVISOR: Professor R.G. Walker

NUMBER OF PAGES: 164

SCOPE AND CONTENTS:

The stratigraphy of the Pleistocene sediments of the Credit Valley of Ontario embraces two large fining-upwards sequences which correspond to the deposition of glacial debris during the retreat of the last two glacial periods in south central Ontario. The fining-upwards sequences have a basal gravel unit, a middle cross-stratified sand unit, and an upper unit containing small coarsening-upward sequences. All three of these sedimentary units are the result of deposition of sediment in a braided fluvial system. The basal gravel unit displays mid-channel gravel bars and side channel deposits. The cross-bedded sand unit exhibits incised bedform deposits such as linguoid bars, dunes and ripples. The upper unit of coarsening-upwards sequences (which in places are interbedded with the cross-stratified sand facies) represents the deposits of bank overflow and consequent reactivation of unused channels on the braided river floodplain.

During the Halton and Wentworth ice advances, till

was deposited on the surrounding plain. With retreat of the ice masses, meltwater and outwash debris built up an alluvial plain in the lower and wider reaches of the Credit Valley near Glen Williams. These alluvial plains or sandurs were built up by deposition from braided streams. Outwash from the Halton Ice built a sandur plain on top of one constructed in Wentworth time. Post glacial drainage has incised these glaciofluvial deposits and leaves them exposed along the banks of the present Credit River.

ACKNOWLEDGEMENTS

The writer wishes to express his gratitude to Dr. Roger Walker, his supervisor, for the encouragement and guidance he has given throughout this project. His suggestions have proved invaluable and have provided insights into the history of sedimentation.

Thanks are also extended to Drs. G.V. Middleton and J.R. Beerbower for their helpful discussions and suggestions. The writer is also indebted to his wife Susan for her able assistence in the field work and also in the drafting. She provided constant encouragement and patience.

Financial support for this study was provided by the Department of University Affairs, Ontario.

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INTRODUCTION

Chapter I

The value of the study of sedimentary structures has been recognized since the early history of geologic investigation (Sorby, 1859). In the present day, the study of primary sedimentary structures assists the geologist to reconstruct the paleogeography in a sedimentary basin by providing a measure of the direction and intensity of the paleocurrents (Potter and Pettijohn, 1966).

Observation and geometrical analysis of sedimentary structures forming in modern environments has improved the state of knowledge about the environmental constraints of these structures (Oomkins and Terwindt, 1960; Allen, 1965; Harms and Fahnestock, 1965; Reineck and Singh, 1967). However with the great complexity and diversity of processes present in most modern environments, the definition of environments solely on the grounds of sedimentary structures has had its limitations.

In an effort to overcome the limitations in the interpretation of primary sedimentary structures, the sedimentologist has turned to the field of hydrodynamics to understand better the processes of their formation and their environments. Experimental studies have resulted in the application of flow regime theory to the analysis of sedimentary structures (Simons and Richardson, 1962; Simons _et al_, 1965). The better understanding of the physics of sedimentation has given the geologist higher resolution in determining deposition environments. Now, sequences of facies defined by sedimentary structures can be analyzed and compared to known, recent, facies sequences and depositional environments whose changing hydrodynamic conditions best suit the investigated sedimentary sequence (Oomkens and Terwindt, 1960; Bouma, 1962; Walker, 1969).

However the geologist, with his incomplete depositional record, faces the limitations of outcrop exposure. Poor outcrops make the analysis of sedimentary structures difficult and prevent much of the micro-detail from entering into the analysis. The extent to which an outcrop lends itself to study largely determines the depth of analysis that a sedimentary deposit may undergo. Such a limitation is considerably reduced in the study of Pleistocene outcrops. The partly consolidated sediments lend themselves to analysis in three dimensions, because the geologist's shovel determines the amount and orientation of the exposure. This ability to work the sediment also permits extensive and selective sampling of the deposit. Therefore the study of Pleistocene sediments has all the advantages of studies of modern sediments, except for direct measurements of hydrodynamic characteristics such as velocity, slope and sediment discharge. The Pleistocene might be considered as a fossil flume.

In the most detailed study to date, Jopling (1966) has attempted a reconstruction of the hydraulic conditions at the time of deposition of a micro-delta in Pleistocene outwash sediments. From the grain size analysis and sedimentary structures, Jopling set limits on parameters such as depth and velocity and extrapolated to calculate other hydraulic parameters. However, reconstruction on such a detailed scale is usually difficult to accomplish with the limited geologic evidence available, and may be restricted to dunes of sand grade sediment. A general reconstruction of a much larger, fluviatile system, with an analysis of the paleoflow, is within the range of the geologic evidence available.

This type of reconstruction (or even a detailed sedimentologic investigation of Pleistocene outwash sediments) has never previously been attempted. Therefore a detailed study of Pleistocene outwash sediments would increase the state of knowledge of fluvial processes during the Pleistocene. Furthermore, sedimentological evidence gathered from the study of a Pleistocene outwash channel could then be used to understand better the pattern of ice movements and the deposition of glacial sediments.

The gross geomorphic features of Pleistocene outwash systems have been described by Fisk (1944), Frodin (1954), Price (1960), Ter Wee (1962) and Andrews (1965). Plumley (1948) undertook a classic study of the Pleistocene terrace gravels of the Black Hills of South Dakota. Limited studies have since been made of the sedimentary structures of Pleistocene outwash sediments (Van Straaten, 1956; Jopling and Walker, 1968). From these studies and from the studies of the outwash of modern glaciers (Hjulstrom, 1952; Krigstrom, 1962; Fahnestock, 1963; Price, 1964) many of the outwash systems appear to essentially be braided river systems. Therefore a study of a Pleistocene braided river system could add to the knowledge of sedimentary processes in braided rivers in general.

For this study an outwash channel with a single source and exit was deemed best, so that the glacial history would not be complex. The lower Credit Valley, extending from Credit Forks south to Norval, was chosen for this reason. The drainage entered the upper Credit Valley from atop the Escarpment at Credit Forks. The exit for the drainage from the Credit Valley is at Norval, where the present drainage stops flowing parallel to the Escarpment and instead flows southeast over the Peel Plain. The sedimentary deposits in this "closed system" channel were mapped, and the sedimentary facies delineated.

PLEISTOCENE HISTORY

Chapter II

Ontario was subjected to four glaciations in the Pleistocene epoch, the Nebraskan, Kansan, Illinoian, and the Wisconsin. With each succeeding glacial advance, the previous glacial deposits were to a large extent eroded and reworked by the ice sheet. Consequently, only deposits of the Illinoian and the Wisconsin glaciations are found today in Southern Ontario. The only outcrop of Illinoian till is found in the deep, glacially filled, bedrock valleys at Toronto (Karrow, 1967). The deposits that are commonly exposed at the surface are the Wisconsin, especially the late Wisconsin, deposits. Table I shows the stratigraphy of some of the till sheets in Southern Ontario. As can be seen, the maximum Wisconsin glaciation occurred in late Wisconsin time, during the four substages - the Tazewell, Cary, Mankato/Port Huron, and the Valders.

Taylor (1913, p. 61) was one of the first to study the moraines of Southern Ontario and he pointed out the roughly concentric pattern that they possessed. These morainic bands circumscribed the highest points of land in Ontario - Dundalk, Orangeville and London. He interpreted the moraines

Table 1: Stratigraphy of the till sheets and glaciofluvial deposits of south-central Ontario.

* (Varved clays, sands and silts)

			1. PORT TALBOT LAKE ERIE	2. GUELPH	3. HAMILTON GALT	4. BOLTON PEEL PLAIN	5. TORONTO	6. C-14 AGES
WISCONSIN	LATE	VALDERS						10,000 yrs BP
WISCONSIN	LATE	TWO CREEKS INTERSTADE			DEPOSITS OF LAKE IROQUOIS	DEPOSITS OF PEEL PONDING	DEPOSITS OF LAKE IROQUOIS	11,500
WISCONSIN	LATE	MANKATO/ PORT HURON		WENTWORTH TILL	HALTON TILL WENTWORTH TILL	HALTON TILL WENTWORTH TILL	UPPER LEASIDE TILL	
WISCONSIN	LATE	CARY	PORT STANLEY TILL	PORT STANLEY TILL	PORT STANLEY TILL		LOWER LEASIDE TILL	
WISCONSIN	LATE	LAKE ERIE INTERSTADE	DEPOSITS OF LAKE ERIE INTERSTADIAL *					
WISCONSIN	LATE	TAZEWELL	CATFISH CREEK TILL DEPOSITS OF PLUM POINT INTERSTADIAL *	CATFISH CREEK TILL	CATFISH CREEK TILL		THORNCLIFFE MEADOWCLIFFE TILL SEMINARY TILL	24,000 39,000
WISCONSIN	MIDDLE		DEPOSITS OF PORT TALBOT INTERSTADIAL *				FORMATION	48,000
WISCONSIN	EARLY		BRADTVILLE TILL			SUNNYBROOK TILL	SUNNYBROOK TILL SCARBOROUGH FM.	>52,000
SANGAMONIAN							DON FORMATION	
ILLINOIAN						YORK TILL	YORK TILL	

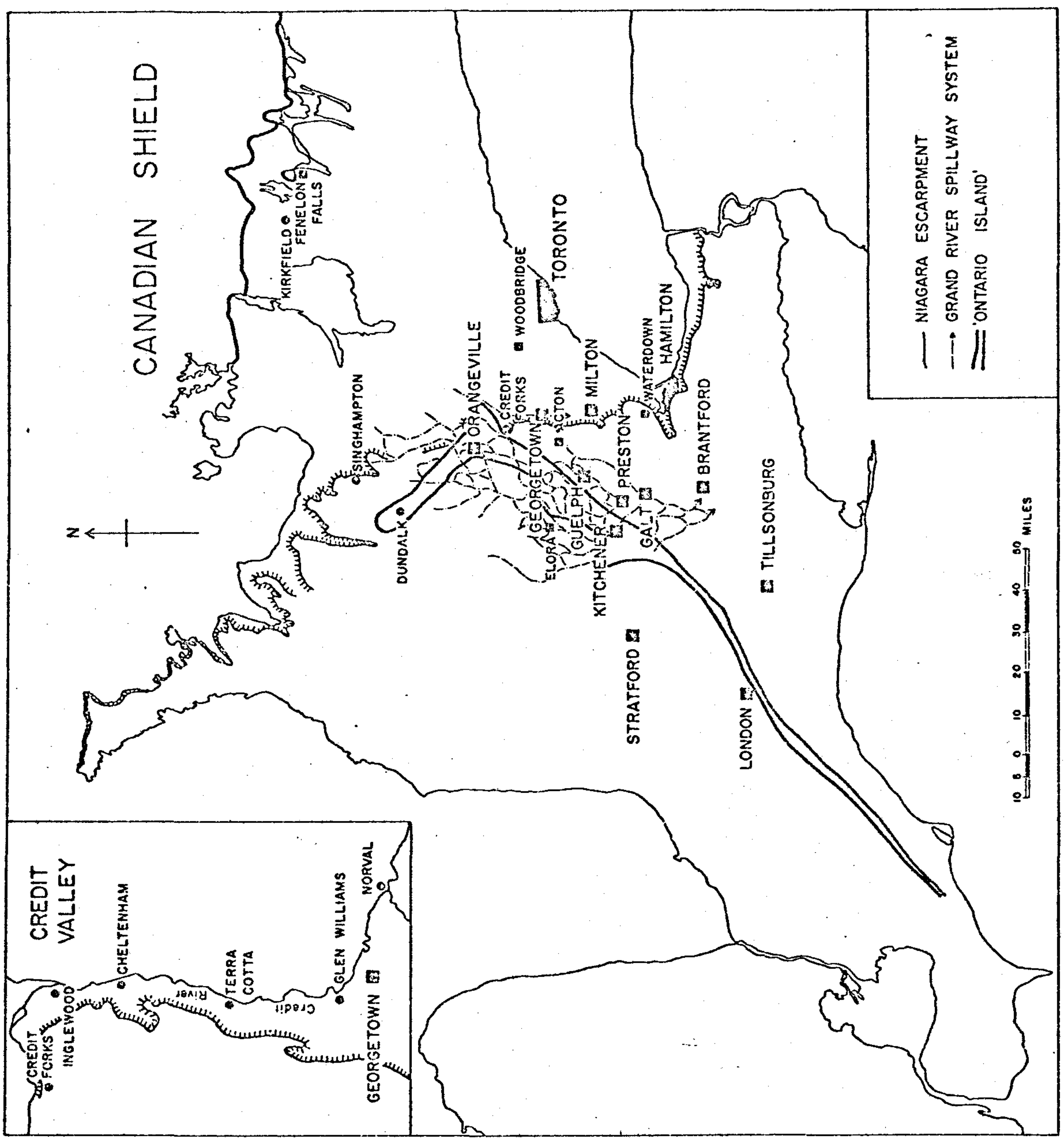


Figure 1: The emergence of the Ontario Island at the Late Wisconsin glacial period. (after Chapman and Putnam, 1966)

to be terminal moraines that were deposited by the ice as it stopped during its retreat away from the high point of land. This ice retreat he referred to as the "uncovering of the Ontario Island". When the ice melted, a strip of land between London in the south and Orangeville in the north, first became exposed (Fig. 1). According to Chapman and Putnam (1966, p. 38), "Drainage flowing into the crease between the lobes brought in sand and gravel and built the Orangeville moraine". With continued retreat, four ice lobes developed; the Huron lobe, the Georgian Bay lobe, the Erie lobe and the Ontario lobe (Fig. 2). The retreat and readvance of these lobes created the concentric shape of the moraine distribution in Southern Ontario.

The area of study, the Credit Valley (Fig. 1), stood directly in the path of movements by the Ontario ice lobe in late Wisconsin time. These movements will now be discussed in detail.

A. Ontario Lobe

An early Wisconsin till, the Sunnybrook Till, has been found in the lower reaches of the Humber River Valley at Woodbridge (White, 1968). This till has been correlated with a till in the vicinity of London (Karrow, 1967), the Bradtville Till (Table 1). It is inferred that at this time, early Wisconsin ice moved out of the Ontario-Erie basins and overrode the Escarpment. By middle Wisconsin time, (50,000 to

35,000 years BP) there is widespread evidence for an interstadial fluvial deposit, which was laid down until the time of the first of the late or main Wisconsin substages, the Tazewell.

This interstadial deposit has been called the Thorncliffe Formation near Toronto (Karrow, 1967) and the Plum Point and Port Talbot Interstade near London (Dreimanis, Terasmae and McKenzie, 1966). The interstadial sediments are composed of varved clays, sand and silt and contain fossil flora that indicate a cold to temporate boreal climate. There were periodic, short lived, ice advances within this interstadial period. During these advances the Seminary and Meadowcliffe Tills were deposited near Toronto.

In the late Tazewell substage, the Catfish Creek Till was deposited by the Ontario-Erie lobe over much of Southern Ontario. The lower Leaside Till at Toronto follows the deposition of the interstadial beds and with the Catfish Creek Till, marks the advance of ice from the Ontario and Erie basins.

Following this ice advance, the Lake Erie basin experienced an interstadial condition, the Lake Erie Interstade. In the following Cary glacial substage, the Port Stanley and the lower Leaside Tills were deposited. All the tills and interstadial deposits so far mentioned have been found in river cuts in deep glacial valleys. None of the tills find a surface expression except where their cover has been eroded off. The next glacial substage, the Mankato/Port Huron, resulted in the formation of the glacial deposits which form the land surfaces seen today in Southern Ontario.

Wentworth Ice Sheet

In the Mankato/Port Huron ice advance, the ice of the Ontario lobe spread westwards into the middle of Southern Ontario. This ice sheet is here referred to as the Wentworth Ice Sheet. During this advance the Paris-Galt-Moffat moraine complex was formed (Fig. 3b). These moraines run as roughly parallel strands from Tillsonberg near Lake Erie to the Credit Forks where they merge into the large Oakridges Moraine. Another series of parallel moraines, the Singhampton-Gibraltar Moraines, were also formed in the early Mankato/Port Huron period by a readvance of the Georgian Bay ice lobe (Straw, 1968, p. 899). These moraines also die out against the Oakridges Moraine. However Chapman and Putnam (1966, p. 41) and Taylor (1913) prefer to place the Gibraltar Moraine construction at a later date, that of the late Mankato readvance.

The drumlins of the Guelph area were formed by the advance of this Wentworth Ice Sheet. The drumlins are found in front of and between the Paris and Galt Moraines (Chapman and Putnam, 1966, p. 42). This is the result of a complex series of retreats and readvances by the ice front. The farthest ice advance deposited the Paris Moraines respectively.

The till which is found deposited in these moraines and drumlins is called the Wentworth Till.

Drainage of water from the ice was funneled along the fronts of the newly created moraines and between drumlin fields above the Escarpment (Chapman and Putnam, 1966). This gave rise to the large gravel outwash channels which occur in the Preston-Galt-Elora areas. These outwash channels coalesced near Brantford to drain into glacial Lake Whittlesey (Fig. 2a). Karrow (1963, p. 56) believes that when the Wentworth Ice Sheet retreated from the Paris-Galt-Moffat Moraine complex, it retreated as far east as Toronto. Assuming a constant rate of melting along the ice front, the ice would first retreat back over the Escarpment in the upper reaches of the Credit Valley near Credit Forks. With further melting, the ice would retreat down the valley.

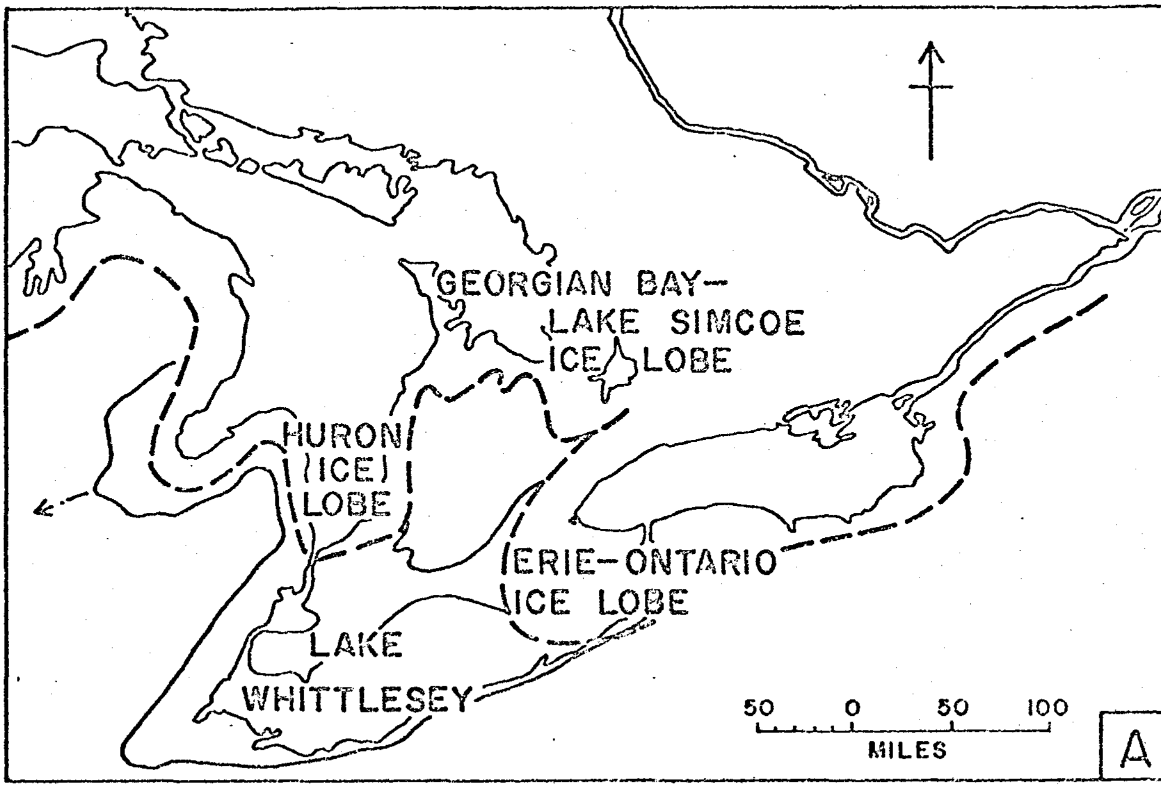
Once the upper end of the Credit Valley became ice free, meltwater that had previously flowed along the top of the Escarpment, could flow down over the Escarpment. Water draining south in the spillway from Mono Mills, past the Credit Forks and down the west side of the Paris Moraine to the Grand River spillway system (Fig. 1) would have been diverted into the Credit Valley. Drainage from the Singhampton and Gibraltar Moraines via the ancestral upper Credit River, near Orangeville, would also be funneled into the Credit Valley rather than the Grand River spillway system. Indeed subsurface

Figure 2: The ice lobes of the Late Wisconsin glaciation.

(a) During the life of Lake Whittlesey.

(b) During the life of Lake Warren.

(after Hewitt and Karrow, 1963)


GEORGIAN BAY-
LAKE SIMCOE
ICE LOBE
HURON
(ICE)
LOBE
ERIE-ONTARIO
ICE LOBE
LAKE
WHITTLESEY
50 0 50 100
MILES
A

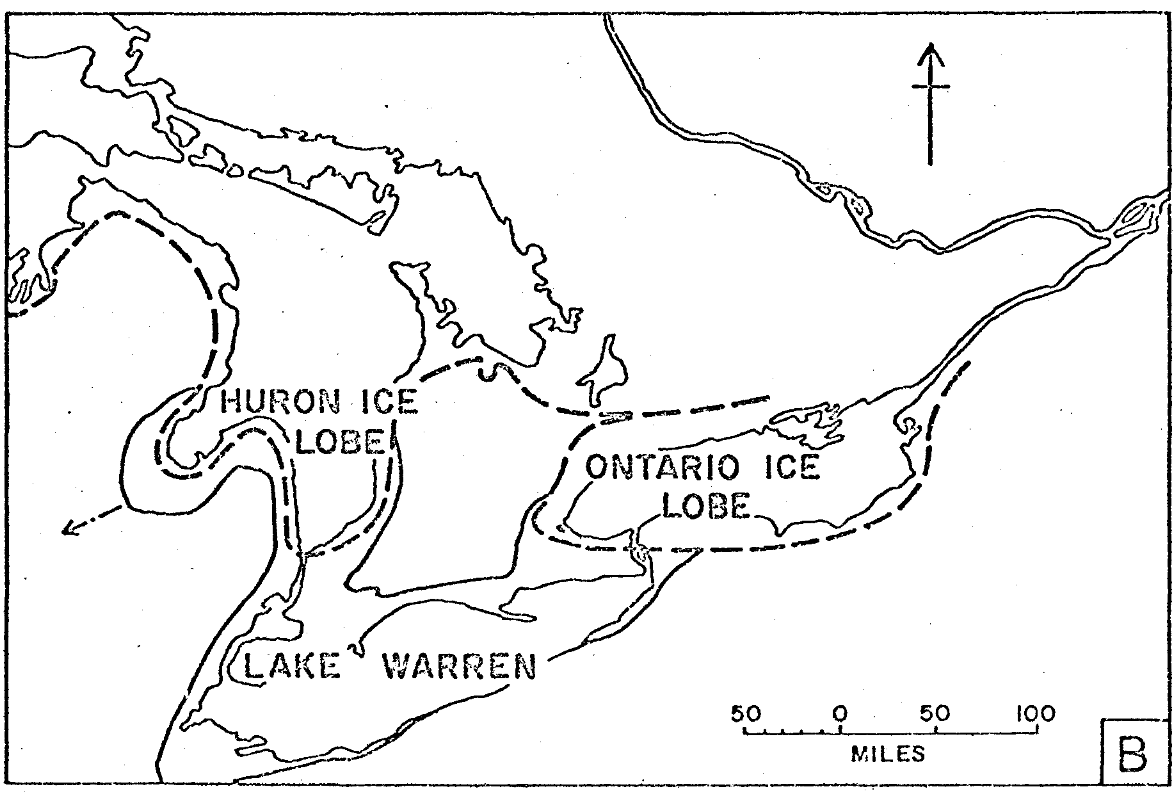

HURON ICE
LOBE
ONTARIO ICE
LOBE
LAKE WARREN
50 0 50 100
MILES
B

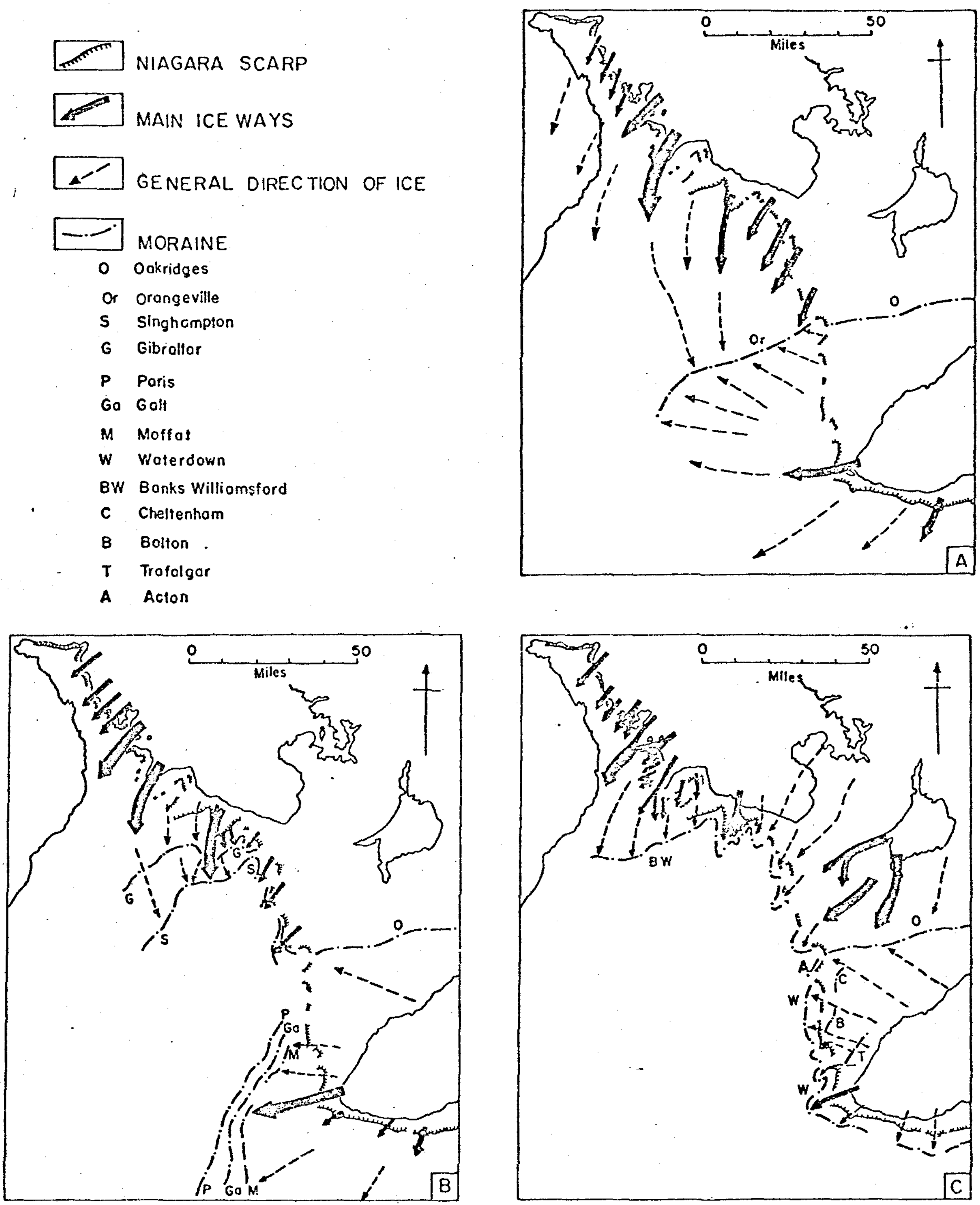
Figure 3: Ice movements and moraines in the vicinity of the Niagara Escarpment of southern Ontario.

(a) During the maximum of the Late Wisconsin Glaciation.

(b) At the Paris-Galt and Singhampton-Gibraltar stage (early Port Huron).

(c) At the Waterdown and Banks-Williamsford stage (late Port Huron).

