National Library of Canada
Bibliothèque nationale du Canada

Canadian Theses Service
Services des thèses canadiennes

Ottawa, Canada
K1A 0N4

CANADIAN THESES

THÈSES CANADIENNES

NOTICE

The quality of this microfiche is heavily dependent upon the quality of the original thesis submitted for microfilming. Every effort has been made to ensure the highest quality of reproduction possible.

If pages are missing, contact the university which granted the degree.

Some pages may have indistinct print especially if the original pages were typed with a poor typewriter ribbon or if the university sent us an inferior photocopy.

Previously copyrighted materials (journal articles, published tests, etc.) are not filmed.

Reproduction in full or in part of this film is governed by the Canadian Copyright Act, R.S.C. 1970, c. C-30.

AVIS

La qualité de cette microfiche dépend grandement de la qualité de la thèse soumise au microfilmage. Nous avons tout fait pour assurer une qualité supérieure de reproduction.

S'il manque des pages, veuillez communiquer avec l'université qui a conféré le grade.

La qualité d'impression de certaines pages peut laisser à désirer, surtout si les pages originales ont été dactylographiées à l'aide d'un ruban usé ou si l'université nous a fait parvenir une photocopie de qualité inférieure.

Les documents qui font déjà l'objet d'un droit d'auteur (articles de revue, examens publiés, etc.) ne sont pas microfilmés.

La reproduction, même partielle, de ce microfilm est soumise à la Loi canadienne sur le droit d'auteur, SRC 1970, c. C-30.

THIS DISSERTATION
HAS BEEN MICROFILMED
EXACTLY AS RECEIVED

LA THÈSE A ÉTÉ
MICROFILMÉE TELLE QUE
NOUS L'AVONS REÇUE

NL-339 (r.86/06)

Canada

LATE QUATERNARY STRATIGRAPHY AND DEPOSITIONAL ENVIRONMENTS IN THE BASIN OF THE RICHARDSON AND RAE RIVERS, N.W.T.

by

Daniel Ernest Kerr

A thesis submitted to the School of Graduate Studies in partial fulfillment of the requirements for the degree of M.Sc. in Geology

UNIVERSITY OF OTTAWA

© Daniel E. Kerr, Ottawa, Canada, 1986.

Permission has been granted to the National Library of Canada to microfilm this thesis and to lend or sell copies of the film.

L'autorisation a été accordée à la Bibliothèque nationale du Canada de microfilmer cette thèse et de prêter ou de vendre des exemplaires du film.

The author (copyright owner) has reserved other publication rights, and neither the thesis nor extensive extracts from it may be printed or otherwise reproduced without his/her written permission.

L'auteur (titulaire du droit d'auteur) se réserve les autres droits de publication; ni la thèse ni de longs extraits de celle-ci ne doivent être imprimés ou autrement reproduits sans son autorisation écrite.

ISBN 0-315-33337-5

UNIVERSITÉ D'OTTAWA
UNIVERSITY OF OTTAWA

ABSTRACT

A detailed investigation of the stratigraphic sections in the basin of the Richardson and Rae Rivers has led to a better understanding of the Late Quaternary history of deglaciation in this part of the Northwest Territories.

Regional lithostratigraphic correlation confirmed the existence of a major glacial lake which evolved through different phases prior to its transition to a postglacial sea following the easterly retreat of a lobe of the Laurentide Ice Sheet from the study area. The glaciolacustrine to marine transition is characterized by an upward-fining trend representing the proglacial lacustrine environment where sedimentation was controlled by a receding ice front, succeeded by an upward-coarsening trend corresponding to the regressive marine sequence resulting from isostatic uplift. No intermediate terrestrial phase or marine transgression is recorded, as the marine incursion was instantaneous since sealevel and the last proglacial lake level were approximately at the same elevation.

The complete sedimentary succession is composed of: a) basal and/or flow till, b) stratified subaqueous sand and gravel, c) glaciolacustrine silty clay rhythmites, d) massive marine silty clay, and e) littoral marine sand. The sudden passage from lacustrine to marine environments is reflected in the abrupt change in depositional mechanisms. Turbidity currents prevailed in the proglacial lake as inferred from thick sequences of varve-like couplets, whereas the salinity of marine

waters, impeding the sedimentation of fines by heavy-density bottom flows, led to the deposition of structureless marine silty clays by flocculation.

Faunal evidence is indicative of near-normal marine conditions in the deeper parts of the sea, and brackish-water marginal marine environments in the upper part of the water mass resulting from the dilution of nearshore marine waters in an estuarine environment.

RÉSUMÉ

Une étude détaillée de la stratigraphie du bassin des rivières Rae et Richardson permet de mieux définir l'historique du Quaternaire supérieur dans cette région des Territoires du Nord-Ouest.

Une corrélation lithostratigraphique régionale confirme l'existence d'un lac glaciaire important et précise l'évolution sédimentologique de ce lac à mesure qu'il se transformait suivant les phases de retrait d'un lobe de glace de la calotte glaciaire laurentidienne. Suite au retrait du glacier du bassin, le lac glaciaire subit une transition à un environnement marin, sans phase intermédiaire terrestre ou de transgression marine puisque ces deux masses d'eau étaient environ à la même altitude. La transition de l'épisode glaciolacustre à la phase marine est caractérisée par une granodécroissance vers le sommet des coupes représentant la phase lacustre où la sédimentation était controlée par le glacier, suivie d'une granocroissance qui correspond à la phase de régression marine résultant du relèvement isostatique.

La succession stratigraphique complète comprend: a) un till de fond et/ou un "flow till", b) des sables et graviers interstratifiés sousglaciaires, c) des rythmites silto-argileuses glaciolacustres, d) des silts argileux marins massifs, et e) des sables marins littoraux. Le passage soudain d'un environnement lacustre à un environnement marin est caractérisé

par un changement brusque dans les processus de sédimentation. Des courants de turbidité sont responsables des dépôts de rythmites glaciolacustres, tandis que la salinité des eaux marines empêcha le dévelopement de courants sous-marins et favorisa la sédimentation d'argile massive résultant de la flocculation des particules fines.

La faune marine indique d'une part des conditions normales pour les eaux profondes de la mer postglaciaire, et d'autre part des conditions marginales d'eaux saumâtres pour la masse d'eau supérieure. Ces dernières résultent de la dilution de l'eau marine littorale dans un estuaire.

ACKNOWLEDGEMENTS

I would like to thank Dr. M.-A. Geurts for her supervision and for the financial assistance (NSERC A 6888) she provided over the course of the study period.

This investigation was made possible due to research agreements granted to the author by the Department of Indian Affairs and Northern Development, and a grant from the University of Ottawa Northern Research Group. Logistical support was provided by D.A. St-Onge (Geological Survey of Canada), W.A. Padgham (D.I.A.N.D.), and Bob Chalifoux (Maple Leaf Helicopters). The continual support of Polar Continental Shelf Project and the field assistance provided to the author (Project 175-85) is greatfully acknowledged.

I also wish to thank A. Mercier, and all those who aided during the research stage of this thesis, for numerous discussions, comments and assistance they contributed in one form or another. Very sincere thanks are also extended to my sister, D. Kerr for having patiently typed this thesis.

TABLE OF CONTENTS

TABLE OF CONTENTS - Continued

TABLE OF CONTENTS - Continued

TABLE OF CONTENTS - Concluded

LIST OF FIGURES

LIST OF PLATES

3 15

LIST OF PLATES - Concluded

LIST OF TABLES

LIST OF APPENDICES

CHAPTER 1

1. INTRODUCTION

1.1 Purpose and Scope

The objectives of this thesis are to investigate the nature of Late Quaternary stratigraphy exposed along river sections in the basin of the Richardson and Rae Rivers, Northwest Territories (Figure 1). This study, providing new data pertinent to the deglaciation history of the region, is based primarily on a systematic investigation of glacial, glaciofluvial, glaciolacustrine and marine deposits. A discussion on the succession of depositional environments is presented, beginning with the retreat of the Late Wisconsin Laurentide Ice Sheet through to the Holocene marine regression. A regional lithostratigraphic correlation shows the extent of the sedimentary facies and their relationships with one another.

A study of the postglacial marine fauna, i.e., foraminifers, ostracodes, and macrofossils, was also undertaken in order to determine palaeoenvironments throughout the marine episode, and provide reference lists of Late Quaternary micro and macrofossils for this part of the Northwest Territories.

1.2 Location of the Study Area

The Richardson and Rae River basin is situated at the western end of Coronation Gulf, District of Mackenzie, Northwest Territories. The study area is deliniated by

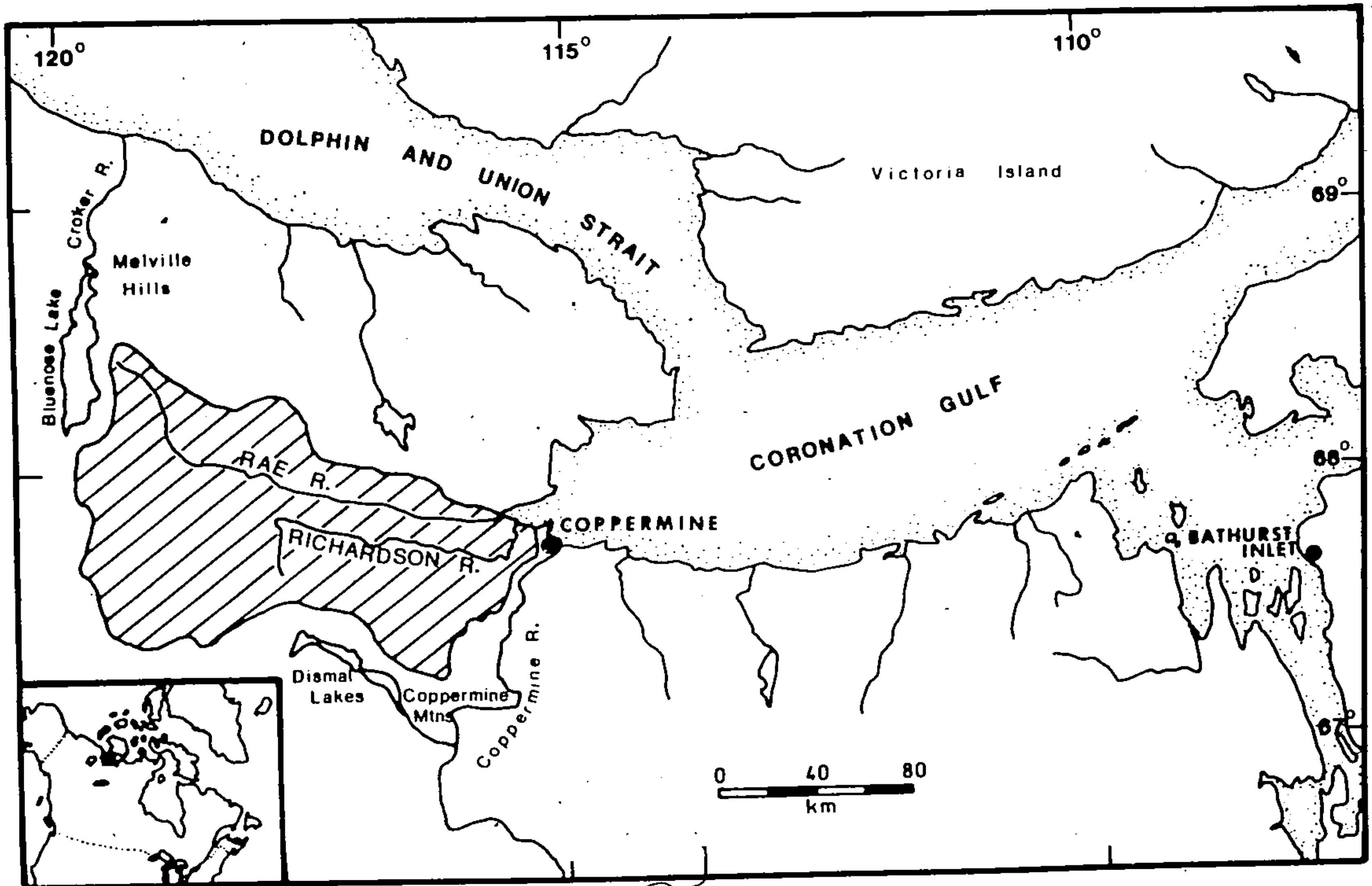


FIGURE 1. Location map of study area.

the limits of this drainage basin, between the longitudes 115° 15' and 119° 45' West, and the latitudes 67° 15' and 68° 30' North. The area covers portions of the N.T.S. topographic map sheets 86 M, N, O and 87 B.

This region was chosen as a suitable study area for a variety of reasons. Firstly, its Quaternary history is poorly known and has received very little attention to date. A regional study of the surficial deposits by St-Onge (1980) in the Coppermine River valley (Figure 1) and by Mercier (1986) in the Richardson River basin, indicated that a stratigraphic study of Quaternary deposits was required, since no detailed stratigraphic investigations had previously been carried out in the basin of the Richardson and Rae Rivers. As a result, the present study was partially undertaken in cooperation with Geological Survey of Canada mapping projects in the area, which facilitated logistical support in this relatively inaccessible region. The presence of numerous, well exposed river-bank sections were also considered to be an asset in the choice of this research area.

1.3 Methodology

The methods employed in this research project can be divided into those carried out in the field, and in a laboratory. Available geological evidence came from a number of sources: regional studies, stratigraphy and sedimentology, paleontology and sedimentary geochemistry.

The field work was carried out from two different camp sites during the summers of 1984 and 1985, Cox Lake and Coppermine respectively. Rotary-wing and fixed-wing aircraft were used to gain access to the most promising sites. In places with little or no slumped material on the cliff face, sections were cleared, measured, described and sampled for grain-size and microfossil analyses. Molluscs were collected where they were readily observed. Samples were carefully chosen so as to be representative of each unit as a whole. Whenever possible, organic matter such as wood fragments and marine bivalves were collected for radiometric dating. During these field seasons, profiles were examined at 40 localities, the majority of them along the Richardson and Rae Rivers.

Grain size analyses were carried out in the Sedimentology Laboratory of the Geological Survey of Canada. Two methods of grain size analyses were employed: dry-sieving and pipette, (Appendix A). These results provide a description of the particle size distribution in each sample. The purpose of these analyses is to identify specific associations between stratigraphic type and grain size and to illustrate the trends in grain-size characteristics.

Identification of clay minerals from certain lithofacies was accomplished by X-ray diffraction at the Mineralogy Laboratory, Geological Survey of Canada, Ottawa.

Wood and marine bivalve collections were submitted to the Radiocarbon Laboratory, Geological Survey of Canada for dating. When such dates were unavailable, stratigraphic principles of superposition were used for relative dating purposes to interpret the depositional history.

For microfossil analyses, approximately 50 g of sediment from certain samples were wet-sieved through a 4.0 Ø mesh wire sieve. The residues on the screens were dried and examined for microfossils. Ostracodes and foraminifers were handpicked with the use of a binocular microscope and identified by the writer using published references. Although it is advisable to obtain at least 100 specimens in order to provide an accurate representation of the fauna, it was not always possible to do so due to the rarity of specimens, notably of ostracodes.

CHAPTER 2

2. REGIONAL GEOLOGIC SETTING

2.1 Physiography

The basin of the Richardson and Rae Rivers covers more than 10,500 km^2. It is restricted to the north by highlands north of the Rae River, to the east by the Coppermine River drainage divide and Coronation Gulf, to the south by Dismal Lakes and the Coppermine Mountains (Figure 1), and to the west by the Melville Hills and Bluenose Lake (Figure 1). It can be subdivided into two distinct physiographic zones; the southern part of the basin is characterized by a relatively unbroken plateau whose altitude decreases eastward, and which is marked by extensive bedrock outcrops with minor, isolated till patches. The central and northern parts of the basin, extending east-west, are relatively flat-lying lowlands which dip towards the south and east. The gently rolling to hummocky topography is due in part to the varying thicknesses of Quaternary sediments of glacial, glaciolacustrine and marine origin which fill depressions between outcrops. The latter, protruding through the surficial deposits, often show evidence of glacial striae and are commonly fluted.

The easterly flowing Richardson and Rae Rivers are the main components of the drainage system. The Rae River descends 600 m over a distance of approximately 180 km, whereas the Richardson River descends 278 m over a distance of 167 km before terminating in the waters of Coronation Gulf.

2.2 Bedrock Geology

The unconsolidated Quaternary deposits in the basin overlie bedrock belonging to parts of the Bear Province, the Arctic Platform, and the Interior Platform of the Canadian Shield (Dept. of Energy, Mines and Resources, 1980). The geology consists of three principal lithologic rock groups, of volcanic and sedimentary origin.

The northern part of the basin is composed of dolomite, limestone and minor gypsum, belonging to the Rae Group, defined by Baragar and Donaldson (1973) as Hadrynian in age (Upper Proterozoic). These are believed to belong to a shallow-water succession of shelf deposits that accumulated in a broad basin (Campbell, 1983). They dip gently north, at 2° to 5°, forming part of the Coppermine Homocline, whose dolomite-dominated unit is intercalated with gabbroic sills. As a result of this structure, numerous cuestas of dolomite overlain by gabbro (Coppermine Sills) were developed, which constitute the major landform unit in this portion of the basin. At the base of some stratigraphic sections, there are the characteristic red and green sandstones of the Rae Group. Unconsolidated sediments directly overlying these two rock types frequently exhibit faint to moderately strong hues of red or green in their matrix.

The southern half of the basin is underlain by a complex sequence of basalt and sandstone, also dipping north, 2° to 15°. These rocks form the Coppermine River Group of Helikian age (Baragar and Donaldson, 1973). Rocks from this group are cross-cut by a series of N.E.-S.W. and N.W.-S.E. trending faults, as well as diabase dykes of various orientations. Topographically, this area is dominated by hills attaining elevations several hundred metres higher than the relatively flat region of the northern part of the basin.

The most northern and western extremities of the basin are underlain by Lower Paleozoic carbonates (Campbell, 1983) which overlie the Proterozoic rocks of the Bear Province.

2.3 Surficial Geology

Late Quaternary and Holocene unconsolidated sediments are found throughout the basin of the Richardson and Rae Rivers. Four principal types of surficial deposits have been defined in the region.

They are:

1. Glacial
2. Glaciofluvial
3. Glaciolacustrine
4. Marine

A till blanket (2 m to 10 m thick) and till veneer (0.5 m to 2 m) (Mercier, 1986) predominate in the southern half of the basin, above 260 m a.s.l. However, scattered till deposits are also present on the surface of more elevated areas in other parts of the basin which were not submerged either by glacial lakes or by the postglacial sea. Varying in thickness from a few decimetres to 10 m or more, the till is also fluted and drumlinized, notably in the southeastern part of the study area.

Glaciofluvial sand and gravel, found at all elevations, are predominantly in the form of eskers, which may locally grade into esker-deltas. Their sinuous forms usually consist of short linear segments tens of metres in length, but larger glaciofluvial complexes extending over 25 km in length exhibit surfaces pitted by kettle-lakes and are frequently flanked by kame terraces. Glaciofluvial ridges may also grade into outwash plains, and have a general east-west orientation.

Glaciolacustrine deposits, consisting of sand, silt and clay, form a uniform blanket of sediments in the northwestern part of the basin. They are present between the altitudes of 150-170 m a.s.l. and approximately 600 m a.s.l. or more to the west of the study area (St-Onge, personal comm., 1986). However, in stratigraphic sections they have been observed at elevations as low as 3 to 5 m a.s.l. along the Richardson River. Such exposures show the true characteristic nature of these silty-clay stratified deposits which cannot be seen when they occur as surficial sediments. Associated with these lacustrine deposits are perched deltas. Situated in the southern, western and northern parts of the basin, they occur between the altitudes of 520 m and 180 m a.s.l., representing the successively decreasing water planes of a series of proglacial lakes.

Below 150-170 m a.s.l., sand, silt and clay of marine origin overlie most other deposits. These sediments vary greatly in thickness throughout the basin, from as little as 20 cm in their western-most location to as much as 20 m in the eastern parts. Sandy littoral deposits have only been recorded below an elevation of 90 m a.s.l., and become increasingly important in distribution below 45 m a.s.l. Raised beaches are conspicuously absent from this area, possibly because of the predominance of fine-grained sediments.

Marsh environments are widespread in topographic depressions. Minor alluvium and fluvial deposits are also present restricted along stream and river beds, although these are of little stratigraphic importance.

2.4 Previous Studies

a) J. O'Neill

The Canadian Arctic Expedition of 1913-1918 led to the geological investigation of parts of the Arctic coast, which included most of the area studied by the present writer. At the mouth of the Croker River (Figure 1), Pleistocene deposits were noted to attain over 30 m in thickness (O'Neill, 1924, p. 29A). These sediments consist of silt and clay, reddish brown to buff in colour and contain well-rounded pebbles and slabs of sandstone and limestone. These widespread deposits were noted all along the coast and as far inland as the head of the Rae River. North of this river, O'Neill noted the presence of three ridges about 61 m high, composed of reddish clay and well-rounded gravel. These are most likely the landforms interpreted by Craig (1960) as end moraines. Southeast of the study area, hills of silty-clay containing marine fossils were also noted over a distance of 14 km south of the mouth of the Coppermine River (O'Neill, 1924, p. 29A).