(modified after Straw, 1968)


NIAGARA SCARP
MAIN ICE WAYS
GENERAL DIRECTION OF ICE
MORAINE
O Oakridges
Or Orangeville
S Singhampton
G Gibraltar
P Paris
Ga Galt
M Moffat
W Waterdown
BW Banks Williamsford
C Cheltenham
B Bolton
T Trafalgar
A Acton
0 50
Miles
A
B
C

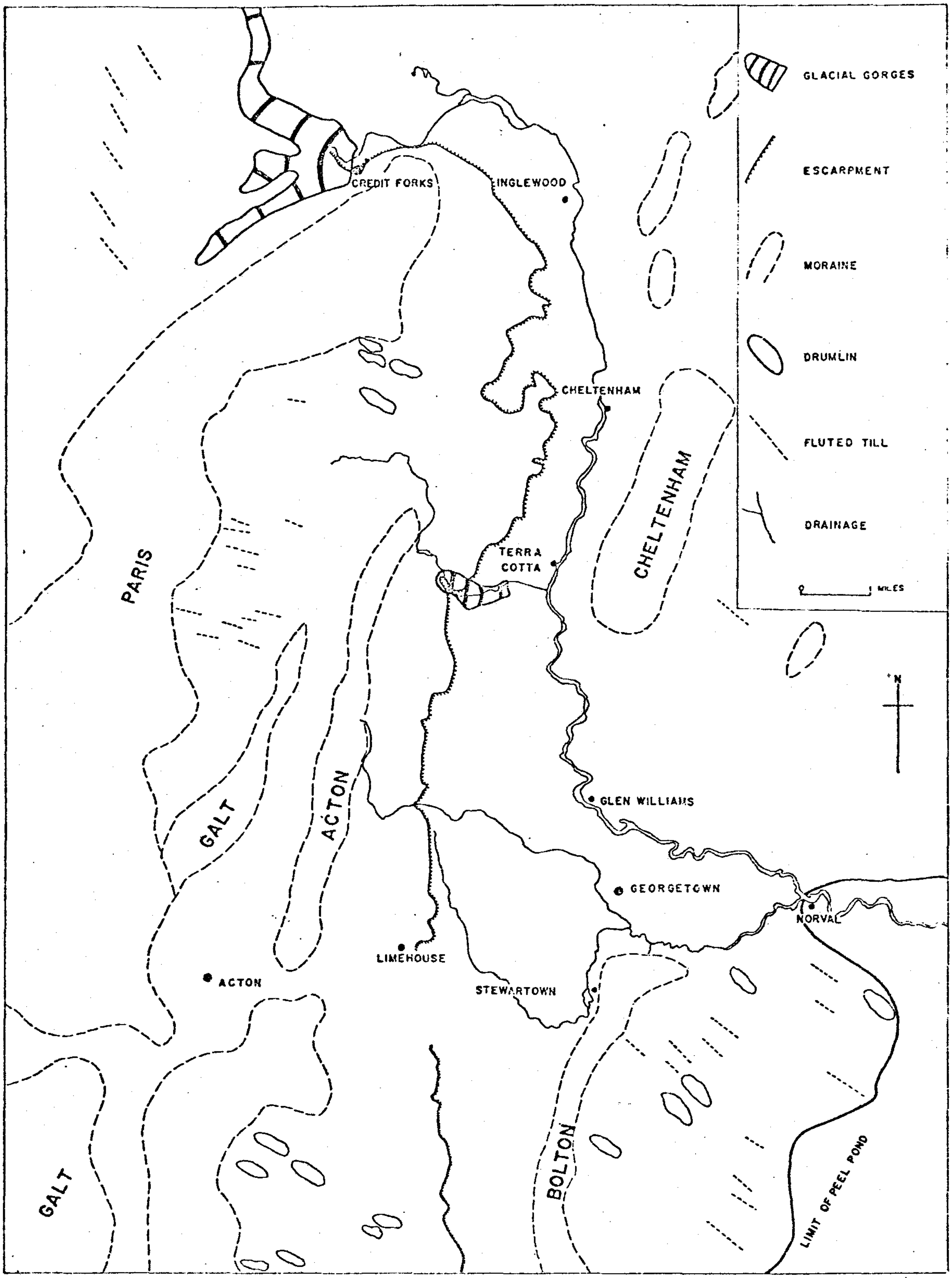


Figure 4: The glacial geomorphology of the Credit River Valley

contours indicate the presence of deep bedrock valleys running from these drainage areas to the Credit Forks. If these bedrock valleys can be taken as indication of earlier flow paths, then it appears that the glacial drainage was similar to the present drainage pattern.

Lateral drainage over the top of the Escarpment supplied more outwash material to the Credit River. At Acton, drainage over the Escarpment must have captured some of the drainage formerly running south via the Blue Springs Creek Spillway to Brantford. Again, subsurface contours indicate the presence of a broad bedrock valley running from Acton east in the direction of Georgetown. Thus with movement of the ice front down the valley in the direction of Norval, the Credit Valley must have become a large catchment area into which glacial debris was washed.

Prior to this time all glacial drainage had flowed south into glacial Lake Whittlesey. However with ice retreat, renewed land emergence occurred in the Ontario Island. A new glacial lake, Lake Warren, was formed (Fig. 2b). This was due to the opening of a new drainage outlet at the Grand River Valley of Michigan (Chapman and Putnam, 1966), which lowered the lake level by sixty feet. The Grand River spillway system still continued to drain along the Escarpment past Brantford and into this lake.

The evidence of both accumulations of outwash and bedrock valley configurations demonstrates that the Credit Valley outwash system did not flow along the edge of the Escarpment, but very closely followed the present drainage patterns. Thus meltwater would be directed southeasterly towards Norval (Fig. 4). If the ice did withdraw to the region of Toronto, as Karrow suggests, this would then permit drainage down the Credit Valley past Georgetown to the ice front near the present Lake Ontario. Here the drainage would continue to flow along the western edge of the ice front and on into Lake Warren.

Halton Ice Sheet

In the late Port Huron/Mankato substage, an ice sheet again advanced to the edge of the Escarpment from the Lake Ontario basin. This readvance is here called the Halton Ice Sheet. Since it derived much of its debris from lake sediments, it deposited the brown, silty, Halton Till. To the east of the Escarpment the depositional topography displays fluted till plains and drumlins, indicating movement from southeast to northwest. The furthest point of advance of this ice sheet is marked by a series of end moraines around Waterdown and along the Niagara Peninsula (Fig. 3c). Chapman and Putnam (1966) visualized the ice as mounting the Escarpment only in those areas south of Milton.

Straw (1968), in his investigation of the moraines sitting on the Escarpment, found that he could trace the Waterdown Moraine northwards to Milton Heights. He further observed that the Acton Moraine, previously interpreted as part of the Galt Moraine (Chapman and Putnam, 1966; Karrow, 1968; Taylor, 1913), actually cuts across the Galt Moraine. In his early study, Taylor (1913) had named small, smooth, morainic strands parallel but below the Escarpment, as the Bolton and Cheltenham Moraines (Fig. 4). Straw (1968, p. 879) offers the opinion that these moraines, the Bolton and the Cheltenham, are contemporaneous and correlative with the Acton and Waterdown Moraines.

If this interpretation of the moraines is accepted, then the Halton Ice Sheet must have surmounted the Escarpment in a region from the Credit Forks south past Milton, to Hamilton (Fig. 3c). From his mapping of the upper Credit Valley, White (1968) states that the Halton Ice Sheet rode up, over the Escarpment near Inglewood but failed to reach the Paris Moraine. However on his map he depicts the area in front of the Paris Moraine as a drumlinized plain of Wentworth Till and it has more water-lain, kame like material associated with it. The possibility arises, therefore, that the Halton Ice Sheet surmounted the Escarpment and reworked the till lying before the Paris Moraine.

Referring to the northern edge of the Credit Valley, White (1968) states that "the outer limit of the ice which deposited the Halton Till is not marked for any great distance by a moraine but by a series of kames and ice block depressions as well as sections of very hummocky topography". These kame deposits sit in front of the northern strand of the Paris Moraine. Therefore there is a great similarity between the deposits of kame material that comprise the Acton Moraine, which sits before the Paris and Galt Moraines, and the deposits of kame sitting before the northern edge of the Paris Moraine (Fig. 4). Both must mark the position of the ice front at its farthest extent in this readvance. Therefore the view of the Halton Ice Sheet overriding the Escarpment and lapping up against existing end moraines, as presented by Straw (1968), appears justified.

With the ice sitting atop the Escarpment once more, the meltwater was constrained to flow south along the top of the Escarpment. The drainage appears to have flowed south along the fronts of the Acton and Waterdown Moraines (Laing, personal communication). The spillways used in the drainage proceeded south to Lake Warren via the Grand River System.

The Halton Ice Sheet left clear evidence that in its lifetime it experienced pulsating advances and retreats. White (1968) and Laing (personal communication) have both found evidence of ice fluctuations such as interbedding of layers of sand and till up to seven times. The sand indicates fluvial deposition in front of a retreating ice mass. The till, which is identical in all layers, records the readvance of the ice. On a larger scale, the retreats and halts of the ice front are clearly seen in the multiple, recessional moraines near Waterdown. Karrow (1959) delineates four bands of parallel, moraine strands in this area.

From its terminal position atop the Escarpment, in front of the Paris Moraine (Fig. 4), the Halton Ice Sheet receeded back over the Escarpment and into the Credit Valley. Here it halted once more and deposited the Bolton and Cheltenham moraines. As in the Wentworth Ice retreat, the removal of ice from the Credit Forks caused meltwater to drain over the Escarpment and into the Credit Valley.

Laing (personal commumication) has recorded the stages of ice retreat by noting the marginal drainage pattern associated with the ice front. As the ice front retreated, the marginal channels receeded to a lower level in the valley. In the area between the Escarpment and Georgetown, these channels are very well developed and are complex in plan. It suggests that as the ice retreated, a large area of ice free ground opened up between Georgetown, Acton and Glen Williams. Into this drainage area flowed much of the meltwater and glacial sediment.

Chapman and Putnam (1966) and Straw (1968) have proposed that the drainage from the melting ice flowed south towards Milton, between the ice front and the foot of the Escarpment. The plan of the meltwater channels and the distribution of the sediment, indicate that the bulk of the drainage gathered at Georgetown and closely followed the present coarse of the Credit River. Small drainage channels are found below the Escarpment, running to the south. However these channels are small and have not eroded the bedrock to leave traces of an established bedrock valley.

The Bolton moraine sits astride the outwash channel at Stewartown, and in the Pleistocene it blocked the drainage down the Credit Valley (Fig. 4). Eventually the water found an outlet to the north of Georgetown and it cut a new course into the Queenston shale. The present Credit River still follows the new route to the north of Georgetown, and bedrock spurs can be seen along the river between Glen Williams, where it was diverted, and Norval. With the opening of this new outlet, the meltwater flowed to the southeast and the river cut down through its earlier deposits.

The ice receeded to the position of the Trafalgar moraine, and the waters impounded before it are now known as the Peel Pond. The outwash from the Credit Valley flowed into the Peel Pond at Norval (Chapman and Putnam, 1966, p. 43). The sheet sand deposits at an elevation of 700 and 725 feet match the elevations of other similar sand bodies on the Humber River to the east. Chapman and Putnam (1966) refer to these as deltas building into the Peel Pond.

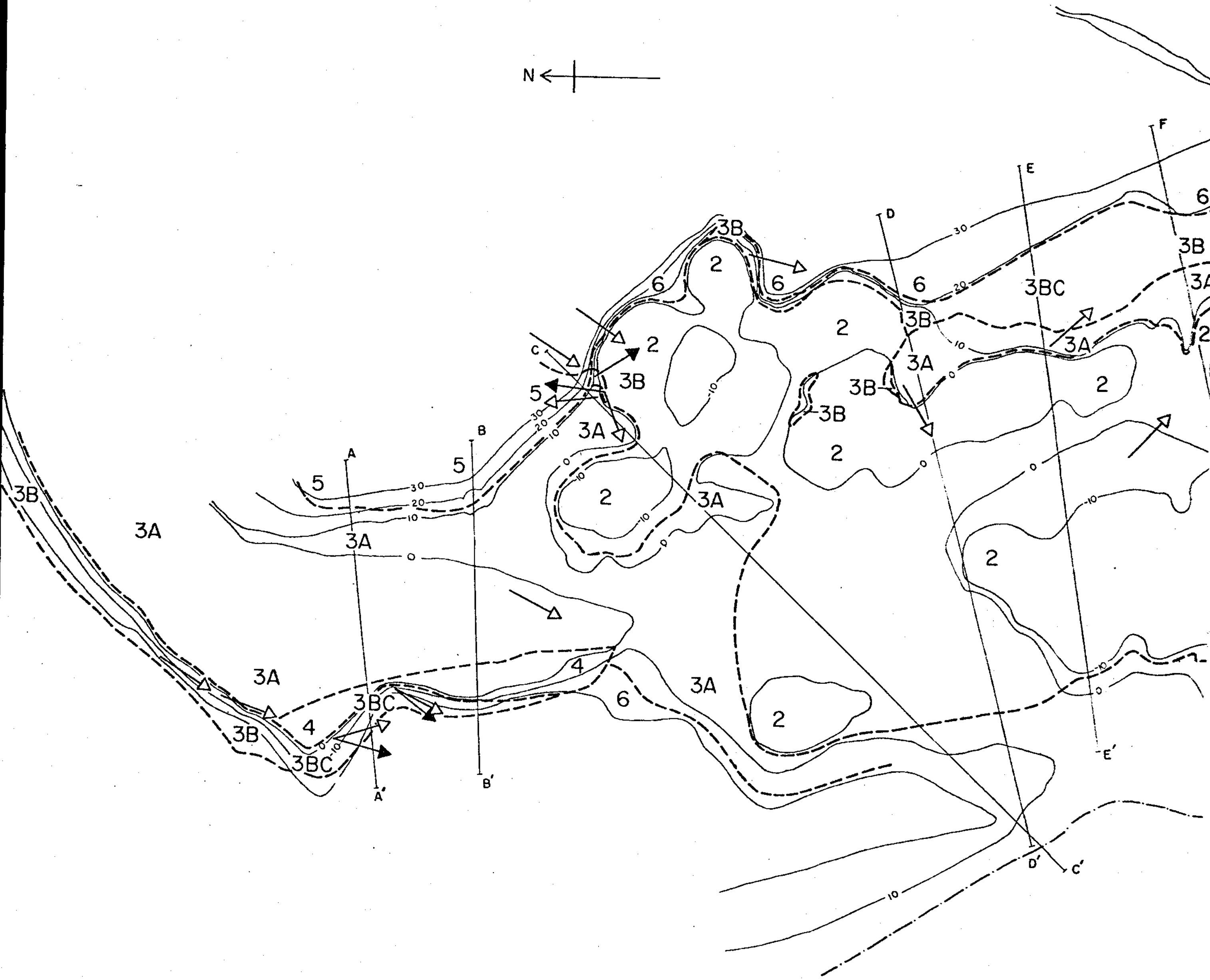
As the ice continued to receed, Southern Ontario became entirely free from ice and Lake Iroquois was formed. The outlet for this glacial lake was at Rome N.Y. This ice recession also opened an outlet for Lake Algonquin at Kirkfield and Fenelon Falls. At a later stage in this recession Lake Algonquin drained down the Ottawa Valley. With retreat of the ice from the south side of the St. Lawrence River, the Champlain Sea developed in Eastern Ontario and the St. Lawrence lowlands. With total ice retreat the Great Lakes, as we now know them, evolved.

STRATIGRAPHY

Chapter III

The area of the Credit Valley that was studied lies between Cheltenham and Norval. To the north of Cheltenham the Credit River runs close along the west side of the valley. From bedrock topography maps (White and Morrison, 1968), it is known that the deep, sediment filled, Pleistocene valley lies several hundred yards to the east of the present river course, directly under the Cheltenham moraine. The Credit River has subsequently cut down into its earlier deposits and has constructed several terraces at different elevations along the river valley. This erosion down into the floor of the valley in late Pleistocene and recent times has exposed extensive deposits of sand and gravel south of Cheltenham.

The largest economic use of the sand and gravel deposits occurs at Glen Williams, where seven gravel pits lie across the width of the valley. It is here, one mile north of Glen Williams, that the largest of these pits is located on the Bishop Farm. This large pit, sitting on a terrace on the west side of the valley, has been worked extensively and is 3500 feet long, 1400 feet wide and 40 to 50 feet deep. Nearly all the lithologies present in the Credit Valley are



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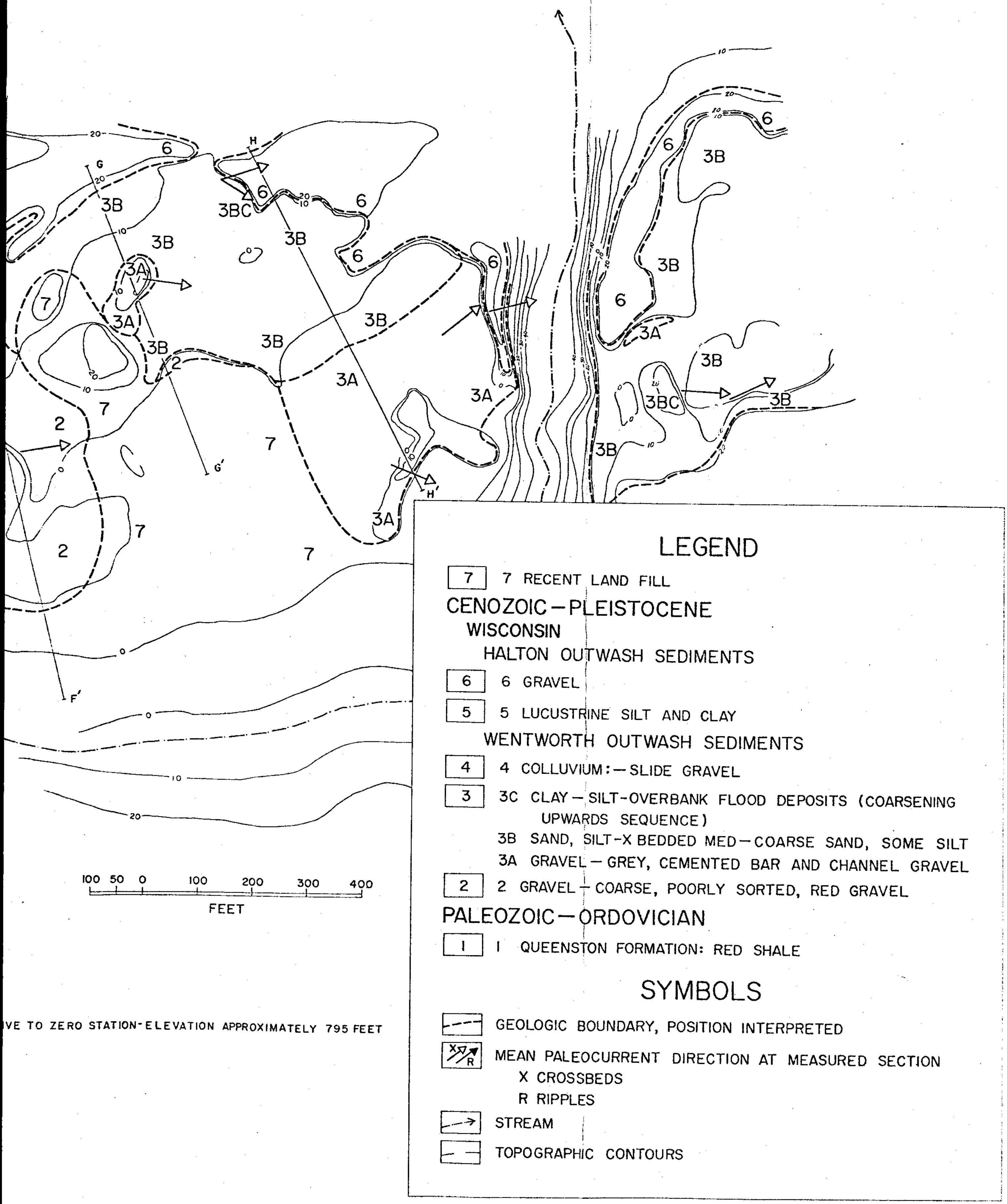

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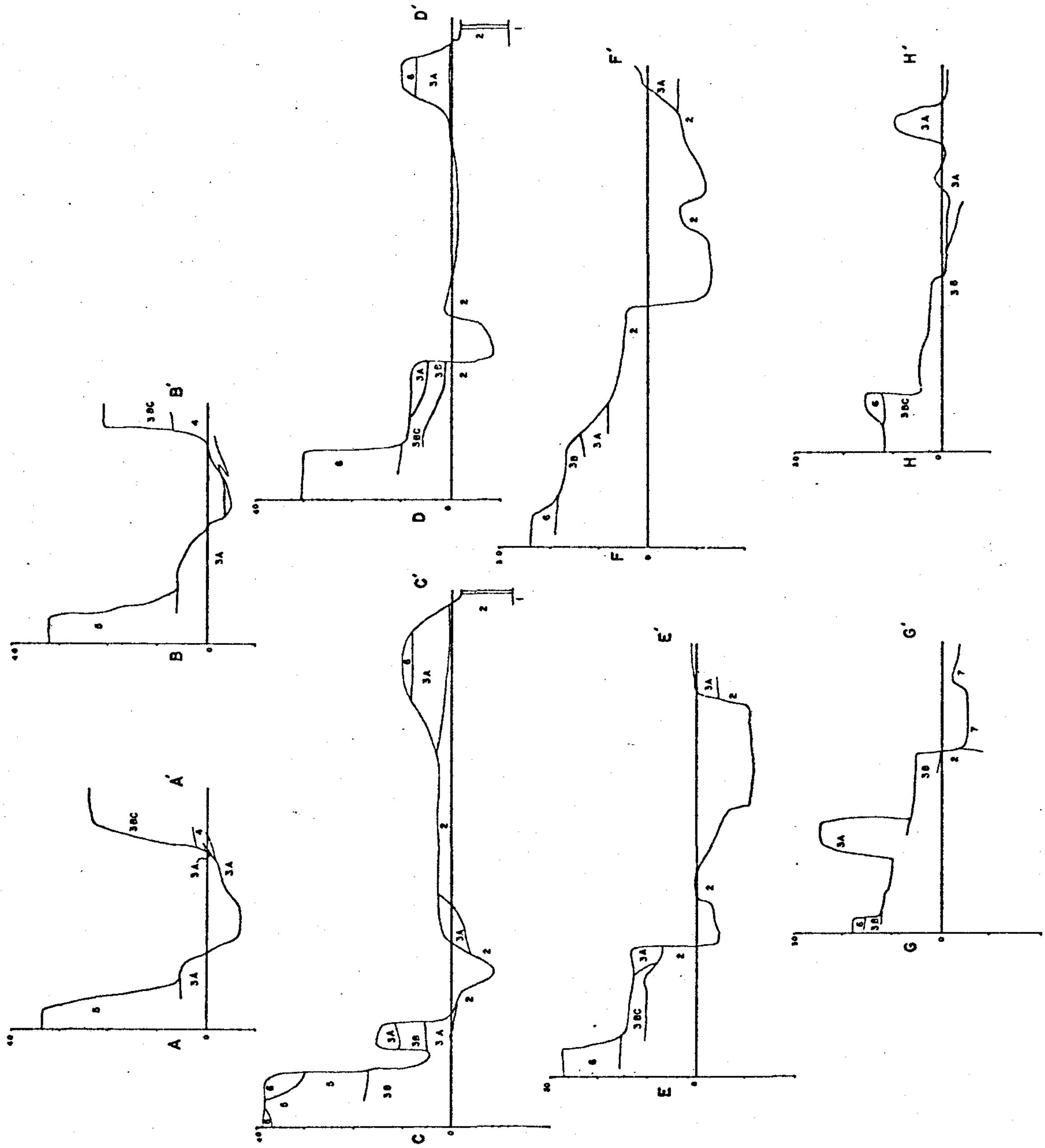

Figure 5b: The stratigraphy of the Bishop Pit, Glen Williams - vertical measured sections. See Fig. 5a for location of sections.

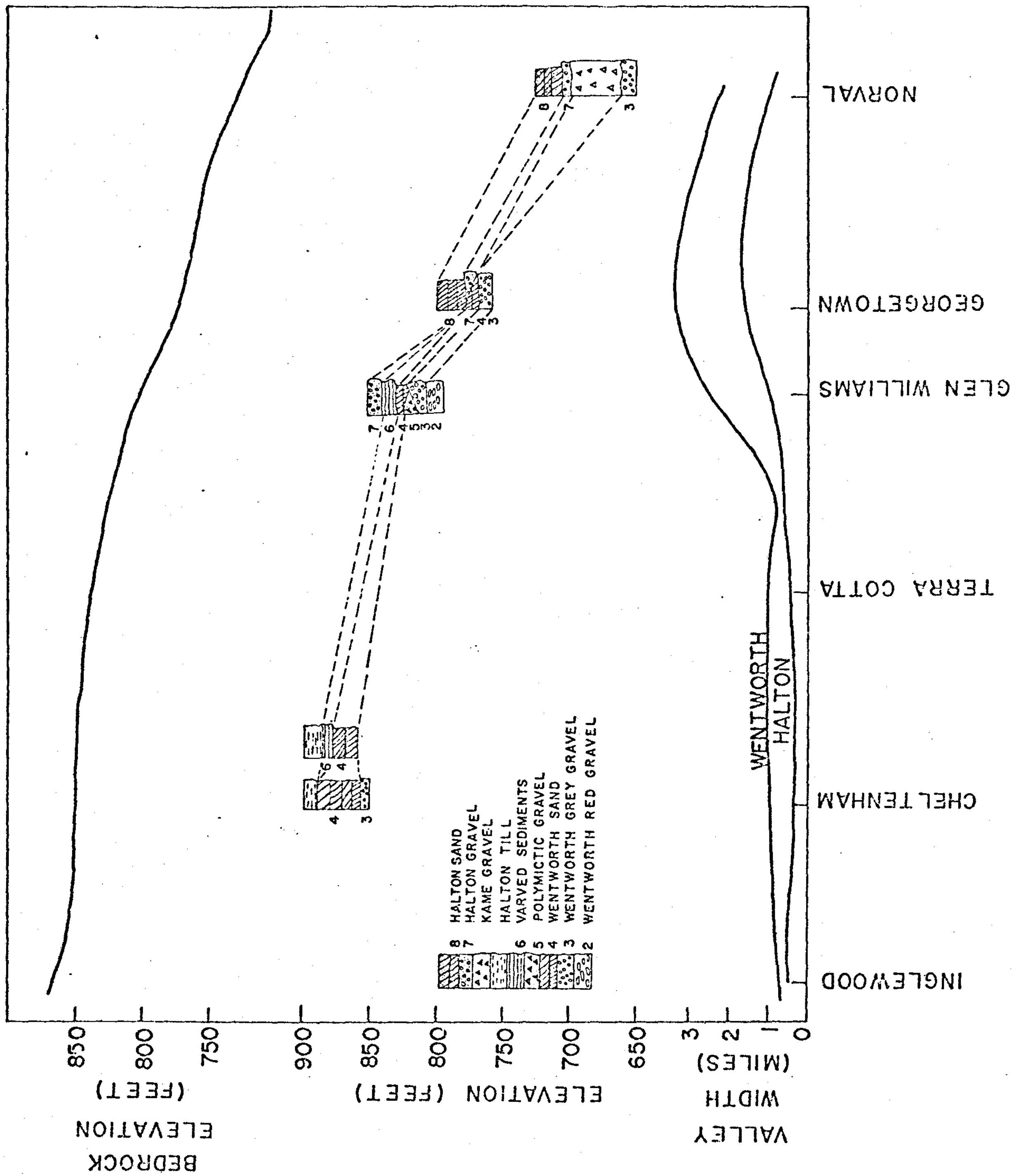

Figure 5c: The stratigraphy of the Bishop Pit, Glen Williams, and its relationship to other outcrops in the Credit Valley.

exposed in this one pit.

The deposits found in this pit were mapped on the scale of 1 inch to a 100 feet. The detailed map is shown in Fig. 5. Twenty-one vertical sections were measured within the pit and representative sediment samples were collected from the different beds. Paleocurrent orientations were also measured.

The beds within the Bishop Pit were correlated with one another and a tentative stratigraphy was constructed. Each set of correlated beds of identical lithology was organized into units. One unit then contained all similar beds that were located at the same stratigraphic level. These units were the basis of the stratigraphy. Ten other gravel pits to the north, west and south of the Bishop Pit were then examined and a vertical section taken at each. These other pits were situated near Cheltenham, Terra Cotta, Glen Williams, Georgetown and Norval.

The stratigraphy that evolved from the Bishop Pit was then compared with that found in each of the other pits so that in a stepwise fashion the different stratigraphic beds could be correlated up and down the valley.

A. <u>Glen Williams-Bishop Pit</u>

Bedrock Unit

The Pleistocene deposits of the Credit Valley rest upon the Queenston Formation of Ordovician age. The formation is composed of red-grey shales and siltstones. Nowhere in the Bishop Pit is the bedrock exposed other than as very large clasts in the gravels. However, a water well sunk 200 feet west of the pit intersects the bedrock at a depth of only 15 feet below the surface (Fig. 5b). From well records (Watts, 1950), the bedrock is known to deepen beneath the present course of the Credit River. Outcrops of Queenston Formation can be found one mile northwest of the Bishop Pit at the base of the Niagara Escarpment.

The bedrock therefore has been scoured into a U-shaped valley. The deepest part of the valley now contains the Credit River.

Basal Coarse Gravel Unit

The initial Pleistocene deposits found in the lower 10 to 15 feet of the Bishop Pit are coarse, red gravels (Fig. 5). These gravels crop out at the center of the pit, where excavations have been carried out to the deepest levels. No other sediment was identified below these gravels in the pit area. However, at the waterwell 200 feet west of the pit, these coarse gravels were found to rest directly upon the Queenston Formation bedrock.

There are abundant clasts of red shale within the coarse gravels and this lends the red color to the gravels. The largest clasts have apparent diameters up to 2 feet. Large sets of cross-bedding with heights up to 10 feet are commonly seen. The foresets display no grading down their lengths. Pockets of unoriented but well sorted pebbles are sometimes found along the large foresets. Many large, cross-sections of channels are present within the gravels. The channels incise one another and have ill-defined outlines due to the coarseness of the sediments. Imbrication of the cobbles and pebbles was not observed in these lower coarse gravels.

The gravels are cemented by a combination of calcite overgrowths and void filling by red clay. The mean paleocurrent orientation for the coarse red gravels in the Bishop Pit is southeast to south-southeast.

The maximum exposed thickness of the red gravels is 22 feet. In the northern area of the pit, the deposit has been largely eroded. Where the red gravel has been eroded, a finer grey gravel is deposited unconformably above it. In most places within the pit the contact between the two gravels is erosive. At other exposures a gradation in color and a reduction in the grain size denotes the change in lithology.

At two outcrops in the Bishop Pit a 3 to 4 foot bed of cross-bedded, pebbly, coarse sand was deposited between the coarse, red gravels and the overlying grey gravels.

Grey Gravel Unit

Deposited above the coarse, red gravel unit is a finer, slightly better sorted, grey gravel (Fig. 5). This pebbly gravel occurs in cross-bedded sets up to 4 feet in height. Numerous cross-sections of gravel filled channels are seen to erode one another laterally and vertically. These channels are 8 to 10 feet in width and 4 to 6 feet in depth. Their outlines are highlighted by alternating gravel and sandy gravel beds which fill the channel.

The grey color of the gravel denotes the predominence of small pebbles of limestone and dolomite rather than shale particles in the gravel. The gravel is partially cemented by calcite overgrowths. This cementation is most advanced in the cross-bedded gravels.

The grey gravels are interbedded with 2 to 3 foot beds of pebbly coarse sand. However, the majority of the cross-bedded sand was deposited above this unit. The contacts between the sand and gravel beds are erosive.

In the northern part of the Bishop Pit the grey gravels are found in the deepest parts of the pit. Large channels have cut through and eroded the coarse, red gravels, and have deposited the grey, better sorted gravels in their place.

The average thickness of this unit is 10 to 20 feet. One outcrop had a thickness of 20 feet of these gravels. The paleocurrent orientation as measured from cross-bed sets was to the southwest.

Cross-stratified Sand Unit

These medium to coarse, pebbly sands are found below, within and dominantly above the grey gravels. The sands are

are a buff color and are for the most part uncemented. The sands have well developed large scale cross-bedding, with maximum foreset heights of 2 to 3 feet. Cross-bedding of the trough or festoon type is also commonly seen. Rib and furrow structure, the plan view expression of sinuous-crested or linguoid ripples, is commonly seen in the finer size sands.

The sands interbedded with the grey gravels are slightly coarser than those found above these gravels. The sands attain their maximum thickness of 25 feet in the northwest corner of the pit. Here, the contact between the sands and grey gravels is not exposed, but the sands do lie at a higher elevation than the gravels.

Extrapolating from the total evidence of outcrop elevations across the pit, it seems appropriate to place the bulk of the cross-bedded sands in the next stratigraphic position above the grey gravels.

Distinct coarsening-upwards sequences of sediment are found interbedded with the cross-bedded sands (Fig. 5). The sequences have an initial deposit of greenish clay, lying upon a sharp basal contact with coarse sands. The clay grades upward into silt and rippled, fine sand. Above the rippled sand the sequence continues up into medium coarse sand with large scale trough cross-bedding. At its top the sequence is cut by sandy, gravel filled channels. The coarsening upwards sequences range from 1 to 10 feet in thickness and are found at many different stratigraphic levels within the cross-stratified sand (Unit 4) in the Bishop Pit. However, the coarsening upwards sequences are not found in those beds which are interbedded with the grey gravels (Unit 3) but are confined to those beds lying stratigraphically above the grey gravels.

Polymictic Coarse Gravel Unit

The very coarse polymictic gravels are found only in one large outcrop along the northwest edge of the Bishop Pit (Fig. 5). Large, angular blocks of shale (Queenston Formation) and dolomite (Lockport Formation) rest in the gravelly sand matrix. The large blocks have an apparent diameter of 1 foot. The deposits are characterized by a complete lack of sorting throughout.

The deposit is fan shaped, with the gravel spreading out laterally down into the valley. The gravels also thin out towards the edges of the deposit. The polymictic gravels interfinger with the grey gravels at the distal edge of the deposit and are in turn covered by a thick blanket of cross-bedded sand (Unit 4). The maximum thickness of this unit is 8 to 10 feet.

On the evidence of field occurrence, the polymictic gravels are placed at about the stratigraphic position of the upper part of the grey gravel (Unit 3) and below the cross-bedded sand (Unit 4).

Varved Clay, Silt and Fine Sand Unit

On the northeastern side of the Bishop Pit there is a 20 to 25 foot sequence of varved clays, silts and fine sands (Fig. 5). In some outcrops the varved sediments lie directly on the grey gravel (Unit 3) while at other outcrops they lie upon the cross-bedded sand (Unit 4).

The varves in the clay are poorly developed and have alternating red and grey colors. The fine sands are frequently rippled and the clays and silts have well developed horizontal lamination. The sand beds near the top of the varved Unit 6 have been contorted into tight folds. Dropstones are deposited among the varves and have clay laminae draped over them.

The surface upon which the varved sediments were deposited is erosive and uneven. In several outcrops the clays were deposited in channel shaped depressions in the grey gravels. The clays and silts were also deposited above an erosive contact with the cross-bedded sands.

By hand augering, it was found that these varved sediments could be traced back as an extensive deposit one hundred yards to the northeast of the pit. This large deposit of hard clay was the reason that the gravel pit was not extended to the east.

Upper Grey Gravel Unit

Occupying the uppermost stratigraphic position in the Bishop pit are grey-brown, sandy gravels (Fig. 5). The gravels have very poorly developed sedimentary structures which are difficult to measure. The deposits of the upper gravel occur as thin sheets, 4 to 5 feet in thickness or within deep channel structures up to 15 feet in depth. Because of the poor sorting, the internal structures of the channels are poorly outlined.

Where there are outcrops of varved Unit 6, the upper grey gravel (Unit 7) can be seen to rest unconformably above them. Where the varved sediments are absent, the upper grey gravels lie directly upon the cross-bedded sand unit. The upper gravel unit is found only along the east side of the Bishop Pit where the elevation is the highest.

Glen Williams Bishop Pit Stratigraphy

The deposits in the Bishop Pit exhibit a great abundance of erosive contacts and a marked lack of continuity of sedimentary beds. Furthermore, the individual units have a great range in preserved thicknesses over a very small distance. This highlights the great predominance and importance of erosion in the overal history of sedimentation.

The sediments do possess similar relations to one another throughout the pit and the Units are distinctive in appearance. The polymictic coarse gravels (Unit 5) and the varved clays, silts and sands (Unit 6) are restricted units in terms of their exposure in the pit. If they are eliminated from the stratigraphic considerations temporarily, the remaining stratigraphic positions are at once simplified. The coarse, red gravel (Unit 2) occupies the basal position in the stratigraphic column and is eroded and covered by a better sorted grey gravel (Unit 3). The grey gravel is interbedded with thin beds of coarse sand which reach their predominance in those horizons topographically higher than the grey gravels (Unit 4). Finally, a new period of erosion occurred, and locally eroded much of the cross-bedded sand (Unit 4). The sand unit was followed by a new, poorly sorted grey gravel (Unit 7).

Because of extensive deep erosion, the complex situation arises where dissimilar older and younger beds are deposited side by side at the same stratigraphic levels. Placing the polymictic gravels back into the stratigraphic column as a restricted event contemporaneous with the deposition of the grey gravels, and likewise placing the deposits of the varved sediments between the cross-bedded sand unit and the upper grey gravel, the following stratigraphy evolves (Fig. 5c):

Unit 7	Upper Grey Gravel
Unit 6	Varved Clays, Silts and Sands (in places eroded completely by Unit 7)
Unit 4	Cross-bedded Sands
Unit 5	Polymictic Gravel (locally present and stratigraphically equivalent to the upper part of the Unit 3)
Unit 3	Grey Gravel
Unit 2	Red Gravel
Unit 1	Bedrock

B. Cheltenham

Two outcrops of Pleistocene outwash sediments were examined in the Credit Valley near the town of Cheltenham. Here outwash and glacial debris from the last ice advance, the Halton Ice Sheet, have buried the original bedrock valley and the present Credit River runs to the western edge of the valley.

The lowermost exposed deposit starts with 5 feet of grey partially cemented coarse sands and gravels. The mineralogy of the gravel is largely carbonate rock fragments. Large scale trough cross-bedding is found within the gravels, ranging from 2 to 4 feet in height.

The grey gravels are in turn covered by 20 to 25 feet of cross-bedded, coarse sands and very fine gravel. The sands are cross-stratified and have many coarsening-upwards sequences within them. The color of the sand is buff and the sands are uncemented.

Above the sands is 3 to 4 feet of silt and clay, which grades upward into 15 feet of dark, blocky, silty till (Fig. 6).

The sequence of grey gravel followed by cross-bedded sand and silt is almost identical to the sequence of grey gravel followed by cross-bedded sands and varved clays that was found at Glen Williams. This similarity suggests that a correlation between the beds at the two locations may be appropriate (Fig. 5c).

The silty till capping the deposits at Cheltenham is

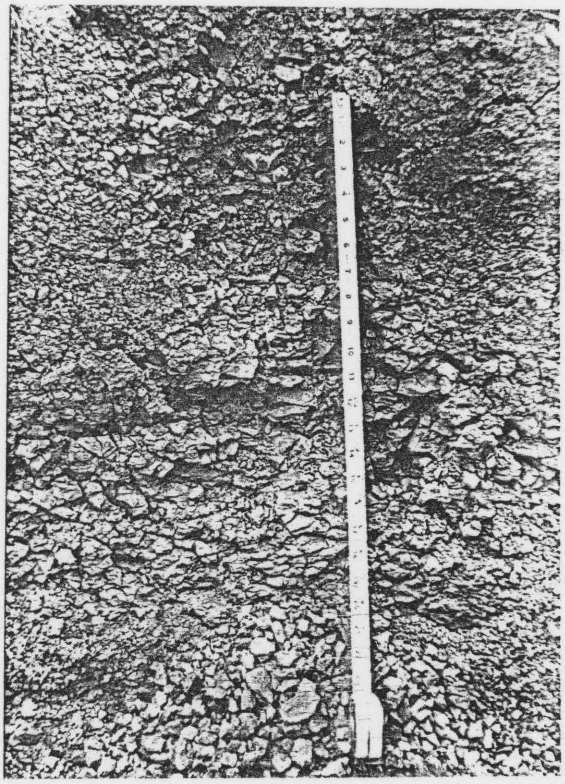

Figure 6: Outcrop of silty, blocky Halton Till.

the Halton Till. The sediments deposited below the till must be earlier than the Halton Ice advance in age and so must be part of the Wentworth outwash sediments.

Because of the similarity of appearance and stratigraphic position, the gravels at Cheltenham are correlated with the lower grey gravel (Unit 3) at Glen Williams. In the absence of an existing formal terminology, they are both here informally termed the Wentworth grey gravels. Likewise the buff colored, cross-bedded sands at Cheltenham and Glen Williams are concluded to be correlative and the sands are here informally termed the Wentworth sands.

The silts and clays at Cheltenham are different from those at Glen Williams in that they have a blue-brown color and they have neither varves nor dropstones within them. However a similar glacial origin for both is argued because the clays at Glen Williams have varves and dropstones while the clays at Cheltenham grade continuously up into the glacial deposit of Halton Till. These sediments are here informally named the Halton varved clays, silts and fine sands.

The Halton Till is not preserved at Glen Williams, probably because the area was subject to extensive erosion by outwash from the Halton Ice front. At Cheltenham, the Halton Till comprises much of the Cheltenham Moraine which buried and preserved the Wentworth outwash sediments deposited below it.

The deposits of Pleistocene sediments in and around the city of Georgetown are not continuous but are found as isolated patches on higher ground. It is in the vicinity of Georgetown that the Credit Valley widens (Fig. 5c).

The lowermost stratigraphic position at Georgetown is occupied by a 5 foot deposit of grey, partially cemented gravel. The gravel contains large scale trough cross-bedding 2 to 4 feet in height. Lying above the grey colored gravel is 1½ feet of rippled fine to medium sand. The top of the sand is everywhere eroded and covered by a 2 foot deposit of cross-bedded fine gravel. Above the fine gravel there is a deposit of 10 feet of small-scale trough cross-laminated sand. This sand unit consists of one single sand-filled channel.

The lower grey gravel and the thin deposit of rippled sand are similar in lithology to the Wentworth outwash sediments. The uppermost deposits of sand and gravel could be either of Wentworth or Halton age. However, no deposit correlative with the large 10 foot deep sand filled channel was found in the Wentworth sediments. This deposit of sand has escaped the extensive erosion that the Halton drainage would have caused were it deposited by the Wentworth outwash. Therefore the possibility exists that these upper sands and fine gravels are part of the Halton outwash sediments. Unfortunately no glacial sediments, such as the varved clay or till, are present to document adequately the age of these sediments.

C. Georgetown

The upper sand and gravel units are tentatively grouped as the Halton sand and Halton gravel respectively. The two sequences of sand deposited above gravel could then be explained as due to two periods of ice retreat and melt-water drainage. The lower sequence depicts the retreat of the Wentworth Ice sheet while the upper sequence depicts the Halton Ice retreat.

D. Norval

At Norval the Credit River ceases to flow south in a narrow valley at the foot of the Escarpment and turns south-west to flow over the broad Peel Plain. The present Credit River has cut deeply into the Pleistocene sediments which are now exposed along the sides of the river.

The lowermost 10 feet of these deposits are grey, well rounded, partially cemented gravels with large trough cross-bedding. The paleoflow was directed to the north. Again, these gravels are identical in lithology and strati-graphic position to the Wentworth grey gravels found at Glen Williams and Cheltenham (Fig. 5c). These gravels are correla-ted with the Wentworth grey gravels at Glen Williams.

Deposited above the Wentworth gravel are 50 to 60 feet of very coarse, unsorted gravels. These gravels have a few large sets of trough cross-bedding, and the paleoflow is again directed to the north. Many of the sediments display contorted laminae. The coarseness of the sediment, its lack of sorting, the contorted appearance and the scarcity of channel cross-section or regular bedding of the sediments suggest that these gravels are of kame or ice-contact origin. If an ice front were situated at Norval at some time in the Pleistocene this would account for the diversion of the drainage to the north, and hence the north pointing paleocurrent indicators.

On top of the kame gravels lies 6 feet of grey poorly sorted and cross-bedded gravel. Paleocurrent measurements give the direction of flow to the south. These gravels were probably deposited by meltwater drainage from the Halton Ice retreat and are correlated with the lithologically similar Halton gravel at Geln Williams.

Upon the Halton gravel lies 14 feet of massive, structureless sand. Within this sand isolated 1 foot thick layers of fine, parallel laminated clays were found. The geometry of the sand deposit is that of a sheet-sand building out to the south and southeast down the Credit Valley from the vicinity of Norval. The structure and topographic elevation of these sands matches descriptions of deltaic sands found in the lower Humber River Valley to the east. The Humber Valley sands built out as a sheet deposit into the glacial lake known as the Peel Pond. The isolated clay beds found at Norval might then be lake deposits from the Peel Pond.

The upper massive sand is correlated with the Halton sands at Georgetown but is described in Chapter 5 under the heading of "Delta" Sand facies.

E. Credit Valley Stratigraphy

Uniting those observations in the entire lower Credit Valley with the stratigraphy worked out in the Bishop Pit at Glen Williams, the following stratigraphic column is proposed for the Pleistocene sediments of the Gredit Valley;

Halton Outwash Sediments

Halton Sands
Halton Gravels
Kame deposits (locally present)

Halton Till

Varved clays, silts and fine sands

Wentworth Outwash Sediments

Wentworth Sands
Polymictic Gravels (locally present)
Wentworth Grey Gravels
Wentworth Red Gravels

Bedrock: Queenston Formation

MINERALOGY

Chapter IV

In order to determine the provenance of the Credit Valley outwash sediments, representative samples of the three major sediment size modes, the gravels, sands and clays, were analysed mineralogically. The sediment samples were all collected from the Bishop Pit at Glen Williams.

The gravels were screened with a -3.0 phi sieve and the retained cobbles and pebbles were examined under a binocular microscope. The sand rich sediments were immersed in Tetrbromomethane (sp. gr. 2.92). The heavy minerals separated out were examined with a petrographic microscope. The clay size fraction was examined qualitatively with a x-ray diffractometer.

Gravel Fraction

Five samples of the gravel size fraction, including samples from the Halton and Wentworth Gravels, were examined to determine their mineralogy. In each sample, 100 pebbles longer than one half inch in diameter were examined for their lithology. The results of the five pebble counts are listed in Table 2.

The gravels are either calcite cemented or contain an interstitial clay matrix. Large pebbles of carbonate or ig-

TABLE 2

Composition of pebble samples. One hundred pebbles were counted in each sample. Bishop Pit, Glen Williams.

Sample	19A	18D	13C	9B	6A
Strat. position	Went. Red Gravel	Went. Grey Gravel	Halton Grey Gravel	Went. Red Gravel	Went. Grey Gravel
Carbonates	94	85	94	84	82
Granite	6	8	3	12	3
Gneiss	-	4	-	-	6
Shale	-	3	-	-	9
Others	-	-	3	4	-

TABLE 3

Point counts (100 grains) of Pleistocene sands in the Bishop Pit, Glen Williams.

Sample	21B	7H	13B	8A	24D
Strat. position	Went. Sand	Went. Sand	Went. Sand	Went. Sand	Halton Sand
Quartz	48	62	58	64	49
Carbonate	37	26	26	20	31
Feldspar	9	8	10	8	13
Rock Frag.	2	2	-	-	4
Others	4	2	6	8	3

neous rock are found covered by small granules and pebbles of carbonate or quartz. The smaller grains are cemented to the larger ones by calcite overgrowths.

Red clays, derived from the shale of the Queenston Formation, fill the interstitial areas of the gravels. On compaction and dehydration, the clays fill the entire intergranular void spaces and bind the gravel particles together.

The Bishop Pit was subject to some investigation by the industrial branch of the Ontario Department of Mines. They sampled the gravels and made a detailed mineralogical examination of them. The constituent rock fragments in order of decreasing abundance were; Black River and Trenton Limestone and Dolomite, Dundas and Queenston Siltstones, Potsdam Sandstone, Queenston Shale, Precambrian acid and basic igneous and metamorphic rocks (Hewitt and Karrow, 1963).

The Trenton and Black River Groups lie to the east of the Credit Valley. The present closest position of this limestone assemblege is at Aurora, north of Toronto, where it lies under Pleistocene deposits. The closest outcropping of dolomite occurs in the Guelph-Lockport-Amabel Formations which are found atop the Escarpment, bordering the Credit Valley. The Queenston Formation underlies the Credit Valley and the Dundas-Meaford Formation is found at depth west of Malton. The closest outcropping of the Potsdam Formation is west of Kingston. Therefore the source region for all these lithologies, with the sole exception of the dolomite, is to the east.

The gravels contain a wide range in grain sizes, from clays up to boulders a foot in diameter. The pebbles and boulders are moderately to well rounded.

Sand Fraction

Five samples of sand from the Bishop Pit were chosen for mineralogical analysis. The entire sand fraction was mounted in transparent, epoxy resin and thin sections were cut from them. The thin sections were mounted on a grid and a point count of one hundred grains was made for each slide. The results of the point counts are listed in Table 3.

According to the classification of Folk (1964), this sand is an arkose or subarkose. The quartz grains show straight extinction with an occasional grain having a wavy extinction or composite nature. A roundness study indicated that the quartz grains have a roundness of 0.3 to 0.4 on the Waddell scale.

The feldspars are predominantly plagioclase, albite and oligoclase. There are feldspars showing microcline twinning and a few have perthitic textures. Few feldspars are fresh in appearance, with many showing alteration to sericite, calcite and kaolinite along the lamellae of the crystals. The altered feldspar looks like the fine textured carbonate grains.

The carbonate grains are well rounded, microcrystalline grains. The rock fragments are fine grained phyllites, polycrystalline quartzite and serpentinized igneous rock.

The straight extinction quartz, common or plutonic quartz, and the plagioclase seem to indicate an igneous source region, the closest being the Canadian Shield. The lesser amounts of undulose extinction quartz and soda feldspars indicate a metamorphic - granitic terrain, again approximated by the Canadian Shield. The carbonate grains are mostly limestone and have their source area in Eastern Ontario.

In the past an attempt has been made to study the provenance of the sands in the tills and outwash sediments in Southern Ontario. These studies (Chapman and Dell, 1963; Dell, 1959) have been based on the calcite/dolomite ratio in sands or the heavy mineral suite. An interesting point was discovered by Chapman and Dell (1963) in their study of sands in the tills and the spillways in the Orangeville, Caledon and Georgetown areas. The ratio of calcite to dolomite in the sands and from the pebbles in the tills suggest that the tills recieved their carbonate sediments from the rocks they locally overrode. However the ratio of calcite to dolomite in the sands of the outwash channels was markedly higher, having more calcite present. The ratio was similar to those ratios found in deposits of the Lake Simcoe-Georgian Bay lobe. This indicates that the source of these sediments was to the north and east.

The amount of carbonates found in the gravels is considerably greater than the amount found in the sands (Dell, 1959). The carbonates are easily reduced by solution. The solution of these carbonate rocks saturates the ground water with carbonate salts and contributes to the calcitic cementation of the sands and gravels.

The sands are well sorted and display evidence of abrasion. This would account for the lack of clay size material in the sands. Because of the lack of clays and the good sorting, the sands of the Credit Valley can be called texturally mature (Folk, 1964).

Heavy Minerals

A sample of well sorted sand was chosed for a representative heavy mineral analysis. Due to the hydraulic ratio effect, heavy minerals of certain specific gravities travel in the company of larger quartz grains. By taking the whole sand fraction it was hoped that the total heavy mineral content would be separated.

As seen from Table 4 the heavy minerals constitute 3.9 % of the sand by weight. From observations in the field it was noticed that there was concentration of heavy minerals along some of the foresets of the cross-bedded sands. The relative amounts of the different heavy minerals are shown in Table 4.

The abundance of hornblende and garnet and the presence of zoisite, serpentine, tremolite-actinolite, diopside, and hedenbergite would indicate a metamorphic terrain such as

TABLE 4

Mineral Composition of Outwash Sands

Percentage by weight	
Heavy Fraction	3.9
Light Fraction	96.1

Percentage by frequency-Light fraction	
Quartz	49.0
Orthoclase	0.8
Microcline	3.0
Plagioclase	9.2
Carbonate grains	31.0
Rock fragments	4.0
Unknowns	3.0
	100.0

Percentage by frequency-Heavy fraction	
Opaques	71.94
Garnet	12.94
Hornblende	7.20
Tremolite-Actinolite	1.44
Hypersthene	0.72
Hedenbergite	0.36
Diopside	0.36
Tourmaline	0.36
Zoisite	0.36
Zircon	0.36
Spinel	0.36
Chlorite-Serpentine	1.44
Muscovite	0.72
Unknowns	1.04
	100.00

the Grenville Province of the Canadian Shield as a source area for these detrital minerals. The zoisite appears to be an alteration product.

The degree of abrasion to which these heavy minerals have been subjected is hard to determine. A single mineral species, such as hornblende, exhibits large numbers of both angular and rounded grains.

Dreimanis et al. (1957) have attempted to use heavy mineral suites to determine the provenance of glacial tills in Ontario. They held that a dispersal of sediment from east to west across Ontario could be traced by the ratio of purple to clear garnets in tills. However Dell (1959) concludes from her study on heavy minerals in till and outwash sediments that no reliable heavy mineral suite exists to differentiate between deposits of the different ice lobes. The heavy minerals only demonstrate that the original source terrain was igneous and metamorphic in character.

Clay Fraction

Six samples of clay rich sediments were X-rayed on a Norelco Diffractometer employing Cu-K alpha radiation. Each sample was prepared as follows; 1) untreated, 2) treated with ethylene glycol vapor for 12 hours, 3) heated to 350 degrees Centigrade for 12 hours, 4) heated to 550 degrees Centigrade for one hour. The techniques for the preparation and treatment of the slides is essentially that outlined by Warahaw and

Roy (1961) and Brown (1961).

All the untreated samples had strong reflection peaks at 14A. On glycolation the peak intensity decreased but no change in position occurred. On heating to 350 degrees Centigrade all samples except 8J lost the 14A peak. Since no 12A peak appeared to indicate a vermiculite shift, the presence of a 14A mixed layer clay was demonstrated.

In sample 8J, the retention of the 14A peak plus the existence of a peak at 7A indicates the presence of chlorite. Third and fourth order chlorite reflection peaks were found at 4.75 and 3.54A.

Strong 10A reflections were also found on all of the samples. This points to the presence of illite. All samples displayed reflection peaks at 7A and 3.58A. Upon heating these peaks were seen to decrease proving the existence of kaolinite. No reflection appeared to indicate the presence of quartz in the clay size fraction. Table 5 summarizes the occurrances of the clay minerals. Included for comparison are two mineralogical analyses performed on the Queenston Formation in Hamilton. Large amounts of the easily eroded shale was found in the sediments. As can be seen, the Queenston Formation could serve as a source of the illite, chlorite and some mixed layer clays. The illite when weathered can lose $\bar{K}$ and be degraded to a mixed layer clay.

Chlorite and kaolinite can be derived as a product of the weathering of the igneous and metamorphic terrain such

TABLE 5

Clay Mineralogy of Pleistocene Sediments

Sample	1E	8J	8B	11E	18B	18A		
Strat. position	Went. Sand	Went. Sand	Went. Sand	Went. Sand	Went. Sand	Went. Sand	Queen.* Shale	Queen.** Shale
Chlorite	none	yes	none	none	none	none	mod.	mod.
Mixed layer	yes	?	yes	yes	yes	yes	none	trace
Illite	yes	yes	yes	yes	yes	yes	abund.	abund.
Kaolinite	yes	yes	yes	yes	yes	yes	none	none
Vermiculite	none	none	none	none	none	none	none	none

* G. Candy - (1963)

** Allen and Johns - (1960)

as the Canadian Shield.

Conclusions

The gravel and clay fractions show evidence of a local source of sediments, the Paleozoic carbonates and shales of Eastern and Southern Ontario. The sand size fraction, showing evidence of more abrasion and reworking, indicates a northern, metamorphic source area.

The fragments of shale of the Queenston Formation in the gravels, and the angular quartz and hornblende grains in the sands, are indicative of relatively little abrasion and reworking. The larger amounts of rounded quartz and hornblende would indicate that some, if not most, of the sand size fraction has been reworked several times.

The variety of mineralogical types, the source of the detrital heavy minerals from the north and the textural inversion all agree with a glacial source for these sediments.

SEDIMENTARY FACIES

Chapter V

The Pleistocene stratigraphy of the Credit Valley encompasses two ice retreat and outwash events. In both of these events, during the Wentworth and Halton Ice retreats, the sequence, cross-stratified sands overlying gravels is found. This repetitive sequence is the result of similar depositional processes acting during both events.