Pleistocene fossils were collected during this early expedition from a number of sites, two of which are located just outside of the study area. From Station 5284 (O'Neill, 1924, p. 30A), 8 km west of the Coppermine River mouth, the following species were recorded: Leda pernula Müller, Cardium ciliatum Fabricius, Macoma calcarea Gmelin, M. balthica Linné, and M. brota Dall. These fossils were found in a clay bank 15 m high. Similar species were noted 5 km east of the mouth of the Coppermine River. Marine fossils were found up to elevations of 152 m a.s.l. at the eastern end of Coronation Gulf and indicate a former submergence of at least this amount, and possibly even more (ibid). O'Neill concluded that as indicated by striae, the ice movement was towards the northwest and that these sediments represent the closing phase of the Pleistocene period.

b) B. Craig

In the north-central District of Mackenzie, Craig (1960) mapped ice-flow indicators in the area (Figure 2) and concluded that ice flow was generally from the southeast to the northwest, as recorded by numerous striae and roches moutonnées throughout the basin. He also concluded that the ice retreated easterly over this area, and proposed that an ice

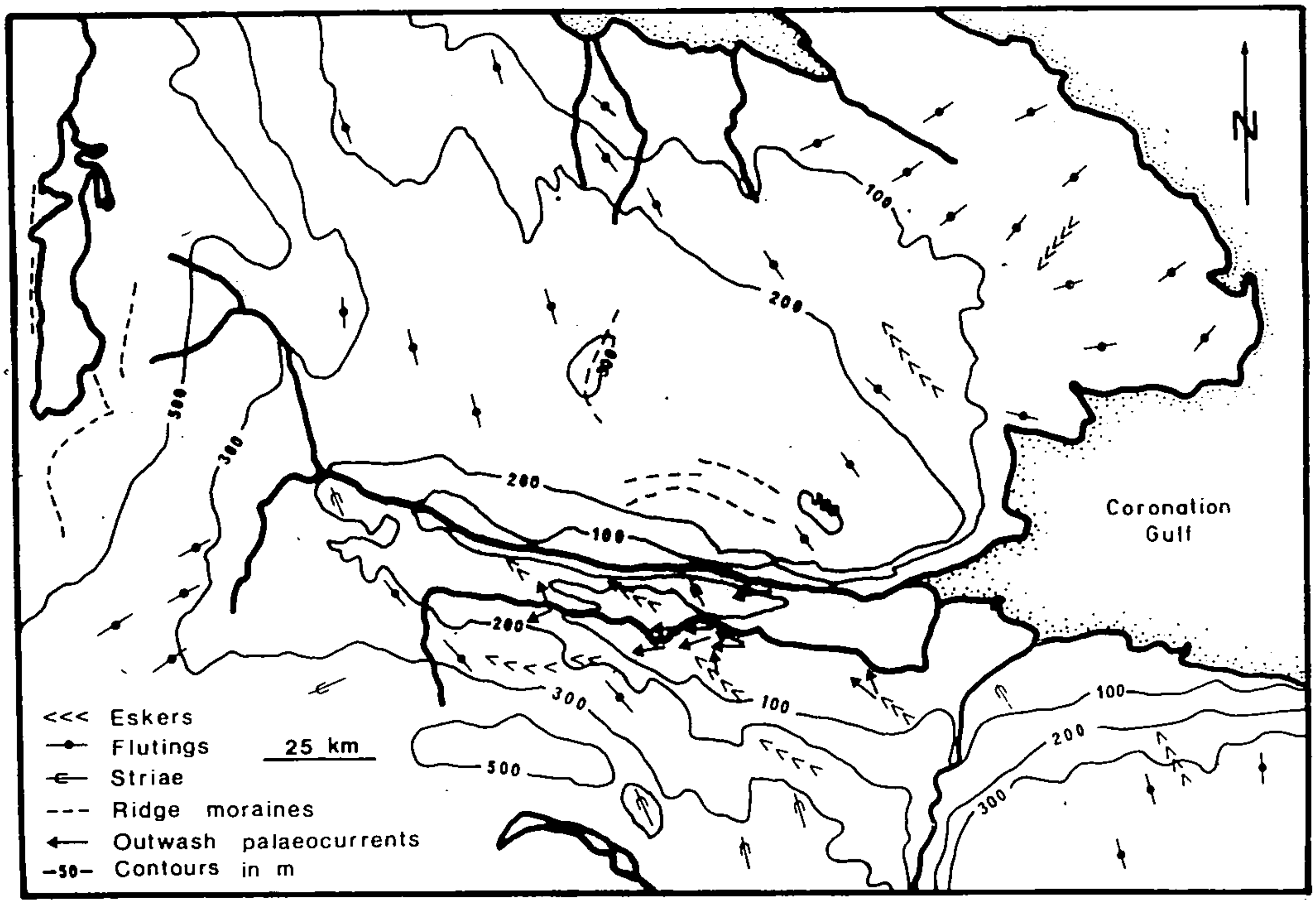


FIGURE 2. Ice-flow and palaeocurrent directional features (modified from Craig, 1960).

lobe from Victoria Island (Figure 1) to the northeast merged with northwestern-moving ice from the continent and continued to flow in a northwesterly direction. The southern lobe became separated into two actively flowing sub-lobes. One of these sub-lobes was directed along the valley of the Richardson River and deflected to the southwest and to the northwest by the highlands north and south of Bluenose Lake. The second lobe stagnated in the Great Bear Lake area, forming large deposits of hummocky moraines (Craig, 1960).

Most glacial lakes encountered by Craig were relatively small and resulted from the blocking of natural drainage by the retreating ice front and by end moraines. Craig (1960) made no reference to the thick sequences of silty-clay rhythmites widely exposed along the Rae and Richardson Rivers. However, given the large area mapped, it is understandable that detailed stratigraphy could not be undertaken over the entire map sheet.

Although end moraines are rarely found in the basin of the Richardson and Rae Rivers, Craig (1960) mapped numerous, large N-S trending moraine ridges immediately east and west of Bluenose Lake, composed of bouldery clay till, rising some 46 m above the surrounding topography.

Abandoned strand lines along the coast and vast deposits of silt and clay are believed to record higher sea levels. Craig used distinct changes in the surficial deposits related to reworking by marine processes to precisely define the extent of submergence. It was sometimes necessary to identify marine sediments by the presence of fossils, as well as by their similarities to known marine deposits. Craig reported that the limit of marine submergence decreases in elevation from east to west: from 146 m a.s.l. just east of Coppermine to 46 m a.s.l. west of the mouth of the Croker River (Figure 1). Blake (1970) attributed this decrease to differential post-glacial uplift. Radiocarbon dates on marine bivalves indicate that the coast was free of ice 10,000 years ago, but, as the present writer has also observed, no fossils were recorded within an altitude of 50 m below the limit of submergence.

Along Dolphin and Union Strait (Figure 1), pebbly mud at the base of thick marine clay and silt deposits was interpreted by Craig (1960) as probably being the result of coarse debris that was dropped from melting icebergs. However, the exact location of these sediments is uncertain.

c) W. Baragar and J. Donaldson, and R. Allan *et al*.

Although the principal interest of study by Baragar and Donaldson (1973) lay with the bedrock geology of the Coppermine and Dismal Lakes map areas, some general observations concerning the Quaternary geology were made. It was noted that Pleistocene surficial deposits are widespread in the area, and that ice-flow was generally from the southeast. They reported that marine deposits of white silt are present along the coast of Coronation Gulf and in the valleys of the Rae and Richardson Rivers. These sediments extend to about 117° 30' W along the Richardson River, at an elevation of about 107 m a.s.l. Baragar and Donaldson (1973) concluded that these marine deposits mark an advance of the sea in the immediate postglacial period.

Allan *et al*., (1972), concerned with geochemical exploration studies of lake sediments in the southern part of the basin mapped the surficial deposits of the area on the sole basis of aerial photograph interpretation. A division of the glacial deposits into four groups resulted in the following classification: till with minor rock outcrop, rock with till, water laid deposits and colluvial deposits. Based on their observations, Allan *et al*., (1972) believed that the till had not travelled far, and may be predominantly local in origin.

d) D.A. St-Onge and H. Bruneau

South of the study area, St-Onge (1980) mapped the Quaternary geology along the Coppermine River valley. This led to the discovery of extensive deposits of a major late glacial lake (St-Onge and Geurts, 1982). Glacial Lake Coppermine was formed when the lower Coppermine valley was dammed by a glacial lobe covering the lowlands north of the Coppermine Mountains and Dismal Lakes, i.e., the region represented by the Richardson and Rae River basin. The lake eventually drained in part westward through the Dismal Lakes system and into a high-level Great Bear Lake. St-Onge's study represents the first attempt to define detailed ice-retreat phases in the region based largely on the occurrence of glacial lakes and outlets.

Systematic mapping of the surficial deposits in the lower Coppermine River valley (St-Onge and Bruneau, 1982) has revealed a series of raised deltas and beaches which mark the postglacial marine limit and successively lower phases of the marine regression. The limit of submergence was determined to be 170 m a.s.l., as evidenced by the presence of an esker-delta built at the ice-sea interface. Further north, four other deltas were identified, one at 140 m a.s.l. a second at 100 m a.s.l. and two more at 70 and 40 m a.s.l.

Up to 40 m of rhythmically bedded sand, silt and clay were deposited in an estuarine environment of the postglacial sea (ibid, p. 51). These deposits have been interpreted as turbidites (Bruneau, 1984), formed as the result of density currents originating from sediment-laden water of the Coppermine River entering a brackish estuarine environment and flowing along the sea bottom, and/or subaqueous flows generated by unstable deltaic deposits. The source of these sediments is thought to have been the deposits of Glacial Lake Coppermine which were supplied by the downcutting of the Coppermine River. A summary of the deglaciation phases in this region is presented by St-Onge (manuscript, 1986).

e) A. Mercier and D. Kerr

A more recent study carried out in the basin of the Richardson River is that of Mercier (1986), who mapped surficial deposits and refined the deglaciation history from Late Wisconsin time to the present. The proposed model established a chronologic sequence of events beginning in the waning stages of glaciation through to the marine transgression and subsequent marine regression.

Mercier (1984) reconstructed the evolution of a series of Late Pleistocene proglacial lakes, identifying four major stages. However, stratigraphic evidence for this lacustrine phase was lacking in certain areas, whilst in others, it raised a number of questions about the extent and duration of the glacial lakes. The most striking of these questions was the restriction of rhythmically bedded sediments to areas below the marine limit. This left some doubt as to their origin, i.e., marine or lacustrine, since similar deposits in the lower Coppermine River valley were interpreted as marine (Bruneau, 1984).

The present writer carried out field work during the summer of 1984 with Mercier and St-Onge. Kerr and Mercier (1985) presented a brief overview of the sedimentary facies and typical stratigraphic sections studied by Kerr; these sections were also briefly discussed by Mercier (1986).

Further research during the summer of 1985 by the present writer led to the discovery of several key stratigraphic sections, included in this thesis, which not only confirm the existence of a series of relatively large proglacial lakes, but also provided new information as to their extent and mode of formation. The associated lacustrine sediments are described herein in detail and a model for the glaciolacustrine to marine transition is presented.

With respect to the marine regression and related deposits, Kerr (1984a, 1984b) analyzed an offlap coarsening-upward sequence, typical of the marine sediments found throughout the basin of the Richardson and Rae Rivers. The present study discusses the evolution of the postglacial marine regression in more detail.

CHAPTER 3

3. SEDIMENTARY FACIES

Several lithological units have been identified within the stratigraphic sections from the basin of the Richardson and Rae Rivers. In order facilitate their description, they have been classified as facies on the basis of the distinguishing characteristics of grain size, primary sedimentary structures, organic content and spatial relationships. These facies not only reflect a wide range of energy levels, from upper to lower flow regimes, but are also representative of a wide spectrum of environments of deposition.

In this chapter, a description of the characteristic properties of each sedimentary facies is presented. The colour of air-dried samples of the different facies was determined with the use of a Munsell Colour Chart. The results of grain size analyses for each lithofacies are listed in Appendix C, as are histograms and cumulative frequency curves plotted with an arithmetic scale. With respect to granulometric parameters, mean refers to graphic mean, sorting refers to graphic standard deviation and skewness refers to graphic skewness. These parameters were derived graphically using the Folk and Ward (1957) formulae (Appendix A), and represent the average data of three samples from each lithofacies. Nomenclature of sand, silt and clay mixtures is after Shepard (1954), as indicated in Appendix B.

A total of nine sedimentary facies has been defined for classification, without implying any genetic connotations. These facies fall into three assemblages:

1. Diamicton facies assemblage
2. Coarse-grained facies assemblage
3. Fine-grained facies assemblage

The term diamicton is used in the sense defined by the INQUA Commission on Genesis and Lithology of Quaternary Deposits (1982): a descriptive term referring to any non-sorted or poorly-sorted sediment that contains a wide range of particle sizes. The distinction between coarse-grained and fine-grained facies is defined as mean sizes above and below the very fine sand-coarse silt boundary of the Wentworth size classification.

3.1 Diamicton Facies Assemblage

a) Facies A

Facies A is a very poorly sorted matrix-supported diamicton (Appendix A) whose silty-sand matrix has a mean grain size of 3.6 Ø. It is strongly coarse-skewed, indicating an excess of coarse material (Appendix A). On the average, this facies contains 6% clay, 42% silt and 52% sand as matrix (Appendix C-1) as well as clasts ranging up to boulders. The dominant class size is coarse to medium silt, representing approximately 30% of the matrix. Its colour according to the Munsell Chart is 7.5 YR 6/2. This corresponds to a light beige, although in the field where it has a high moisture content, it often appears as dark gray. However, where Facies A is observed directly overlying bedrock, its colour may vary from reddish brown to grayish green. The matrix-supported clasts consist of well-rounded gravel, cobble and boulders of various lithologies, through to angular slabs of sandstone and limestone, indicating that some material is derived from local Precambrian rocks. This diamicton is relatively compact and no preferred orientation of clasts was noted. Facies A is predominantly massive, as no internal structures were observed. It varies in thickness from 1 m to 6 m, forming distinct structureless beds, interstratified with Facies B. Lower contacts with Facies B and bedrock are sharp and irregular, whereas upper boundaries can be either sharp, planar to irregular or more rarely, gradational with Facies B, and sharp with respect to Facies H. Weathered surfaces of Facies A do not appear to differ in colour or texture from fresh exposures. Analysis of the clay-sized fraction from unweathered samples indicated the presence of the clay minerals chlorite and illite, as well as minor traces of dolomite, quartz and feldspar.

Shilts (1978) noted that chlorite and illite are the predominant phyllosilicates detectable in unweathered glacial sediments from the Canadian Shield.

b) Facies B

Facies B consists of a very poorly sorted heterogeneous mixture of variable sized-clasts with random orientation (Plate 1). The sandy silt matrix contains an average of 11.5% clay, 60% silt and 29.5% sand, but gravel, cobbles and boulders up to 2 m in diameter are common. The matrix mean grain size is 5.08 Ø and its dominant classes are very fine sand to medium silt, representing 60% of the matrix. The latter is fine-skewed, with an excess of fine material and has a light tan colour, 10 YR 6/1. This matrix-supported diamicton incorporates a wide range of particle sizes, from angular clasts of fragmented bedrock (limestone, dolomite and sandstone) to well-rounded clasts of variable composition, some of which are striated. Although not as compact as Facies A, this facies exhibits both massive texture and poorly developed bedding, with thickness of beds ranging from less than 0.5 m to 15 m. Occasionally, the upper zone of this facies is characterized by small lenses of sandy gravel which do not exceed 20 cm in thickness and 2 m to 3 m in length (Plate 2). When

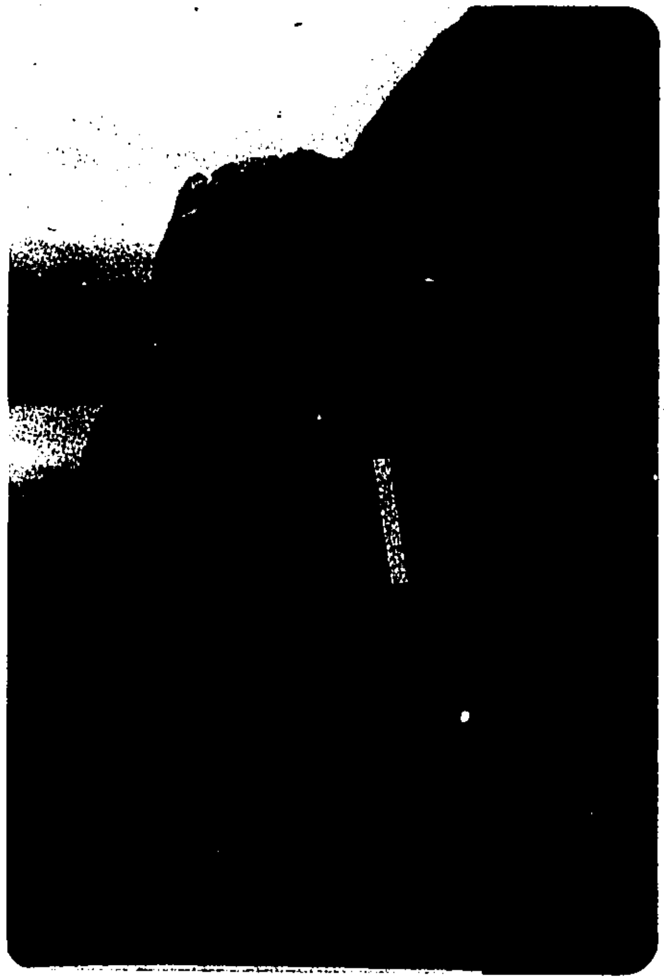


PLATE 1. Facies B exhibiting the poorly sorted nature of the diamicton.

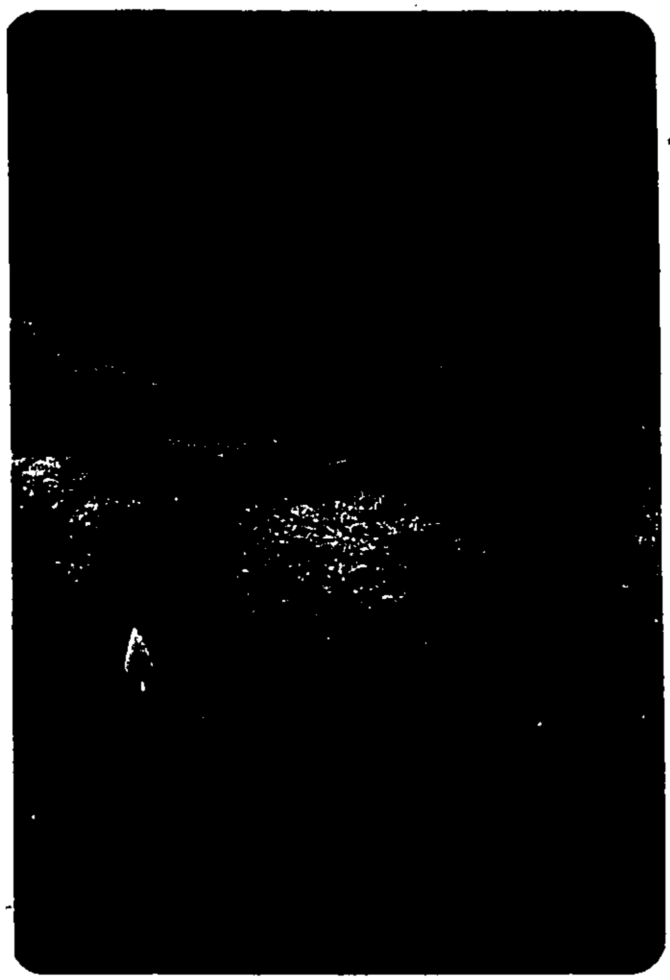


PLATE 2. Facies B overlain and underlain by units of Facies A. Note sandy gravel lenses in Facies B.

interbedded or occurring as lenses in Facies A, lower and upper contacts are sharp and planar. Upper gradational boundaries exist over a vertical distance of 0.3 m to 1 m where it is overlain by Facies H. Facies B can also occur interbedded with Facies H in which strata vary in thickness from as little as 5 cm to 4 m. In this context, internal laminations may be deformed by small rounded gravel clasts, and cobbles at the lower contact of Facies B disturb the underlying rhythmites of Facies H. There is no evidence for erosion of these rhythmites, nor is there any indication of flow in the diamicton unit. Weathered and fresh surfaces resemble each other and show no obvious differences.

3.2 Coarse-Grained Facies Assemblage

a) Facies C

Facies C is a moderately sorted cross-stratified sand, with a mean grain size of 1.13 Ø. It is generally fine-skewed, with grain sizes ranging from 0.9% clay, 1.4% silt and 97.7% sand (Appendix C-3). The dominant class sizes are medium to coarse sand, making up 80% of the facies, and it is light gray (7.5 YR 4/0) in colour. Well rounded gravel sized clasts up to 15 cm in diameter may be present, but only as a minor component, and are restricted to one site where they occur at the base of the unit (Section 25, Appendix D-25). Facies C is composed of poorly to well-bedded 10 cm to 40 cm thick tabular cross-bedded cosets that commonly dip from 18° to 30° towards 280° and 300°, as well as very minor parallel bedded sands. The lower boundary surfaces of the cosets form planar erosional contacts. Upper surfaces may be sharp or transitional to another coarse unit (Facies D). Occasionally these cross-bedded strata alternate between medium and coarse sands, 10 cm to 20 cm thick. Only rarely is horizontal parallel bedding observed, between cosets of cross-bedded sand. Where this occurs, the latter units are truncated by planar pebbly sands, 5 cm to 10 cm thick. Units of Facies C vary in thickness from 2.0 m to an observed maximum thickness of 4.5 m in Section 40 (Appendix D-40).

b) Facies D

Facies D is a moderately well sorted sand which consists of 0.8% clay, 0.9% silt and 98.7% sand with rare granules. The dominant class size is fine to medium sand (84%), with a mean grain size of 1.53 Ø. It is near symmetrical with respect to skewness (Appendix A), and light gray (5 YR 4/1) in colour.

It is generally composed of massive structureless sands, but locally may exhibit faint planar bedding and horizontal laminations, although bedding definition is poor and shows no well defined structure. This facies is variable in thickness, from 0.4 m to greater than 12 m (Section 24, Appendix D-24). The nature of the upper and lower contacts with the horizontally bedded sands of Facies C and E is gradational. At one location (Appendix D-40), it contains what appear to be sandy gravelly mud rip-up clasts derived from the underlying deposits of Facies A.

c) Facies E

Facies E is a moderately sorted silty sand, which is light gray (10 YR 6/2) in colour. Grain sizes range from 1.4% clay, 30.8% silt to 67.8% sand (Appendix C-5), but the dominant grain size is very fine sand to coarse silt, representing approximately 78% of the sample. Facies E is strongly fine skewed and has a mean grain size of 3.79 Ø. Its minimum and maximum unit thicknesses are 30 cm and 16 m respectively. This facies is composed of alternating cosets of planar laminations, ripple-drift cross-laminations and sinusoidal ripple laminations as described by Jopling and Walker (1968). Cross-laminated cosets are vertically stacked, and are separated from one another by horizontal boundaries which may be either erosional or transitional. A typical sequence consists of gradational type A and type B climbing ripples commonly found superimposed on one another, and overlain by draped undulating laminae which may in turn grade into or be truncated by horizontally laminated fine to medium sands.

The least common bedform is entirely composed of climbing sets of well-defined lee-side laminae with no preservation of stoss-side laminae (Plate 3). These are type A cosets, which are developed in the coarser sand fraction and have low angles of climb. Those observed were found to be not more than 10 cm thick.