Because of the similarity in aspect of the Wentworth and Halton sands and gravels, the descriptions given below are in terms of sedimentary facies rather than informal stratigraphic units. The facies are defined by the lithology and the assemblage of sedimentary structures and are stratigraphically unrestricted. The facies recognized are:

1) Basal Red Gravel Facies-This facies is composed of the coarse, red, unsorted Wentworth gravel which was deposited directly upon the Paleozoic bedrock in the Credit Valley. It does not occur in the Halton deposits.

2) Grey Gravel Facies-Both the Halton and Wentworth grey gravels are included in this facies.

3) Cross-stratified Sand Facies-The Halton sands and the Wentworth sands are grouped into this facies.

4) Coarsening-Upwards Facies-This facies includes

the coarsening upwards sequences found interbedded with the Wentworth and Halton sands.

5) Varved Sediment Facies-This facies consists of varved clays, silts and fine sands. The facies is found deposited between the Halton and Wentworth Outwash sediments.

6) Polymictic Gravel Facies-The coarse polymictic gravels, that are found in small scattered outcrops in the Wentworth deposits, constitute this facies.

7) "Delta" Sand Facies-The massive sheet sands in the lower Credit Valley make up this facies. The sands are outwash sediments that built as a delta out into the Peel Pond, and are included in the Halton deposits.

1) Basal Red Gravel Facies

Description

This facies occupies the basal position in the infilling of the glacially scoured Credit Valley. Because of this position, the gravel contains large amounts of the red shale bedrock upon which it was deposited. The gravel is an outwash sediment deposited during the retreat of the Wentworth Ice Sheet.

The gravel has two dominant sedimentary structures, large delta-like foreset beds and festoon or trough cross-stratification. The large delta-like foreset beds have heights up to 10 feet and contain the coarsest sediment particles to be found in the gravels. Deposited on the foresets are very large clasts of red shale, the largest clast having an apparent length of 2 feet. These large shale fragments are mainly unweathered. The dip angle of the large foreset beds is from 10 to 15 degrees. The foreset beds have an asymptotic bottom contact. Coarse and uncemented cobbles, pebbles and sand are deposited along the foresets (Fig. 7). The pebbles in these clusters have no preferred orientation.

The large delta-like structures laterally interfinger with trough cross-bedded gravels and cross-sections of gravel filled channels. The gravels in these channels are red in color but are slightly finer than those in the delta-like structures. The channels range from 3 to 5 feet in depth, and from 15 to 20 feet in width. Parallel sections through the gravel filled channels display trough cross-stratification within the channels. The foresets of the trough cross-bedding are composed of sandy gravels. Several of the foresets, however are composed of entirely of pebbles, (Fig. 8). The gravel filled channels are numerous and commonly cut into one another laterally and vertically.

Grain Size Analysis

Grain size analyses were made for one sample of each of the gravels having a delta-like structure and a channel structure. The size fractions were separated by sieving the sediment at quarter phi intervals as outlined by Folk (1964). These gravels have a high proportion of shale within them which helps bind the sediment together. Attempts to disaggregate these gravels can not help but break down some of the shale which was originally deposited as small lumps, as in armoured clay balls. This sample preparation then slightly increases the amount of fines in the sediment.

The weights of the different size fractions from the sieving operation were entered into a digital computer and the mean, standard deviation, skewness, and kurtosis were calculated by the method of moments. The computer also produced a cumulative plot of the size fractions (phi scale) on a probability scale. The results of the size analysis are shown in Table 6.

The grain size distributions are polymodal with the majority of the sediment in the gravel fraction (larger than -1.0 phi). The mean grain size measurements indicate that the delta-like structure is coarsest. But in polymodal sediments the mean grain size is a rather meaningless statistic. It is of greater value to note the sizes of the major modes in polymodal sediments. Both samples have modes at -2.75 phi (pebbles) and at -1.0 to -1.25 phi (granule). Sample 9B has another mode at 0.0 phi (very coarse sand). The sample from the large delta-like structure has its major mode in the pebble size range whereas the channel structure has its major mode in the granule size range. So again the delta-like structure is shown to be the coarser sediment.

The standard deviation is a measure of the sorting of a sediment. Both samples display poor sorting with values between 1 and 2 phi intervals (Folk, 1964). The channel gravels have slightly better sorting values.

Skewness measures the non-normality or asymmetry of a sediment size distribution. The gravel samples are positively skewed, especially sample 3A. This indicates that the size distribution is asymmetric in that it has a fine "tail" in its size distribution. Kurtosis is a parameter which expresses the ratio between the spread of the central part of the distribution and the spread of the tails (Folk, 1964, p. 85). The gravel sample from the delta-like structure has a slightly leptokurtic value which is a measure of its major gravel mode. The trough cross-laminated gravels have a platikurtic value indicating the presence of two large modes, a -1.25 and 0.0 phi.

The standard deviations and kurtosis values indicate the poor sorting of the sediments. The trough cross-laminated gravels are however slighty better sorted even though they have a larger proportion of sand within them.

Interpretation

The coarseness and poor sorting of the sediments as well as the preservation of the large shale clasts suggests that these gravels were deposited close to their source area, without extensive transportation. The currents which deposited the gravels were both strong, since the sediment load is

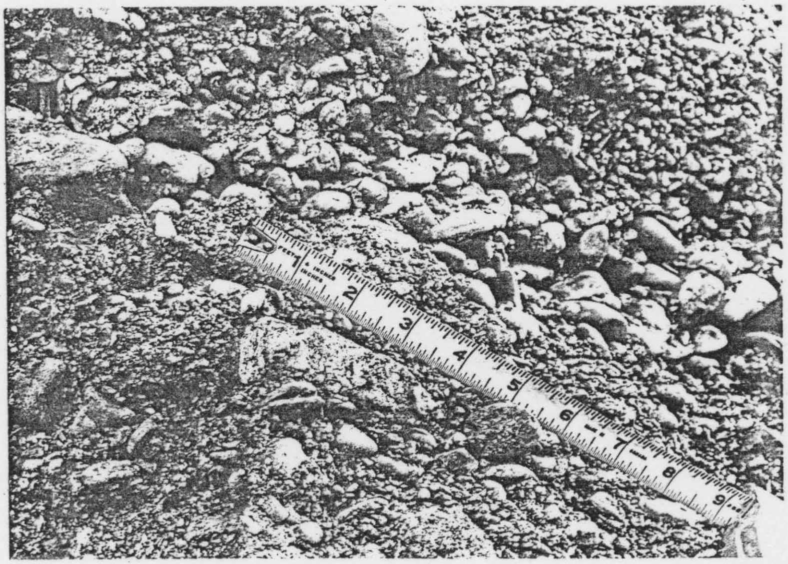

Figure 7: Coarse, uncemented pebbles along the foresets of the Basal Red Gravel.

Figure 8: Large trough cross-stratification in the Basal Red Gravel. Field of view approximately 3 feet high and 4 feet wide.

TABLE 6

Grain Size Analysis of Basal Red Gravels

Outcrop at Bishop Pit	3A	9B
Sedimentary Structure	delta-like foreset beds	trough cross-stratification
Size modes (phi intervals)	-2.75 -1.00	-2.75 -1.25 0.00
Mean size (phi intervals)	-1.54	-0.84
Standard deviation (phi intervals)	1.73	1.25
Skewness	0.62	0.23
Kurtosis	1.86	0.78
Percent sand content	36.00	49.00
Percent gravel content	64.00	51.00

coarse, and narrowly confined, since the channel shaped scours are everywhere abundant in cross-sections. These observations thus imply that the depositional environment was one where there were numerous, fast flowing streams cutting through a floodplain composed of coarse, unsorted gravels. The frequent erosion, vertically and laterally, indicated by the channel shaped scours further suggests that the fast flowing streams were in a constant state of instability, and frequently migrated back and forth over the floodplain.

At extremely high river discharges coarse gravel would be transported in the river channel and would not be deposited. As the velocity decreased and the river became incompetent to transport the coarsest fraction of its total load the coarsest load fraction would initiate channel bar development. Once the initial deposit of coarse sediment is present on the river bottom, sediment swept over its surface could be trapped on the lee side of the initial bar deposit. Here the sediment would be protected from erosion because of the flow separation over the bar form. Thus the bar would grow by addition of sediment to its downstream end as in the manner of an advancing wedge of sediment such as a delta. The bar would also build sideways by the sweeping of sediment obliquely across the bar surface by irregular components of the flow. These sediments would avalanche down the side slope of the bar constructing foreset slopes which are oriented obliquely to the river flow direction rather than perpendicular to it.

As the bar builds upward more and more water would be forced into small channels to either side of the bar. Because of the narrowness of the sidechannels, caused by the growth of the mid-channel bar, the velocity of flow increases and would tend to trim the bar sides and inhibit further widening of the bar. If the banks are easily eroded then the side channels would erode the banks and allow further widening of the bar.

The coarse, delta-like accumulations of sediment are therefore concluded to be the result of rapid deposition by avalanche of coarse gravels down the lee slope of an initial bar deposit. The poor sorting of the deposit and the pebble clustering support this view. The bar built upward and downstream. The delta-like foresets represent the downstream component of the growth.

The trough cross-stratified gravels, which are deposited within the large channel shaped scours, interfinger laterally with the river bar deposits. The channel shaped geometry of these slightly finer gravels suggest that they were the site of active channel flow alongside the gravel bars.

At times of high discharge the flow velocity within the channels would be large enough to sweep the channels clean of any gravel accumulations beside the larger mid-channel bar. With lower periods of discharge or increased aggradation within the flow system, gradually these side channels would start to be filled up with gravel. Since the flow velocity would be substantial, the infilling would take place by filling in and shallowing of the channels with gravel from the upstream direction. However the gravels would be subject to reworking by the channel flow and would tend to be better sorted than the rapidly deposited and quickly buried bar gravels.

The basal red gravel facies thus consists of sequences of fluvial gravels. The gravels are the result of bar development within alluvial channels and also are the result of channel filling on either side of the mid-channel bars.

This facies is found only in the lower Credit Valley, where it widens in the vicinity of Glen Williams. Because the gravels rest directly on the Paleozoic bedrock, and because they are found only in the lower, wider reaches of the valley, the red gravels are interpreted to be the initial deposits in the growth of an alluvial plain which was constructed by the meltwaters of the Wentworth Ice Sheet. The floor of the valley would be covered by a coarse ground moraine and glacial debris derived from the underlying shales of the Queenston Formation. This very coarse sediment would be the primary load of the initial meltwater drainage. Meltwater flowing down the Credit Valley would experience a decrease in its velocity where the valley widens and there would be a resulting decrease in the competence of the flow. Therefore at this wider part of the valley, sediment would be deposited and an alluvial plain initiated.

2) Grey Gravel Facies

Description

Two similar deposits of sandy, pebbly gravels are present at different stratigraphic levels in the Credit Valley. The upper gravel is the Halton Grey Gravel while the lower one is the Wentworth Grey Gravel. Since the mineralogy, size and structures of the two gravels are identical they are inferred to have a common origin and so are combined into this one facies.

The gravels have two dominant sedimentary structures, coarse gravelly, angular foreset beds and trough cross-stratified gravel beds. The deposits are quite simialr to the preceding Basal Red Gravel but have a finer size and much larger areal extent.

The angular foresets are 2 to 4 feet in height and are composed of pebbly gravels which have been later partially cemented by calcite overgrowths. The cross-stratification stands out well due to the postdepositional cementation. The pebbles on the foresets are well rounded Paleozoic carbonates, sandstones and siltstones. Because of the coarse nature of the sediment, the foreset thicknesses range up to 4 inches. The sorting in these deposits is better than that seen in the Basal Red Gravel Facies (Fig. 9).

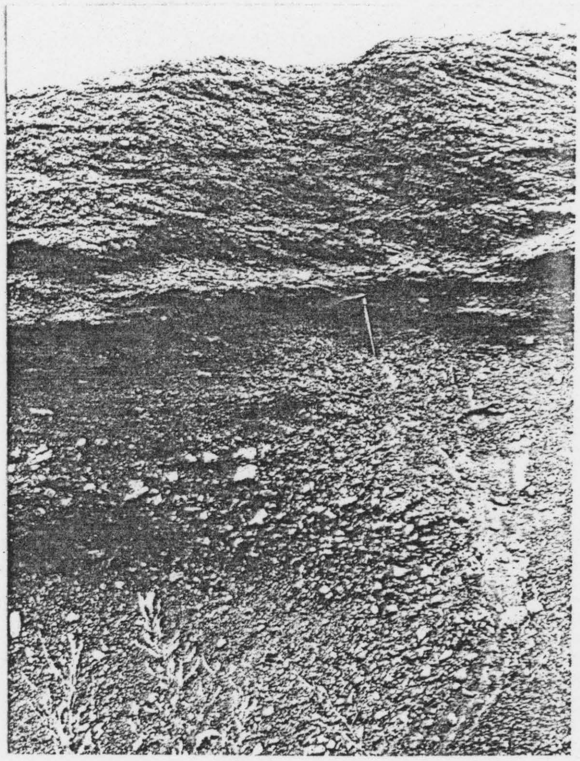

Figure 9: The contact between the Basal Red Gravel and the Grey Gravel.

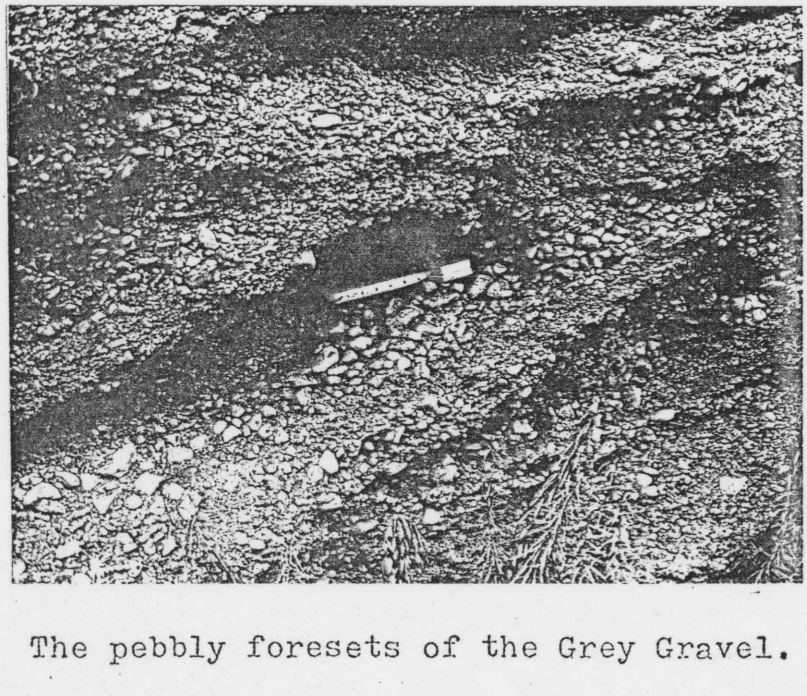

Figure 10: The pebbly foresets of the Grey Gravel.

The foreset bed deposits again pass laterally into the trough cross-stratified gravels which fill in channel shaped scours. Cross-sections of gravel filled channels are abundant in these gravels. Most channel deposits range from 2 to 4 feet in depth. One complete cross-section of a gravel filled channel was 6 feet deep and 40 feet wide.

Isolated areas of the lower Wentworth gravel have a red color due to the presence of incorporated red shale particles. However, unlike the Basal Red Gravel Facies in which the red shale occurred as numerous large clasts, the red shale in these grey gravels is very fine, a product of disintegration of clasts from the Queenston Formation. A few large clasts are preserved in the lower portions of the Halton Gravel.

These gravel channel deposits erode one another laterally and vertically. Where one channel has eroded part of the deposits of an earlier channel, a lag gravel deposit of large cobbles can be observed at the bottom of the more recent channel. These cobbles were the coarsest sediment fraction within the eroded portion of the earlier channel. The latter channel flow was not competent enough to remove these cobbles but did however winnow out and carry away all the intergranular fines. The effect of the current on these cobbles is demonstrated by the excellent upstream imbrication that they now possess. The cobbles are 4 to 6 inches in diameter and have steeply dipping (30 to 36 degrees) long axes

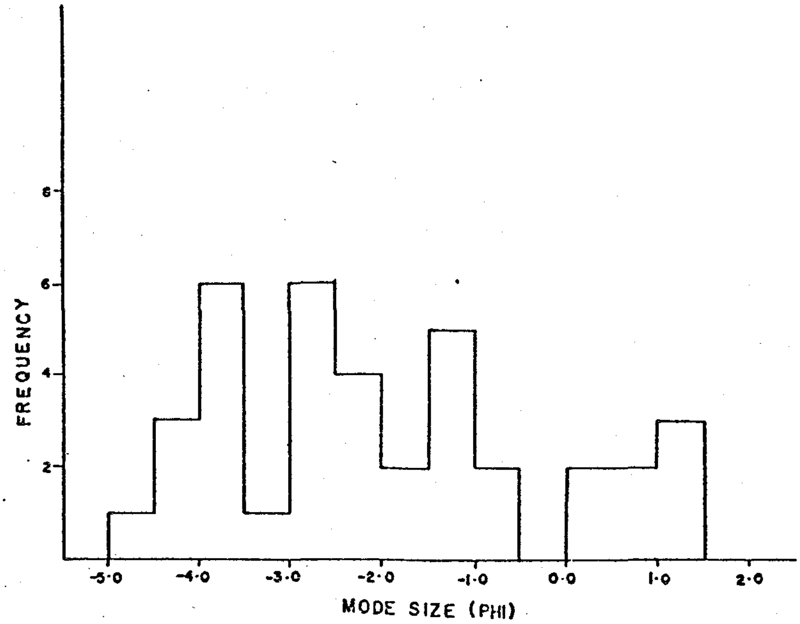

Figure 11: Histogram of the sediment modes in the Grey Gravel.

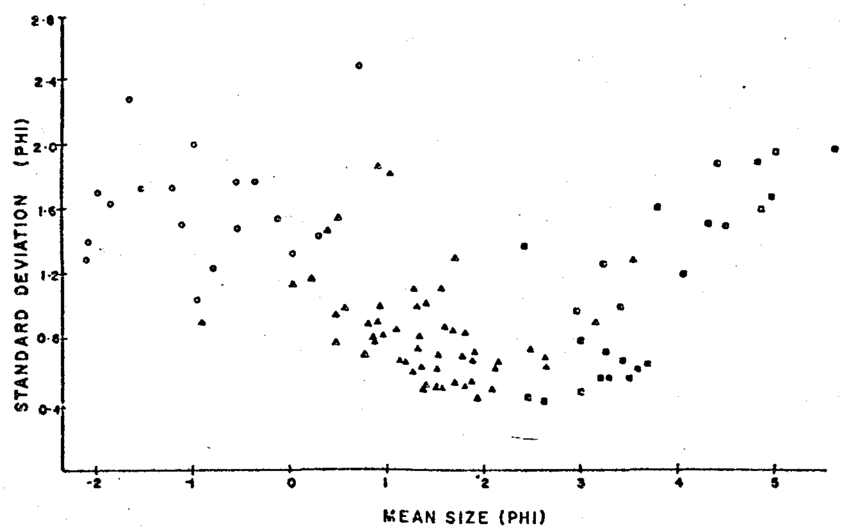

Figure 12: Plot of the mean sediment size vs. the standard deviation in the Pleistocene outwash sediments

O - cross-bedded gravel
▲ - cross-bedded sand
□ - clays, silts and fine rippled sands of the Coarsening Upwards Sequence

directed upstream.

Where sections parallel to the channels are available the large trough cross-stratification that fills the channel can be seen. Frequently the pebbles along the foresets will display an alignment roughly parallel with the dip (Fig. 10). The channel gravels are also partially cemented by calcite overgrowths.

Both these gravel deposits are interbedded with thin beds of coarse cross-stratified sand. These sands are members of the Cross-stratified Sand Facies.

Grain Size Analysis

Analysis of the sediment size distributions for 13 Wentworth and Halton Gravels indicates that these sediments are polymodal. In Fig. 11, a frequency distribution of the major sand and gravel modes for the gravel samples is shown. The polymodality of the gravels is clear and the modes can be divided into four groups. There are two pebble size modes (-3.5 to -4.5 phi and -3.0 to -2.0 phi) and a granule size mode (-0.5 to -2.0 phi) in the gravel fraction. A further mode occurs in the range coarse to medium sand (0.0 to 2.0 phi). The limit of the gravel size fraction is usually taken as -1.0 phi, but in these sediments the significant break in the distribution occurs between -0.5 and 0.0 phi. In all references to the amount of gravel in these sediments, the gravel content is taken as that quantity having a size in excess of 0.0 phi.

The amount of gravel in these sediments ranges from 36 to 91 percent by weight. Greater than two thirds of the samples have gravel contents greater than 70 percent. In the Lafayette Gravel of Kentucky, Potter (1955) found the gravel content to be from 60 to 80 percent. In the terrace gravels of South Dakota, Plumley (1948) found an average of 80 percent gravel.

From studies of the tightest and loosest packing configurations of gravels, Plumley (1948) proposed a test to determine whether the sand fraction present in gravels was the result of similtaneous deposition with the gravels or else a result of later infiltration with groundwater. Gravel with the loosest packing could accomodate up to 32 percent sand within its void spaces, whereas gravel with the tightest packing could accomodate only 22 percent sand. Since two thirds of the grey gravels of the Credit Valley have less than 32 percent sand within them, they might have had the sand washed into the interstitial void spaces.

The mean grain size of the gravels ranges from -2.25 to 0.75 phi. The standard deviation values range from 1.14 to 2.60 phi. The high value of the standard deviation indicates a very poor sorting. This is a reflection of the many sediment size modes present in the gravels. The values of the mean and standard deviation are plotted in Fig. 12. There is wide scatter in the data but there is still a trend to better

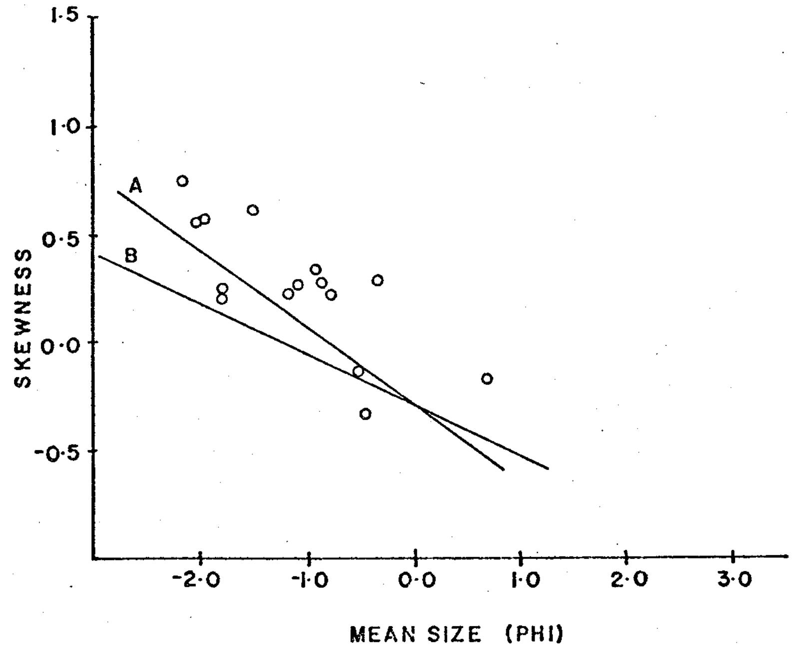

Figure 13: Mean size vs. skewness in the Grey Gravel.
(A) Plumley (1948)
(B) Folk and Ward (1957)

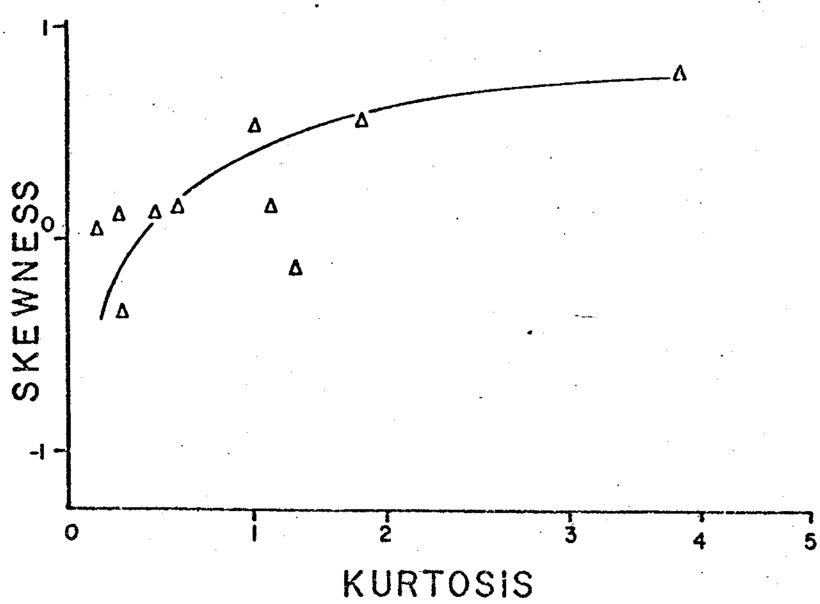

Figure 14: Skewness vs. kurtosis in the Grey Gravel.

sorting with increased grain size. Those sediments having a low mean value are characterized by major size modes in both the gravel and sand fractions. This equivalence of modes ensures a wide sediment size distribution and consequently poor sorting. Those sediments with a high mean value are characterized by a dominant pebbly gravel mode or modes. This mixing of several modes within the gravel fraction again produces poor sorting.

In these polymodal sediments, the values calculated for the skewness do not represent the characteristics of a single mode but the interrelation of several modes. In these gravels the skewness values range from -0.3 to 0.8. These values are plotted with the mean size in Fig. 13. A trend of reducing skewness with reducing grain size is apparent. This trend was also observed in the studies of other polymodal sediments (Plumley, 1948; Folk and Ward, 1957). The positive skewness represents an asymmetry in the distribution toward the finer sediment sizes. Those sediments having skewness values near zero have equally large modes in the gravel and sand fractions making the distribution symmetrical. This combination of two modes, pebbly gravel and medium sand, explains the poor sorting observed in these sediments.

The values of kurtosis vary from 0.0 to 3.45 phi. A plot of skewness and kurtosis (Fig. 14) displays a trend of decreasing kurtosis with decreasing skewness. Those samples with positive skewness values have a fine tail in their domin-

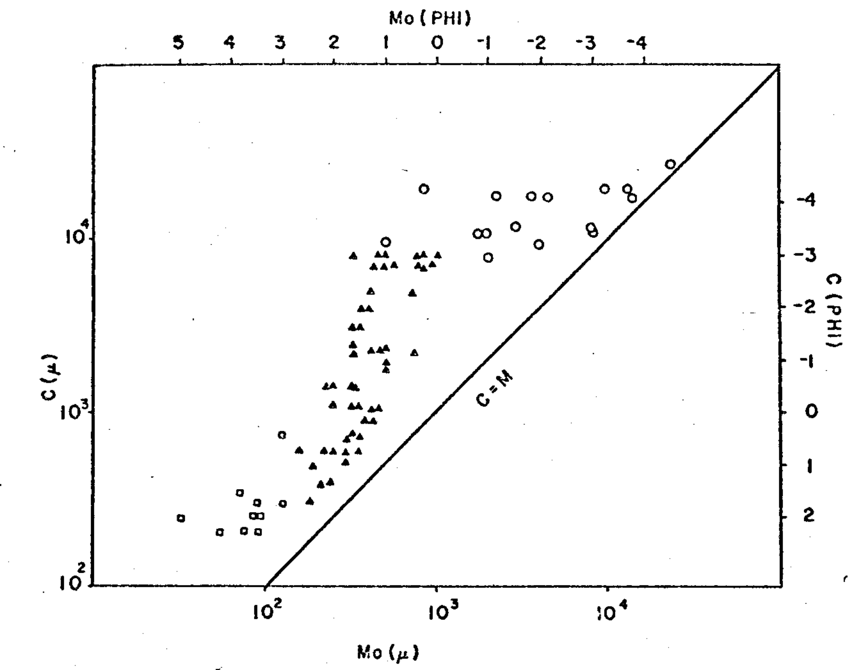

Figure 15: Plot of Mo vs. C for the Pleistocene outwash sediments.
O - cross-bedded gravels
▲ - cross-bedded sands
◻ - clays, silts and rippled fine sand of the Coarsening Upwards Sequence

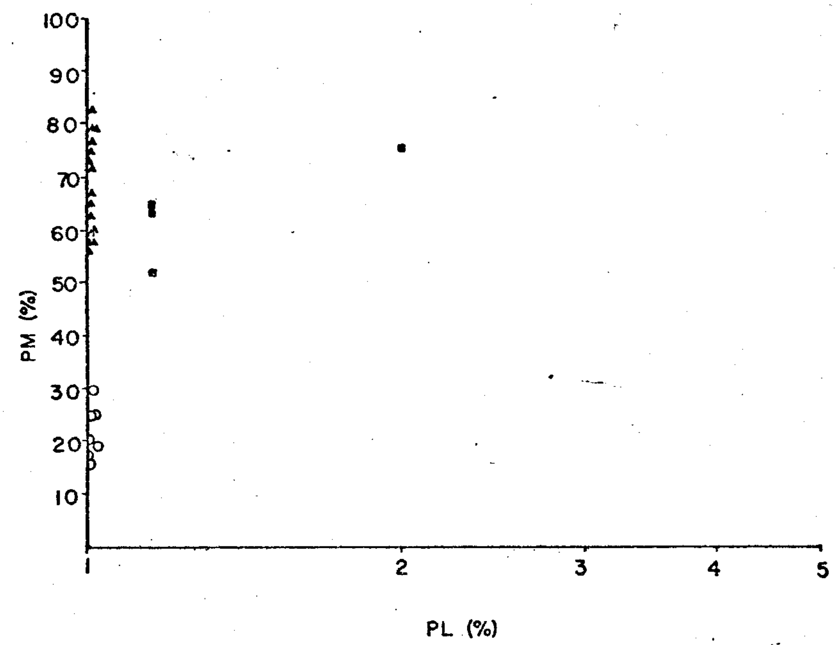

Figure 16: Plot of PL vs. PM for the Pleistocene outwash sediments.

antly gravel size distribution. The dominant gravel mode gives the sample its leptokurtic value. Sediments with low values of kurtosis have both positive and negative skewness values. These sediments have either a fine or coarse tail to their distributions and have a wide range in grain sizes.

The high values of the standard deviation and the platykurtic values of kurtosis indicate that these sediments have experienced little effective sorting.

Rizzini (1968) advocates the use of C-Mo and PL-Pm diagrams to describe sediment size distributions. The coarsest one percentile of the distribution (C), is a measure of the competency of the depositional current. The major modal size of the sediment (Mo), represents the range of sediments deposited by the same process. In very coarse sediments C falls within the range of Mo. PL is the percentage of the size distribution which is less than 30 microns in size. PM is the percentage of the total sediment that is in the major mode and was deposited by the same process. This then is a measure of the sorting of the main mode. Plots of C vs. Mo and PL vs. PM are shown in Fig. 15 and 16 respectively.

The lack of appreciable clay size sediment within the gravel is expressed in the near zero values of PL. The low values of PM represent the poor sorting, as only one-third of the sediment sizes are included in the major mode. This is also an indication of the several sediment modes present in the gravel fraction. The degree of variation in the values of PM underlines the variability in the sorting of the different samples.

In Fig. 15 the variability of Mo is the result of the multiple modes present in the gravel. The values of Mo range from very coarse sand to pebble sizes. The value of the coarsest percentile (C) is fairly constant in the pebble size. The coarsest percentile can lie far from the major mode or within it. Sediments with C and Mo lying far apart represent sediments deposited by two different methods, saltation of the finer sizes and rolling or sliding of the coarsest percentile. Sediments whose values of C and Mo lie close together have been deposited by one method, that of rolling or sliding of pebbles along the bottom.

Paleocurrent Orientation

(a) Wentworth Gravels

The paleocurrent orientations measured from cross-bedding in the Wentworth gravels are shown in Fig. 17 and Table 7. The paleocurrent measurements were made at exposures where a full cross-section of the trough cross-bed could be seen. A three dimensional area was cut into the gravel at the axis of the cross-bed and a dip indicator and compass used to measure the orientation on the clearly recognizable laminae. There is a pronouned north to south trend, which follows the Pleistocene bedrock valley to Glen Williams. At Glen Williams the paleocurrent orientations diverge as the

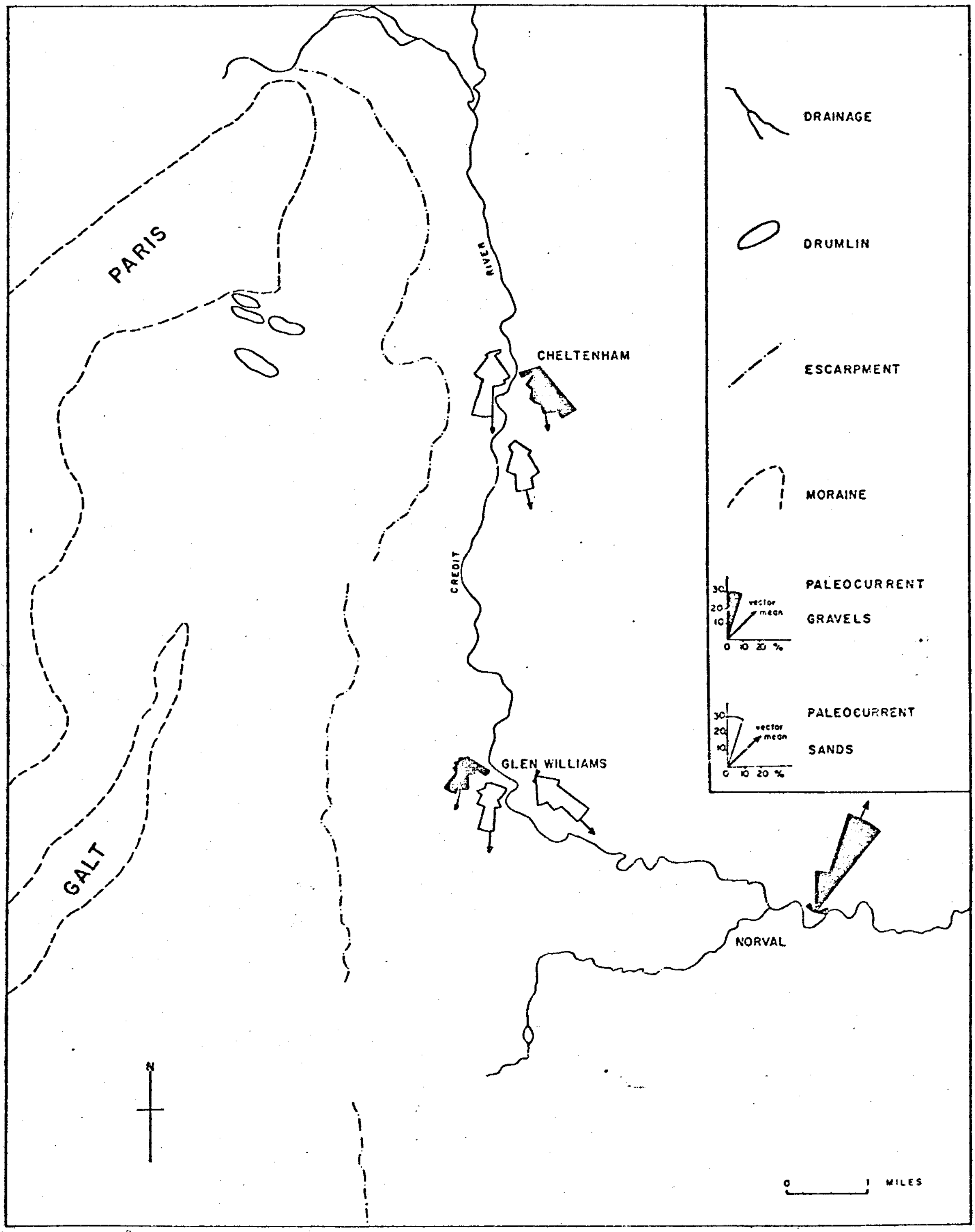

Figure 17: Paleocurrent orientations of the Wentworth Gravel and Sand.

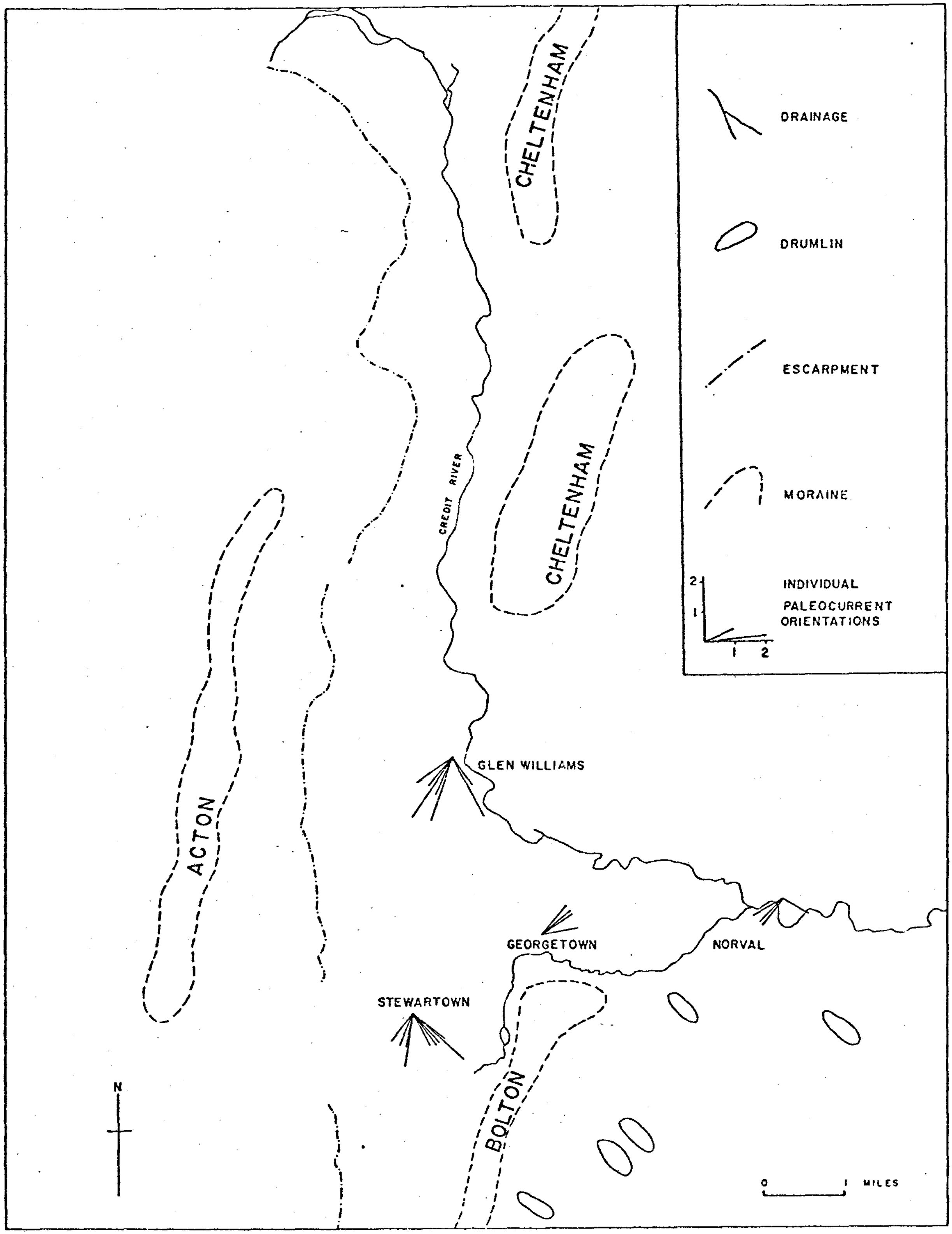

Figure 18: Paleocurrent orientations of the Halton Gravel.

bedrock valley widens. To the south and west of Glen Williams, the deposits of Wentworth Gravels are small and isolated. Erosion of these gravels by the Halton Ice Sheet, which advanced through this area, accounts for this lack of preservation.

At Norval, the paleoflow was reversed and flowed to the north.

(b) Halton Gravels

Outcrops of Halton Gravel are found in the Credit Valley to the south of Glen Williams. The paleocurrent direction is south past Georgetown to Stewartown (Fig. 18 and Table 7). At Stewartown, there is a gradual shift in the paleocurrent flow to the east. The gravel deposits west of Georgetown have been found in small bedrock valleys that are directed in a semi-circular fashion around Georgetown.

The barrier that earlier existed at Norval was no longer present and the drainage flowed to the south-southwest over the Peel Plain.

Interpretation

The similarity of structure between the Basal Red Gravel Facies and the Grey Gravel Facies leads to the conclusion that the grey gravels were also deposited as mid-channel bars and channel fill sequences.

The angular foreset structures represent mid-channel bars whose growth proceeds as sediment, swept over the top,

TABLE 7

Summary of directional current data from the Wentworth and Halton Gravels of the Credit Valley.

Location	G.V.M.	Vect. Mag.	Chi Square	P	Rayleigh P
Wentworth Gravels:					
Cheltenham	176.4	88.2	20.7	$>.001$	$<10^{-9}$
Bishop Pit	196.0	74.4	49.5	$>.001$	$<10^{-10}$
Glen Williams	93.8	96.5	13.2	$>.01$	$<10^{-3}$
Norval	27.4	94.3	35.2	$>.001$	$<10^{-5}$
Halton Gravels:					
Glen Williams	197.0	88.6	14.5	$>.001$	$<10^{-3}$
Stewartown	168.9	86.6	18.2	$>.001$	$<10^{-3}$
Georgetown	56.1	96.7	7.3	$>.05$	$<.01$
Norval	213.6	69.6	4.3	$<.10$	.10

avalanched down the lee slope. The bars were from 2 to 4 feet in height and received better sorted material than those mid-channel bars in the Basal Red Gravel Facies. Either the gravels had been transported for a large distance prior to deposition or else the depositional process was such as to cause deposition of only the coarser sizes.

The gravel bar deposits display the lowest proportions of sand and have pebble clusters on their foresets due to sediment avalanching down the lee slope. Sand and pebbles would be moving along the upper bar surface, the sand by suspension or saltation and the pebbles by rolling. The pebbles would avalanche down the steep lee slope of the bar and come to rest on the foreset. The sand fraction would be entrained by the flow and carried in various trajectories to the lower foreset areas (Jopling, 1967, p. 296). Sand deposited on the foreset slope, upstream from the point of flow reattachment in front of the bar face, would remain free from further current action and would be buried. Sand carried beyond the point of flow reattachment would continue to be entrained by the flow and would not be deposited.

The presence of channel lag deposits, imbrication of pebbles, and absence of large shale clasts indicates that the sediments have been subjected to reworking.

The grey gravels are for the most part polymodal and poorly sorted. This poor sorting is apparent in the high values of the standard deviation, the low values of PM, and the values of kurtosis.

It is interesting to note that there is a major mode in these sediments in the size range -0.5 to -2.0 phi. Russell (1968) has commented on the notable deficiency of this size range in most fluvial deposits. He accounts for this deficiency by pointing out that this size range has a hydrodynamic instability in that the sediment grains "are more easily entrained and more readily transported than grains of larger or smaller sizes". A combination of rapid deposition and abundant supply could account for their presence in the grey gravels.

The Grey Gravel Facies does not seem to have been built by a fluvial system aggrading as fast as that which deposited the Basal Red Gravel Facies. In fact, the erosion and incorporation of the red gravel by the finer grey gravels indicates that the drainage was cutting down into older deposits and to some extent was incorporating reworked older outwash deposits as part of its load.

(a) Wentworth Gravels

The Grey Wentworth Gravels were built as a broad deposit from the upper parts of the Credit Valley above Cheltenham down to Norval. These gravels represent mid-channel bar and channel fill deposits in the drainage channels of the Wentworth meltwater system.

The many gravel-filled channel deposits signify the shifting and aggrading nature of the river system which rapidly migrated over the broad floodplain. At Georgetown and Glen Williams this floodplain was 1 to 2 miles in width. At different periods the river ceased to aggrade and reworked the earlier deposits. At these times, deep erosive channels were cut in the grey gravels and even down to the lower red gravels. The thin cross-bedded sand beds that are infrequently found interbedded with the grey gravels were deposited during these periods of reworking. The sand was derived from the gravels during a reduction in the flow velocity of the river. The sand migrated down the bed of the river as dunes. The passage of the dune bedforms is recorded in the cross-bedded sand layers.

The paleocurrent data suggest that at one time a barrier was situated at Norval. This barrier may have been an ice sheet, perhaps a temporary readvance of the Wentworth Ice Sheet or else an early advance of the Halton Ice Sheet.

(b) Halton Gravels

The Halton Gravels are situated in the wide part of the Credit Valley south of Glen Williams. The channels have a complex distribution and are related to a semi-circular drainage pattern around the town of Georgetown.

Laing (personal communication) has interpreted these channels as marginal drainage channels running parallel to the front of an ice sheet, which occupied the area from Glen

Williams to Milton along the Escarpment. As the ice retreated, meltwaters gathered at the ice edge and flowed to the south and southeast. The area between Glen Williams, Georgetown, and Stewartown became ice free first, and the meltwaters, with their coarse glacial debris, gathered here. While the ice maintained its position atop the Bolton and Cheltenham Moraines, a drainage corridor opened along the present route of the Credit River.

These outwash channels contain the mid-channel bar deposits (large angular foreset beds) and the interfingering channel fill deposits (trough cross-laminated beds). Again, the periodic channeling indicates that the drainage system experienced periods of degrading.

The large number of channels and their erosive contacts further indicates that the drainage system was in a constant state of flux and that it migrated rapidly back and forth across the floodplain between Glen Williams, Georgetown and Norval.

3) Cross-stratified Sand Facies

Description

This facies is composed of red to buff colored, cross-stratified sands of Wentworth and Halton age. The sands are found dominantly above the Grey Gravel Facies. Both the upper and lower contacts between cross-stratified sand and grey gravel are erosional.

The Halton and Wentworth sands contain trough and tabular cross-stratification, and parallel lamination (Harms et al., 1963; Harms and Fahnestock, 1965). The trough cross-stratification is of large and small scale. The large scale trough cross-beds range from 1 to 6 feet in height. The larger cross-beds occur as isolated erosional scours. The smaller cross-beds occur as multiple sets which erode one another laterally and vertically.

Where the cross-bedded sands are interbedded with the grey gravels, the trough cross-stratification is found infilling channel shaped depressions in the gravels. The gravels have angular foresets and the previously described mid-channel bar structure, with gravel filled channels to either side. The top of the bar is eroded by several small, 2 to 3 foot deep, channel shaped scours. These scours run obliquely across the bar. The scour is filled by medium to coarse sand. In cross-section the sand has trough cross-bedding. In long section the sands have long foresets with asymptotic lower contacts with the underlying gravel. Directly beneath the sand lies a lag gravel.

The small scale trough cross-stratification is found in sets which are truncated by large scale trough and tabular cross-stratification. The amplitudes of the individual small scale trough sets are from 1 to 3 inches in height. The sets are found stacked one on top of another with a total height of 1 to 2 feet, and having little lateral extent, commonly

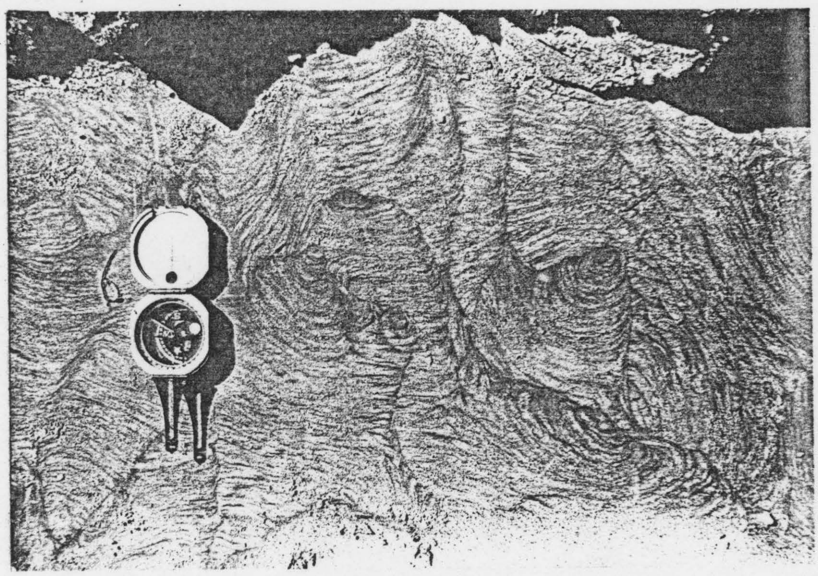

Figure 19: Plan view of rib and furrow structure.

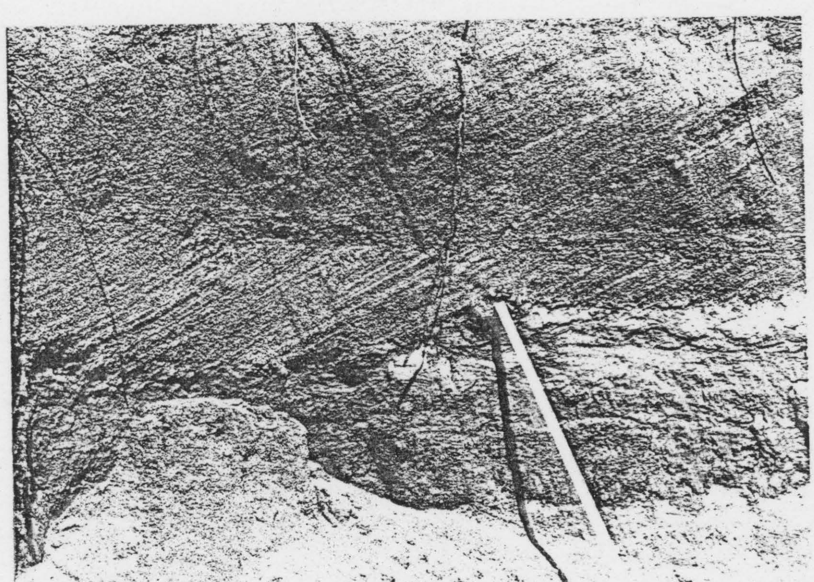

Figure 20: Tabular cross-stratification infilling scour holes in fine sand and silt.

less than 10 feet. In plan view the cross-lamination frequently shows a rib and furrow structure (Fig. 19). In cross-sections the cross-stratification shows no stoss side preservation or clay drapes. At several outcrops ripple drift cross-stratification is observed overlying tabular cross-stratification. The ripple drift cross-stratification is the Type A of Jopling and Walker (1968, p. 973), which has climbing sets of lee side laminae, with no preservation of stoss side laminae.

The tabular cross-stratification is found in sets 6 inches to 3 feet in height. The cross-stratified sets erode one another leaving the tops of the foresets planed off. The bottomset deposits of the tabular cross-stratification have angular, and less commonly, asymptotic contacts with the underlying deposits. At one outcrop a 1 foot high tabular set, having long shallow foresets and composed of fine sand, displayed small scale trough cross-stratified bottomset deposits. The orientation of the small scale cross-stratification showed a greater variance than the orientation of the tabular set but maintained the same general orientation. The foreset beds were largely eroded by an overlying tabular cross-bed set. This structure is identical to the Type 4 megaripple foreset deposit of Boersma (1967, p. 226), described from the Rhine fluvial system.

Commonly, tabular foreset deposits are found within small scours in the underlying bed (Fig. 20). The scours have a 7 to 8 inch maximum depth at their upstream end and gradually shallow out downstream. The scour holes are 2 to 4 feet in length and are cut in silts and fine sands. The foresets have angular lower contacts in the deeper part of the scour hole and asymptotic contacts in the shallower regions.

Tabular cross-stratification is found in sediments ranging from fine sand to granules. In many exposures the foresets of the cross-stratification are alternately sandy and pebbly. At one outcrop a single tabular cross-bed set had built contemporaneously with the cross-bed set below it. The upper set built forward faster than the lower set because it is seen to catch up with the lower set. At the capture point, the lower set ceased to build forward any more and the upper set continued to build forward depositing foreset laminae that were twice as high as they had been before.

Parallel lamination is rarely observed in these sands. Where it is observed, the sands are medium to coarse in size and frequently pebbly. The thickness of the parallel laminated deposits is from 1 to 6 inches and the lateral extent of the deposit is rarely more than 3 to 4 feet. The parallel lamination is deposited on top of eroded large scale cross-stratification.

Grain Size Analysis

The beds of the Wentworth and Halton Sands were samp-

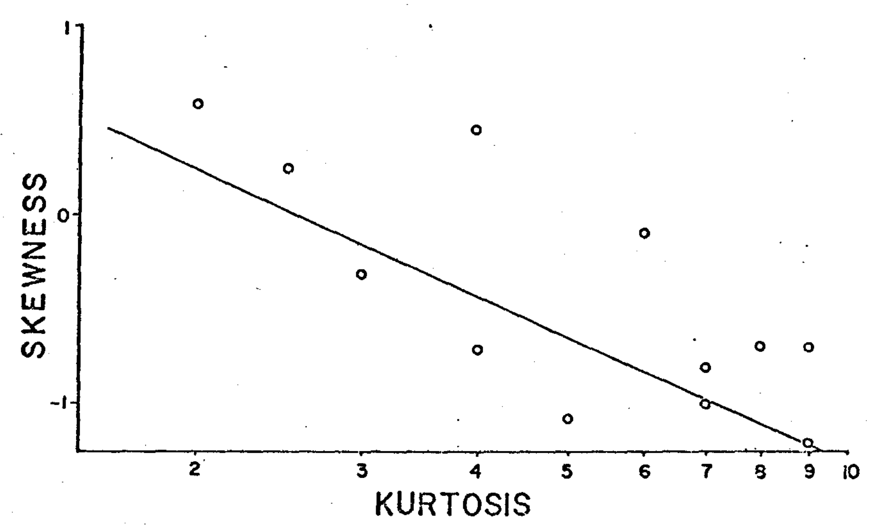

Figure 21: Skewness vs. kurtosis in the Cross-stratified Sand facies.

led and analyzed by sieving at $\frac{1}{4}$ phi intervals. The mean grain size of the sands was observed to range between 0.0 and 2.5 phi (coarse to medium-fine sand). The standard deviation ranged from 0.4 phi (well sorted) to 1.9 phi (poorly sorted). More than two-thirds of the samples were in the well to moderate sorting range however. A plot of the mean grain size versus the standard deviation is shown in Fig. 12. There is a noticeable trend of increased sorting with decreased mean grain size. This trend is the opposite to that trend found for the grey gravel facies. The same opposing trends in the gravel and sand fractions has been observed in fluvial sediments by Folk and Ward (1957, p. 17). The best sorting of the sediments is observed in the mean size range, 2.1 to 2.7 phi. This size minima and sorting maxima is identical to that found by Folk and Ward (1957).

The values of the skewness for these sands ranged from 0.6 to -1.25. The negative values of the skewness indicate that a dominant coarse tail is present in the sediment size distribution. Those sands having a negative skewness had isolated small pebbles along their foresets. Those sands with a positive skewness and a fine tail in their distribution are medium to coarse sands with no pebbles. The fine tail in their distribution is the result of the presence of fine sand within the deposit. The rippled fine sands have a slight negative skewness due to the presence of medium sand within the deposit.

The kurtosis values of the cross-stratified sand facies range from 0.0 (very platykurtic) to 20 (leptokurtic). More than two-thirds of the sands were grouped in the leptokurtic range indicating that they were well sorted. The skewness is plotted against the kurtosis in Fig. 21. A trend of increasingly positive skewness with decreasing kurtosis is observed in these sands.

The values of the sediment size of the main mode (Mo), the size of the coarsest one percentile (C), the percentage of the sediment included in the main mode (PM), and the amount of sediment having a size below 30 microns (PL) were calculated by the method outlined by Rizzini (1968). The main mode groups into the size range 0.0 to 2.0 phi. This suggests that the sediments were deposited by a single depositional process. However the value of the coarsest one percentile ranges from 1.0 to -4.0 phi. This in turn indicates that there was a great spread in the sizes of the coarsest fraction of the sediment size distribution. This agrees well with the observed negative skewness values in the size distributions. The plot of C versus Mo is shown in Fig. 15. The measurements indicate that the cross-bedded sands were deposited by a single process, probably saltation. The coarse tail of the distribution is the result of rolling and sliding of granules and pebbles along the bottom of the bed. These pebbles avalanched down the foreset slopes of the megaripples and were buried. Those sediments with high values of C were those which were gravelly sands. They contained isolated pebbles along their foresets.

At two outcrops of cross-bedded sands, the size of the coarse tail increased up through the succession of beds. It would then appear that at these outcrops the sands were deposited by an increasingly strong current.