The second type of ripple-drift cross-lamination, type B, is composed of climbing sets of lee-side laminae with complete preservation of stoss-side laminae (Plate 3). These laminae are continuous from one asymmetrical ripple to another. Each set, varying from 1 cm to 20 cm in thickness, may occasionally be overlain by a draped layer consisting of parallel clay laminations ranging in thickness from 0.2 cm to 0.5 cm which mimic the form of underlying ripples. Sequences of uninterrupted cosets of type B cross-laminations do not exceed 1 m in thickness, and have higher angles of climb than those of

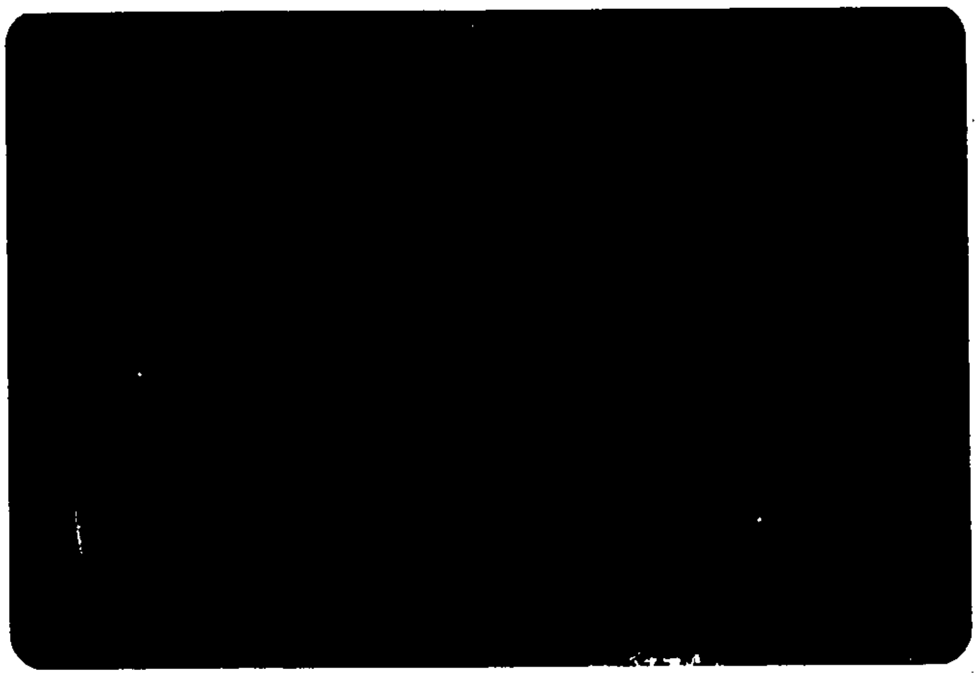


PLATE 3. Sequence of type A ripple-drift cross-laminations overlain by type B ripple-drift cross-laminations with clay draped laminations.

type A. Closer examination of some of these sets reveals that they are composed of alternating light and dark parallel laminae as evidenced by mica flakes which are distributed equally both on stoss and lee faces. For the most part, the crests and troughs of ripples are gently rounded, although sharp-crested peaks sometimes occur at the base of the individual cosets where there is a transition from type A to type B ripples. Palaeocurrent measurements of type B ripple-drift cross-laminations indicate variable flow directions towards 240° through to 315°.

Undulating laminae or sinusoidal ripple laminations are also present, but only as a minor component. They consist of superimposed sinusoidal laminations of silt and clay, which are laterally continuous. Both lee-side and stoss-side of the ripples are equal in thickness.

This facies is not particularly common in the stratigraphic sections studied, except in Section 22 (Appendix D-22). In all cases, upper and lower boundaries are gradational with both Facies D and G. The only effect of weathering on these sediments is differential erosion which emphasizes the clay-rich laminations overlying the sandier ripples.

d) Facies F

This facies consists of 1.2% clay, 1.5% silt and 97.3% sand, and with 95% of the components falling into the fine to medium sand classes, Facies F is the best sorted of the lithofacies in this study (Appendix A). Mean grain size is 2.02 Ø, and skewness is close to zero. Units of Facies F range from 1 m to 5 m in thickness, and consistently overlie Facies I. A gradational boundary always exists between these two facies. Although Facies F is generally composed of structureless light brown sand (10 YR 5/2), it may locally change upwards to discontinuous laminations, small pebbly lenses, and more rarely grade into alternating layers of silty sand and sandy silt, overlain by another unit of massive sand. The silty sand beds average 10 cm thick and do not show any internal stratification. The upper and lower contacts are gradational with the sandy silt beds. The latter contain no internal laminations and range in thickness from 2 cm to 10 cm. These rhythmites occur over a vertical distance of no more than 2 m within a section and have previously been described by Kerr (1984b).

This predominantly massive sandy facies is fossiliferous, containing both marine micro and macrofauna. Although articulated bivalves, some in life position, and disarticulated bivalves are equally common, no bioturbation features were observed. Certain horizons within these sands are more oxidized than others, suggesting preferred horizons of groundwater flow or an increase in the amount of groundwater.

3.3 Fine-Grained Facies Assemblage

a) Facies G

Facies G is a moderately sorted light beige (10 YR 6/1) silt with grain size consisting of 3.5% clay, 94.4% silt and 2.1% sand. Mean grain size is 5.35 Ø, as indicated by the predominance (78%) of medium to coarse silt. It is moderately fine-skewed with an excess of fine silt and clay. Structure varies from massive to horizontal finely laminated silts, to type A and type B ripple-drift cross-laminations, overlain by silty sinusoidal laminations. This vertical sequence, whether partial or complete, forms individual beds 0.2 m to 3 m thick, that are separated by 1 cm to 8 cm thick red-brown to gray clay layers which may have either sharp or transitional upper and lower contacts. Although massive to finely laminated strata dominate the succession, type B cross-laminations are nevertheless relatively common (Plate 4), and only rarely does type A occur.

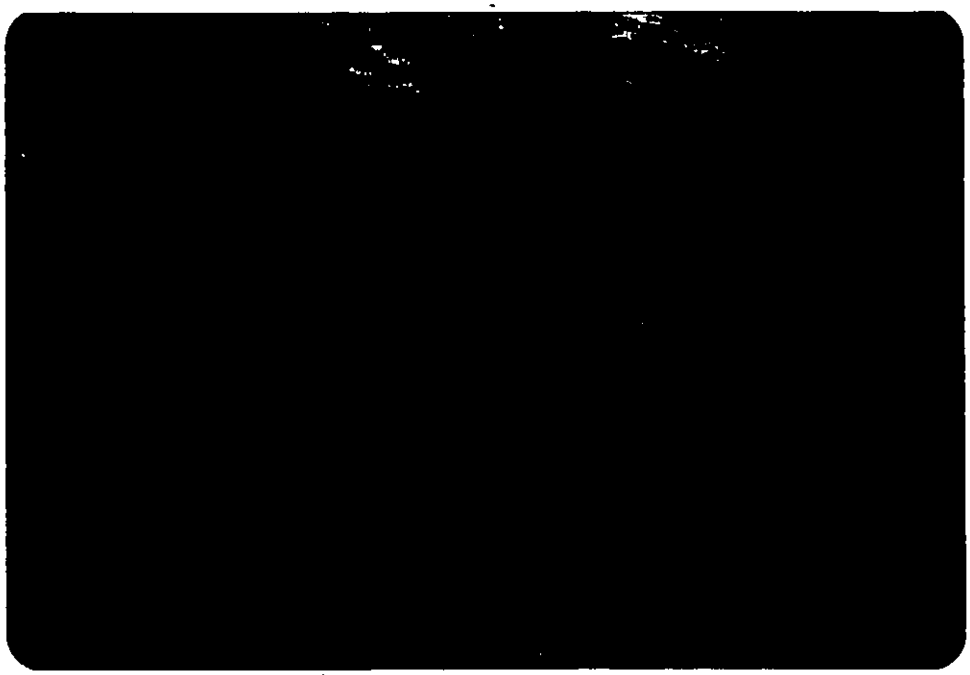


PLATE 4. Type B ripple-drift cross-laminations of Facies G.

Type B climbing ripples are restricted to 20 cm to 40 cm thick cosets which are usually overlain and underlain by transitional undulating laminae. Some of these ripples appear to have sharper crests and steeper angles of climb than those of Facies E. Neither draped clay laminations nor alternating dark and light laminae were observed within the ripple cosets. Palaeocurrent directions towards 250° through to 340° have been recorded at various sites. The thickness of Facies G varies from 0.4 m to 10.5 m. In all sections, this facies is overlain by Facies H and underlain either by Facies H or Facies E. The nature of the boundaries with respect to Facies G is always gradational.

b) Facies H

This facies is a major component of most of the sections studied. It is characterized by couplets consisting of two texturally distinct layers: a light gray (10 YR 6/1) silty clay (Hb, Appendix C-9) and a darker brownish gray (10 YR 8/1) clay (Ha, Appendix C-8). The silty clay layers are poorly sorted, and on average consist of 57% clay and 43% silt. The mean grain size of Hb is 8.2 Ø, and the dominant size classes are fine silt to clay (68%), indicating a fine-skewed distribution. The clay-sized fraction consists largely of illite and chlorite, as well as minor dolomite, feldspar and quartz.

The silt layers, being consistently thicker than the clays, are composed of discrete horizontal laminations of varying grain size and thickness, numbering up to 45 per layer (Plate 5), which may show normal and reverse grading. These internal laminae are characteristic of the silt layers within the basal couplets of sections; only on occasion are such laminae found in the rhythmites stratigraphically higher. Sandy laminations are also present within some of these layers, and more rarely couplets are separated by thin diamicton beds (Facies B), as seen in Plate 6, frequently less than 20 cm thick. Although the thickness of the silty clay layers is relatively constant within a given vertical range, the samples measured within the basin varied from less than 1 cm at the top of sections to 40 cm thick at the base of sections.

The thinner, darker moderately sorted clay layers have a mean grain size close to 9.3 Ø. The particles consist on average of 91.4% clay and 8.6% silt. These sediments are strongly coarse-skewed, having an excess of relatively coarse material. Clay minerals consist of chlorite and illite, although

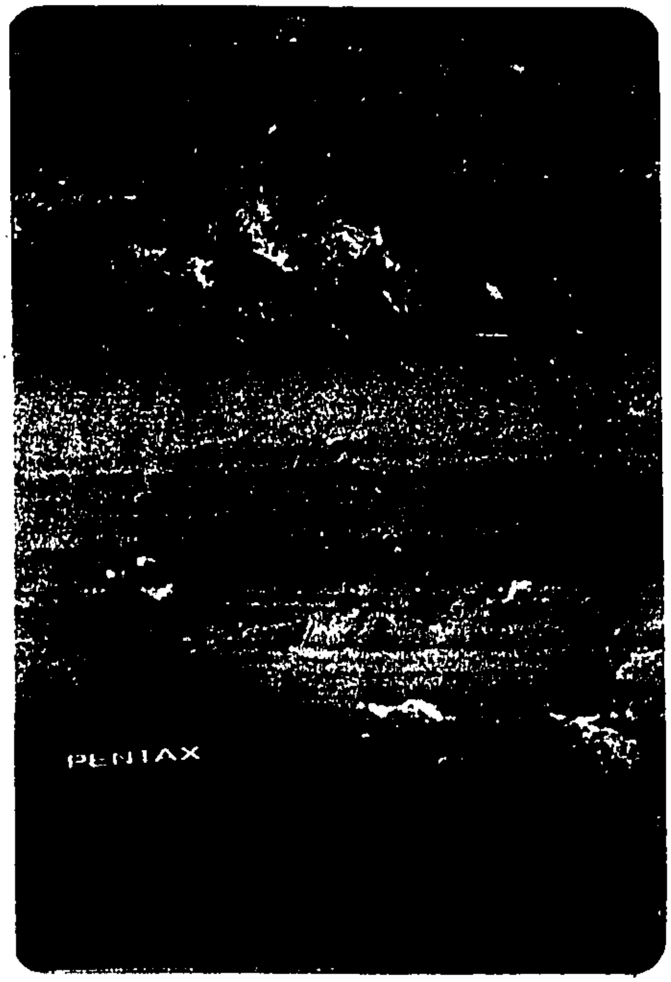


PLATE 5. Proximal glaciolacustrine rhythmite showing multilaminated nature of the summer layers.

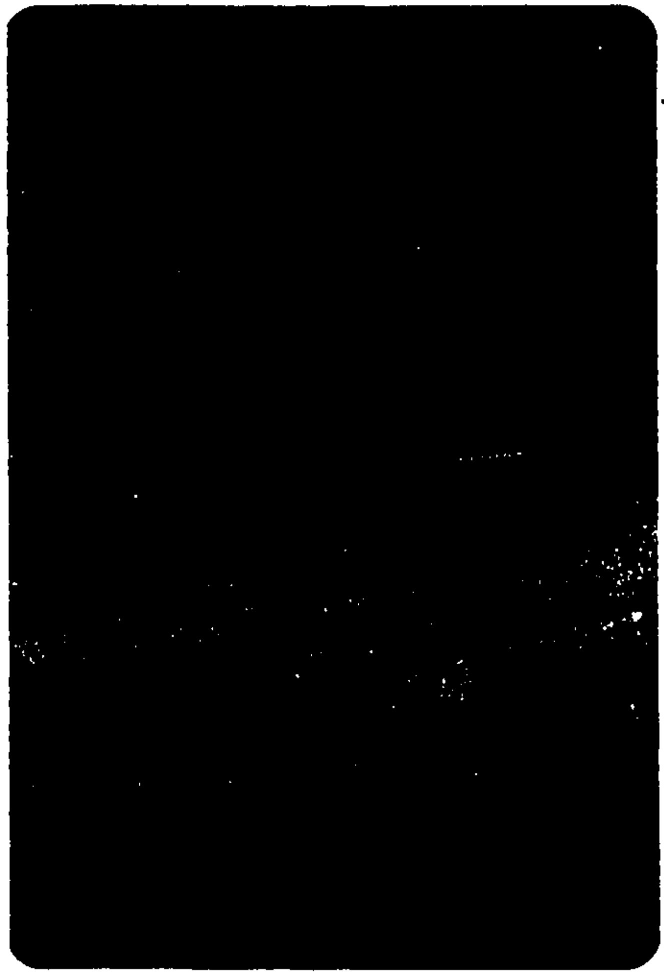


PLATE 6. Thin diamicton bed (Facies B) interbedded with glaciolacustrine rhythmites.

clay-sized quartz, feldspar and dolomite are also present. The clay layers vary in thickness from a few mm in the rhythmites near the stratigraphic top of sections, to 10 cm in basal couplets. They appear to be predominantly massive, but may exhibit 1 to 5 internal laminations 1 mm to 2 mm thick within the thicker clay layers and occasionally, contain thin sand laminae with rare pebbles.

Within a couplet, the contact between silty clay and overlying clay can be either sharp or gradational over a very short vertical distance. Boundaries separating clay layers from overlying silty clay layers are sharp and planar, but irregular erosional contacts have also been observed in some localities.

Generally, there is an overall trend for the couplets to decrease in thickness upwards (Plate 7), although there is locally considerable fluctuation within this trend. The rhythmites do not appear to thin or thicken laterally over a distance of up to 20 m within a given section, although on a regional scale the thickness of Facies H varies considerably, from approximately 1 m in the eastern part of the basin to at least 26 m in the central part.

Other features present in Facies H are intra-formational deformation structures and faulting. There are several cases of contorted bedding showing

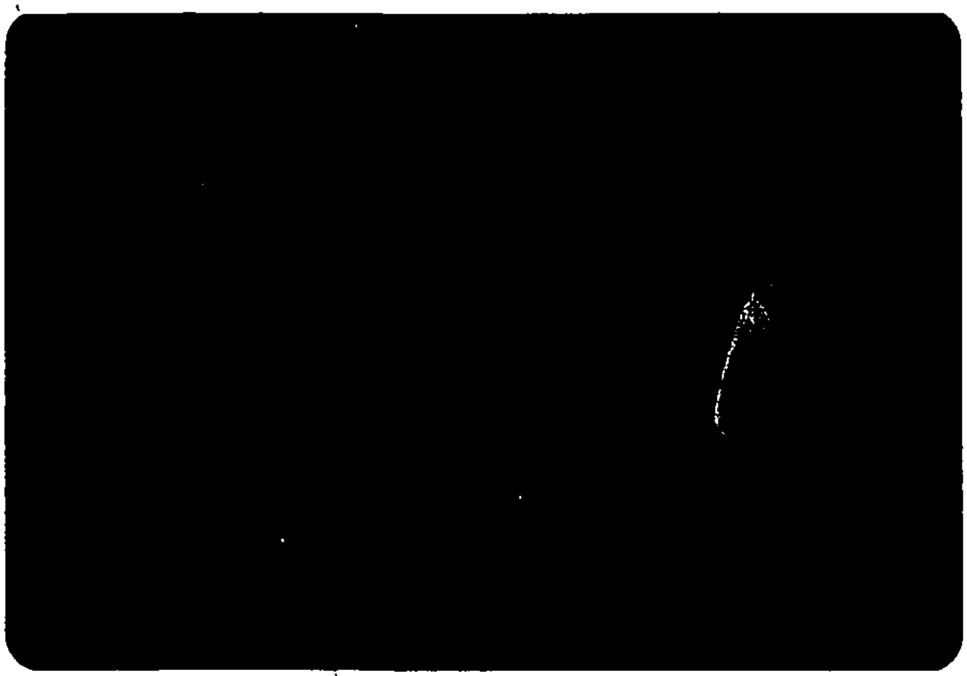


PLATE 7. Distal varved couplets averaging 2 cm thick.

small-scale folding and diapiric structures, 0.2 m to 1.5 m in amplitude, overlain and underlain by undisturbed horizontally bedded rhythmites (Plate 8). Present in only two sections is high-angle normal micro-faulting of the finely laminated basal couplets, with a displacement of no more than 5 cm. Also commonly found in Facies H are isolated dropstones which disturb laminae, depressing the underlying rhythmites and which are overlain by draped rhythmites. There is a distinct difference in colour between the silty clay and clay layers which does not appear to be due to any mineralogical variation, but rather to changes in grain size. The colour difference appears also to be dependent upon the water content; layering becomes apparent in samples with a high moisture content, but sections which have dried out completely do not show any noticeable layering. In addition to this, some rhythmites directly overlain by Facies I are salty, i.e., have a salty taste.

Facies H is overlain by either Facies B, G or I and the nature of the upper contact is sharp with respect to Facies B and G, and gradational with Facies I. It is underlain by Facies A where lower contacts are sharp, by Facies B where contacts are sharp or gradational, and by Facies G where contacts are transitional.

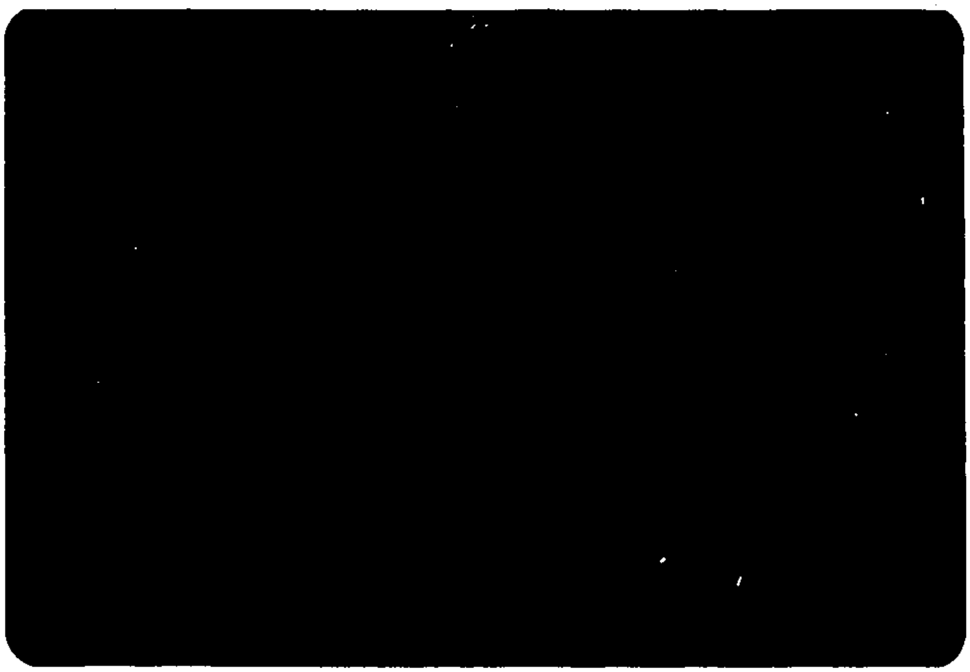


PLATE 8. Soft sediment deformation consisting of intraformational crumplings and diapir structure.

c) Facies I

This facies is composed of poorly sorted dark (10 YR 6/2) silt, and on average contains 20.3% clay, 77.6% silt and 2.1% sand. Mean grain size is 6.74 Ø, and the dominant class sizes are fine to medium silt, representing 56% of the sediment. It is strongly fine-skewed, indicating an excess of fine material. Composition of the clay minerals is identical to that of the other lithofacies: chlorite, illite and clay-sized accessory minerals such as dolomite, feldspar and quartz. Facies I is for the most part massive and structureless. It varies in thickness from 4 m to 17 m in the eastern and central parts of the basin, and decreases to as little as 0.4 m further west. Upper contacts are gradational with overlying facies, whereas the lower boundaries are sharp and planar to irregular with Facies B and H (Plate 9). Like Facies F, this unit is also fossiliferous, containing both marine micro and macrofossils. Pebbly sands with gravel sized clasts up to 4 cm in diameter were observed at the base of this facies at the contact with Facies H, but only in one section (Section 35, Appendix D-35).

PLATE 9. Massive marine silt of Facies I overlying the silty clay rhythmites of Facies H.

CHAPTER 4

4. STRATIGRAPHY

4.1 Facies Relationships

In order to describe the spatial and temporal distributions of lithofacies, a number of stratigraphic sections were measured in different parts of the basin of the Richardson and Rae Rivers. The location of these sections is shown in Figure 3. The stratigraphic data were projected onto a single base line oriented E-W throughout the basin, as seen in Figure 4 (in pocket).

A total of 40 stratigraphic sections have been studied. The characteristic properties of each are presented in Appendix C. They are numbered from Section 1, representing the most elevated and western location, to Section 40, the eastern-most site along a tributary of the Richardson River. Elevations were determined from 1:50,000 scale topographic map sheets of the area.

As noted by Karrow (1984), individual lithofacies are sometimes difficult to interpret when considered alone, but the recognition of lithofacies associations and vertical sequences is a useful analytical tool. It is therefore necessary to define specific associations which group together different sedimentary facies that are environmentally or genetically related. Any preferred vertical position of a facies may also have an

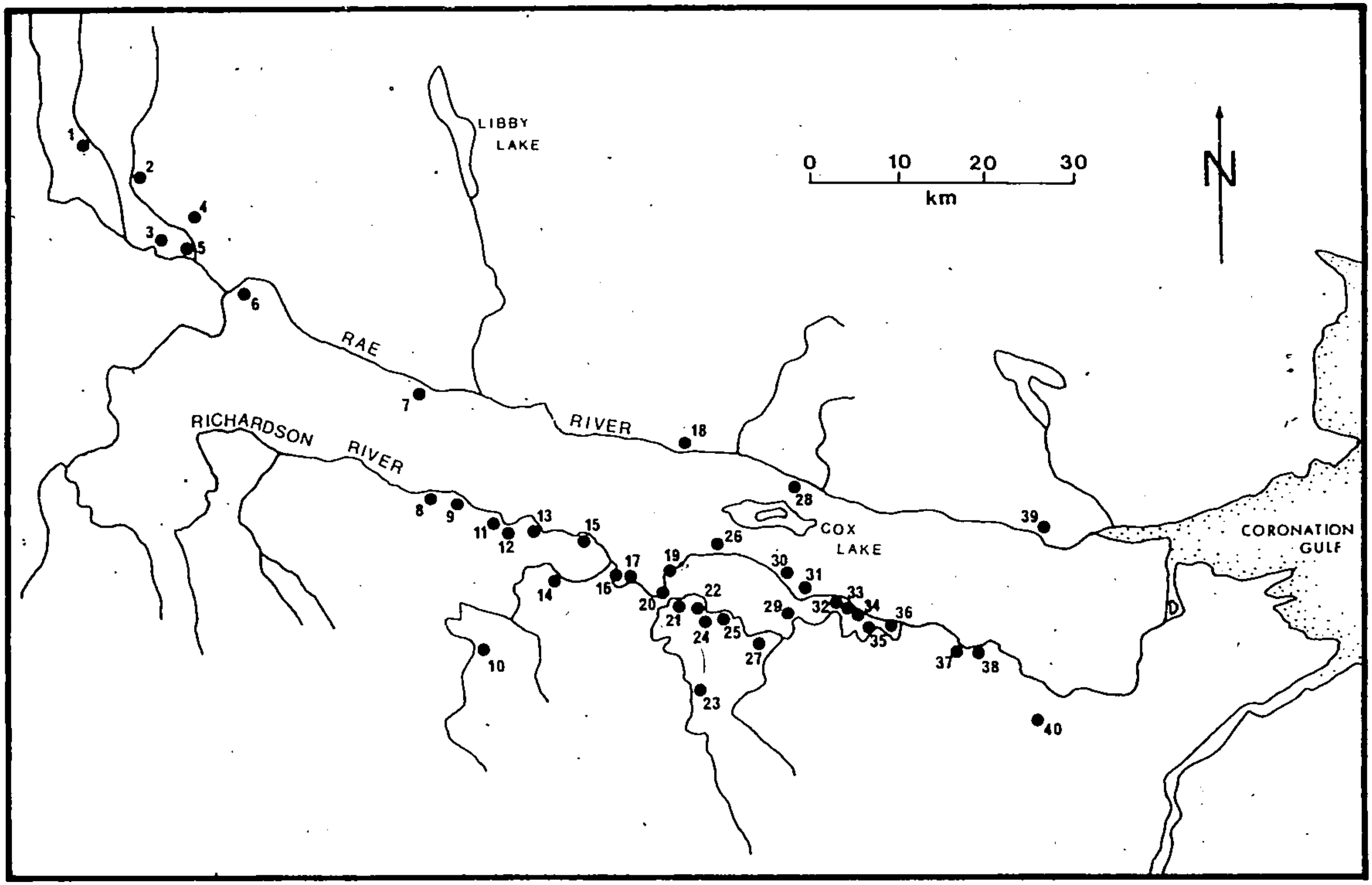


FIGURE 3. Location of stratigraphic sections.

important bearing on the interpretation of a sequence (Shaw, 1975), and must be considered. According to Boulton (1972), sequences of deposition can be established based on patterns of sediment distribution when they occur repeatedly.

In this study, five different lithofacies associations have been identified from sedimentary successions:

1. Facies A and B
2. Facies B with interbedded subaqueous outwash
3. Facies C, D and E
4. Facies G and H
5. Facies I and F

The first four are the result of deglaciation processes. During the easterly down-gradient retreat of a lobe of the Laurentide Ice Sheet from the basin of the Richardson and Rae Rivers, ponding of meltwater between the ice front and the surrounding highlands led to the formation of a series of proglacial lakes. Both till and glaciofluvial sediments were deposited individually and simultaneously along the ice margin. As the ice liberated the highlands to the northeast of the basin, the marine incursion took place, permitting seawater to mix with the freshwater of Glacial Lake Richardson. The fifth lithofacies relationship is the product of the marine regression resulting from postglacial uplift of the land. This period is characterized by a coarsening-upward off-lap sequence. Uniform rates of deglaciation over a wide area of similar topography may produce a widespread single style of deposition (Boulton, 1972). However, this does not seem to have been the case in the present study as three styles of deposition have been inferred:

1. Glacial till, occasionally lying on smooth and striated bedrock, is the only sediment deposited by the retreating glacier.
2. The retreat of a glacier can be represented largely by beds of glacial drift of different origins, with minor stratified beds of sand and gravel within them. Outwash deposits may also occur above and below glacial drift, although the latter case was not observed in the study area.
3. In some areas, subaqueous outwash deposits dominate the environment of deposition.

In the first case, there is an association between Facies A and B, as suggested by their bedding relationships and the nature of the various types of contacts.

In Sections 6 and 34 (Appendices D-6 and D-34), these two diamicton facies occur as interbedded units of different thicknesses. Although abrupt boundaries are more common (Plate 10), vertical gradational contacts are also possible between these facies, as seen in Plate 11. No lateral variations or transitions in the composition of the glacially-derived diamictons were detected.

A second sedimentary association is expressed in Section 32, where Facies B contains a sand and gravel lens (Plate 2), and in Section 40, where Facies A appears to grade up vertically into Facies D. This suggests that the deposition of both diamicton facies alternated, at least locally, with periods of subaqueous glaciofluvial activity.

In his model of glaciolacustrine sedimentary environments, May (1977) stated that various types of till can be found in stratigraphic sequences in close relationship with lacustrine deposits, notably varved silts and clays. As observed in the stratigraphic record from the basin of the Richardson and Rae Rivers, such associations are relatively common. Three types of transitions occur between Facies A and B, and Facies H. Sharp contacts frequently separate the silty clay rhythmites of Facies H from the diamicton of Facies A (Plate 12). In other cases, Facies B grades directly into Facies H, as seen in Plate 13. A third possibility is the passage

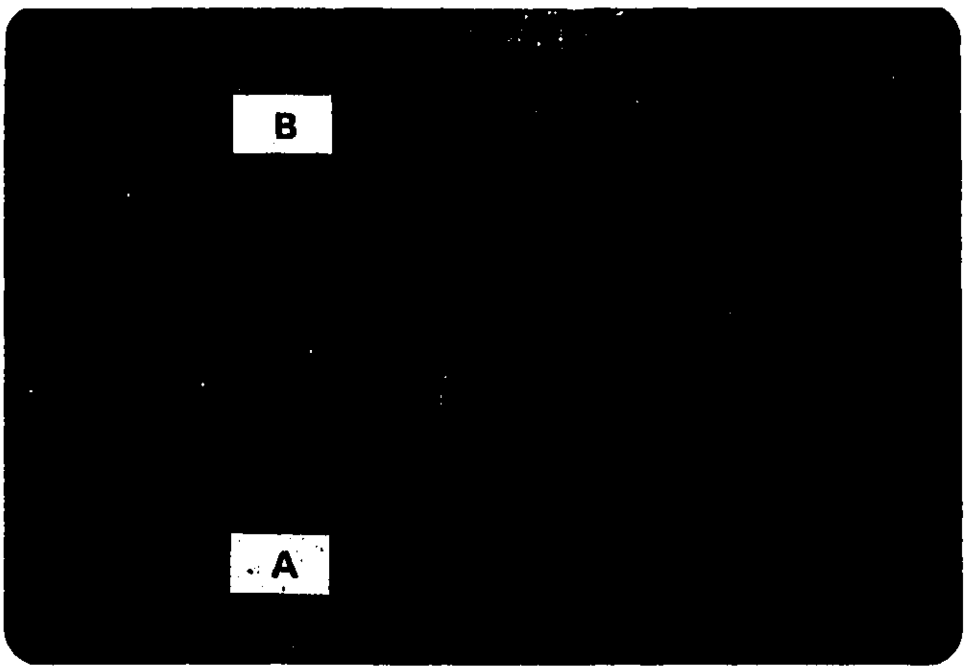


PLATE 10. Poorly sorted diamicton of Facies A sharply overlain by silt-rich diamicton of Facies B.

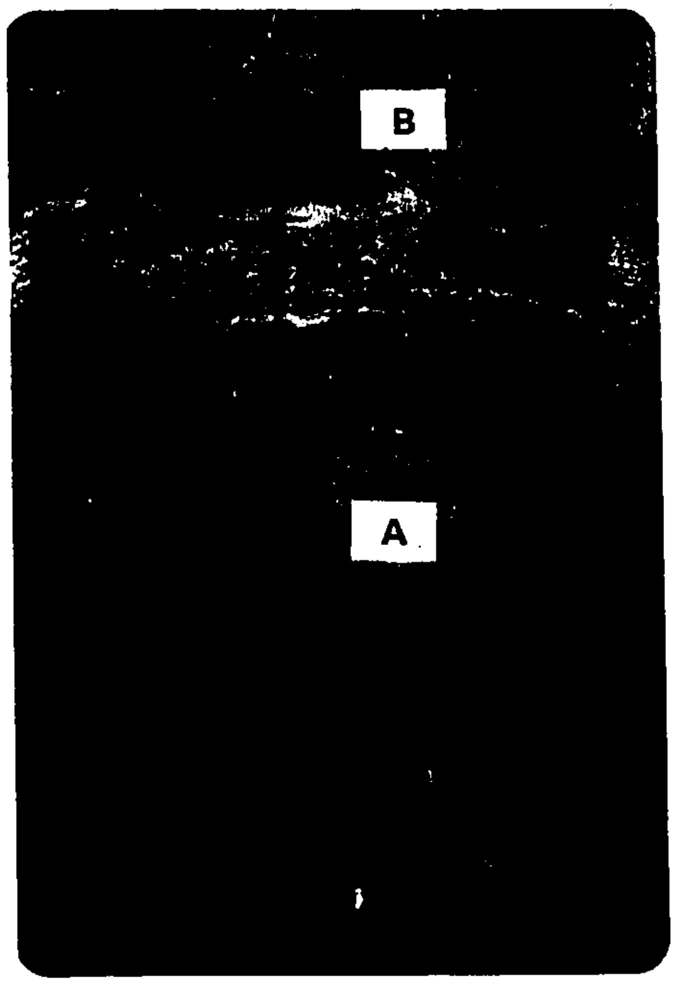


PLATE 11. View of Facies A grading upward into Facies B.

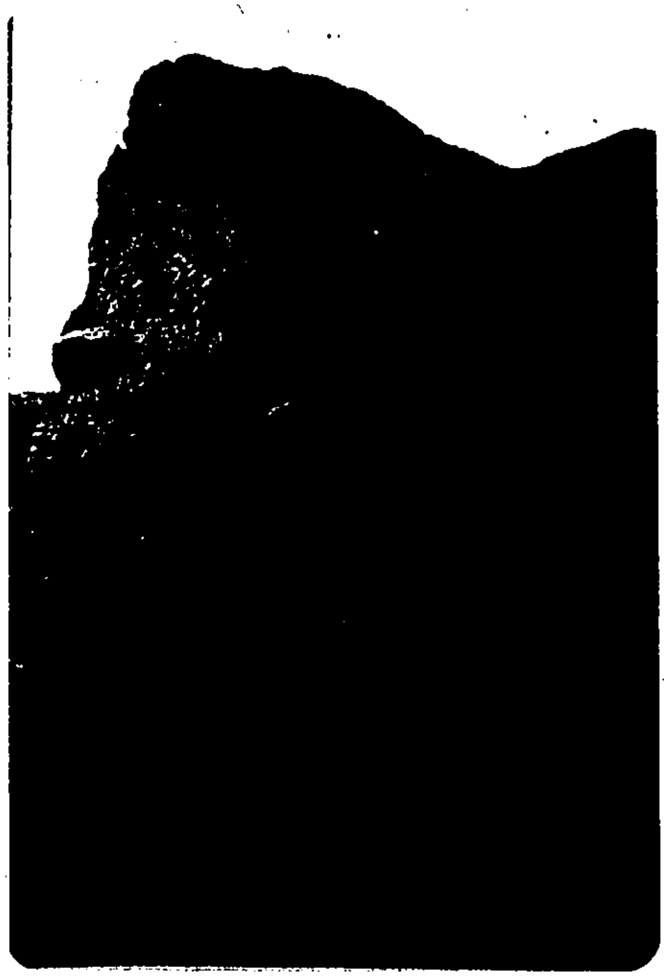


PLATE 12. Diamicton layer of Facies A sharply overlain by glaciolacustrine rhythmites of Facies H.

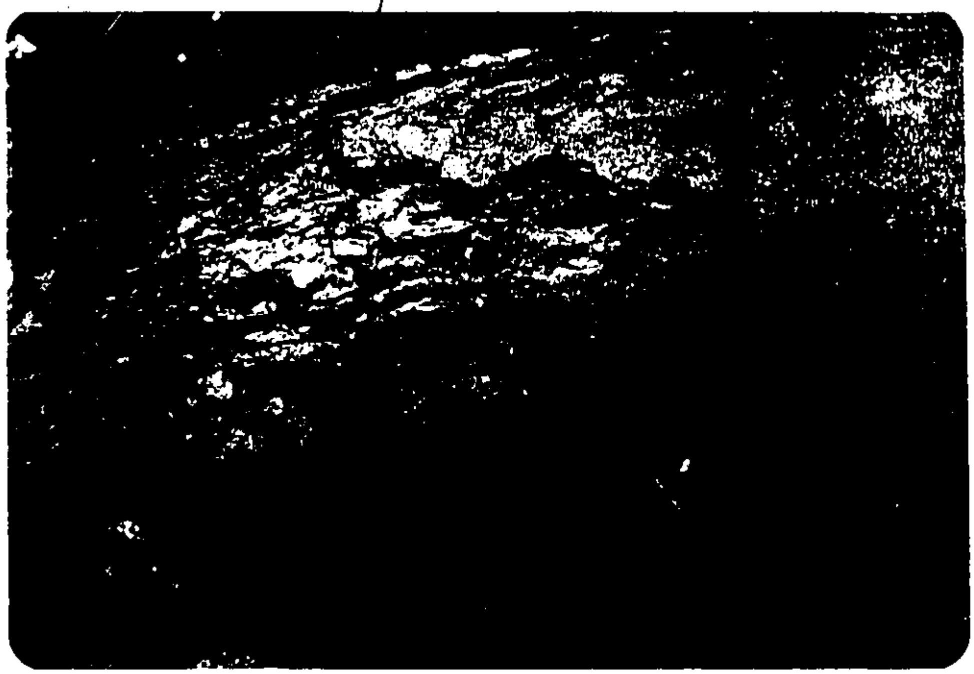


PLATE 13. Direct gradational contact between rhythmites of Facies H underlain by diamicton unit of Facies B.

of Facies B into Facies H by an intermediate process of alternating horizons of thinly bedded silty layers and diamicton layers over a short vertical distance (Plate 14). However, the latter transition was observed only in Section 8.

As noted earlier, distinct beds of Facies B are found interbedded with Facies H. They occur throughout the basin and vary in thickness from 0.1 m to 4.25 m. The relationship of Facies B with Facies A and H is indicative of an environment dominated by glacial influence.

A third lithofacies association is inferred from the occurrence of Facies C, D and E over short vertical distances in a number of stratigraphic sections. These are all basically coarse-grained sediments characterized by the presence of Facies C at the base (Section 25), overlain by Facies D which is in turn followed by Facies E, a transitional phase between Facies D and G. Several variations of this basic succession may exist, as shown in Sections 11, 16, 22, 28 and 40. It is however, the vertical position of these three facies within the sections studied which has interpretative importance. They occur consistently throughout the basin, underlying thick successions of silty clay rhythmites (Facies H), forming part of a general fining-upward sequence.

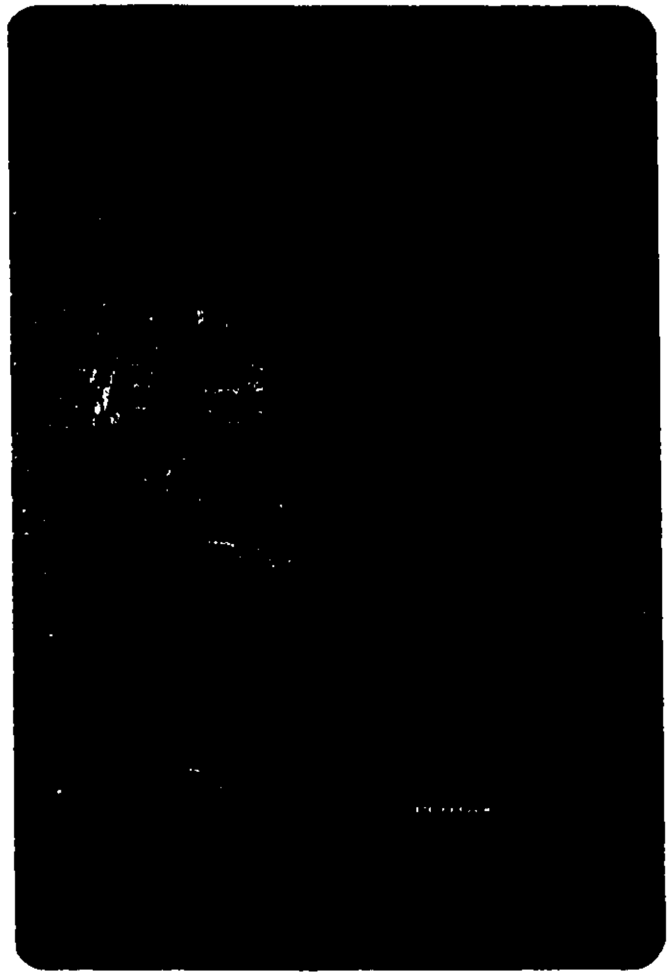


PLATE 14. Thinly interbedded diamicton and rhythmite layers at contact between Facies H (top) and Facies B (bottom).

Cheel and Rust (1982) described a similar vertical succession of sediments from a retreating ice front, comprising a fining-upward sequence, with gravelly sands at the base and increasingly distal fine-grained sediments at higher stratigraphic levels, passing into a marine environment, as opposed to a lacustrine water body as is the case in the present study.

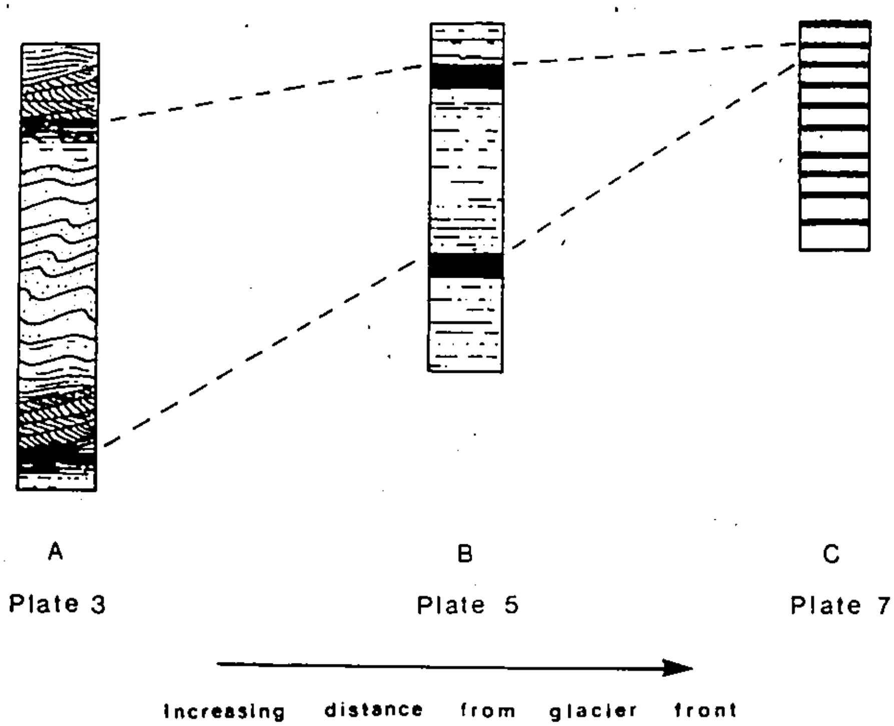
Lithofacies associated with proximal positions close to an ice margin are transitional between glaciofluvial and glaciolacustrine environments. For example, at Sections 25 and 40, thick bodies of cross-bedded (Facies C) and massive beds (Facies D) are predominant in the lower parts, and are developed in the coarser sediments. The stratigraphic position of these sands suggests that they were deposited in a proximal lacustrine environment.

Each lithofacies association can be related to changes in sedimentary processes which are a function of seasonal variation and distance from the ice front (Agterberg and Banerjee, 1969). Distance from the ice front has a significant influence on sedimentary associations, and a single depositional process may generate a number of different lithofacies in the downcurrent direction. Although lateral variations between Facies C, D and E were not directly observed, field evidence suggests that lateral transitions are probable in which Facies C grades downstream into Facies D and eventually to Facies E in a more distal environment.

According to Banerjee and McDonald (1975), the downstream relationship of relatively coarse-grained esker sediments with rhythmically bedded silt and clay illustrates the important genetic relationship between glaciofluvial (Facies C, D and E) and glaciolacustrine (Facies G, H) sediments.

The fourth sedimentary relationship is that of Facies G and H, the latter dominating the majority of the sections in the basin. The silt-rich sediments of Facies G appear to be gradational from Facies E to Facies H. In this context, they are laterally extensive, occurring at the base of Facies H at many sites throughout the basin (Sections 11, 16, 24, 25, 28, 29 and 40). An exception is found in Sections 19 and 20 where Facies G is underlain and overlain by Facies H, and is of considerable thickness. This poses a particular problem with respect to the interpretation of these sequences and will be discussed in Chapter 5. Figure 5 illustrates an ideal downcurrent transition between lithofacies resulting from a dominant depositional process; these transitions have been previously described by Ashley (1975) and Smith and Ashley (1985), and are largely attributed to the action of turbidity currents over a one year period.