That the currents which deposited these sands were fairly strong is again displayed in the negligible amounts of clay size sediment present in the cross-bedded sands (PL has a zero value, Fig. 16). The currents winnowed the deposits and sorted them moderately well. This sorting is shown in the high values of PM, which show that 80 percent of the sediment size distribution is grouped within the main mode. This agrees well with the observed values of the standard deviation.

Paleocurrent Orientation

(a) Wentworth Sands

The deposits of Wentworth Sands are present along the entire length of the lower Credit Valley. The sands have their greatest lateral extent at and below Glen Williams where the valley widens and where a large, wide alluvial plain existed. The paleocurrent orientations and vector magnitudes are listed in Table 8. Paleocurrent determinations were made on large and small trough and tabular cross-stratification.

TABLE 8

Summary of directional data for the Wentworth Sands.

Location	G.V.M.	Vect. Mag.	Chi Square	P
Cheltenham	180.0	90.2	78.5	<.001
Brickyard (south of Cheltenham)	170.0	92.0	39.4	<.001
Glen Williams (Bishop Pit)	186.0	77.9	51.0	<.001
Glen Williams (in town)	87.0	89.0	19.4	<.001
Georgetown	122.0	99.2	7.3	<.05

The grand vector means indicate that the paleocurrents were directed to the south past Cheltenham and Glen Williams. However on the east side of the Credit Valley at Glen Williams there is a thick (60 foot) accumulation of Wentworth Sands whose paleocurrent orientation is to the southeast (139 degrees). It is at this position that the bedrock valley widens and opens out to the east and south (Fig. 17). The Wentworth Sands spread out across the valley after having been funneled through the narrow, straight corridor of the upper Credit Valley, between Inglewood and Glen Williams. The outcrop of Wentworth Sands at Georgetown is very thin because of the heavy erosion in that area during the Halton advance and retreat. The few paleocurrent determinations obtained coincide with the shift in the paleocurrent orientation to the southeast.

(b) Halton Sands

The Halton Sands are found only at Georgetown, Norval and in isolated channel deposits west of Stewartown. At Norval the sands contain no current structures at all. The few good paleocurrent orientations measured at Georgetown and Stewartown are:

Georgetown - 55, 55, 280, 315, 320 degrees

Stewartown - 48, 75, 115 degrees

These few orientations can not be used to deduce any general paleocurrent orientation. However they do highlight the variability of paleocurrent orientation that the different isolated outwash channels had. This is to be expected as the drainage channels of the Halton Ice Sheet are themselves arranged in a complex semicircular geometry around Georgetown (Laing, personal communication).

Interpretation

The large scale cross-stratified sands were deposited by migrating dunes and bar-avalanche faces upon the river bed (Harms and Fahnstock, 1965, p. 103-105). Sand moving over the bed form by traction or saltation, avalanches down the lee slope and is buried there. The flow separates over the bed form and suspended sediment is transported to the bed in front of it. If suspended sediment is deposited in the zone of reverse flow behind the point of flow reattachment, the sediment can be carried up onto the foreset bed (Jopling, 1964). If the suspended sediment is transported to the bed downstream of the point of flow reattachment it can be incorporated into ripple deposits on the bottomset beds of the dune in the coflow zone (Boersma, 1967, p. 222). Such coflow ripple forms have been observed in the dune cross-stratified sands of the Credit Valley.

If the sand was deposited in a depression in the bed which was below the general level of erosion, the cross-lamination would be preserved (Harms and Fahnestock, 1965, p. 111). The infilling of scoured depressions on the river bed by migrating dunes has been observed by Frazier and Osanik (1961) and Harms et al. (1963), in modern fluvial environments. Allen (1968) points out that these migrating dunes infill the trough shaped scours and result in the formation of trough cross-stratification.

In the cross-stratified sands of the Credit Valley the trough cross-bed sets are seen to erode one another laterally and vertically. This association is indicative of the river bed area where there is active turbulence for constant scouring and where there is a bedload sediment supply to fill in the troughs.

The tabular cross-stratification is the result of the forward migration of bar avalanche forms along the river bed (Harms and Fahnestock, 1965; Allen, 1968; Collinson, 1970). Those ripple-like bedforms of large size (greater than dune size) have been called linguoid bars (Allen, 1968). Collinson (1970, p. 48) has excavated the deposits of some linguoid bars and has found the basic structure to be isolated tabular cross-bed sets with straight foresets having an angular bottom contact.

Small scale trough cross-stratification is the result of the migration of ripple forms over the bed surface (Allen, 1963, p. 107-108). The small thickness and extent of the ripple cross-lamination deposits in this facies indicates that the occurrence of ripple bed forms was limited. Ripples develop in a lower flow regime than dunes. Therefore the ripple cross-lamination must have been produced during a period of reduced flow velocity and/or in a sheltered position of the river floodplain.

Commonly, ripples are found on the backs of dunes or bars. Such an occurrence is suggestive of a transition in the regime of flow intermediate between that flow velocity and shear for either bed form. Collinson (1970, p. 55) has found ripples upon the tops of linguoid bars and interprets their existence as evidence of a high water stage. With a large depth of water moving over the bar surface, the shear stress acting on the bar top would be less than for a shallow flow. Hence the ripple form is generated above the bar and its migration supplies sediment to the foresets of the bar face. Therefore ripples are found not only in sheltered scours on the river floodplain, but also as thin sets intimately related to the linguoid bars. The latter ripple sets are generated by the separation eddy before the bar face and by the shear stress on the upper bar surface.

Parallel lamination can be either an upper or lower flow regime deposit. In this facies, the parallel laminatiion is composed of coarse pebbly sand and is deposited above eroded dune and tabular cross-stratification. The deposits are thin and have a very small lateral extent.

The strong current responsible for erosion of the dune tops may also have deposited the overlying parallel laminated sands. The flow velocity may have increased, and/or the flow depth may have decreased, causing local increases in the flow regime. An alternative to the lowering of the water level could be a local increase in the height of the river bedforms. When the bedforms attained a critical height, the depth of water above them would be small enough to produce an increase in the local flow velocity. Such a hydraulic setting would suit the limited thickness and lateral extent of the parallel laminated deposits.

Another sedimentary sequence, that of large trough-shaped scours cutting into the tabular cross-stratification, is indicative of a lowering of the water level. The depth of erosion is from 1 to 3 feet and must have been caused by stream erosion. The scour was later infilled by migrating dunes. If the water level dropped, the drainage would have to be localized around the bars rather than over them. Such channel scouring at low water stages has been observed on linguoid bars by Collinson (1970, p. 44). The drainage channels can cut through the bar face and construct a small delta lobe at its lee face.

(a) Wentworth Sand

These sands are deposited in a narrow valley from Cheltenham to Georgetown. The paleocurrents indicate that the drainage was confined to a narrow channel until it reached Glen Williams. Here the drainage spread out over the coarse floodplain built up by the Wentworth Gravels.

The drainage system carried a smaller and finer load than that which deposited the Wentworth Gravels. The incising of the gravels by sand filled scours indicates that the river system was actively eroding the earlier deposits for some periods of its history. The reworked gravels were a source of some of the river's load.

The numerous river channels again indicate that the river system constantly shifted over the floodplain. This lateral migration helped spread the outwash across the valley. Although the river deposits are better sorted than the gravels, which they frequently incised, the river system was aggrading as is seen by the thick sedimentary record left by the Wentworth Sand.

(b) Halton Sand

These deposits are thin and without much continuous lateral extent. The semicircular drainage system around Georgetown was still in effect at the time of deposition of the sands. However the ice had retreated enough so that the drainage flowed south east past Norval and into the Peel Pond. The limited extent of these deposits means that there was reduced flow at this time. The drainage was flowing in smaller channels across a large alluvial plain constructed by the earlier outwash deposits. The rivers actively eroded the earlier Halton Gravels and were not aggrading and building up the flood plain as had happened in the earlier outwash history.

4) Coarsening-Upwards Facies

Description

This facies is composed of the coarsening-upwards sequences that are observed in the Wentworth and Halton sands of the Credit Valley. The coarsening-upwards sequences are interbedded with cross-bedded, medium to coarse sands and their top and bottom contacts are erosional. The base of the sequence is sharp and lies upon cross-stratified coarse sands or upon an earlier coarsening-upwards sequence deposit. The sequences most often start with clay and grade upwards into silt, fine sand, medium sand, and coarse sand and gravel. Where the basal clay is not present, silt or fine sand initiates the coarsening-upwards sequences.

The cross-bedded, coarse sands, upon which the sequences rest have small scale trough cross-lamination developed in the upper 2 to 3 inches of sediment. Upon this coarse sand lies a deposit of massive to parallel laminated clay 1 to 2 inches in thickness (Fig. 22a). The clay mantles the eroded upper surface of the cross-stratified, coarse sand, and acts as an impermeable boundary to rising groundwater. The groundwater gathered at this interface is saturated with carbonate from the surrounding Paleozoic rocks, and as a result the upper 1 to 2 inches of the coarse sand is often completely cemented by calcite overgrowths.

The clay grades upwards into silt (Fig. 22A). Irregular, wavy laminations are found within the silts together with anomolous, sandy, small scale trough cross-lamination.

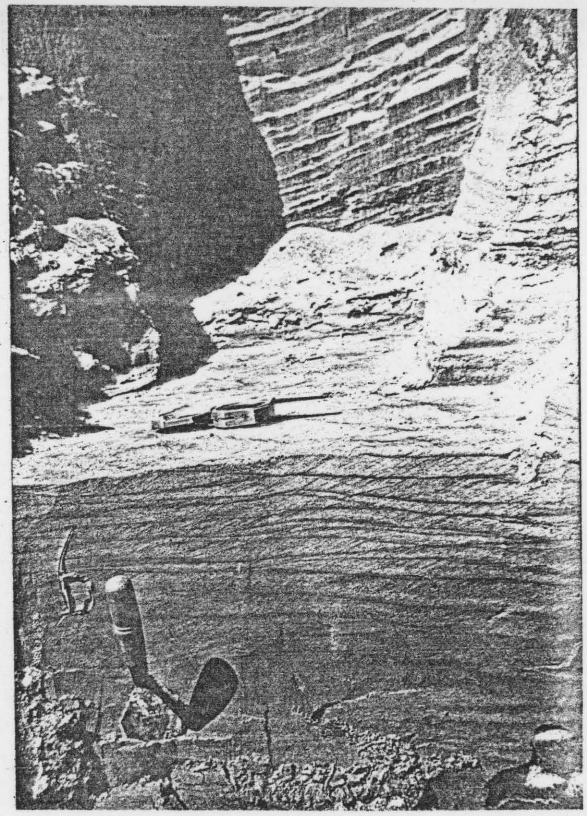

Figure 22b: Clay, silt and fine rippled sand unit of Coarsening-upwards Sequence - rippled fine sand unit with trough cross-stratification above, in background.

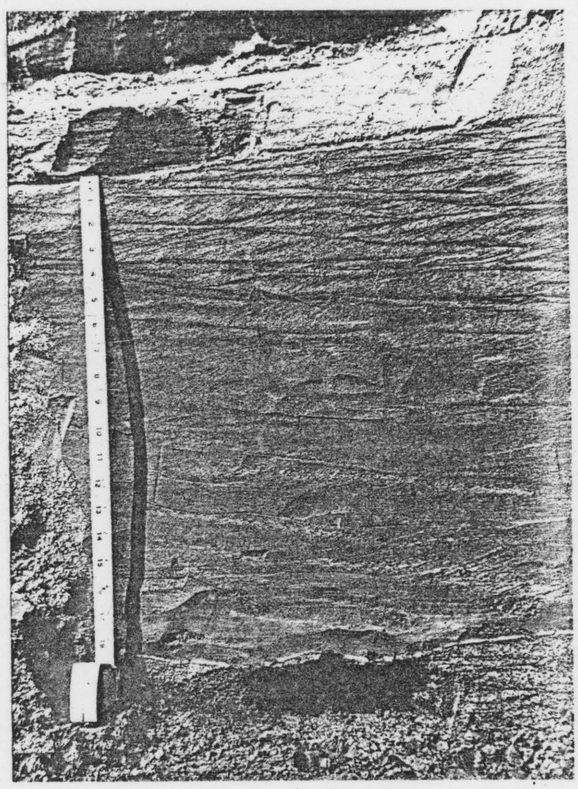

Figure 22a: Base of the Coarsening-upwards Sequence on rippled medium sand.

These small scale trough cross-laminations have clean, fine sand deposited on their lee sides in varying amounts (Fig. 22a). The sand accumulates along the middle and lower foreset while silt is deposited on the bottomset. A thin topset deposit of silt is preserved over the stoss side of the cross-stratification.

The silty lee and stoss side laminations and the silty horizontal to wavy laminations above and below the small scale cross-stratification suggests a high rate of sedimentation from suspension. If the sediment was not deposited so fast, the current would have swept the silt from the exposed stoss side of the cross-stratification. The structures, with their varying amounts of fine sand deposition, suggest the presence of irregular, ephemeral currents carrying coarser material to be deposited. The position of the fine sand on the foresets indicates that it was deposited here after being rolled along the top of the sedimentary structure then avalanched down its lee face. The small scale cross-stratification, $1\frac{1}{2}$ to $2\frac{1}{4}$ inches in height, is in turn covered by wavy to parallel laminated silts.

The silt grades upwards into fine sand, where the small scale trough cross-lamination is better developed. The cross-stratification has little or no stoss side development. The amplitudes of the sedimentary structures range from 1 to $1\frac{1}{2}$ inches. The sets are separated clearly by horizontal layers of dark clay. This clay layer is the result of clay de-

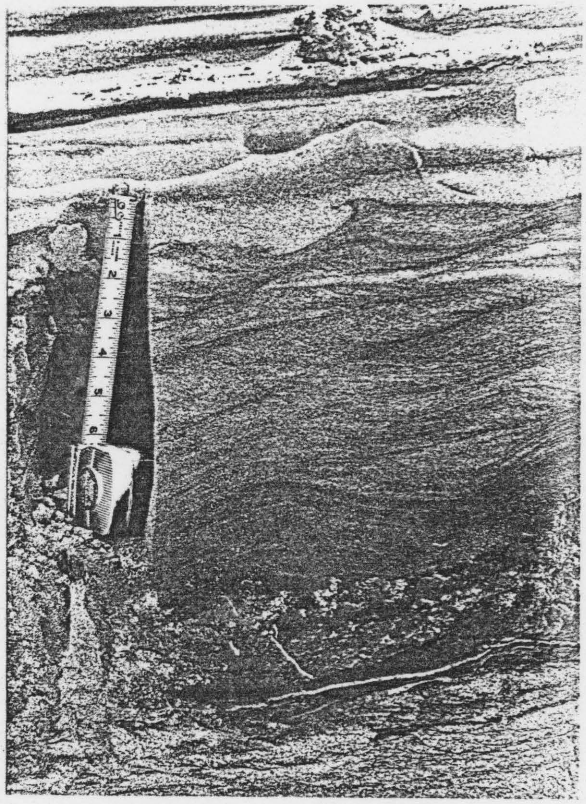

Figure 23a: Ripple drift cross-lamination in the coarsening-upwards sequence - Sinusoidal, Type A, and convolute structure in the ripple drift lamination.

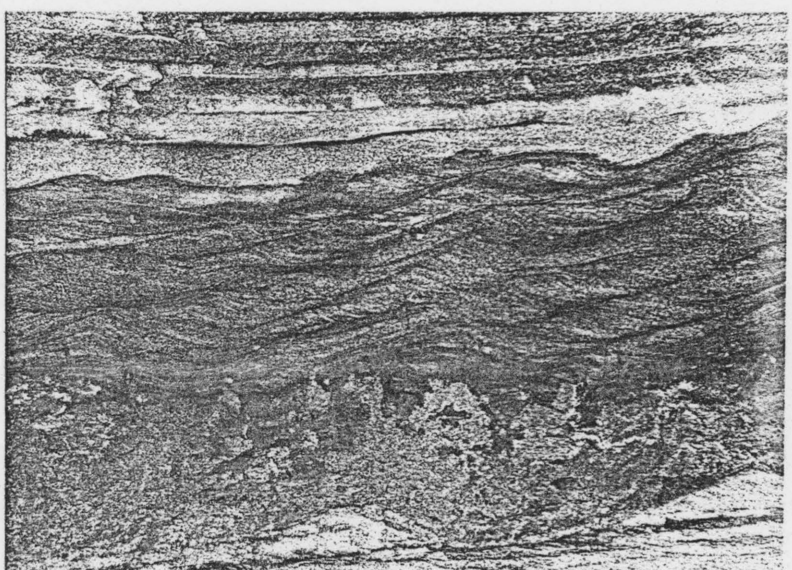

Figure 23b: Ripple drift cross-lamination in the coarsening-upwards sequence - Type A and B ripple drift.

postion along the bottomsets of each cross-lamination form. As succeeding sandy foresets migrate forward over earlier bottomset deposits, the clay is covered by the fine sands.

At several outcrops of coarsening-upwards sequences there is well developed ripple drift cross-lamination in the silt and fine sand. The silts have sinusoidal ripple lamination (Fig. 23a,b). The fine sand above the silts has A type ripple drift covered by B type ripple cross-lamination (Jopling and Walker, 1968, p. 973). Ripple drift has been shown to occur when the rate of deposition from suspension exceeds the rate at which sediment is moved by bed load (Walker, 1969, p. 388). The sinusoidal ripple laminations imply rapid sedimentation from suspension of cohesive silt. The change from sinusoidal to A type ripple drift indicates an increase in the rate of tractional movement during deposition. The change to B type ripple drift marks the renewed importance of suspension rather than traction in sedimentation of the bed. Convolute lamination is found above the ripple drift and further indicated high rates of deposition.

Therefore the lower silt and fine sand parts of the coarsening-upwards sequences appear to be the result of rapid deposition predominently by suspension but also partly by traction. The ripple drift structures and the convolute bedding are indicative of high rates of sedimentation.

The top of the rippled fine sand is eroded by a sand filled channel, which has trough cross-bedding within it.

One complete channel was measured to be 3 feet deep and 10 feet wide. Deposited above the channel were sets of large scale tabular cross-bedded, medium sands. The height of a cross-bedded set is a maximum of 1 foot. Where the channel has not cut into the rippled fine sand, these tabular cross-bedded sands sit discordantly upon the small scale cross-stratified fine sands.

The bottomsets of the tabular cross-beds have both angular and asymptotic contacts with the underlying deposits. Pebbles are frequently found on the lower foresets of these cross-bedded medium sands.

Above the large scale cross-stratified medium sands, the coarsening-upwards sequences are truncated by large gravel or coarse sand channels. In some outcrops, the sequences are followed by deposition of a new coarsening upwards sequence.

Plant remains were found in several of the coarsening-upwards sequences. The plant material was found in the lower clay unit as washed in debris. Plant remains were found in an upright position at the top of the fine sand unit. Where this occurred, no sand filled channel cut into the fine sand unit. This represents a halt in deposition and in the formation of an ephemeral pond. This pond was later filled in by large scale cross-stratification.

From field observations these coarsening-upwards sequences were seen to be filling depressions on a coarse sand or gravel erosion surface. The depressions were channel shaped and had a maximum width of 20 feet. At one outcrop however, the coarsening-upwards sequence filled a depression which could be traced for 200 feet perpendicular to the flow direction.

Grain Size Analysis

The sediments constituting the coarsening-upwards sequences were analysed by sieving and pipette techniques. The only variation in the trend of coarsening upwards is found in the megarippled medium sands.

The mean grain size of the sediments constituting the coarsening-upwards sequences ranged from 1.5 to 5.5 phi. The mean size of the megarippled sand, the rippled fine sand and the clays and silts, fell between 1.25 to 2.0 phi, 2.0 to 3.5 phi, and 4.0 to 5.5 phi respectively. The changes in the sorting up through a coarsening-upwards sequence are from poorly sorted (standard deviation, 2) in the basal clays, to moderately or well sorted (standard deviation, 0.4 to 1.0) in the small scale cross-stratified fine sand and the tabular cross-stratified medium sand.

The mean size is plotted versus the sorting in Fig. 12. The initial deposits of the coarsening-upwards sequence display a trend to increased sorting with increased mean size. This mirrors the decrease in fines upward through the sequence. With the development of large scale cross-lamination, coarser sediments are being deposited as avalanche deposits. This coarse tail of sediment, increases the value of the standard deviation and lowers the degree of sorting. Therefore the cross-bedded sands have a trend to increased sorting with decreased mean size. A similar V-shaped trend has been observed by many authors (Folk, 1964, p. 84).

The value of the skewness for these sediments ranges from -2.0 to 2.5. Where subsampling of the coarsening-upwards sequence was carried out, the fine sand was found to have the higher skewness values (2.0 to 2.5). The large scale tabular cross-stratified sands have a negative skewness. The negative skewness indicates the presence of a coarse tail in the distribution. Observations of pebbles along some of the cross-stratification foresets explains these skewness results. The high positive values of the skewness of the fine sand is indicative of a fine tail in the size distribution. This fine sediment is present as the horizontal clay and silt layers separating the small scale cross-stratified sets.

The kurtosis values of these sediments ranges from 0 to 70. Again the fine sands have the highest, leptokurtic values of the kurtosis which indicates good sorting. Some of the fine sands, the clays and silts and the medium sands have low kurtosis values. A plot of skewness versus kurtosis (Fig. 24) displays a trend of decreasing kurtosis with decreasing skewness until a value of zero is reached for both the skewness and kurtosis. The addition of coarser sediment to the

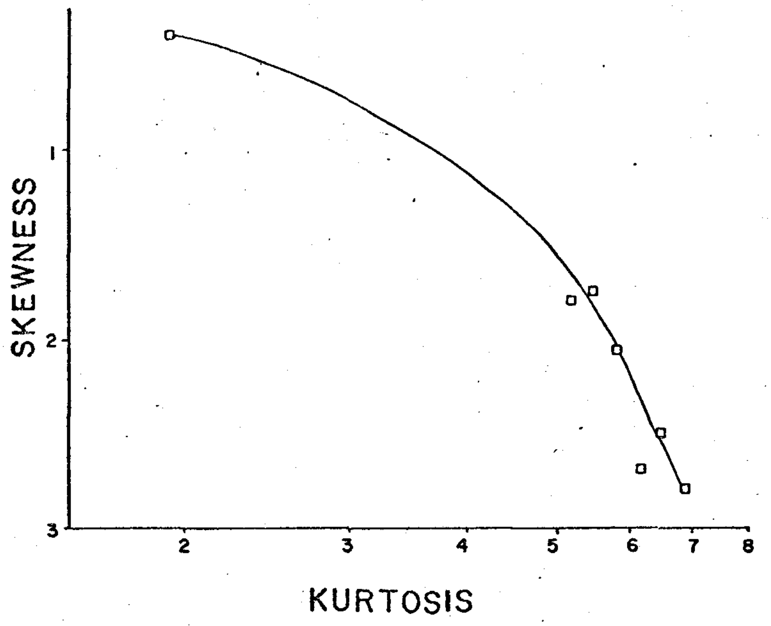

Figure 24: Plot of skewness vs. kurtosis in the Coarsening Upwards Facies.

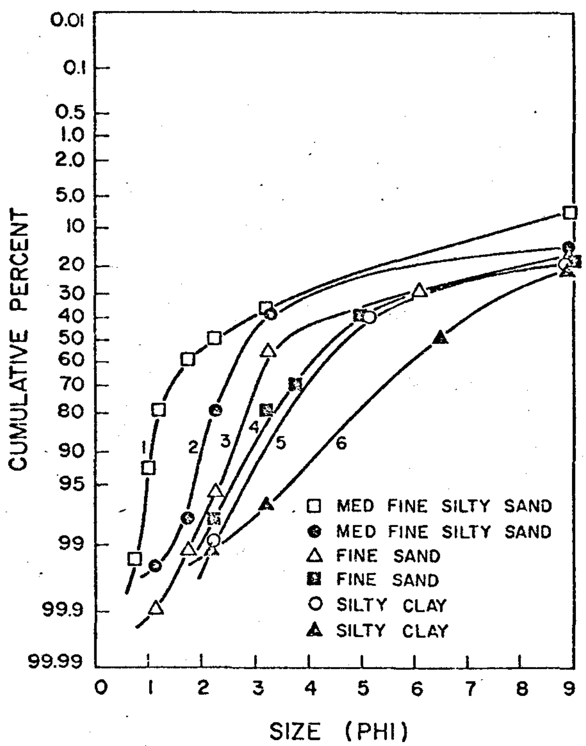

Figure 25: Sediment size distribution in the sediments of the Ardeche River.

fine sands reduces the fine tail of their size distribution. This lowering of the skewness is accompanied by a reduction in the kurtosis values since the sorting becomes poorer as more sediment classes are added to the size distribution. When a sediment has an equal mixture of fine and medium sand, the distribution has no tail (skewness is zero) and the distribution has a wide range of size classes (kurtosis is zero). With continued addition of coarse particles and the removal of finer particles, the medium sands develop a coarse tail in their size distribution (negative skewness values). The sorting action of the current is now increased as is the kurtosis.

The coarest one percentile of the size distribution (C) and the size of the major sediment mode (Mo) and values of PM and PL can be determined in the method outlined by Rizzini (1968). On a plot of Mo versus C (Fig. 15) three major fields may be delineated. The clays and silts show a spread in their modal size indicating the coarsening nature of these sequences. The coarsest percentile is constant since the current is not increasing its competency. The lack of a change in the current strength suggests that the silts and clays were deposited from suspension in a flow system where the sediment supply was gradually getting coarser.

The fine sands have a narrow range of modal sizes, 2.25 to 3.75 phi. Most of the sediment was deposited by the same process, that of suspension. The variation in the coarest percentile is due to addition of coarser sediment by another process, saltation. The greatest variation in the value of C came in that sample collected from an anomolously coarse small scale cross-stratified set in the silt unit. The fine sands were then deposited mainly by suspension with minor amounts of sediments supplied by saltation, when a current carrying coarser sediment particles acted in the depositional area.

The medium sands also have a narrow range in the modal grain size, 1.5 to 2.0 phi. This grouping is due to the sedimentation being primarily by the process of saltation. The deviation in the coarsest percentile is again marked. This is due to pebbles and granules being rolled over the bed forms and deposited by avalanche down the lee slope. The pebbly nature of the foresets was very noticeable in the field. It is also of interest to point out that the increase in the value of the coarsest percentile sometimes occurs as samples are taken from higher up in the coarsening-upwards sequence. This is a further indication of the coarsening nature of the sediments.

A plot of PL versus PM is shown in Fig. 16. The PM values are a measure of the sorting of the major mode. The medium sands and the fine sands display good sorting with high PM values. The clays and silts have slightly poorer sorting with high PM values. The clays and silts have not been subject to complete winnowing since their PL values are above the zero value.

From the sediment size analyses, the sediments of the coarsening-upwards sequences can be seen to become coarser (mean, Mo, C increases), and to become better sorted (lower standard deviation, increasing kurtosis, increasing PM), upwards through the sequences. The increase in size and sorting is a result of the action of a current which is becoming stronger and which is receiving progressively coarser sediments. From the C-Mo diagram (Fig. 15) a change in the means of sedimentation, from suspension to saltation then rolling, occurs upwards through the sequences.

Paleocurrent Orientation

The paleocurrent orientations at different levels in a coarsening-upwards sequence were measured. The silt and fine sand unit was sampled at 10 equally spaced levels, 2 inches apart, to determine any variance in the paleoflow direction in the small scale cross-stratified section. The vector means (θm) and the vector magnitudes (L) of the paleocurrent orientations at these different levels are listed in Table 9.