FIGURE 5. Hypothetical lateral facies variation of an annual couplet in a glaciolacustrine environment.



A. Subaqueous outwash deposits dominated by current-bedded structures characteristic of Facies E.

B. Proximal (basal) rhythmites of Facies G that consist of multiple normal-graded flat beds composed of silt and very fine sand, grading upward into clay.

C. Thin silty-clay couplets of Facies H are the more distal lateral equivalent of ripple-drift laminae of Facies E.

Facies I and F comprise the fifth lithofacies association inferred from the Late Quaternary deposits in the study area. Together they form a coarsening-upward cycle in which clayey-silt grades upward through to a thin silty sand horizon and into well sorted sands. In two sections studied, the fine-grained sediments of Facies I are overlain by a coarsening upward sand sequence which is not characteristic of Facies F. These two sections, 20 and 23, are described in further detail in Chapter 5.

4.2 Correlative Stratigraphy

In order to discuss depositional processes and environmental conditions, it is first necessary to consider proportional thicknesses of facies and their vertical position. This approach provides a more complete overview of the stratigraphic record which may then be incorporated with specific discussion in Chapter 5 on sedimentary properties and processes in various glacial and marine environments. As noted earlier, lateral variations in lithofacies must be considered in a stratigraphic context when interpretation of several profiles are made from widely separated sections. Therefore, profiles in Figure 4 (in pocket) were placed in the context of a stratigraphic datum so that proper relationships could be established. A lithostratigraphic correlation of facies between sections was undertaken in order to establish a more accurate reconstruction of the depositional environments which evolved during the retreat of the last glacier and in postglacial time.

Correlation of time equivalent units between sections has not been attempted as it was not possible to define specific lateral variations in facies based on field observations. The lack of appropriately situated sites where significant facies gradations were expected did not permit a chronostratigraphic study of the relatively widely separated sections. In addition, such a study would have been complicated by the fact that at any one locality, aspects of both proximal and distal facies can be expected, and that local variations in depositional style could result from topographic diversity, a disorderly retreat, and an irregular configuration of the ice sheet.

The spatial distribution of the principal lithofacies is present in Figure 4. In the western part of the basin, the rhythmites of Facies H are dominant in most sections and have an overall high proportional thickness. The present study shows that Facies H can be considered the most widely distributed lithofacies in the Richardson and Rae River basin. Rhythmites were previously reported only between 30 m and 145 m a.s.l. and said to be restricted to the western part of the Richardson River basin (Mercier, 1986, p. 84, 87). However, the present writer has recorded Facies H in all parts of the basin ranging in altitude from as low as 3 m a.s.l. (Section 35) to greater than 205 m a.s.l. (Section 1), 50 m above the marine limit. The fine grain size and lateral continuity of Facies H suggests that the silty clay couplets resulted from turbidity currents, a process which transported fines over several kilometres, enabling them to cover relatively wide areas. The thickest deposits of Facies H occur in depressions in what would have been low-lying regions of the lake floor, whereas the thinner deposits normally are found over topographic highs in the basin.

In the eastern part of the basin, the marine silts of Facies I are the most important deposits observed in stratigraphic sections. They are thickest in Sections 35, 36, 37 and 38, and thin towards the west in the form of a wedge. Their presence has been noted at elevations ranging from 4 m a.s.l. (Section 36) to 76 m a.s.l. (Section 22) in stratigraphic sections, but are commonly found up to 150 m-170 m a.s.l. as surficial deposits. This facies is overlain by the sands of Facies F which occur between 18 m a.s.l. (Sections 37 and 38) and 80 m a.s.l. (Section 22). These sands have also been found up to 90 m a.s.l. as surficial deposits. Thus, the upper zone of most sections in the eastern part of the basin are characterized by a sedimentary succession dominated by two facies, namely Facies I and F. These sequences illustrate the close association between the silts and overlying sands.

In the central part of the basin, there is an overlap of the two most volumetrically significant lithofacies, i.e., marine deposits of Facies I overlying the glaciolacustrine rhythmites of Facies H, either directly or they may be locally separated from each other by Facies B.

The oldest unconsolidated deposits in the region are probably the diamicton facies which lie directly on bedrock, as seen at the base of Sections 32, 33, 34 and 39 (Figure 4). Facies A at the first three sections can be laterally correlated, although the continuity of this facies east of Section 34 and west of Section 32 cannot be determined. Depending on the site, Facies A is covered by either rhythmites (Facies H) at Sections 32 and 33 or sands (Facies C, D and E) in Section 40. Facies B appears to have more of a regional distribution as it occurs at a number of sites throughout the basin, at the base of Sections 2, 3, 5, 6, 7, 8, 9, 32 and 34. It can be correlated almost continuously in the western part of the basin over a distance of over 40 km (Figure 4).

Where it is found interbedded with Facies H (Sections 21 and 26), its spatial distribution cannot be established. However, in Sections 30 and 31, Facies B can be correlated over a distance of ten's of metres, decreasing in thickness towards the west.

Three commonly associated lithofacies, Facies C, D and E, are also relatively widespread. Sections such as 24, 25 and 40 clearly show the preferred positions for these three facies, and provide important evidence as to the nature of sedimentation and the environment of deposition for a large part of the basin. Their proximity to eskerine surficial deposits suggests a close relationship between eskers and Facies C, D and E. These facies are also found as more isolated occurrences such as in Sections 11, 16 and 28 which make it impossible to trace them laterally to sections on either side of them. In contrast, Section 25 passes from Facies C which can be correlated with Facies D in Section 24 and Facies E of Section 22 (Figure 4). This type of correlation is based on the assumption that Facies C grades into Facies ~~D~~ and E in a downcurrent direction to the west. Basinward, this variation in lateral facies changes is related to differences in slope and energy conditions in the lacustrine environment, as material was transported upgradient away from the glacier. The generally thick sequences are indicative of a proximal, major source of sediment.

These three coarse-grained facies (C, D and E) grade up vertically into Facies G and H, as observed in a number of stratigraphic sections throughout the basin. In most sections, Facies G contributes only relatively small thicknesses (Sections 11, 16, 24, 25, 28 and 40) and is not considered to be a laterally extensive unit, although it can be correlated between Sections 24 and 25 over a short distance. An exception to this can be found in Sections 19 and 20, where Facies G occurs as a comparatively thick unit. This deposit may be the distal equivalent corresponding to the coarser grained outwash sediments at the base of Sections 22, 24 and 25, since Sections 19 and 20 are generally aligned with the palaeocurrent measurements of Sections 24 and 25. However, the presence of rhythmites under Facies G in Sections 19 and 20, and their apparent absence under Facies C, D and E in Sections 24 and 25 is problematical, and will be further discussed in Chapter 5.

As noted earlier, Facies H dominates in successions deposited in distal positions in the lake. Noticeable regional variations in thickness of this facies have been observed in the eastern, central and western parts of the Richardson and Rae River basin. In general, the thickness is greatest in the central area and becomes thinner to the east and to the west. In the east, this facies attains a maximum thickness of 2 m (Section 40), whereas to the west, a minimum thickness of 5 m was noted (Section 1), although this section has undergone a certain amount of surficial erosion which cannot be determined. Thickness increases gradually towards the basin center, where 24 m to 26 m thick sequences are common (Sections 12, 13 and 17).

Facies H is sharply overlain by Facies I (Plate 9), a contact which is readily visible in Sections 20, 22, 25, 27, 29, 35 and 40. Facies I has been correlated between Sections 35, 36, 37 and 38 (Figure 4). East of this last site, it has been eroded in more elevated areas, but reappears in Section 40. West of Section 35, Sections 32, 33 and 34 have also been eroded down to Facies H. However, Facies I reappears in Section 31 and can be correlated almost continuously (Figure 4) through to Section 20, over a distance of 14 km. West of this site, it was not observed in stratigraphic sections as most river-bank sections form part of a "badlands" topography in which the tops of sections are frequently missing due to erosion. Other than in areas where surface erosion has taken place, Facies I and F overlie all others, completing a diverse sedimentary succession.

CHAPTER 5

5. ENVIRONMENTS OF DEPOSITION

In this chapter an interpretation of the lithofacies is presented, based on a comparative study of modern sedimentary environments as well as other well-studied similar deposits of Pleistocene age. In the context of the regional setting, the principal sedimentary environments have been classified into four phases. These constitute a succession of depositional environments which accounts for both lithofacies relationships and spatial distribution. As a result of deglaciation of the Richardson and Rae River basin and the subsequent marine transgression and regression, a distinct relationship was developed between environments of deposition and the resulting stratigraphic sequences.

An eastward retreating ice front of a lobe of the Laurentide Ice Sheet, blocking the natural drainage system of the basin, led to the ponding of glacial meltwater between the ice front and surrounding highlands. As a result, a dammed glacial lake evolved throughout the deglaciation of the basin. This series of glacial lakes, called Glacial Lake Richardson (Mercier, 1984), provides a framework around which different depositional environments developed.

5.1 The Glacial Phase

In the stratigraphic record, ice deposited material is restricted to Facies A and B, which are closely associated in stratigraphic sequences, and have been interpreted as tills. In some areas the retreat of a glacier is initially represented stratigraphically by till containing minor stratified beds and lenses which tends to lie on smooth or striated bedrock (Boulton, 1972). Similar sequences are recorded in Sections 32, 33, 34 and 39, where either Facies A or B directly overlies bedrock, and in Section 6, where Facies B is underlain by Facies A. Such multi-till sequences have frequently been misinterpreted in the past as due to multiple glaciations, as noted by Boulton (1972), Clayton (1973), Shilts (1981), and Eyles et al., (1983). However, in the present case a study of the individual facies shows that they were deposited in a similar environment but by different processes.

Facies A has been interpreted as basal (lodgement) till on the basis of sedimentological and stratigraphic evidence. Basal till is usually a dense matrix-supported diamicton with locally high clast concentrations in the lower portions (Eyles et al., 1983), which can appear compact and may not show any preferred orientation of clasts (Kruger and Marcussen, 1976). Facies A does not occur as successive tabular stratabound units and therefore is probably not the product of mudflow processes. The similarity of Facies A to surficial till within the basin that is fluted and drumlinized also suggests that this lithofacies is a subglacial lodgement till deposited from actively moving ice, as opposed to stationary dead ice.

With respect to temperate glaciers, Boulton (1972) noted that basal melting almost always deposits lodgement till, which is generally massive, with an absence of stratified horizons. In addition, as mentioned earlier, the mineralogy of Facies A reflects the composition of the bedrock in the drainage basin. Thus, Facies A generally represents a lower, structureless unit which has not been reworked during or subsequent to its deposition.

The close association of Facies B with Facies A and minor stratified sands and gravels is an important indicator of the environment of deposition. The occurrence of interlayered massive diamictons and stratified resedimented diamictons has been noted by a number of researchers (Shaw, 1982; Haldorsen _et al_., 1982; Eyles _et al_., 1983). Sections 32 and 40 show the close stratigraphic association between Facies A, B and outwash sediments which are indicative of an ice-proximal environment. The characteristics of Facies B suggest that it could be a flow till, although it is difficult to attribute its widespread deposition to a specific mechanism on the basis of the available evidence. Boulton (1968, p. 391) stated that flow tills may be derived from englacial or supraglacial origin, and were defined by Shaw (1982, p. 1549) as tills deposited by gravity flow of debris subsequent to its release from ice. In this context, the definition of till approaches that of Dreimanis (1982), described as a sediment that has been transported and deposited by or from glacier ice, with little or no sorting by water. Francis (1974) stated that flow tills (supraglacial tills flowing off glaciers) commonly flow down on to marginal or proglacial glacioaqueous sediments, and also reported that glaciolacustrine tills should generally resemble the associated regional tills in content, which is the case in the present study with respect to Facies A and B.

May (1977) proposed the term lacustrotill for a till-like sediment deposited in a lacustrine environment by flow mechanisms, and which corresponds to Boulton's (1972) flow till. May (1977) also defined the term waterlaid till as a sediment deposited beneath a floating glacier terminus, where the water depth does not allow any major amount of size-separation during settling. This deposit consists of glacial drift dumped in standing water at the snout of a glacier and which may be massive or exhibit a crude stratification. May's definition resembles that of Dreimanis (1976), a crudely stratified variety of till deposited in water containing glacially abraided and striated clasts, clay and silt pebbles, with a random alignment of clasts.

Therefore, Facies A could be considered as a lodgement or basal till, or a waterlaid till, whereas Facies B can be classified as a flow till or lacustrotill. Francis (1974) noted that the latter, when deposited largely by slumping off the ice front, is unlikely to possess any preferred orientation of fabric elements, a feature also observed by the writer in Facies B.

It is believed that flow till may be released directly into a proglacial lake by sliding off the ice front (Boulton, 1968; Francis, 1974; May, 1977; Shilts, 1978; Smith and Ashley, 1985) or, if the glacier is floating, by dropping from the subglacial surface. Massive waterlaid tills up to 30-m thick are thought to have been deposited in this way below the floating Lake Huron ice lobe (Quigley, 1980). Although such thicknesses were not encountered due to the smaller scale of the study area, Boulton (1968, p. 391) noted that flow tills may attain a thickness of 5 m, which is comparable to occurrences in Sections 3, 5, 6, 7, 14, 33, 34 and 39. Boulton (ibid) concluded that flow tills may build up to considerable thicknesses in topographic depressions. Crudely stratified flow tills have also been reported overlying more massive, compact basal till (May, 1977), a common occurrence in the study area where Facies A is overlain by Facies B. A poorly developed fining upward sequence in which Facies A grades up into Facies B (Plate 11) may represent a deposit which melted out from beneath the floating ice front and which settled-out fairly rapidly before another layer of till accumulated above it.

In ice-proximal environments (Sections 32, 33, 34 and 40) where subaqueous outwash activity at a particular locality is sporadic, interstratification of till and outwash may occur (Boulton, 1972; Shaw, 1982), producing multi-till sequences with sharp planar bases and no disturbance of ~~the~~ underlying sediments. Meltwater currents are believed to be responsible for the gravel lag deposits found at contacts between Facies A and B in Sections 32 and 40. Thin sand lenses or discontinuous layers of sand may have accumulated in a channel on a subaqueous flow till surface and then were covered by a further unit of flow till.

In Sections 17, 21 and 26, Facies B is found interbedded in distal environments with the rhythmites of Facies H. Four possibilities have been envisaged to interpret the occurrence of Facies B in these sections. It may have originated as ice rafted debris (Domack and Lawson, 1985), it may be a till deposited during a minor glacial readvance (surge), it could be the result of slumping of deposits accumulated along emerged lake shorelines (Gilbert, personal comm., 1984), or it could be the product of slumping of a semi-fluid till off the ice front. Upon closer consideration, the first three models appear unlikely for a number of different reasons. In the first case, ice rafted debris does not form continuous tabular units which extend ten's of metres laterally. Sediments released (dumped) suddenly from icebergs tend to form conical mounds of poorly sorted diamicton, 1 m to 5 m in length and 20 cm to 2 m in height (Thomas and Connell, 1985). Therefore, this mechanism does not satisfactorily explain the presence of relatively thick, stratabound diamicton beds.

In the second case, a readvance also seems doubtful, as there is no stratigraphic evidence for such an event. A glacial advance into the lake would be expected to cause shearing or deformation of the lacustrine sediments if they were associated with tills (Shilts, 1981), but this was not observed in the study area.

The third possibility of slumping of till from the sides of the basin could result in till-like material flowing out over normal rhythmic lacustrine sediments. Although this interpretation is more plausible, it remains difficult to accept as no equivalent of Facies B exists as a surficial deposit in the basin of the Richardson and Rae Rivers.

An alternative explanation is that flow till melted from the glacier front and spread over the lake bottom, where some of the larger clasts may have settled into the soft underlying rhythmites shortly after deposition or later during compaction (Plate 15). Following this interruption, normal lacustrine sedimentation resumed. Thus, this appears to be the most probable model of emplacement, as flow tills may extend for considerable distance, from 0.5 km to 1 km or more from their source (Francis, 1974). As noted by Rust (1977), initiation of slumping and flow tills could result from wave activity or underwater turbulence resulting from calving of icebergs along the submerged ice front.

Boulton (1972), Shaw (1982) and Haldorsen *et al.*, (1982) noted that a third type of glacial deposit can be a relatively common constituent in ice-proximal stratigraphic sequences. Melt-out tills are released either supraglacially or subglacially from stagnant ice beneath a confining overburden (Boulton, 1972, p. 379) and are defined as having a preferred fabric parallel to the direction of ice flow (Haldorsen *et al.*, 1982, p. 276). However, the present writer has not attributed any of the stratigraphic units to a melt-out till origin.

PLATE 15. Rhythmites of Facies H sharply overlain by diamicton unit of Facies B.

5.2 The Glaciofluvial Phase

Numerous eskers and glaciofluvial complexes with a predominant N.W.-S.E. orientation (Figure 2) were developed during ice retreat. These deposits are readily visible throughout the basin as esker ridges composed of glaciofluvial sediment forming raised, sometimes sinuous landforms. Stratigraphically, they are represented by Facies C, D and E.

Although there is evidence of outwash activity in Sections 32 and 34 associated with basal and flow tills, these sands and gravels represent only a relatively minor component. More representative sites such as Sections 11 16, 22, 24, 25, 28 and 40 show the full spectrum of energy levels, sedimentary structures and sediments associated with subglacial meltwater activity.

Stratigraphic associations show that Facies C, D and E could only have formed in an ice-proximal environment of a relatively deep glacial lake. These facies are believed to represent subaqueous outwash sediments, a term proposed by Rust and Romanelli (1975). They consist of stratified sand and gravel removed from a glacier front by subglacial meltwater streams and deposited in the tunnel, or in front of and beyond the glacier margin into a standing body of water. Boothroyd (1984) described similar deposits as lacustrine fans. As meltwater streams entered the ice-dammed lake which occupied the basin of the Richardson and Rae Rivers, the coarser esker sediments were deposited near the mouth of a subglacial tunnel and as the competency of the flow decreased in the distal direction, sediments became progressively finer-grained in a glaciolacustrine environment. Deposition of most of the sediment load took place just beyond the ice front and downstream, the flow energy was rapidly dissipated by turbulent mixing with standing water (Banerjee and McDonald, 1975; Harrison, 1975; Saunderson, 1975; Rust, 1977; Smith and Ashley, 1985). The resulting sedimentary succession consists of Facies C which may have been deposited at or near the apex of the fan in the form of small-scale foreset bedding (Mode et al., 1983) which progressed basinwards as long as the glacier was relatively stationary and abundant sediments were being introduced. When stream currents flowing across the fan surface reached the front, their high density made them flow down the foreset face as turbidity currents. Massive structureless sand units (Facies D) were then deposited by rapid deposition by turbidity currents (Banerjee and McDonald, 1975). However, interpretation of Facies D is difficult to determine with certainty because parallel bedding in coarse sediments may be produced both in lower and upper flow regimes (Southard, 1971). Nevertheless, in view of its association with Facies C, poor sorting and poorly defined parallel bedding suggest that rapid sedimentation in upper flow regimes may have existed. The low stratigraphic position and coarse grain size also indicate that deposition of Facies C and D occurred under relatively high flow velocities. In Section 40, cross-bedded sand (Facies C) alternates with flat-bedded and massive sand (Facies D) and cross-laminated sand (Facies E). This variation in grain sizes and sedimentary structures within alternating beds indicates that it is possible to have lateral variations between these types of sequences due to a change in the flow direction of meltwater, retreat of the tunnel mouth or to fluctuations in discharge during accumulation, diurnally to seasonally, a characteristic feature in glaciofluvial environments (Arnborg, 1955; Cohen, 1979).

The absence of exposures at certain sites did not permit a three-dimensional assessment of these outwash deposits. Possibly as a result of this, channel deposits were not observed, although they are the most environmentally significant feature in subaqueous outwash deposits from the Ottawa area (Rust, 1977). Whether deposition resulted from an individual ice tunnel in the form of a fan extending out over the lake floor or a series of ice tunnels producing overlapping forms or a sheet-like deposit as described by Rust and Romanelli (1975, p. 190) cannot be determined. Isolated occurrences of

Facies C, D and E at the base of Sections 11, 16 and 28 which are widely separated from each other could be interpreted as evidence for broad subaqueous outwash blankets covering large areas of the lake floor.

Pulsating turbidity currents generated by variations in stream discharge and sediment load distributed finer-grained sediment by traction and suspension fallout from decelerating flow across the lake floor away from the ice-front where bottom topography controlled its path. In this manner, parts of the lake-bottom were covered locally by a subaqueous fan regressing from west to east.

In areas where glaciers are retreating down gradient, gravels and varves have commonly been found located adjacent to each other (De Geer, 1940; Saunderson, 1975). These gravels may change laterally into varved couplets, becoming finer-grained in a distal direction. In the Richardson and Rae River basin, a similar sequence occurs at 4 sites (Sections 11, 22, 24 and 40) in which a vertical equivalent to a single time-stratigraphic unit was deposited. The succession of Facies C, D, E and G can be explained by a gradual increase in the ratio of suspended load to bed load in the downcurrent direction due to retreat of the ice front from the point of deposition. Saunderson (1975) stated that this particular facies sequence is diagnostic of the transition that occurs between eskerine deposits and sediments that are deposited in relatively deep water. As discussed earlier, a deep glacial lake (up to 200 m) was in contact with the ice front during deposition of outwash deposits and proximal, as well as distal rhythmites. Field observations reveal that in a vertical sequence, there is a distinct lithofacies change which is thought to represent an esker to varve transition:

a) Basal planar and cross-bedded coarse sand and gravel (Facies C)
b) Massive to bedded, poorly sorted medium to coarse sand (Facies D)
c) Cross-laminated fine sands and silts exhibiting type A and B climbing ripples (Facies E)
d) Upper rhythmites consisting of sandy silt and clay couplets (Facies G) which grade to distal silty clay couplets (Facies H).

Although eskers deposited in water show rapid downstream transitions into fine-grained lacustrine facies (Eyles and Miall, 1984), well defined complete sedimentary successions are not always visible. Ripple cross-laminations occur relatively frequently as part of a subaqueous fan, whereas there is usually an absence of a well defined foreset slope which makes it difficult to precisely define true bottomset and topsets. However, Gustavson (1975) noted that some prodelta slope deposits may consist largely of fine sands and silts deposited as sequences of ripple-drift overlain by draped laminations, and Jopling and Walker (1968) observed that foreset-bottomset beds can have a gentle inclination and transitional character in which a clear distinction is not always apparent.

Facies E is an important member of this succession because of its position in the transition from proximal bedload deposition to more distal suspension deposition. This lithofacies consists primarily of type A and type B ripple-drift cross-laminations, as well as sinusoidal laminations. These structures originate from density underflows of sediment-laden meltwater transported over the gently sloping subaqueous fan surface during the summer melt season (Leckie and McCann, 1982). Thus, large amounts of coarse to fine sediment are transported onto the lake floor as fan-shaped deposits at the tunnel mouth or in the form of esker ridges. The origin of these forms may be explained as a function of the rate of suspended load to bed load during sedimentation (Jopling and Walker, 1968). As the suspended load contributed increasingly to sedimentation, there was a change in the ripple type from cross-laminations with stoss side erosion, through cross-laminations with stoss side preservation, to sinusoidal laminations.

Type A ripple-drift cross-lamination (Jopling and Walker, 1968) reflects deposition from suspension which was not rapid enough to bury the grains moving on the bed, and therefore did not preserve stoss-side laminations. These erosional-stoss climbing ripples are associated with higher current velocities (Ashley et al., 1982).

Type B ripple-drift cross-laminations (Jopling and Walker, 1968) represent conditions where bedload movement and deposition from suspension were equally important in its formation. They are referred to as depositional-stoss climbing ripples by Ashley et al., (1982) which are produced by aggradation rates that are large relative to ripple migration rates, and are associated with lower current velocities.

Sinusoidal laminations represent an end member in which traction load is virtually absent, and suspended load is dominant (Jopling and Walker, 1968). The structures produced are also known as draped laminations resulting from continued fall-out of fine-grained sediment from suspension after ripple migration ceases.

There are very few scoured or erosional surfaces between cosets of ripple cross-laminae, suggesting that even during acceleration of the currents where type A overlies type B ripples, deposition rather than erosion was taking place. Sequences of climbing ripples overlain by clay layers (Plate 3) are probably the result of repeated flow events in one season and are not considered to be annual deposits. Because turbidity currents are frequently the result of a more or less continuous, although pulsating density underflow, sequences of Facies E may show intermittent influx of coarser sediment, followed by a succession of sedimentary structures indicating a gradual deceleration, resulting in the deposition of fines between coarser layers. However, since the deposits originated from glacier-fed streams, it is possible that the cyclicity is seasonal in certain sections, with the coarser units accumulating during periods of high discharge during late spring and summer. Whether seasonal or not, alternating cosets generally reveal fluctuations in flow due to weather conditions during the ablation season (Shaw, 1975).

Palaeocurrent directions determined from cross-bedding and ripple-drift cross-laminations show a general divergent flow pattern (Figure 2). The broad relatively flat basin floor provided an ideal environment of deposition with few topographic constrictions, except in the southwestern part of the basin. Several palaeocurrent measurements towards the northwest are approximately parallel to many esker ridges, showing that the general direction of ice retreat was also parallel to glacial striae in the basin.

Subaqueous outwash sediments were deposited directly on the lake floor and were not built over buried, stagnant ice. No evidence of ice-cored esker ridges or collapse structures was observed in stratigraphic sections, although kettle-lakes are associated with ice-contact deltas and esker-delta complexes along the periphery of the drainage basin.

5.3 The Glaciolacustrine Phase

Where subaqueous outwash sediments are associated with lacustrine facies, the transition between these two environments is a gradual one. This may be seen in sections where eskerine deposits can be traced vertically into glaciolacustrine rhythmites. A characteristic end member of this trend is Facies G. This lithofacies is interpreted as a distal fan sediment deposited from waning high sediment concentration flows originating from a subglacial tunnel. Facies G separates the climbing ripples of Facies E from the typically distal lacustrine rhythmites of Facies H (Sections 11, 16, 22, 24, 28 and 40). It is also found at many other sites underlying Facies H and has been interpreted as proximal rhythmites.

However, in Sections 19 and 20, Facies G is found interbedded between relatively thick deposits of silty clay distal rhythmites. This is accompanied by an increase in grain size in the silt layer, a change in the dominant sedimentary structures and a significant increase in the thickness of the couplets, from 10 cm-20 cm to 3 m (Plate 16).

This comparatively sudden upward increase in thickness can either be caused by an ice readvance or by an increase in the rate of retreat (Agterberg and Banerjee, 1969). However, neither of these models takes into account the effects of a change in direction of meltwater discharge issuing from a subglacial conduit. This change in flow direction causes a lateral shift of lithofacies which can account for a temporary reversal in the upward-thinning trend, until the predominant flow direction resumes. Thick sequences of climbing ripples record palaeocurrent directions derived from type B ripples consisting of greater than 90% silt with a few percent each of sand and clay. Harrison (1975) obtained similar granulometric results from this type of ripple in glaciolacustrine sediments. However, as type A and sinusoidal laminations are also present, it can be inferred that a gradual increase in the ratio of suspended load to bed-load in the downcurrent direction played an important role in sedimentation. During construction of subaqueous fans, sites of deposition are likely to shift rapidly and often (Ashley, 1975). Therefore, it seems improbable that late-glacial oscillations of the retreating

PLATE 16. Proximal rhythmite of Facies G approximately 70 cm thick.

ice sheet are responsible for the occurrence of Facies G between more distal deposits, because as discussed earlier, there is no evidence of the lacustrine sediments being overriden by ice.

The change from thick sandy silt proximal rhythmites (Facies G) to thinner distal lake rhythmites (Facies H) is a gradual one, and as noted by Smith and Ashley (1985) in lakes where the principal source of sediment is glacial meltwater, bottom deposits thin and fine from proximal to distal portions of the basin.

Considering the large amount of sediment involved in lake-wide deposition of each couplet of Facies H, the consistent repetition of sedimentological processes in thick sequences (up to 24 m, Sections 12, 13, 17) is best explained by the annual climatic cycle. Although the annual nature of the Glacial Lake Richardson rhythmites has not been proven conclusively, it seems likely that the couplets are varves. Varves are rhythmically layered sediments deposited in glacial lakes; a single varve is a couplet composed of two distinct layers: a summer silt unit and a winter clay unit, each varve representing one year of deposition (De Geer, 1912). Each couplet represents seasonal variation in melting and is the result of two separate sedimentary processes.

Turbidity currents (sediment flows in which the grains are suspended by turbulence, Middleton and Hampton, 1973; Lowe, 1979), are generally accepted as the principal depositional mechanism for the relatively coarse-grained current-bedded summer layers whereas the clay-rich winter layers are deposited by fallout from suspension when most currents cease to flow (Kuenen, 1951; Banerjee, 1973). These turbidity currents may be the product of a continuous flow (several days to weeks) or shorter, sporadic pulses related to short-term climatic oscillations (hours), leading to high meltwater discharge. These cycles may also be generated by a variety of processes associated with an ice front, i.e., calving, a sudden change in flow direction of meltwater, a glacial burst or slumping. Lambert and Hsü (1979) demonstrated that graded laminae of silt and clay are not necessarily deposits of annual cycles, however, the scale of the sediments with which they worked cannot compare to that of the present study.

In the glacial environment, turbidite bedding can result from a continuous undercurrent. This may fluctuate during the melt season to produce the alternations of fine and coarse-grained laminae observed in the summer layers of proximal (basal) varves (Shaw, 1975). Short pauses in the density underflow currents permitted fines to settle out of suspension forming thin graded beds. The proximal summer layers are composed of discrete horizontal laminations of varying grain size and thickness (Plate 5). This multilaminated texture is thought to be the result of these intermittent lake-bottom flows whose greatest influence is felt near their point of origin, that is close to the glacial source.

According to Smith and Ashley (1985), the thickness of the silt layer depends upon the effectiveness of the sediment dispersal mechanisms, particularly underflows. Bottom turbidity currents may flow for several kilometres, depositing a graded bed of sand and silt, and represent the major mechanism for sediment distribution (Quigley, 1980). Deposits which accumulate from underflows are more commonly found in topographic lows of the lake basin where they are thickest. However, as the basin depressions were slowly filled, the turbidity currents travelled further west over a more gentle sloping lake bottom. Some turbidity currents have been known to flow upslope 400 m in elevation over a distance of up to 40 km (Damuth <u>et</u> <u>al</u>., 1979, p. 825). However, eventually they decelerate and are stopped by a rising lake floor at the distal end of the basin or on steeper sloping sides where the momentum of the underflows is dissipated (Kenney and Chan, 1972; Pickrill and Irwin, 1983). For this reason, varves are absent on topographic highs. Their absence in the southern, more elevated areas of the basin may also be a function of the energy level of the environment of deposition. This area remained largely an ice proximal environment where marginal lakes persisted throughout most of the stages of ice retreat. Locally, washed till and glaciofluvial esker systems attest to an environment where most of the fine-grained outwash was transported out of this area, into the western and central regions of the basin. The latter region represented a more open lacustrine phase where deposition occurred in a more tranquil environment.

Each varve couplet was completed by the deposition of a clay unit from suspension when continuous underflows ceased during the winter months. During this period, the clay layer accumulated gradually from the suspended fine silts and clays dispersed throughout the lake by overflow and interflow currents. Interflows occurred during the melt season when the densities of meltwater discharging into the lake were less than the maximum lake-water density, but greater than the minimum lake-water density (Gustavson, 1975). Distal areas may also have been supplied by surface spreading of overflows which gradually mixed with lake-water. Therefore, the thickness of the clay layer is essentially a function of settling time and basin depth.

Any inflowing water during the winter would be comparatively deficient in sediments, and consequently its density would be less than that of the lake waters. This difference produces overflows and interflows whose effect upon slow settling of clay particles deeper in the water column is minimal. However, recent studies by Grosvenor et al., (1985) show that bottom-flowing currents may persist throughout most of the winter, intermittently depositing sands during a brief period of higher flow. If the environment of deposition remains undisturbed the clay layer itself will show evidence of an upward-fining trend (Ashley, 1975). Shaw (1977) noted occasional sand partings in the clay layers of varves and attributed these features to slope failure at a delta front as a result of oversteepening. Thin sand laminae in the winter layer of varves from the study area may have originated from a similar process. Similarly, sediments may have slid or slumped down subaqueous fan slopes producing the same effect.

Slumping of a metastable delta front which has become too steep to support itself can occur at any time. This process may create accumulations at the foot of the delta slope as slump mounds which may be misinterpreted as evidence of ice-contact in a glaciolacustrine environment (Leckie and McCann, 1982, Fig. 6).

Occasionally, diamictic varves (Banerjee, 1973) are present, although they were observed only at the base of Section 29. They consist of a poorly sorted silty diamicton with numerous granules to small cobbles which are covered by draped laminations of the same material (Plate 17). They differ from Facies B in that they show evidence of internal laminations and flow structures. These coarser units grade up to regular winter clay layers, and are believed to be deposited by a combination of dense mudflow slurry, high density turbidity currents and possibly ice-rafting.

Although boundaries separating clay layers from overlying silt layers are generally sharp, irregular erosional contacts may also occur due to relatively powerful and erosive early spring turbidity currents of diverse origin.

According to Shaw (1977), where the glacial retreat rate is high compared to the rate of sedimentation, varved silts and clays are expected to dominate the succession. This phenomenon is most noticeable in the central part of the basin where 24 m sequences composed entirely of glaciolacustrine rhythmites are located (Sections 12 and 13). The upward decrease in thickness of these rhythmites is attributed to deposition in increasingly distal positions as the ice

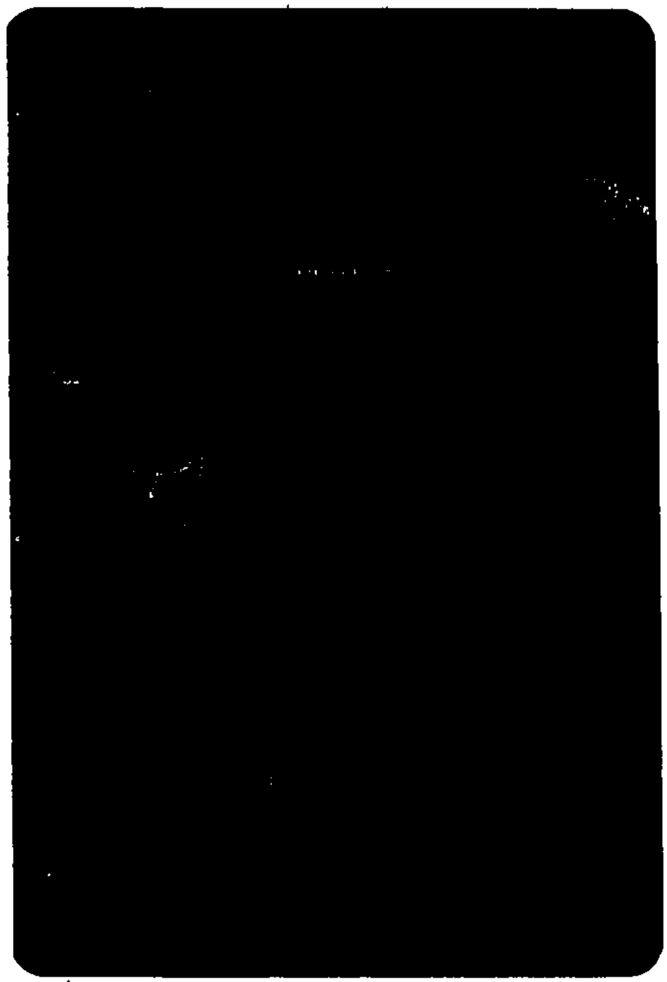


PLATE 17. Diamictic varve showing poorly sorted nature of the summer layer.

front retreated. This trend is characteristic of classical glacial lake sediments deposited during ice retreat (Shaw, 1977). Although Agterberg and Banerjee (1969) concluded that thickness of varves decreases exponentially upwards in a section, it was not possible to determine whether this trend applies to the varves of Glacial Lake Richardson.

Soft sediment deformation, occurring in several sections from the study area, is relatively common in Pleistocene varved clays. Two forms of contorted bedding are found associated with Facies H: convolute zones which are bounded above and below by undisturbed bedding (Plate 8), and diapirs which deform surrounding horizontal rhythmites (Plate 8). These structures could have resulted from either overriding of glaciers, overriding turbidity currents or postdepositional movement during compaction. Pettijohn and Potter (1964, Plate 107B) attributed décollement structures in overturned folds to thrust of overriding glaciers, however, no such features were observed. Harrison (1975) suggested that convolute bedding can result from overriding turbidity currents which would likely erode the tops of the deformed zones, and Morgenstern (1967) noted that subaqueous slumping can occur on lake bottom slopes as low as 1°, but the deformations in Facies H do not appear to have undergone any erosion. They are also not the result of slumping due to buried ice masses, a widespread feature in glaciolacustrine sediments (Reineck and Singh, 1980). These deformation structures are more likely to have formed in response to differential downward pressure exerted by the weight of overlying sediments (vertical loading). Compaction and dewatering of oversaturated sediments may result in diapirs and mass movement down very gentle slopes between horizontally bedded deposits (Banerjee and McDonald, 1975; Rust, 1977).

Dropstones are a second feature common to Facies H. Where coarser grains are present in what are usually fine-grained sediments, more than one source may have contributed (Hooper, 1975). As it is highly improbable that these clasts, ranging up to 1.5 m in diameter, were transported by current action, it seems likely that they are ice-rafted debris and provide evidence for floating ice. Icebergs and floes can transport material in quantity over great distances and on melting, deposit isolated stones to till (Harland <u>et al</u>., 1966; Anderson <u>et al</u>., 1980; Drewry and Cooper, 1981; Keys and Williams, 1984). The most common top and bottom contact associated with dropstones found in Facies H is bending (Thomas and Connell, 1985, Fig. 7), where the varves are simply downfolded and conformable with the base of the clast. Ovenshine (1970) observed that sediments within icebergs are released continuously from the submerged part as it melts while drifting. Sediments above the waterline may leave the iceberg by tilting or overturning of the iceberg, mudflow and slumping, and by meltwater streams.

More rarely, microfaulting of the varves was observed. It is believed that these normal faults are the expression of a readjustment within the section (Smith, 1959), formed as a result of slumping along a riverbank undergoing erosion, and are in no way related to melting of buried ice.

Varve deposition of the glaciolacustrine phase persisted throughout most of the various stages of Glacial Lake Richardson. Figure 6 summarizes the principal periods of lake evolution, showing the relationship between deltas and proximity to the ice front, as well as between sedimentary facies and the relative extent of the proglacial lake at the time of deposition. Figure 7 illustrates the geographic extent of seven consecutive phases of Glacial Lake Richardson. Although it is not the purpose of this thesis to give a detailed description of the history of the lake, this synthesis of its development through time facilitates the interpretation of stratigraphic successions in the study area. Each major

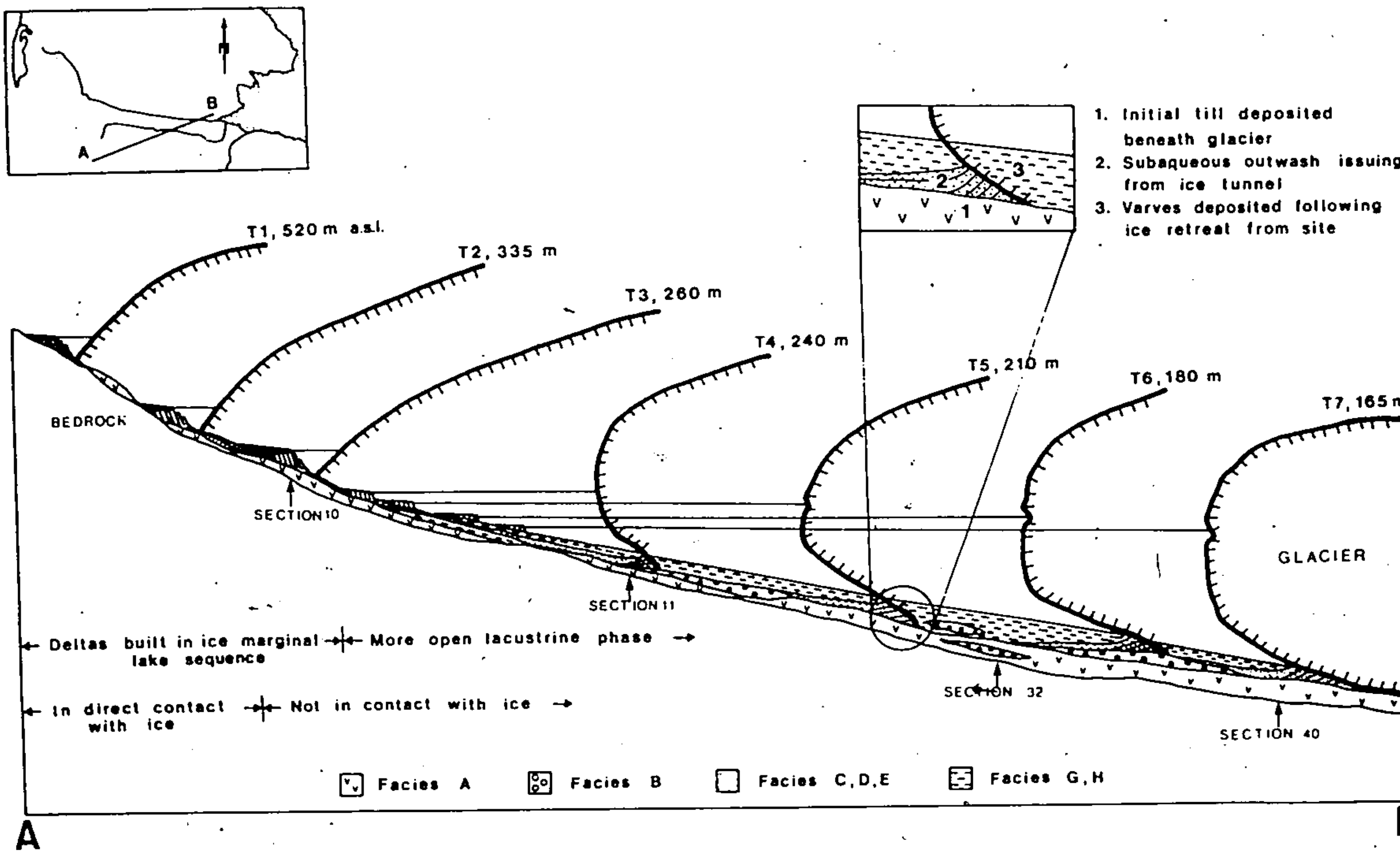


FIGURE 6. Facies model of glaciolacustrine sedimentation during evolution of Glacial Lake Richardson. (Not to scale)