The Tukey Chi-Square test was used in addition to the Rayleigh test to determine the significance of the orientation results. Tukey's Chi-Square indicates a distinct departure from random orientation at the 95% level for all the measurements. The probability values determined by the Ray-

TABLE 9

Preferred Ripple Orientation at 11 levels up through a Coarsening-Upwards Sequence

Sample No.	θm (degrees)	L%	Tukey Chi Square	P	Rayleigh P	
8A ripple	221.5	98.0	19.16	.001	$<10^{-4}$	μ_1
8B 1	200.0	96.8	10.75	.001	$<10^{-3}$	μ_2
8B 2	201.0	95.5	8.87	.01	$<10^{-2}$	μ_3
8B 3	200.0	97.7	7.06	>.05	$<10^{-3}$	μ_4
8B 4	202.0	95.0	8.88	.0	$<10^{-2}$	μ_5
8B 5	195.0	97.3	12.40	.01	$<10^{-3}$	μ_6
8B 6	185.0	96.1	10.20	.01	$<10^{-3}$	μ_7
8B 7	184.0	98.3	16.00	.001	$<10^{-4}$	μ_8
8B 8	184.0	97.9	11.20	.01	$<10^{-3}$	μ_9
8B 9	202.0	94.9	8.80	.02	$<10^{-2}$	μ_{10}
8B 10	178.0	99.7	8.00	.02	$<10^{-3}$	μ_{11}

TABLE 10

Test of Variance between Paleocurrent Orientations at different levels in a Rippled Silt and Sand Unit

Hypothesis:

Vector Mean	t	n	t =0.05	accept	reject
$\mu_1 = \mu_2$	4.000	14	2.145		x
$\mu_2 = \mu_3$	-0.103	9	2.262	x	
$\mu_3 = \mu_4$	0.100	7	2.365	x	
$\mu_4 = \mu_5$	-0.100	7	2.365	x	
$\mu_5 = \mu_6$	0.885	10	2.228	x	
$\mu_6 = \mu_7$	1.330	11	2.201	x	
$\mu_7 = \mu_8$	0.143	12	2.179	x	
$\mu_8 = \mu_9$	0.000	12	2.179	x	
$\mu_9 = \mu_{10}$	-1.970	9	2.262	x	
$\mu_{10} = \mu_{11}$	2.500	7	2.365		x

leigh test lend further evidence to support the preferred orientation of the ripple orientations.

To test whether there were significant differences between the vector means of the ripple orientations at the different levels in the coarsening-upwards sequence, the "t" test (Dixon and Massey, 1957, p. 121-122) was applied. The results of the "t" test are given in Table 10. The level of significance that was taken was 95%.

Statistically significant differences in the vector means were found in only two parts of the sampled sequence. The small scale cross-stratified coarse sand lying immediately below the clay base of the coarsening-upwards sequence had a statistically significantly different orientation from the first layer of cross-stratification within the sequence. The uppermost cross-stratified layer in the coarsening-upwards sequence was also statistically significantly different in its paleocurrent orientation than the layers below it.

The small scale cross-stratified coarse sand deposited below the coarsening-upwards sequence was deposited by an earlier flow system. Cross-bedded sands which underly the coarsening-upwards sequences frequently have small scale cross-lamination on their upper surface. This cross-stratification is interpreted to be a falling water stage feature, constructed as the earlier flow system declined in activity and strength.

The difference in the paleocurrent orientation of the uppermost fine sand is not large. An increase in the turbulence and complexity of the currents acting at the site of deposition might cause a slight change in the local flow direction. The approach of a channel with its stronger currents, as is recorded by the trough cross-bedded channels, must have had considerable effect on the previous depositional conditions and could cause a slight paleocurrent change.

The strength of the vector magnitudes and the close relation of the vector means that are found in the small scale cross-stratified parts of coarsening-upwards sequences are the result of the action of a continuous, unidirectional current. A fluvial channel would be one environment for such a paleocurrent stability.

Though the cross-stratification below the coarsening-upwards sequence is significantly different in orientation from that in the coarsening-upwards sequence, there still exists a general similarity in flow direction between the two flow systems. This similarity of paleocurrent direction is a marked characteristic at all outcrops. In view of the channel shaped lower contact of the basal clay part of the coarsening-upwards sequence, these sediments are interpreted to be deposited in older, unused fluvial channels. The trough cross-beds deposited above the small scale cross-stratified fine sands have a very similar paleocurrent orientation (210 degrees). The large scale tabular cross-stratified, medium sands deposited either above the trough cross-beds or directly upon the fine sand, have the same paleoflow orientation. This uniformity of paleoflow direction up through the coarsening-upwards sequence further supports the view of continued deposition of sediment within a fluvial channel.

Interpretation

To allow the discrete deposition of clay size sediments on the floodplain of a river which is capable of transporting coarse sand and gravel, the presence must be postulated on the floodplain of a sheltered depression which can receive fine sediment. Doeglas (1962, p. 175) described such a depression on the floodplain of modern braided rivers. The depression was a channel, cut off from the river flow by a natural levee.

The cut off channel was blocked at its upstream end while water and fine sediment still entered the cut-off channel at the lower end and deposited silty clay in the quiescent water (curves 5 and 6, Fig. 25). With a slight rise in the water level, water ran over the levee at the upstream end and deposited fine sand in the cut-off channel (curves 3 and 4, Fig. 25). Doeglas (1962, p. 175) states, "At the highest level of overflow in the secondary channel, medium to fine silty sand is deposited and large current ripple marks are formed", (curves 1 and 2, Fig. 25). As the water level drops, fine sands are again deposited followed by silty clay from the downstream end of the channel. Doeglas (1962) finds this sequence, clay-sandy silt - medium to fine sand - sandy silt - clay, repeated up to six times. Ripple marks are also found in the silty clay.

The sequence described by Doeglas (1962) has significant differences from the coarsening-upwards sequence described here. Firstly, the sequence of Doeglas coarsens upwards from clay to medium sand but then fines upwards to clay once more. Secondly, the cut-off channel sequence indicates deposition by a growing then waning current, whereas the deposits of the coarsening-upwards sequence indicate a continuous strengthening of the current. This increasing current strength is demonstrated by the increase in the mean grain size up through the deposit and by the change in sedimentary structures upwards, from ripples to megaripples or dunes. Lastly, since the initial deposits of Doeglas's sequence enter from the downstream end of the cutoff channel and later deposits enter from the upstream end, a difference in the paleocurrent orientation up through the deposit would be expected. No such difference between the silt and sand parts of the coarsening-upwards sequence are evident.

Coarsening-upwards sequences in the geologic history have been interpreted as a result of marine regressions and the advance of deltaic sediments. On a braided river floodplain a regression model would describe the building of a small delta into a pool of water. The expected sequence due to the approach of this delta would be clay and silt (bottomset deposits), rippled sand (prodelta sands), cross-bedded sand (delta foresets) then channels filled with trough cross-lamination (streams on delta). However in the coarsening-upwards sequence, the trough cross-laminated, sandy channels erode the rippled fine sand before the cross-bedded medium sands are deposited. That a river channel could scour the pool floor before the delta was deposited is unlikely. Also if a small delta were building into a pool of quiet water a greater variation in the ripple orientations would seem probable.

To account for only the very finest sediments being initially deposited upon a floodplain, a process of decantation of surface waters from a turbulent river is proposed. By this it is meant that a river experiencing a stage of high discharge, such as a flood, would over flow its banks and fill adjacent, unused channels on the floodplain. At an early stage of flood only the surface waters, which transport fine sediment, would flow into the lower depressions on the floodplain and deposit sediment. The coarse sediment is concentrated near the river bed and therefore can only be carried over the river banks at a higher stage of flooding. Consequently with increased overbank flooding, it can be expected that there will be an increase in the size of the sediment deposited.

The process of decantation would produce the increased mean size of sediment upwards in the sequence as more water overflowed the river banks. As the initial overbank flood waters entered the floodplain depressions, the velocity of flow of the water would decrease and sedimentation would occur rapidly. This rapid sedimentation is responsible for the ripple drift cross-lamination. The channel shaped depressions that the coarsening-upwards sequences occupy are deserted, unused channels that the river formerly occupied. Since the floodwaters entered these old channels, the flow was restricted and unidirectional paleocurrent orientations resulted. With increased overbank flow, the floodwaters would increase the discharge and velocity of flow in the older channels. This increase in velocity causes the increased sorting and the changes in the process of sedimentation, from suspension to saltation, up through the rippled fine sand unit of the coarsening-upwards sequence.

The channels with trough cross-bedded sands that erode the rippled fine sand unit record the breaching of the river bank and the entry of the river into the once abandoned, now reactivated channels. The following cross-bedded, medium sand deposits are the river bed load structures, the migrating dunes and bars. Because the river has breached its banks it supplies coarser sediment and greatly increases the flow velocity in the reactivated channel. Therefore changes occur in the mean size of the sediment deposited, the sedimentary structures (ripples to dunes) and in the processes of sedimentation (saltation to rolling of the sediment particles).

5) Varved Sediment Facies

Description

This facies is composed of varved clays, silts and fine sands. The sediments are found above an eroded surface of Wentworth gravel or sand. Commonly, the clays and silts are found within channel-shaped scours cut into gravels. These 6 to 8 foot deep channels appear to have been active up to the time of deposition of the clays because the channel bottoms are only partially filled with sands or gravels derived from the bed load movements. It is in these eroded channels, rather than on the flat eroded surface, that evidence of a gradational current change is found. This evidence consists of a 1 to 2 foot bed of rippled sand deposited above cross-bedded, coarse sand or fine gravel, which constituted the bed load of the channel. The rippled fine sand grades up into silt and clays, Fig. 26. The ripple structure is the ripple-drift cross-lamination type. The lamination is composed of climbing sets of lee side laminae, without stoss side laminae. This type of ripple-drift cross-lamination has been called Type A ripple-drift by Jopling and Walker (1968). Paleocurrent orientations from the ripples give a varying paleocurrent direction.

The clays and silts range from grey to blue in color. The clays are compact and very hard. Where the clays are water saturated they are viscous or plastic in nature. A typical varve sequence is shown in Fig. 27.

The varves are characterized more by color changes than by changes in grain size. The grain size in the varves ranges from coarse silt to clay. However dropstones are pre-

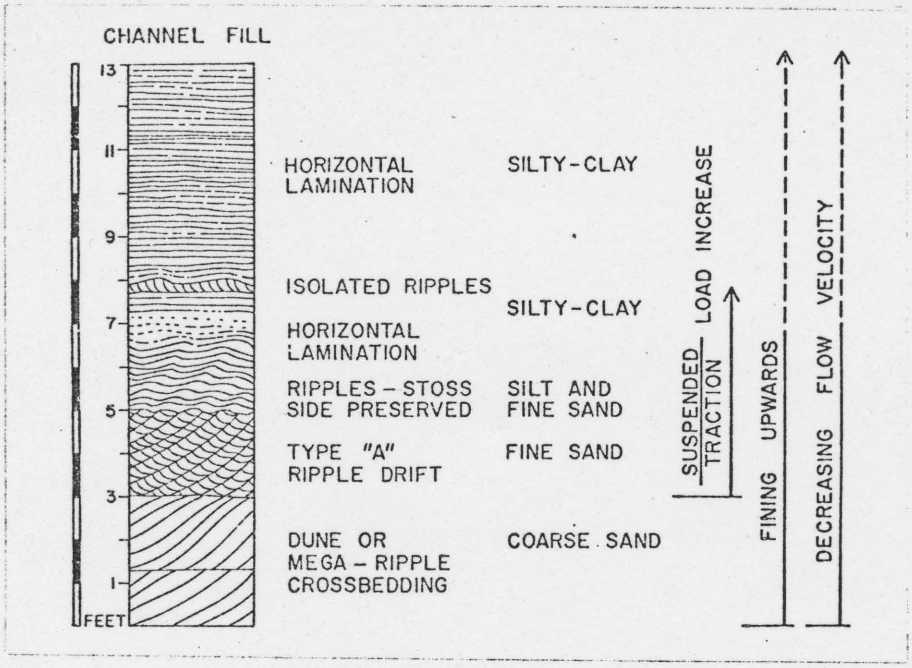

Figure 26: Fining upwards sequence in the Varved Sediment Facies.

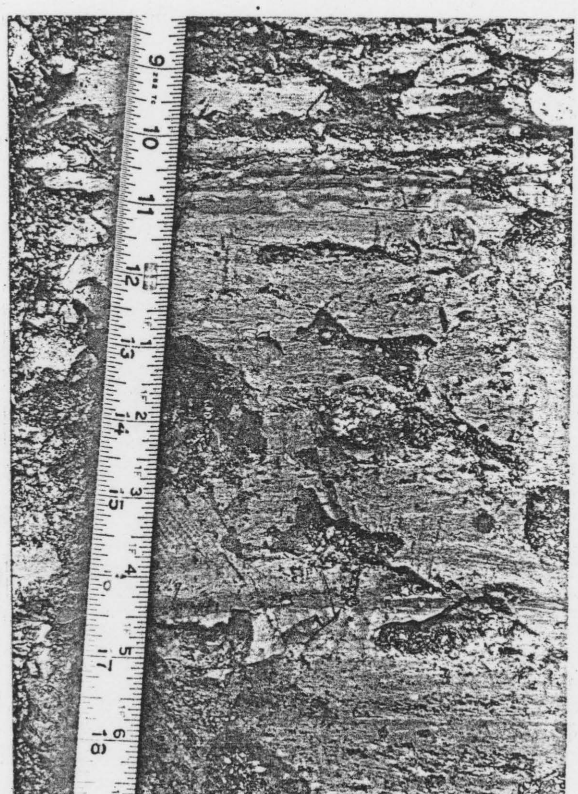

Figure 27: Typical varve sequence in varved clay.

dominant in the thicker grey varves.

The dark red varves are 3 to 4 cm. (1½ to 2 inches) in thickness whereas the grey varves are 25 to 30 cm. (10 to 12 inches) in thickness. All the varves have well developed flat lamination. The dropstones in the grey varves occur as layers of granules in the lower 8 to 10 cm. (3 to 4 inches) or the varves.

Thin layers of sand with a maximum thickness of 5 cm. are occasionally found among the clays and silts. At the Bishop Pit, these sand layers are contorted into tight folds near the top of the clay deposit.

Outcrops of varved clays, and silts were found at Glen Williams, on the western side of the valley southwest of Terra Cotta, and just south of Cheltenham on the east side of the valley. White (1968) has mapped stratified lacustrine clays and silts on both sides of the Credit Valley between Cheltenham and Inglewood. In all cases the lacustrine deposits lie below the Halton Till.

Interpretation

The presence of varved and parallel laminated sediments with dropstones within them indicates deposition in a glacio-lacustrine environment. The red colored varves represent deposition in a colder season when the melting of the ice is decreased and freezing of the surface waters occurs. With this reduction in meltwater drainage, a reduction in the amount of sedimentation also occurs and thinner varves are deposited.

In a warmer season there would be increased drainage due to ice melting. Therefore there would also be an increase in the amount of sedimentation. The initial layers of clay, which contain the layers of dropstones indicate the melting of frozen floating ice blocks which are laden with sediment. These dropstones settle to the bottom where the clay laminae are deposited over them. The larger dropstones found scattered through the grey varves record the melting of individual ice blocks.

The deposits of lacustrine clays, silts and fine sands suggest that the Credit Valley was at one time occupied by a proglacial lake. With the northwestward advance of the Halton Ice the lower end of the Credit Valley became ice dammed and a backup of drainage occurred. As the water backed up within the valley, the currents in the drainage channels decreased and sedimentation took place. This decline in the currents is indicated by the fining upwards sequence; cross-bedded coarse sand - ripple-drift cross-laminated fine sands - silt - clay (Fig. 26).

As the ice front advanced up the Credit Valley, the proglacial lake continued to exist at the ice front. Hence the lacustrine deposits are found along the length of the Credit Valley.

6) Polymictic Gravel Facies

Description

The coarse and angular polymictic gravels comprise this facies. These gravels are found in an isolated outcrop on the west side of the Credit Valley at Glen Williams (Fig. 5). The gravel contains clasts from all the local bedrock lithologies - shale, siltstone, sandstone, and dolomite. The clasts are large with a maximum diameter of 20 inches, and are very angular (Fig. 28). The relatively easily weathered shale of the Queenston Formation exhibits the same angularity as does the more resistant dolomite of the Lockport Formation.

The matrix is composed of a gravelly sand and constitutes up to 70 percent of the outcrop by volume. Because of the great amount of matrix, the sand and the gravel must have been deposited similtaneously. The angularity of the clasts and the very poor sorting of the gravel suggests that the sediment has not been reworked in a fluvial environment. The absence of sedimentary structures other than a few subhorizontal bands of cobbles within the gravel further indicates that the deposit has not undergone any fluvial transport.

The gravel deposit has been covered by dune cross-bedded sand. The influence of this fluvial activity above the gravel is demonstrated by a layer of lag gravel at the top of the polymict gravel. The top 1 to 2 feet has had all the

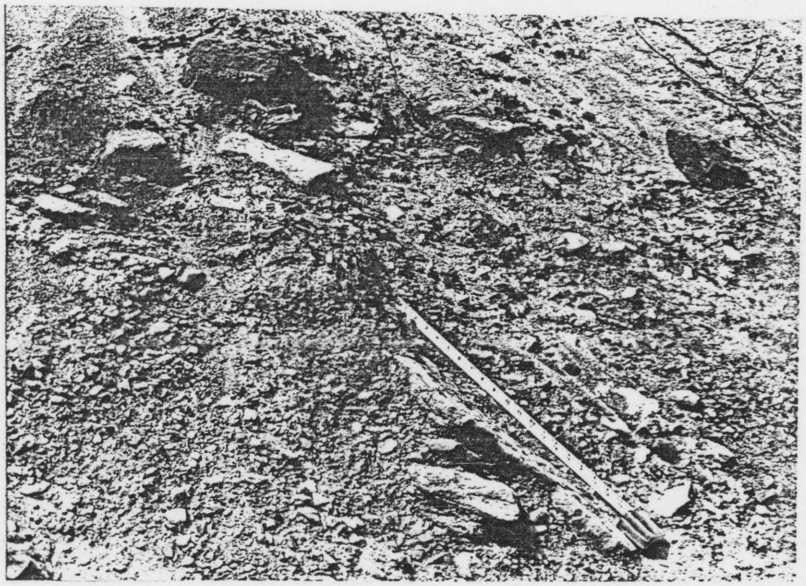

Figure 28: Polymictic gravels in the Bishop Pit.

fine sediment removed from it and the cobbles remain as a lag deposit.

The geometry of the deposit is that of a fan-shaped wedge that thins out in the direction of the river valley. At its distal edge the polymict gravels interfinger with the grey Wentworth Gravels. Along its exposed front the gravel deposit is seen to be thickest at its center and it thins out at the edges. The maximum exposed thickness is 9 to 10 feet and the width is 350 feet. The deposit has been subject to erosion at its edge and upper surface.

The apex of this fan-shaped deposit is up on the west side of the Credit Valley. Within one mile of the outcrop of polymict gravels, outcroppings of all the local Paleozoic rocks are present. The inferred apex of the gravel deposit lies very close to these bedrock outcrops.

Interpretation

From the wedge-shaped geometry, the coarse poorly sorted texture, and the lack of sedimentary structures, the polymictic gravel deposit is concluded to be a mass movement slide or slump deposit. The source of all the constituent clasts in the polymictic gravel are exposed within one mile upslope of the deposit. Large clasts of rock which had come only this short distance would be expected to be very angular in shape.

Lattman (1960, p. 279) describes similar wedge-shaped deposits of coarse debris and terms them colluvium deposits. These deposits are in part alluvium and in part angular fragments of rock. The colluvium is found at the foot of a slope and was transported by gravity sliding.

The colluvium is interpreted to be the result of a series of small slides or slumps of Paleozoic bedrock down into the river valley from the Escarpment. The matrix was incorporated from glacial debris on the valley sides or from eroded outwash sediments in the river valley.

7) "Delta" Sand Facies

Description

The Halton Sands at Norval are 15 to 20 feet in thickness. The fine to medium sands are well sorted and do not have any sedimentary structures developed. Within these sands there are one foot layers of parallel laminated clays and silts.

The massive sand deposits are found along the Credit Valley to the southeast of Norval. The sands are deposited as a tongue of sediment building down the river valley from Norval. To the north of Norval the Halton Sands are not found extensively but do display large, well developed, trough cross-stratification. At Georgetown these sands have deposited in eroded channels with large trough cross-bedding. No coarsening-upwards sequences were found inter-bedded with these sheet sands.

The Credit Valley bedrock channel lies below the present course of the Credit River for most of its extent. However at Georgetown the ancient bedrock channel cuts to the south along the present course of Rogers Creek. In Fig. 18 the paleocurrent direction of the Halton Gravels displays this flow pattern. During the retreat of the Halton Ice, the meltwaters cut the present channel to the north of Georgetown. Therefore the meltwaters would utilize both these channels during the retreat of the Halton Ice. The sheet sands are found just below the confluence of these two channels. The upper elevation of these sands is at 750 feet. Clay sediments of the Peel Pond have been mapped by White (1968) within two miles of this outwash tongue. The elevation of the glacial lake sediments is at 750 feet.

Interpretation

These sands have been interpreted to be deltaic sands building into the Peel Pond (Chapman and Putnam, 1966). Because of the lack of sedimentary structures and the typical foreset geometry of advancing deltas, these deposits would better be described as sand sheets. The basis of the past interpretation has been the geographic setting of the deposit. The deposit lies at the confluence of two drainage channels. Both the deltaic sands and the lake sediments have the same maximum elevation and the sands build out over the lake deposits.

The term delta has been applied in the past to mean a large river-transported deposit built out into a lake. The author agrees with this geographic interpretation. However in the strict sense of a delta being a sequence of sedimentary beds deposited as topsets, foresets and bottomsets, the author can not apply this interpretation. Rather these sands are interpreted as a sheet sand deposited at the junction of the Credit Valley drainage system and the Peel Pond.

THE BRAIDED RIVER MODEL

Chapter VI

Facies Relationships

From the stratigraphy of the Pleistocene deposits of the Credit Valley, it is concluded that the sedimentary facies are organized into two major fining upwards sedimentary sequences. The vertical and lateral facies relationships are summarized in Fig. 29.

The Wentworth fining upwards sequence includes the Basal Red Gravel facies up to the Varved Sediment facies. Each of the vertical contacts is erosional. The Polymictic Gravels and the Coarsening-Upwards facies are laterally interbedded with the Grey Gravel facies and the Cross-stratified Sand facies respectively.

Above the Varved Sediment facies the stratigraphy displays two differing trends. In the upper end of the valley the varved sediments grade upwards into glacial till (Halton Till). In the lower valley the sediments start a renewed fining upwards sequence (Halton) ranging from a deposit of the Grey Gravel facies up to the Cross-stratified Sand facies. Again the Coarsening-Upwards sequence is interbedded with the Cross-stratified Sand Facies. The Delta Sand facies is a down valley equivalent of the Cross-stratified Sand facies.

Large deposits of sediments in a fining upwards sequence

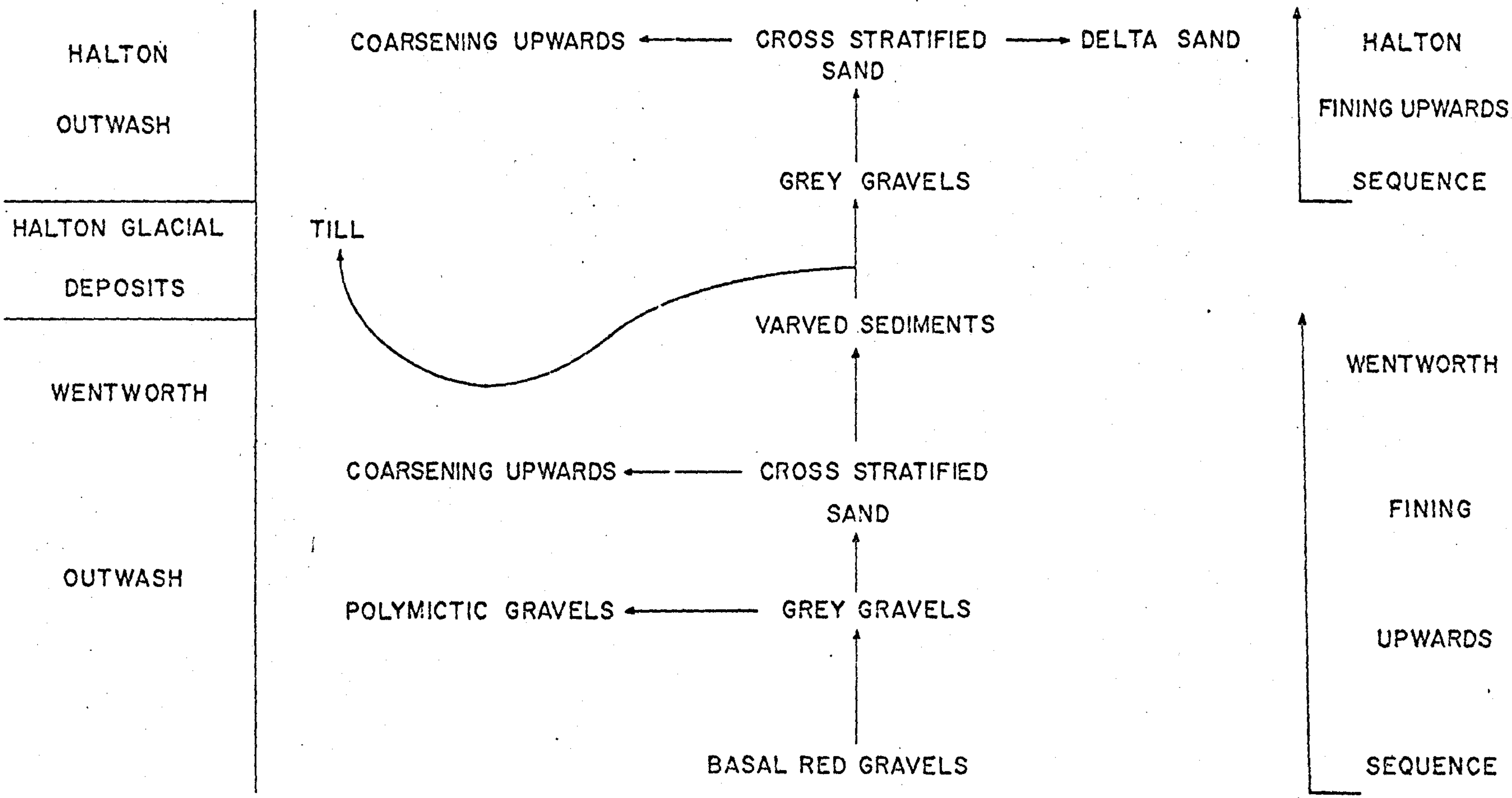

Figure 29: The vertical and lateral facies relationships in the sediments of the Credit Valley.

have been interpreted as the infilling of a valley by fluvial processes. The processes of such fluvial or valley fill models are well described in the literature (Happ et al., 1940; Sundborg, 1956; Leopold and Wolman, 1957; Visher, 1965). The two fining upwards facies sequences found in the Credit Valley are therefore concluded to be valley fill sediments deposited by a fluvial system.

Braided River Model

In the two cycles of fluvial valley fill, sequences of dissimilar deposits such as the grey gravels and the cross-stratified sands are found. Such dissimilar deposits are the result of sedimentation by one fluvial system.

The valley fill sequences, with the mid-channel braided bars and absence of sandy point bar deposits similar to those described by Harms et al., 1963, and McGowan and Garner, 1970, are unlike those described in the literature. Because the meandering river model does not appear to be appropriate, a model of braided river deposition is presented. There are three distinctive and extensive deposits in the fluvial valley fill sequences of the Credit Valley. From the generalization of Fig. 29, it appears that the fining upwards sequences can be essentially characterized by basal gravels, cross-stratified sands, and small coarsening-upwards sequences interbedded with the cross-stratified sands. Each of these deposits is interpreted as being the result of a different braiding process with-

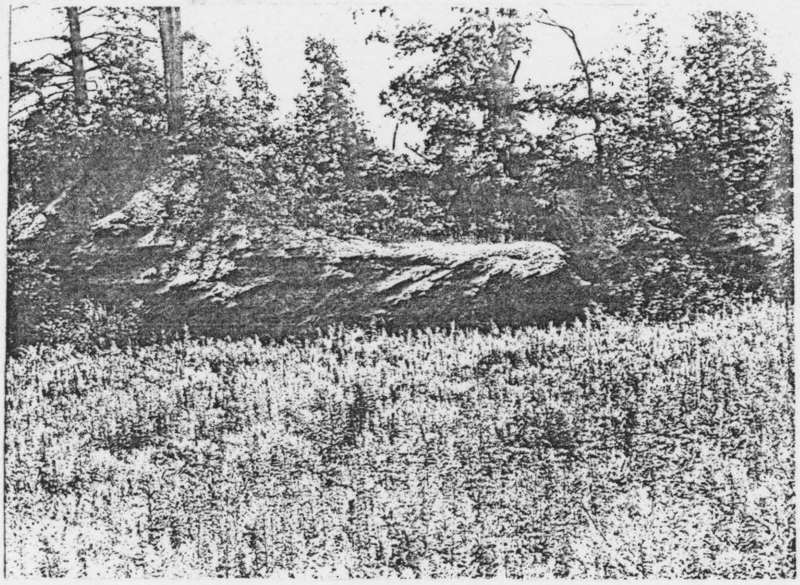

Figure 30: Mid-channel gravel bar.