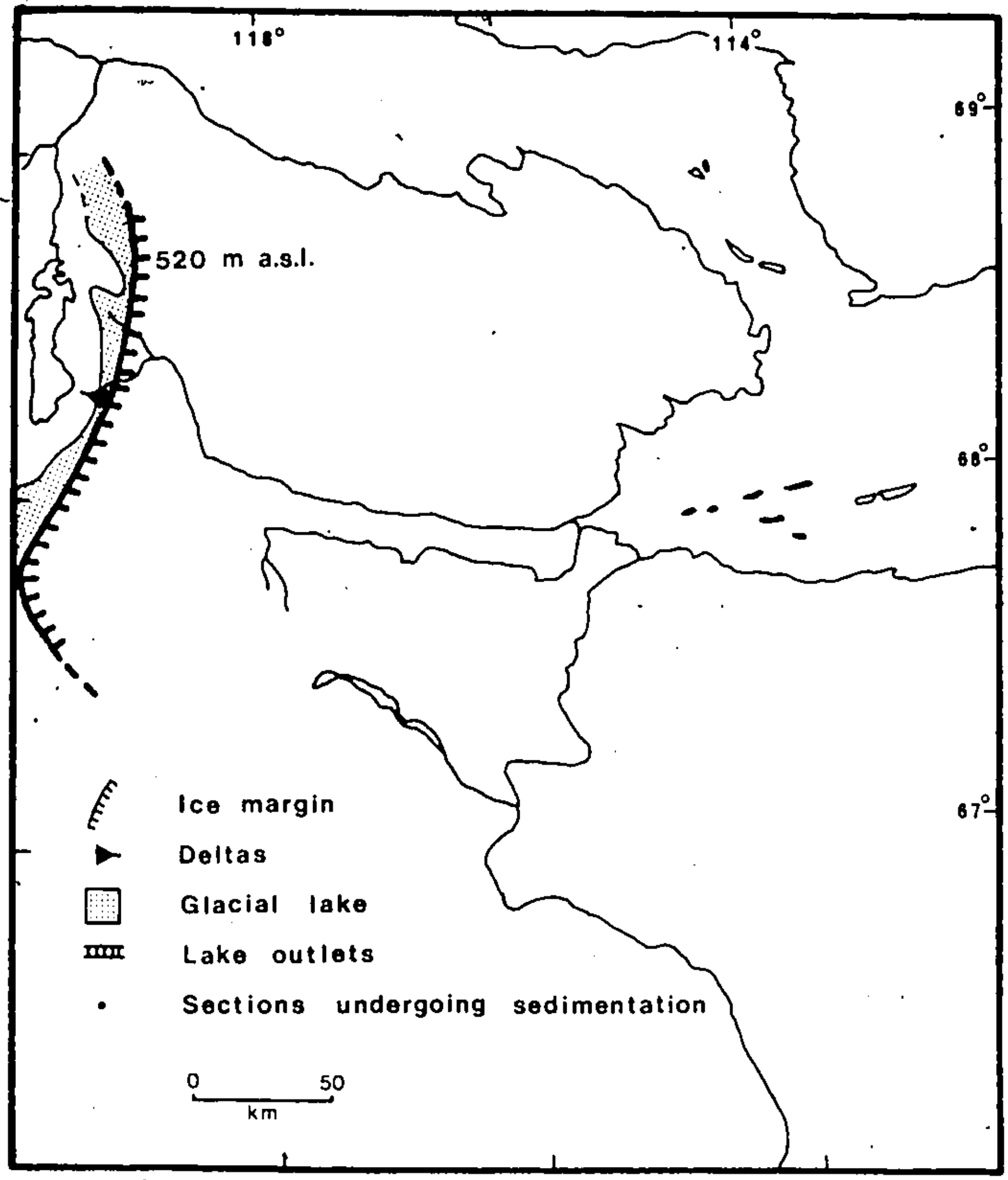
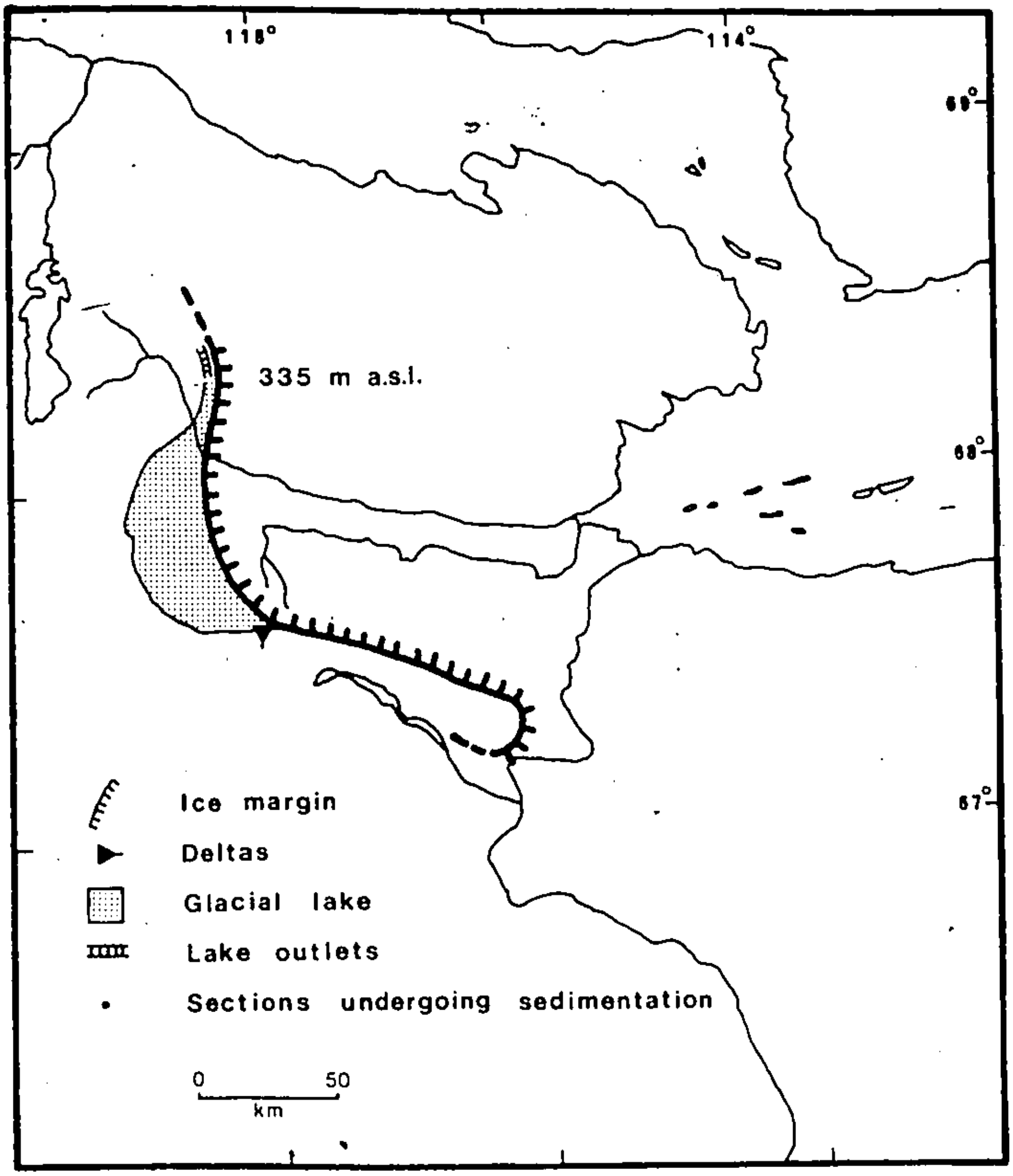
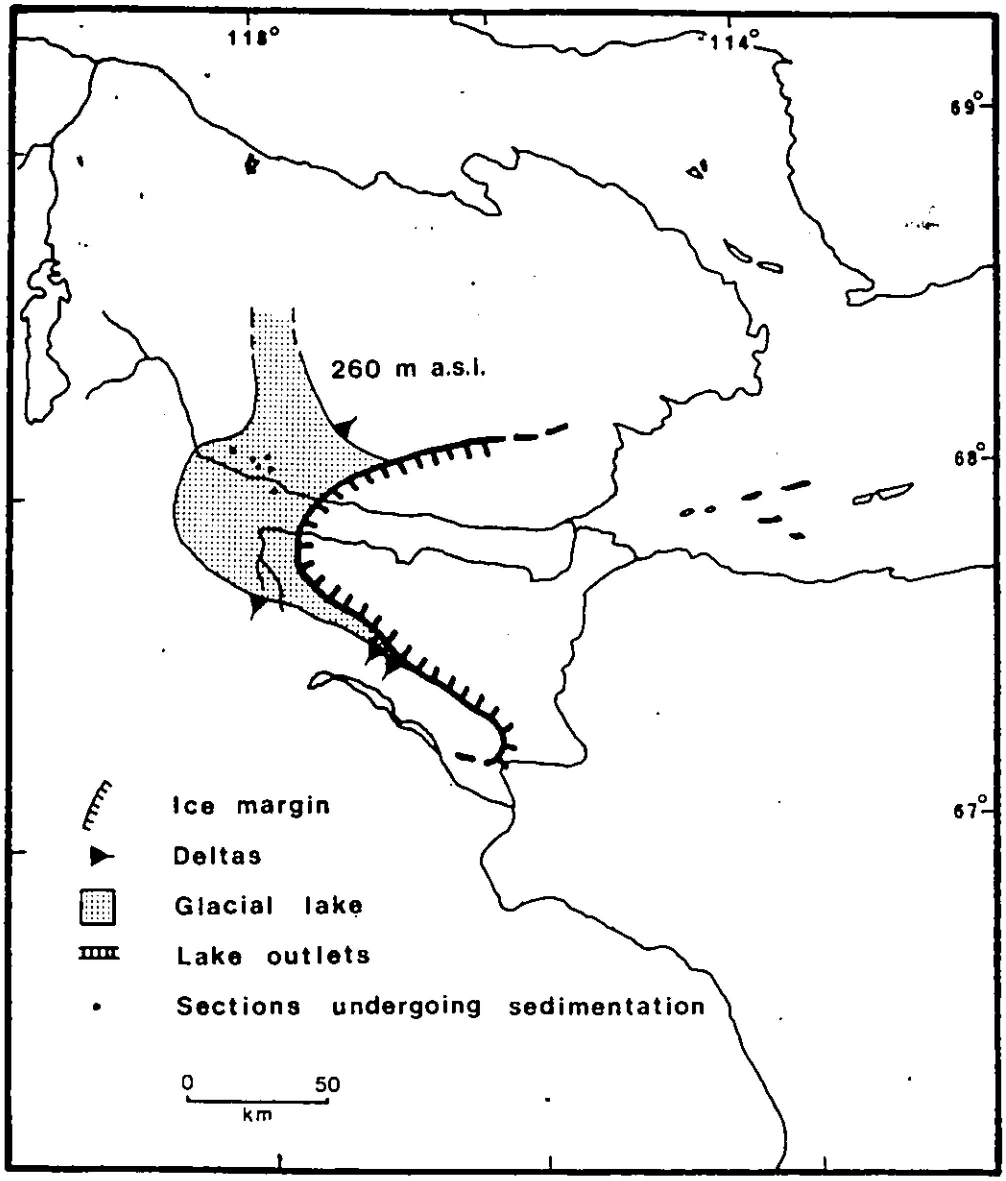


FIGURE 7. Evolution of Glacial Lake Richardson during deglaciation (modified from Mercier, 1986).

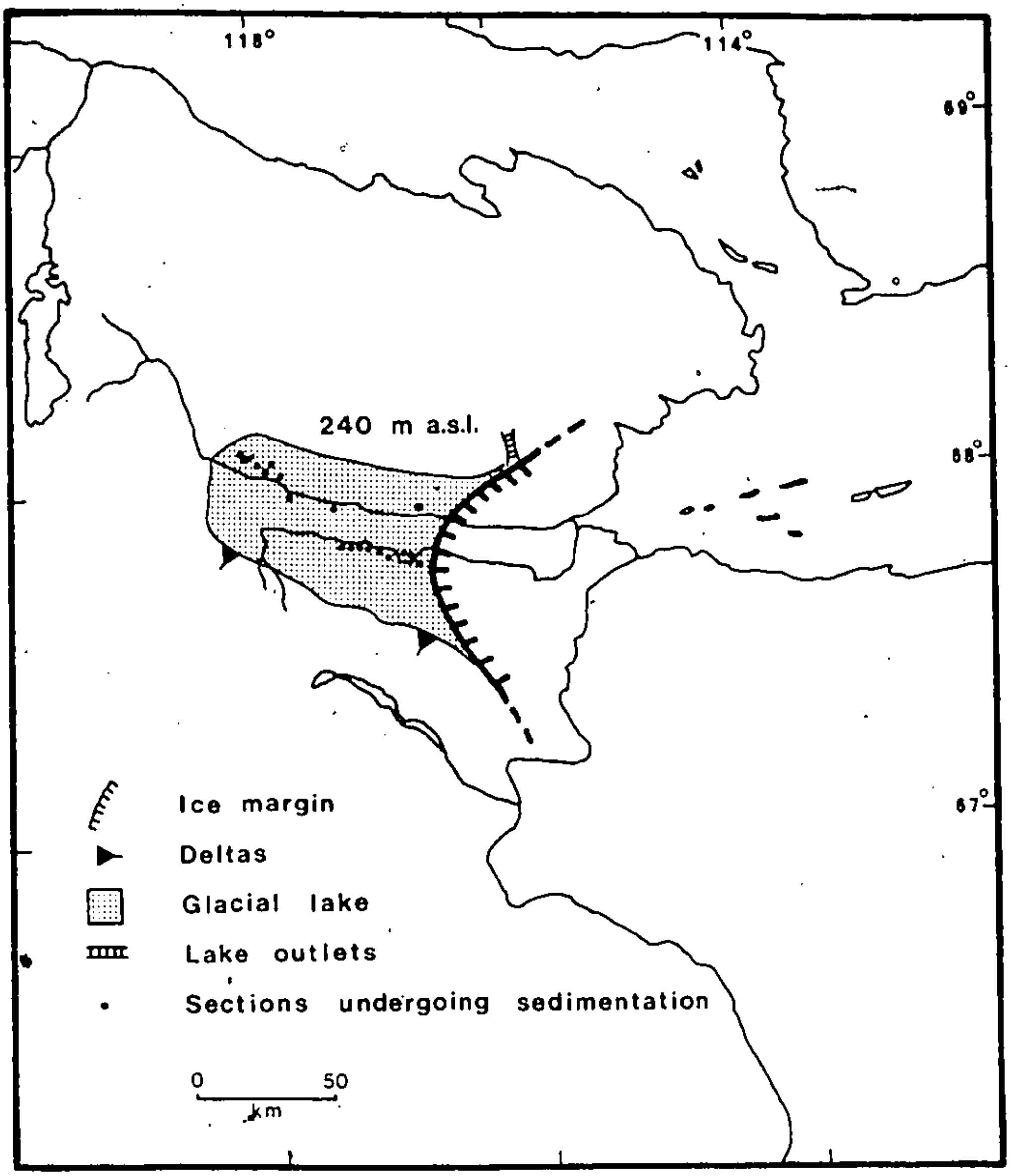
Phase 1, (T1)



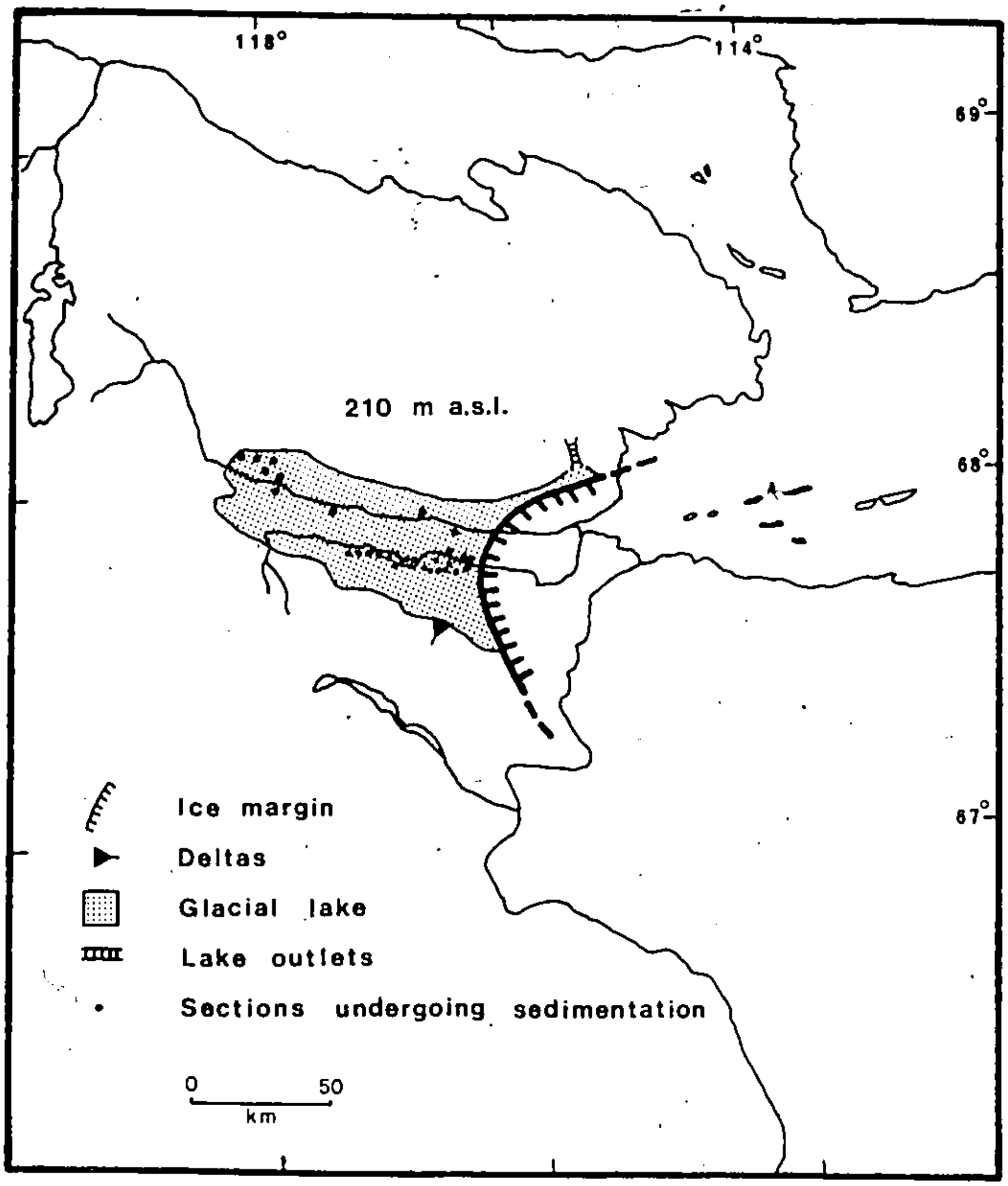
Phase 2 (T2) of Glacial Lake Richardson.



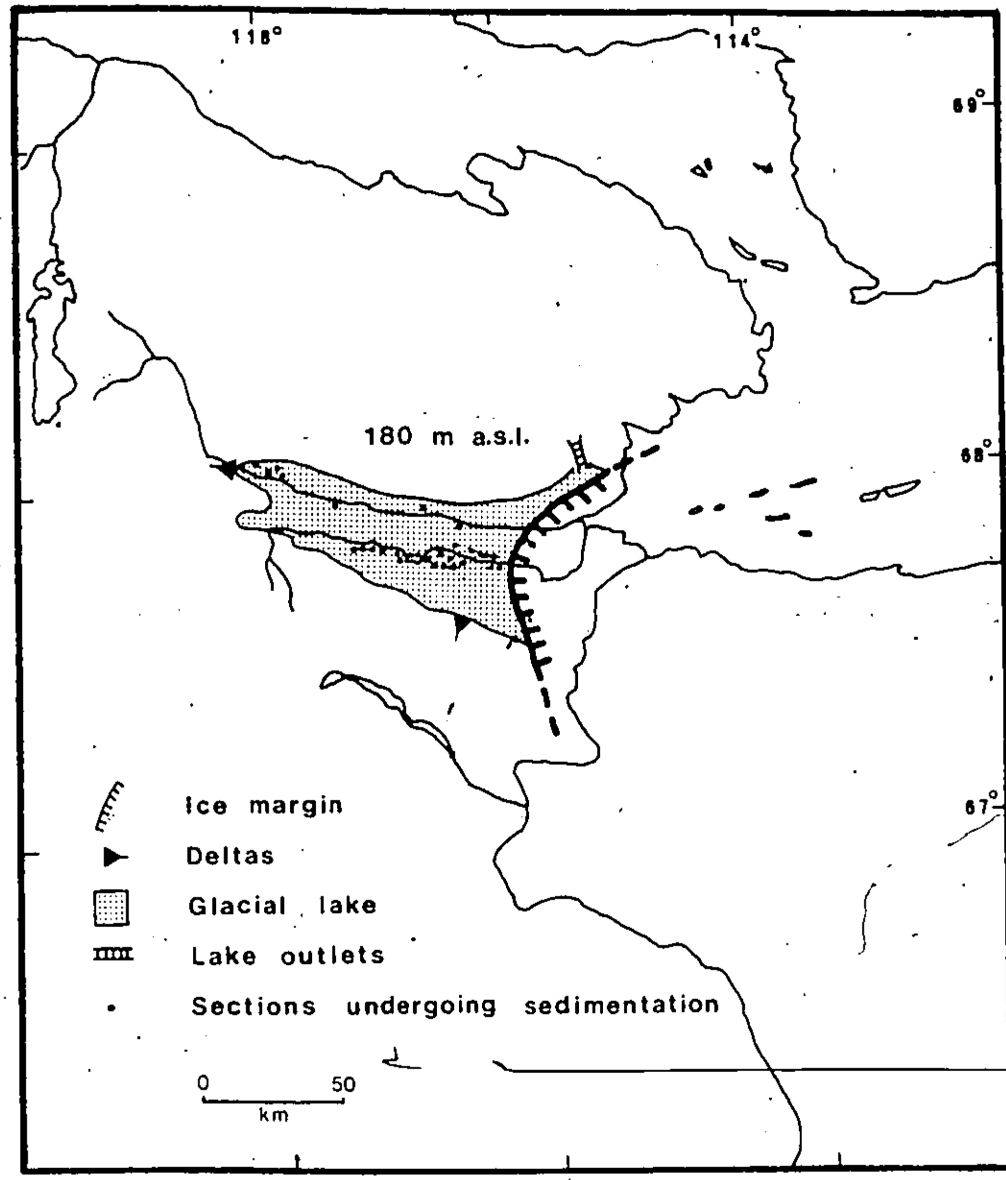
Phase 3 (T3) of Glacial Lake Richardson.