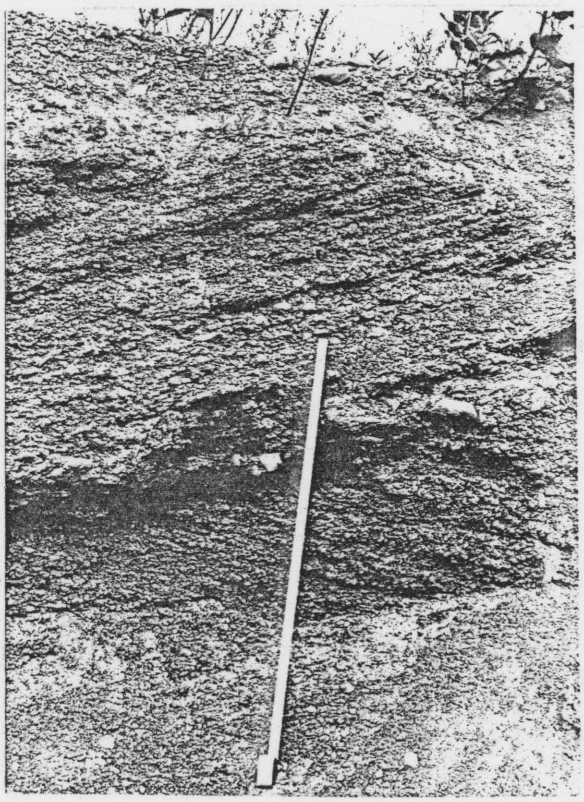

Figure 31: Two gravel filled channels lying unconformably upon one another.

in the braided river system.

Longitudinal Mid-Channel Braid Bar

The Basal Red Gravel and Grey Gravel facies occupy the lower-most stratigraphic position in the sequences of valley fill deposits. The characteristics of the gravel deposits are many. The dominant structures are gravel bar deposits 2 to 10 feet in height (Fig. 30) which laterally interfinger with gravel channel fill deposits to either side of the bar. The channel fill deposits have large festoon cross-stratification. The gravel bar deposits are coarser than the gravel channel deposits and both are poorly sorted. The bar deposits built upwards by accretion and built downstream by avalanche over the lee slope. The tops of the bar deposits are eroded by small channel shaped gullies which have an oblique orientation to the bar. The contact beneath the initial bar deposits is erosive. There are numerous cross-sections of gravel filled channels having lateral and vertical erosional contacts with one another (Fig. 31).

The presence of a mid-channel bar causes a river to anastomose or braid. Such river characteristics as the erodibility of the banks, rapid discharge variation, river slope increases, abundant load and local incompetence have been cited as reasons for the onset of mid-channel bar deposition and river braiding.

The processes leading to the development of mid-channel bars have been described by numerous authors (Hjulstrom, 1952; Leopold and Wolman, 1957; Fahnestock, 1959, 1963; Krigstrom, 1962; Ore, 1963, 1964; Coleman, 1969). From observations of natural braided streams and stream table experiments, Leopold and Wolman (1957, p. 43-45) pointed out that braid bars are initiated by deposition of the coarsest bed load sediments in the center of the river. This deposition of only the coarsest load is due to the local incompetence of the river.

The coarsest bed load can only be moved at high flow velocities. The mid-channel bars of most natural rivers are composed of gravel which is transported during flood discharges. The sedimentary load of these rivers varies from clay to gravel. At flood discharges all the sediment size fractions are transported. With a reduction in the flow velocity aggradation takes place by deposition of those sizes which the river is incompetent to transport. This hydraulic effect is documented by Hjulstrom (1935) and Shields (1936). For a river with a wide range of grain sizes only the coarsest would be deposited. Therefore the deposition of coarse gravel mid-channel bars is a selective aggradation (Krigstrom, 1962) of only the coarsest sediment sizes.

In the two gravel facies evidence of the selective aggradation and incompetence is seen in the coarser size of the bar deposits as compared to the gravel channel deposits. The initial gravel bar deposits trap finer sediments that are being transported by the river and incorporate them into the bar. This results in a polymodal size distribution and poor sorting. The gravels build upwards by vertical accretion in the center of the channel. As the bar builds upwards the depth of water above it decreases and the velocity of the flow over the bar increases. This increased velocity and shear stress cause sediment particles to be carried along the bar surface and not to be deposited until the sediment reaches the downstream end of the bar where the depth of water is increased and the flow velocity is reduced. The downstream end of the bar builds forward by avalanching of sand and pebbles. The gravel facies display this foreset geometry with avalanche pebble clusters on the foresets. As the bar builds in size vertically and horizontally, the stream width to either side of the bar decreases and the flow velocity increases. This causes bank erosion as well as channel deepening. In the gravel facies the channel fill deposits display numerous lateral and vertical erosional contacts. With the cutting downwards of the side channels, the surface water level may drop and the bar can emerge. During this falling water level stage or at a later highwater stage, when the water can again flood the bar surface, small channels or gullies can be eroded in the bar surface (Krigstrom, 1962; Ore, 1963, p. 3). These sand filled channel scours have been observed on the tops of the gravel bars in the Grey Gravel facies.

The floodplains of braided rivers have an abundance of stream channels upon their surface. Hjulstrom (1952) observed this characteristic on the Icelandic sandur plains where innumerable abandoned channels which were dry or partically water filled covered the floodplain. The rapid migration of the braided channels and the complexity of the channel plan was described by Doeglas (1962, p. 171) from the River Durance of France. From cross-sections through the floodplain deposits Doeglas describes the effect of the channel migration on the sedimentary deposits, "As the various channels form an intricate pattern, any section made through the river bed, even parallel to the average current direction will cut bars and channels. Accordingly the large macro-structures of deposits of braided rivers hardly show any continuous horizontal bedding. Channel structure, channel or festoon cross-lamination, cross-sections of filled channels, will nearly always be observed". The bar and channel deposits and the lack of horizontal lamination within the gravel facies of the Pleistocene deposits of the Credit Valley are analogous to this description of the modern deposits of the River Durance.

Another analog to the Credit Valley gravel deposit is the Lafayette Gravel of Kentucky. Potter (1955) found that the gravels had a predominance of large trough cross-stratification which could be as high as 20 feet. The poorly sorted gravels were deposited in numerous channels. These deposits were interpreted to be foresets of small deltas or bars which were migrating downstream within a channel. The interfingering cross-bedded relationships were described as due to stream channeling and deposition by a series of fast flowing, laterally migrating braided streams.

The essential characteristics of the Lafayette Gravel, the trough cross-stratification, the large bar or delta structures and the numerous gravel filled channels are again the characteristic of the Pleistocene gravels of the Credit Valley.

The Basal Red Gravel and Grey Gravel facies are therefore interpreted to be deposits of longitudinal mid-channel gravel bars and lateral stream channels in a braided river system.

Transverse Braid Bar

The Cross-stratified Sand facies is deposited above the Grey Gravel facies. It is differentiated from the gravel facies by its finer size and moderate to good sorting. The cross-stratification types include ripple, dune and tabular cross-lamination, as well as parallel lamination. The tabular and dune cross-stratification is the most abundant with lesser development of the ripple, ripple-drift and parallel lamination.

The tabular cross-stratification is the result of the migration of linguoid sand bars down the bed of the river. The smaller scale dunes and ripples which are also present within the river channel and even superimposed upon the linguoid bar form (Collinson, 1970, p. 38), produce trough and small scale trough cross-stratification respectively above and below the tabular cross-stratification. Therefore these different stratification types are found in many different sequences. Besides being found as thin isolated sets between tabular and dune cross-stratification, the ripple lamination is found as isolated beds in channel shaped depressions 1 to 2 feet deep.

At several outcrops the tabular cross-stratification is cut by channel shaped scours several feet in depth. The scour is infilled by large scale trough cross-stratification. This sequence of structures indicates a lowering of the water level and localizing of the drainage into channels which subsequently erode parts of the linguoid bars. The presence of parallel lamination above eroded megaripple or dune cross-stratification also points to a lowering of the water level during this period of sedimentation.

Tabular and trough cross-stratification are the dominant sedimentary structures of the point bar environment of meandering streams. Point bar models include fining-upwards sediment sequences and a vertical succession of cross-stratification types ranging from trough cross-stratification to parallel and ripple lamination (Visher, 1965). Coarse grained point bar deposits have been described by McGowen and Garner (1970, p. 82-91). They found no fining upwards size sequence but found the sequence of structures; large trough cross-stratification, thin foreset (tabular) stratification and small trough cross-stratification, thick foreset (tabular) stratification and above, parallel lamination.

In the cross-stratified sands of the Credit River Valley there is no fining-upwards trend of the sediment sizes nor is there any parallel laminated floodplain clays and silts. The large trough cross-stratification cuts into tabular cross-stratification which is in direct contrast to the point bar sequence. Therefore it is concluded that these sediments are not of point bar origin.

Returning to the evidence of a lowering of the water level and the isolated channeling of the linguoid bars, it appears that these transverse bars must have emerged as mid-channel braid bars disected by anastomosing low water channels.

Such a braiding process was first described by Ore (1963, p. 21) in well sorted river sediments. "During extended periods of high discharge aggradation is by large tabular bodies of sediment with laterally sinuous fronts at the angle of repose migrating downstream . . . stabilization of discharge or decrease in load after establishment of these transverse bars results in their disection by anastomosing channels".

Collinson (1970, p. 42) has studied the mid-channel linguoid bars of the Tana River in Norway and has noticed the same process of bar disection during a falling water stage. "On the stoss sides of some linguoid bars, falling stage currents may cut very shallow channels." Popov (1065, p. 190-191) has also pointed out the same process and calls it the "mid-stream bar" process. He finds that if there is a large quantity of bed load in a channel, a broad channel will result with a series of separate large dunes within it. With a lowering of the water level these dunes emerge and become mid-channel bars.

Trough, tabular and ripple cross-stratification are commonly observed upon the floodplains of braided rivers (Hjulstrom, 1952; Ore, 1963; Williams and Rust, 1969; Coleman, 1970; Collinson, 1970). Williams and Rust (1969) and Coleman (1970) have both reported ellipsoidal scour depressions being infilled by migrating bar avalanche faces as is envisaged in the Cross-stratified Sand facies. In the River Ardeche of France, Doeglas (1962, p. 181) has reported migrating dunes at periods of flood discharge in those parts of the river with a sandy bed.

Coleman (1970, p. 179) has reported migrating dunes and sand waves on the bed of the Bramaputra River of India. The sedimentary load is well sorted with a mean size of fine sand. The river channels are constantly being built up by aggradation which causes the channels to become wider and shallower. At a low water stage, the river has an excessive width and it incises the bed deposits (dunes and bars) and forms pools and riffles.

This braiding process naturally results as a consequence of the good sorting and small size range of the sediments. The size range of sand (0.06 to 2.0 mm) is an area of inflection on the Hjulstrom diagram. Any decrease in the velocity below that needed to transport the sand size load will result in deposition of all the sand size sediment. In a well sorted sand load, a velocity reduction would cause a general aggradation of all the size classes. The sand would then become the bed load of the channel and would assume that bed form corresponding to the boundary shear stress. Such a bed form would be a ripple, dune or bar structure. Also, since the aggradation is general rather than selective, the deposition would take place in all the areas of the channel, not just at its center. In the cross-stratified sands the sediment is sand sized and well sorted. The river system either did not have the competency to transport gravel or else no gravel was available to become the bed load or a combination of these two events occurred.

Ore (1963, p. 22) produced both transverse and longitudinal bars in stream table experiments. Using the same slope and discharge but with well sorted rather than poorly sorted sediment, longitudinal bars could not be formed after 20 hours, despite having previously formed them with well sorted sediment.

The Cross-stratified Sand facies is therefore concluded to be the deposits of a braided river system. The migratory bedforms, ripples, dunes and bars, are recorded by the cross-stratification. With a lowering of the water level the transverse, linguoid bars emerged. The water drained them and flowed between them rather than over them. This led to dissection of the transverse bars and to the creation of many mid-channel bars and the onset of the braided pattern.

Overbank Flooding and Bank Breaching

The Coarsening-Upwards facies has been found interbedded with deposits of the Cross-stratified Sand facies and has been interpreted as an overbank flooding deposit.

Since the Coarsening-Upwards facies is interbedded with braided river deposits, the question arises as to whether the overbank flood-decantation process is part of the sedimentation pattern in a braided river environment. From the observations and reports of Hjulstrom (1952), Chien Ning (1961), Krigstrom (1962) and Fahnestock (1963), the answer is yes.

Hjulstrom (1952) and later Krigstrom (1962) made studies of the braided drainage systems of the Icelandic sandur plains. Both geologists noted two braiding processes. The first process was the deposition of longitudinal mid-channel bars. The second process was the breaching of the river banks due to rapid sedimentation and aggradation within the river channel. The channel built itself above the surrounding floodplain and the river waters overtopped and broke through the river banks to flood the abandoned channels lower on the floodplain.

Fahnestock (1963) studied the valley train deposits and the braided White River. He observed that the braided channel built itself up as a ridge higher than the surrounding floodplain. The ridge was formed by natural levees which could block off channels and divert the river flow. Breaching or overtopping of the levee then causes a braiding process as Fahnestock (1963, p. 61) describes, "Observations on the White River show that a braided pattern may also result from a re-occupation of numerous old channels due to the deposition within the main channel as well as high flows which raise the water surface."

The lower Yellow River of China is a braided river with a sediment load that is dominantly sand and silt. Chien Ning (1961, p. 739) describes the braiding process as follows, "There exists a main channel which carries a larger portion of the flow. As time goes on, the channel is gradually filled up with sediment. This induces the river stage to rise and makes the river spill more water into another fork which is situated at a lower elevation and has a better alignment with the upper river course. After the passage of a higher flood, the main flow completely shifts over towards the new course and the old channel is rapidly silted up."

The gradual aggradation and building up of the river bed must force the water level to rise in the main channel so that at first surface waters and then later water from increasing depth would overflow the river banks. This then is the decantation mechanism that causes only the very finest sediment to be deposited on the floodplain at first, followed by increasingly coarser sediments. Also this decantation mechanism causes increasingly larger amounts of water to be carried over the river bank. This increasing flow causes the progressive change in sedimentary structures (horizontal lamination - ripple lamination - megaripple lamination) and the increased sorting of the sediments deposited.

Besides the slow process of aggradation within the river channel, the river may overtop its banks during seasonal flood periods. The floods would again raise the channel water level above the banks and cause the decantation process to begin as before. If a flood continued for a sufficiently long time to enable erosion of the banks to occur, the main channel would breach the banks and be diverted to a topographically lower, unused channel on the floodplain.

The Coarsening-upwards facies deposits are deposited in channel shaped scours and have a unidirectional flow orientation. Hjulstrom (1952), Krigstrom (1962) and Fahnestock (1963) have described numerous unused channels on the braided river floodplain. It is into these unused channels that the decanted river waters flow. The large volume of flood waters enter the disused channels and reactivate them.

The decantation process can continue to flood the lower unused channels or else the water overflowing the banks can cut through them and divert the entire flow to a lower level on the floodplain. This bank breaching is evidenced by the deposition of the trough cross-stratified sand beds upon the eroded tops of some of the rippled fine sand units of the Coarsening-upwards Sequence. Other outcrops of Coarsening-upwards Sequences display only the horizontal clay lamination and the ripple cross-lamination in the silts and fine sands. Evidently the unused channel recieved overbank flood deposits but was not reactivated by bank breaching. Coarsening-upwards Sequences are also found stacked one on top of another with the underlying sequences having the overbank flood deposits and the top sequence having overbank flood deposits and the trough cross-lamination of the reactivated channel.

The coarse sand to fine gravel deposits that are found above the trough cross-laminated sands are the bedload deposits of the newly captured river.

Therefore the decantation of surface waters and the eventual bank breaching and channel reactivation is responsible for the deposition of the Coarsening-upwards facies. This process of channel diversion is the third braiding process.

Conclusions

The central characteristic of these three braiding processes is that the fluvial system is locally aggrading. In the development of longitudinal mid-channel bars the sediment load had a wide distribution of grain sizes available. With a lowering of the flow velocity the aggradation is selective in that only the coarse tail of sediment load size distribution is deposited.

At the time of development of transverse braid bars only well sorted sand sized sediment comprised the sediment load. With a resulting loss in flow velocity, all sand sizes are equally affected and general aggradation follows. With a further lowering of the water level the bed form deposits will be incised to form the transverse braid bars.

The overbank flood-decantation process is dependent upon aggradation and building up of the channel floor. This then causes the river to seek a lower level by bank breaching and overbank flooding.

This aggradational characteristic has been observed long before by Hjulstrom (1952, p. 310), "A fundamental fact for understanding the braiding of rivers is the great sedimentary load which they carry." However this aggradation may be a local ephemeral event in an otherwise degrading system. Mackin (1956) and Williams and Rust (1969, p. 677) have both described braided rivers which are degrading. These rivers do have unstable banks and channels which allow local aggradation and the development of the braided pattern.

The Sandur Plain

The two sequences of valley fill were deposited by braided rivers whose sediments were supplied by glacial debris and outwash. These sequences occupy glacially scoured valleys indicating deposition following ice retreat.

The Halton and Wentworth glaciers advanced into the Credit Valley from its eastern side and retreated in the same direction. Ground moraine and coarse glacial debris derived from the underlying Queenston Formation were left in the bottom of the valley. This coarse, red, shale rich debris which became the Basal Red Gravel, was the initial valley fill deposit and was reworked by braided rivers. The red gravels are found only at Glen Williams where the river valley widens out. This deposit initiated the growth of the Wentworth alluvial outwash plain.

Outwash plains are found in wide valleys or plains, such as the sandur plains of Iceland, or in narrow valleys of steep gradient where they are referred to as valley train deposits. In most instances the dominant drainage pattern of the outwash plain is the braided stream. The northern portion of the Credit River Valley is often one mile or less in width but lacks the steep slope of a valley train. Below Glen Williams, the Credit Valley attains a width of two miles or more. Embleton and King (1968) suggest that a slope of 5m/km. is characteristic of the slopes of sandurs. The slope of the bedrock surface in the Credit Valley is approximately 6m/km.

As the ice retreated to the east, the source of the sediment supply became more distant and the grain size of the deposits decreased. The meltwater actively eroded and reworked the initial deposits and left the sandur with an uneven surface covered by used and unused stream channels. The uneven surface of Icelandic sandurs has been described by Krigstrom (1962). However the many laterally migrating braided channels distributed sediment over the outwash plain surface, filling in the scours and depressions on the plain, and building the outwash plain laterally and vertically.

This aggradation is evidenced by the braided river deposits of the Grey Gravel, Cross-stratified Sand, and Coarsening-upwards facies. Each of these deposits are widespread indicating a rapid building up and outwards of the alluvial plain surface. The discharge of meltwaters on outwash plains varies enormously over the year due to fluctuating climatic conditions (Krigstrom, 1962; Fahnestock, 1963). This fluctuating discharge rate causes periodic flooding of large portions of the sandur surface.. This flooding and bank overflow resulted in the deposition of the Coarsening-upwards facies in the Credit Valley sandurs.

The two sequences of valley fill deposits were then the result of outwash plain construction. These outwash plains took the form of modern sandur plains with their characteristic braided stream patterns, high rates of sediment discharge and seasonally varying meltwater discharge. The Wentworth sandur plain was constructed by the meltwaters of the Wentworth glacier. The upper Halton sandur plain was constructed by the Halton glacier.

PLEISTOCENE SEDIMENTARY HISTORY

Chapter VII

The Credit Valley and the adjacent ice ways from the valley onto the top of the Escarpment have suffered extensive bedrock erosion. As has been pointed out there is a similarity between the events in the last two ice advances of the Late Wisconsin period. Such similarities in the manner and direction of ice flow might have extended to earlier Nebraskan, Kansan and Illinoian glaciations. The deep bedrock erosion might then be the result of repeated ice advances through this region.

The oldest Pleistocene sediments that were mapped were Mankato/Port Huron (Late Wisconsin) in age. These sediments were deposited during the advance and retreat of the Wentworth and Halton Ice sheets. The Wentworth glacier overrode the Escarpment and extended as far west as at least the Paris and Galt moraines. It has been suggested that when the ice retreated, it receeded to the vicinity of Toronto (Karrow, 1963, p. 56). The glacial till deposited by this ice advance is found atop the Escarpment. It is found nowhere in the Credit Valley and is covered to the east by later glacial deposits. As the Wentworth Ice sheet retreated, the meltwaters flowed down into the Credit Valley from atop the Escarpment at Acton, Credit Forks as well as in a meltwater channel running in front of the Escarpment, south of Mono Mills.

The initial deposit of the meltwaters in the glacially scoured valley was the coarse bedrock debris. The debris was largely red shale of the Queenston Formation and it was transported only a short distance before its deposition and burial. This shale rich basal red gravel was deposited by a braided river system. As the ice continued to retreat, meltwaters from further to the north and the west flowed into the Credit Valley.

The meltwaters reworked the red gravels and brought additional gravels from areas to the north. These gravels became incorporated into the grey gravel, and the deposit is localized for the most part in the wider reaches of the Credit Valley at and below Glen Williams. The red and grey gravels were the basal units of the Wentworth sandur plain. The aggradation of the sandur plain was continued with the deposition of the cross-stratified sand and coarsening-upwards units.

The fining upwards character of the sandur deposit was due to the continued withdrawal of the Wentworth Ice front. As the ice moved further away, the sediment source was further removed and the meltwater discharge lowered in this area.

The beginning of the last ice advance, the Halton Ice sheet, is evidenced by the deposition of varved lacustrine clays on top of the Wentworth sandur deposits. The varved sediments were deposited in a proglacial lake which occupied the Credit Valley. The ice advanced from the south and east and dammed the southerly flow of the drainage. With continued advance the ice sheet surmounted the Escarpment west of the Credit Valley. The Halton Till was deposited atop the Escarpment and directly before the Paris moraine. In the valley, the lacustrine clays grade continuously upward into silty Halton Till.

The Halton Ice retreated to the base of the Escarpment during a fluctuating period of retreat and readvance. A series of small moraine strands as well as the broader but very shallow Bolton and Cheltenham moraines were deposited. Straw (1968) interprets these moraines as correlatives of the multiple series of Waterdown moraines. This ice halt ponded meltwater between the ice front and the Escarpment. The shallow and smooth Bolton and Cheltenham recessional moraines indicate deposition in a body of water. The ice front was fluctuating as there are exposures of till and sand interbedded several times. At Georgetown a kame delta deposit is covered by ten feet of fine sand. The meltwaters of the Halton Ice constructed the kame delta as the ice stagnated and started to receed. The ice then readvanced and deposited the ten feet of fine sand.

At Norval the presence of the Halton Ice sheet is indicated by a deposit of 60 feet of coarse poorly sorted kame gravels. The kame gravels were deposited as small mounds by the meltwaters of the wasting Halton Ice sheet. As the ice wasted back it slowly receeded to the southeast over the Peel

Plain. The ice receeded south east to the Trafalgar moraine where it once again halted. The meltwaters flowing southeast from the Credit Valley were ponded before this moraine and became the Peel Pond.

The meltwaters in the Credit Valley had once again constructed a sandur plain with the grey gravel, cross-stratified sand and coarsening-upwards sequences. The sandy units of the sandur construction were not as extensive as they were in the Wentworth sandur. This outwash period was probably one of degradation rather than aggradation. Much of the Wentworth and Halton outwash sediment was removed by erosion from the area around Georgetown.

The sandy outwash built a sheet sand deposit into the waters of the Peel Pond at Norval. This deposit is identical to other so-called deltas (Chapman and Putnam, 1966) on the Humber River. Clay horizons interbedded with the sheet sand were the deposits of this short-lived glacial lake.

The waters of the Peel Pond drained to the west in front of the Trafalger moraine until they reached the present Lake Ontario. This they did by flowing over the moraine in that area where it intersects the Escarpment. When the ice finally retreated from the area around Toronto and from its halt along the Trafalger moraine the Credit River drained directly into Lake Ontario. With complete removal of the ice the river dissected its deposits and cut down into the valley floor to expose the older Pleistocene deposits.

APPENDIX I

GRAIN SIZE MOMENTS

BISHOP PIT, GLEN WILLIAMS

Sample No.	Mean (phi)	St. Deviation (phi)	Skewness	Kurtosis
1a	0.813	0.793	0.14	2.31
1b	0.834	0.764	0.17	1.67
1c	1.235	0.596	0.06	1.75
1d	1.503	1.089	-0.41	1.62
1e	4.783	1.891	1.31	8.32
1f	3.614	1.316	1.81	21.71
1g	2.509	0.767	0.11	-0.73
1h	-0.357	1.770	0.29	-0.12
1ht	1.412	0.634	0.05	3.09
2a	1.406	0.527	0.90	7.59
2b	-1.491	1.441	-0.36	0.16
2c	-1.097	1.538	0.27	0.01
2d	-1.807	2.326	0.22	-0.69
3a	-1.537	1.728	0.62	1.86
3b	0.860	0.914	0.11	1.20
3c	-2.045	1.407	0.56	1.74
6g	1.268	0.995	-0.02	0.83
6l	0.052	1.163	-0.34	1.32
6m	2.169	0.655	0.27	1.16
7d	2.644	0.684	-1.12	11.40
7e	1.088	1.815	-0.61	0.13
7g	2.656	0.423	0.17	0.20
7h	1.670	1.303	-0.86	2.12
8a1	1.315	1.094	-1.12	5.70
8a2	2.068	0.483	-0.07	3.76
8b1	5.165	1.961	0.99	5.25
8b2	4.414	1.874	1.55	10.43
8b3	4.400	1.494	1.52	13.76
8b4	3.730	0.647	2.55	64.11
8b5	3.652	0.612	2.45	67.19
8b6	3.550	0.577	1.86	55.55
8b7	3.078	0.817	0.22	19.15
8b8	3.353	0.573	2.11	58.95

Sample No.	Mean (phi)	St. Deviation (phi)	Skewness	Kurtosis
8b9	3.437	0.686	2.66	68.05
8b10	3.317	0.577	1.91	54.96
8b11	3.390	1.001	2.32	30.49
8c1	1.344	0.780	-1.29	11.56
8c2	1.508	0.612	-1.32	19.10
8d	1.328	0.718	-0.79	7.07
8f	1.587	0.492	-0.08	6.42
8g	1.798	0.505	0.11	2.39
8h	1.489	0.492	0.44	3.28
8i	1.808	0.838	0.43	0.16
8j	1.432	0.477	-0.14	6.25
8l	0.512	0.774	0.59	5.02
9a	-0.839	1.247	0.23	0.78
10a	-2.183	1.316	0.75	3.44
11a	0.569	1.517	-0.88	2.72
11b	0.797	0.902	-0.40	4.16
11c	1.228	0.644	0.22	2.15
11d	1.456	0.987	0.30	0.03
11e	2.992	0.478	-0.16	2.73
11f1	1.691	0.535	0.30	3.22
11f2	1.500	0.690	0.15	3.98
12a1	0.541	0.994	-0.41	3.37
12a2	0.412	1.454	-0.61	0.64
12a3	0.853	0.763	-0.48	8.42
12c	0.975	1.838	-0.47	1.18
13a	0.925	0.832	0.27	3.51
13b	2.670	0.650	-0.51	6.21
13c1	1.691	0.868	0.03	0.61
13c2	1.961	0.678	-0.16	1.12
13d1	3.776	1.632	1.42	12.38
13d2	3.288	1.260	1.61	15.93
13d3	2.989	0.986	1.89	28.76
14a1	1.584	0.882	-0.07	0.05
14a2	1.844	0.545	-0.06	3.52
14d	0.144	1.335	-0.03	0.93
14e	-1.198	1.742	0.22	0.21
15b	-0.536	1.771	-0.12	-0.68
15c	0.221	1.175	-0.01	1.68

Sample No.	Mean (phi)	St. Deviation (phi)	Skewness	Kurtosis
16a1	-0.234	1.500	-0.17	-0.78
16a2	0.473	1.444	-0.85	2.41
16a3	0.535	0.937	-0.39	2.95
16c	0.930	1.003	-0.93	5.19
16d	2.422	0.456	-0.25	3.55
16e1	1.924	0.421	-0.19	3.28
16e2	1.784	0.684	-0.75	3.96
16f	1.858	0.713	-0.60	3.72
16g	1.452	0.744	-0.95	8.75
16h	1.180	0.676	-0.66	9.60
16i	1.113	0.827	-1.01	9.23
17b	-0.935	1.999	0.34	-0.13
18d	-1.987	1.698	0.57	0.96
18f	0.698	2.468	-0.16	-1.27
19a	-1.796	1.619	0.22	-0.49
21b	2.125	0.611	0.44	1.06
21c	-0.916	1.143	0.29	1.46

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