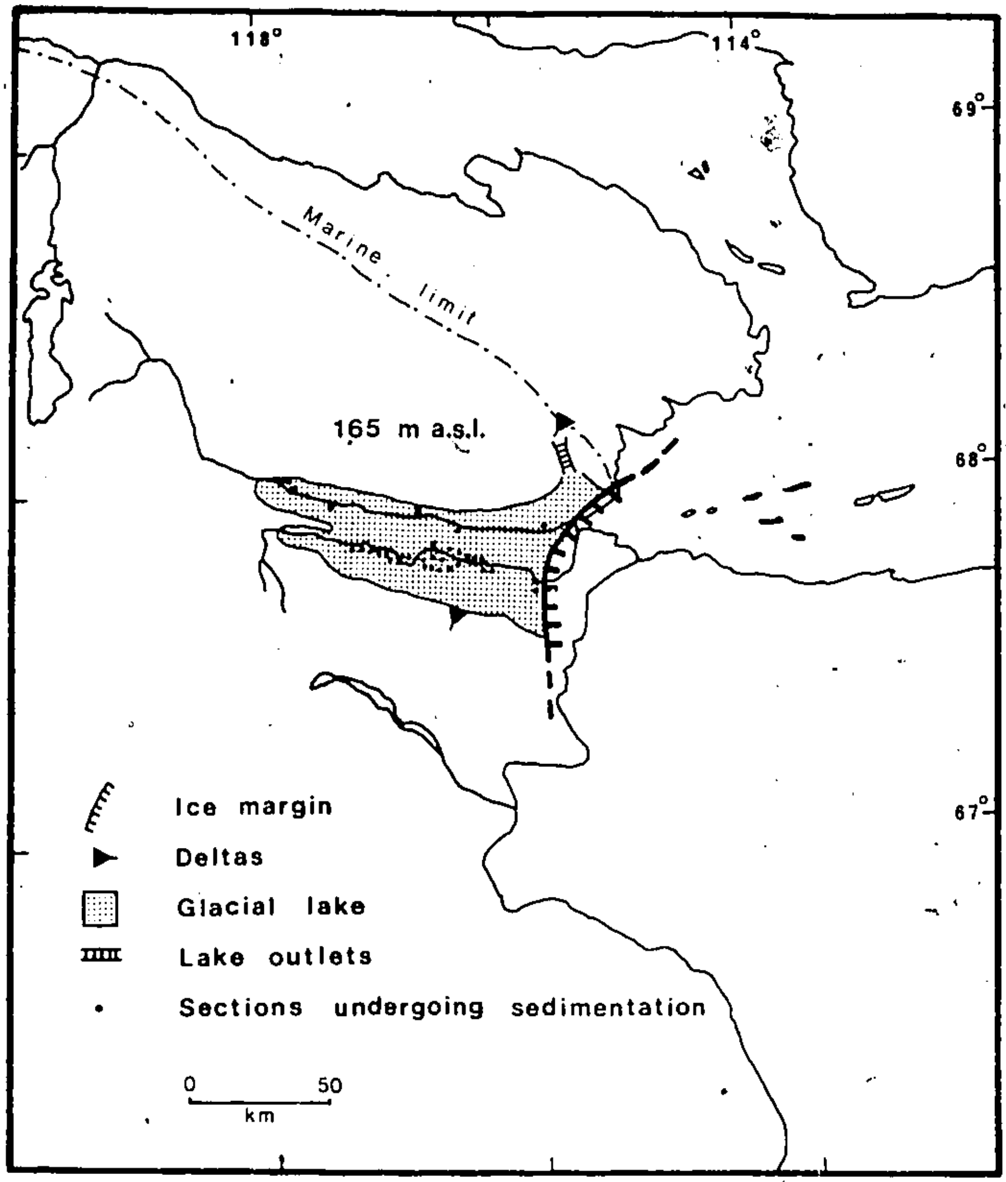
Phase 4 (T4) of Glacial Lake Richardson.



Phase 5 (T5) of Glacial Lake Richardson.



Phase 6 (T6) of Glacial Lake Richardson.



Phase 7 (T7) of Glacial Lake Richardson.

phase was controlled by successively lower outlets. All seven phases can be related to perched deltas in or just beyond the basin of the Richardson and Rae Rivers, indicating that each lake level was stable for some unknown length of time. It should be remembered that a number of other different phases probably existed between those indicated in Figures 6 and 7, since the retreat of the glacier was a continuous process.

Section 10 (Appendix D-10) is representative of the many deltas built at the various stages of Glacial Lake Richardson. This delta was built in an environment where there was an absence of tides, effective waves and longshore currents which commonly redistribute deltaic sediments. As a surficial deposit, it is fan shaped, with its apex to the south. Many of the glaciolacustrine deltas are of the lobate type, with an arcuate front. This form is typical of deltas composed of coarse material, whereas digitate forms such as the Mississippi delta tend to be composed of finer-grained sediments (Selley, 1970; Miall, 1984).

Most commonly, topsets (Unit 5, Appendix D-10) consist of crudely planar beds. They are fluvial sediments which represent braided stream deposits. This interpretation is based on the presence of braided palaeochannels preserved on the subaerial surface of the delta.

Foreset bedding (Units 2, 3, 4, Appendix D-10) in the lower and middle sections of the delta consist of subparallel beds of sand and gravel, showing consistent dips in the order of 23° to 30°, typical of classic Gilbert type deltas (Shaw, 1977). The absence of ripples and the coarse nature of the sediment suggest the rapid deposition of a large amount of sediment over a short period of time (Cohen, 1979). This seems likely as braided streams with large bedloads tend to occur in arctic environments where precipitation is erratic and vegetation sparse (Miall, 1984). The contact between topset and foreset beds in glaciolacustrine deltas is erosional and could correspond to the elevation of the lake into which the delta was built (Gustavson et al., 1975, p. 269).

Delta growth is influenced by the position of distributary mouths along the delta front. Each distributary supplies a lobe-shaped mass of sediment to the delta front and lake bottom. As the position of the distributaries changes, the lakeward progradation of the delta front proceeds by overlapping lobes of sediment. This process is responsible for the growth of arcuate-front glaciolacustrine deltas in modern and Pleistocene glacial lakes (Gustavson et al., 1975).

The distribution of the sediments is the result of a decrease in current velocity as sediment-laden fluvial waters enter a standing body of water. This decrease in velocity is accompanied by a decrease in the transport capacity of the water which consequently deposits its sediment load (Reineck and Singh, 1980). This depositional process is responsible for the coarsening upward sequence resulting from the progradation of the delta (Coleman and Prior, 1981). The streams and rivers originating from the initially deglaciated areas (highlands surrounding the basin) are not likely to have been major sediment supplying agents. The large volume of glaciolacustrine sediments and palaeocurrent data suggest that the bulk of the sedimentation took place when glacial ice occupied the eastern part of the drainage basin. This assumption is also based on the relatively small number of deltas and their limited extent in comparison to the length of the lake shore bordered by land. Another indication is the absence of palaeocurrent data and more complex sedimentary patterns which would be expected if streams and rivers draining ice-free terrain were primary sources of sediment. These streams probably entered the lake as interflows, overflows and minor underflows.

Figure 8 illustrates the associations between the glacial, glaciofluvial and glaciolacustrine phases which may have existed at almost any stage in the evolution of Glacial Lake Richardson.

5.4 The Marine Phase

a) The Marine Incursion

The continuous eastward retreat of the ice front permitted lake levels to fall, as lower glacial lake outlets were freed of ice. Shortly after Phase 7 (Figure 7, T7), the ice front retreated still further east and eventually permitted the incursion of the sea to its maximum elevation, which marked the end of the Glacial Lake Richardson sequence. It is believed that just prior to this stage, the glacial lake and the sea were approximately at the same elevation, because the marine limit was established at 170 m a.s.l. in the Coppermine River valley. However, this region has experienced considerable differential isostatic uplift, and consequently, a single relatively stable water plane at a certain point in time can in fact be associated with several different elevations over a wide area. For example, the marine limit at 170 m a.s.l. in the Coppermine area decreases to approximately 160 m a.s.l. near the mouth of the Rae River, and to 150 m a.s.l. in the northwestern part of the basin of the Richardson and Rae Rivers. It is known

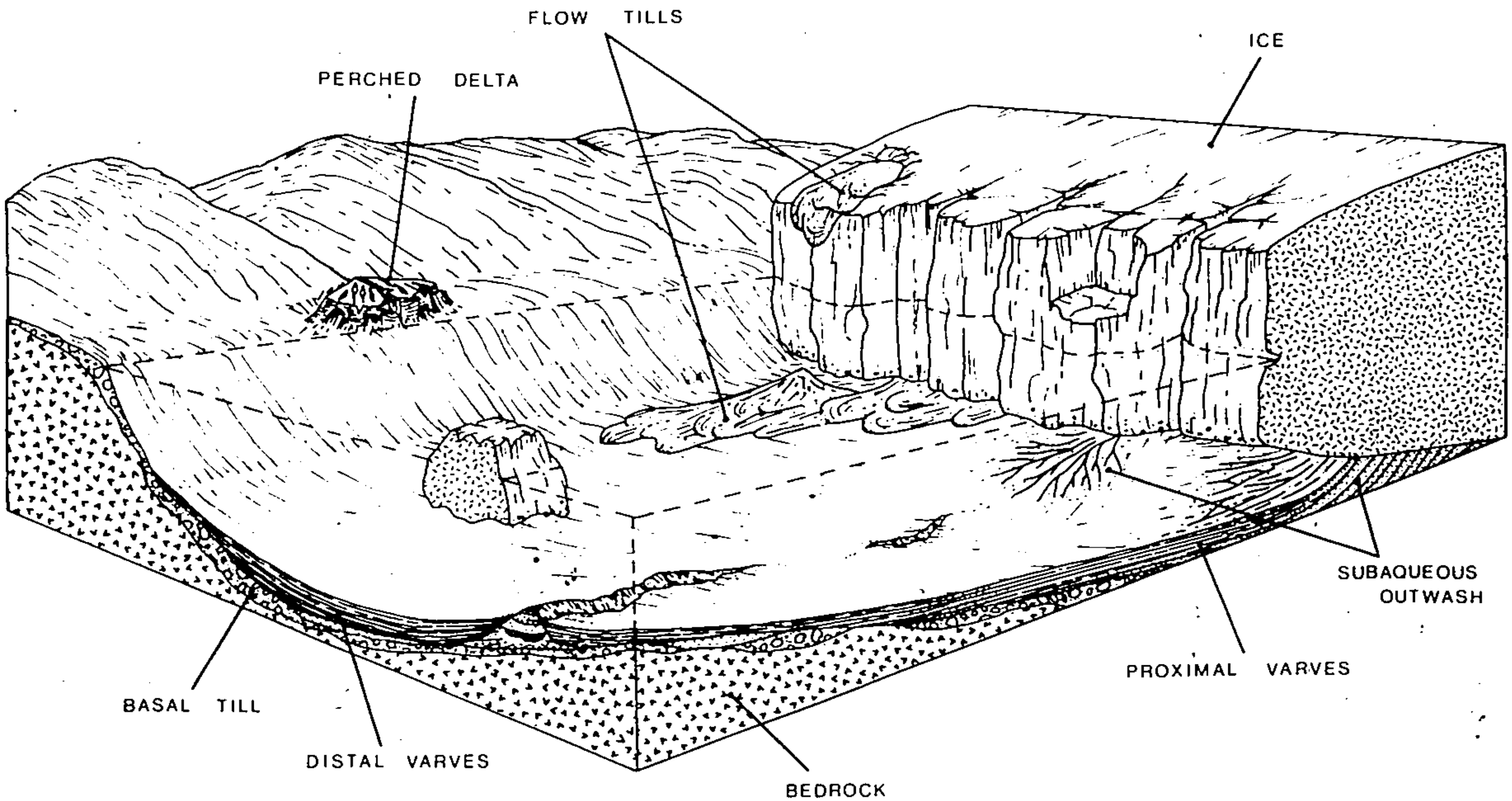


FIGURE 8. Schematic block diagram illustrating the deposition of various sedimentary facies associated with proglacial and distal glaciolacustrine environments.

to be approaching the present-day sea level to the west of Croker River (Figure 1). Because of this, there is always some degree of uncertainty in chronostratigraphic correlation of features over wide areas on the basis of elevation alone. To further complicate matters, the marine limit cannot always be directly observed in the field and must be extrapolated in certain areas. However, this does not pose any serious problems in the present study.

The relatively sudden marine incursion is reflected in the stratigraphic record by two features. The first is the presence of a thick bed of Facies B in Sections 30 and 31, lying between the varves of Facies H and the massive silts of Facies I. This deposit is interpreted as a flow till which may have originated from the ice front or a large iceberg during a possible break-up of the ice margin or accelerated melting and retreat of the glacier following the invasion of the sea. A discontinuous gravel layer at the base of Facies I (Section 35) may also have resulted from a similar process.

Secondly, the marine incursion is characterized by an abrupt change in the mechanism of sedimentation which resulted in the subsequent deposition of a lithofacies totally dissimilar from that which was deposited in the glaciolacustrine environment. This can be seen in Sections 20, 22, 25, 27, 29, 35 and 40, and in Plate 9 where the structureless fossiliferous marine silts of Facies I overlie the glaciolacustrine rhythmites of Facies H.

When the marine waters entered the basin and mixed with the freshwater of Glacial Lake Richardson, flocculation occurred. Flocculation is the process by which small clay particles form aggregates known as flocs or floccules (Kranck, 1981). The latter contain grains of all sizes from the suspension and consistently lie within the silt range (4 Ø to 8 Ø). Silt and flocculated clays (which behave more like silt-sized particles) settled together to form massive silt deposits (Torrance, 1983).

Locat (1982) reported that flocculation occurs at salinities above 0.5 g/l (0.5 ppt). Previous studies (Kranck, 1975; Syvitski, 1978) have shown that in salinities above 3 ppt, fine particles in a suspension are unstable and flocculate readily upon contact, and that salt promotes flocculation. Kranck (1975) noted that sediments in suspension that are thoroughly flocculated will contain particles that have all the same settling speed. As smaller particles have a large relative surface area, they tend to flocculate first and most readily. Larger grains (silt-sized) are not sufficiently surface-active to flocculate with other grains, but will adhere to flocs composed of many smaller particles. Kranck (ibid) concluded that particle size rather than mineralogy becomes a controlling factor in flocculation, although this claim has been disputed (Whitehouse et al., 1960).

In addition to inorganic floccules, organic agglomerates are relatively common in some marine environments. Agglomeration of particles by the feeding of organisms in the form of fecal pellets can represent a major source of sediment aggregates (Reineck and Singh, 1980). This process is also responsible for the large amount of clay-sized particles deposited in summer layers of a glacier-fed lake (Smith and Syvitski, 1982). Due to mixing of seawater and river water, precipitation and flocculation of fine colloidal organic particles are also likely to take place (Sholkovitz, 1976) and can contribute significantly to sedimentation rates.

Studies undertaken at Howe-Sound, B.C. (Syvitski, 1978) show that low-salt flocculation of clay particles begins within the first 5 m where mixing of freshwater overflows and saltwater proceeds by diffusion and turbulence. Below this zone, biological activity modifies the clay flocs by ingestion, a process which is responsible for some types of mottling commonly associated with marine clays (Quigley, 1980).

The salt dissolved in normal seawater (salinity of 35 ppt) gives it a very high density, 1028 g/l (Kuenen, 1951). Bearing in mind that near normal salinities of 35 ppt are to be expected in the deepest part of the sea, and that suspended sediment concentrations in meltwater streams are probably 1 to 4 g/l and rarely exceed 12 g/l (Gilbert, personal comm., 1984), then all freshwater interflows and underflows should rise to the surface after entering the sea at the ice front. Consequently, bottom turbidity currents appear to be a remote possibility, even under brackish water conditions, since abnormally high sediment concentrations (greater than 40 g/l, Gilbert (1983)) are required to overcome the buoyant effect of salinity in normal marine waters. However, density underflows may easily occur if they develop from an accelerating submarine slump (or similar process) already containing saline pore fluid (Quigley, 1980). Figure 9 shows the relationship between salinity, and suspended sediment concentrations and water density. It can be observed that even in dilute seawater (5 to 10 ppt), an above average sediment load is required to generate an underflow originating directly from glacier meltwater.

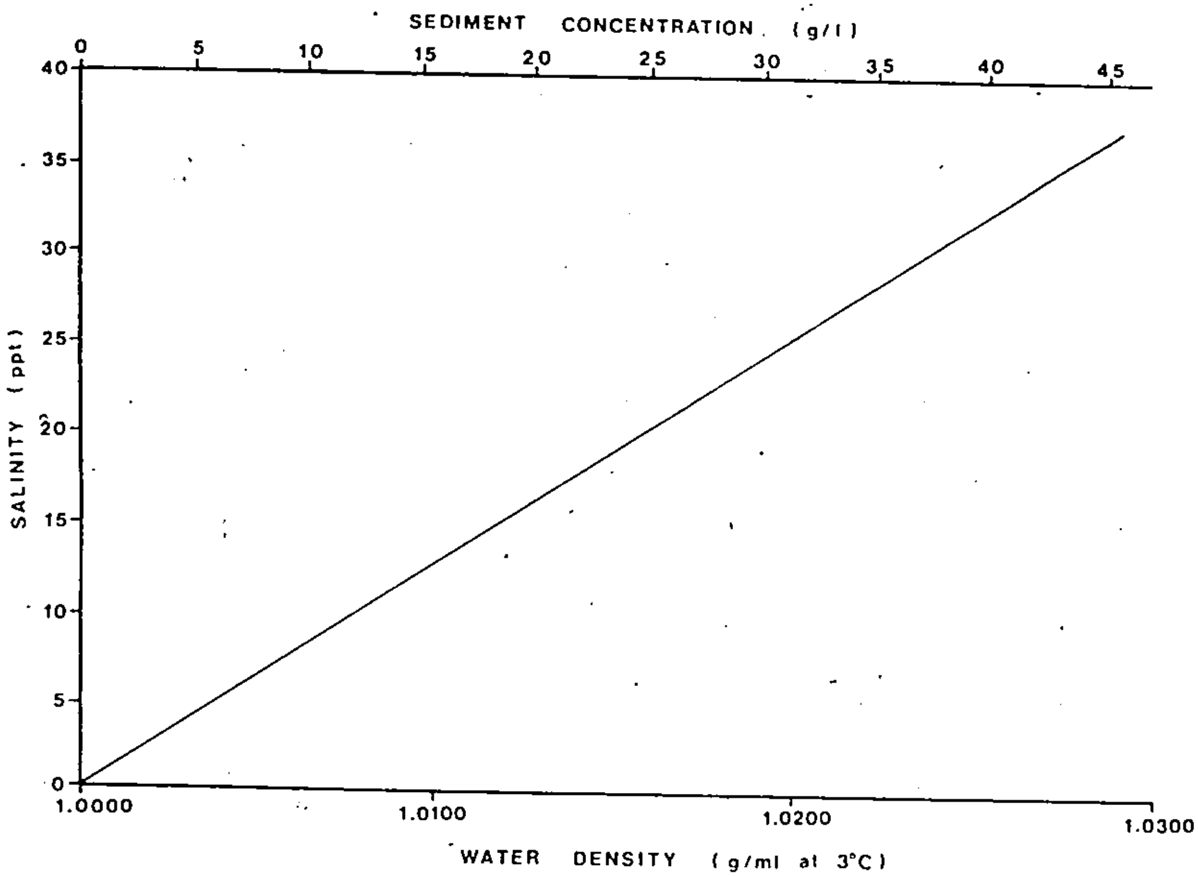


FIGURE 9. Water density vs salinity and suspended sediment concentration (after Gilbert, 1983).

In the present study, a rapid desintegration of the ice sheet is inferred immediately following the marine incursion (St-Onge, personal comm., 1986), as there is little evidence that the glacier had a significant influence on marine sedimentation. Buoyancy exerted by the sea is known to reduce friction at the base of ice sheets and can cause rapid thinning (Edwards, 1978). Thus, it is believed that the ice sheet or possibly an ice shelf over Coronation Gulf receded steadily south towards the mainland where a number of ice-contact glaciomarine deltas record the limit of marine submergence along the coast to the east of the study area.

This model differs from more conventional glaciomarine sedimentation models due to the absence of glaciomarine deposits, which is interpreted as an indication of a rapidly thinning and floating ice margin, or possibly a desintegration of the ice sheet itself along shear planes and crevasses due to the buoyancy of seawater.

Most published studies on modern glaciomarine sediments deal specifically with fjord environments (Powell, 1981; Gilbert, 1982; Mode et al., 1983), although some of the sedimentary processes discussed by these authors may apply to most ice-marginal marine environments. In general, proximal glaciomarine zones are characterized by lodgement and meltout till where ice is grounded, and by sand and gravel deposited as ice-contact fans by subglacial meltwater streams (De Geer, 1912; Rust and Romanelli, 1975; Mode et al., 1983). The changing positions of the ice margin and meltwater tunnels produces non-correlative but repetitive patterns at various sites along the ice front. Underflows may be generated if sediment loads are sufficiently high; they carry away sand and form thin laminae within mud which is deposited during periods of low stream discharge when underflows are suppressed. Domack (1984) reported glaciomarine rhythmites of silty sand and pebbly mud, resulting from turbidity currents, flocculation and ice rafting, and attributed these to seasonal fluctuations of meltwater discharge. However, the environment of deposition was a brackish submarine fan in water less than 10 m deep.

Stevens (1985) stated that glaciomarine varves may develop in ice-proximal environments if density stratification occurs which delays flocculation and allows a greater separation of silt and clay fractions. The meltwater flowing out from a submarine front forms a fresh to brackish surface layer where suspended sediments eventually begin to settle as the meltwater spreads out and its velocity decreases.

Meanwhile, flocculation takes place as the salinity in the lower part of the surface layer increases to 0.5 to 3.0 ppt.

In the outer part of the proximal zone, massive sediment consisting of silt and very fine sand accumulates, although the interaction of tidal currents and sediment-laden plumes of freshwater which flocculate on meeting more saline water, produce finely laminated and graded silty mud and sand lithofacies (Mode et al., 1983; Miall, 1984).

The distal zone which is characterized by massive sandy silts and clays with occasional isolated dropstones, is supplied by surface currents (Kuenen, 1951).

A number of factors have been proposed for the absence of varved sediments in most glaciomarine environments: the rapid settling of flocculated muds and the density contrast between marine waters and inflowing meltwater which leads to the suppression of density current underflows in the marine environment (Gilbert, 1983). Unlike the lacustrine environments, more open marine systems may be affected by various types of currents which may remove clays (Powell, 1981), so seasonal influence is greatly reduced. Furthermore, bioturbation may also play an important role in destroying any primary bedding in marine sediments (Gilbert, 1982).

b) The Marine Regression

The marine limit of 150 m-170 m a.s.l. in the basin of the Richardson and Rae Rivers results from the maximum subsidence of the earth's crust under the weight of the continental ice sheet. During and following ice retreat, isostatic uplift caused the gradual regression of sea level. During this period, massive silts accumulated over the glaciolacustrine rhythmites of Facies H. Under these conditions, it is believed that saline waters penetrated the already oversaturated lacustrine varves by diffusion. The oversaturated nature of clay layers of the varves decreased their impermeability, permitting a relatively free circulation of water (Torrance, 1983). Although diffusion through such sediments is probably a slow process, it can penetrate downward up to 4 m deep over several hundred years during the prolonged marine submergence. The fact that these sediments remain salty to this day following their emergence indicates that little or no leaching has occurred, since the majority of these sediments remain frozen throughout the year due to the presence of permafrost. It is unlikely that the salt content is original, i.e., deposited simultaneously with the varves, as the rock types in the region are not particularly abundant in halite or gypsum.

The spatial distribution and thickness of Facies I are a function of the relative length of time sedimentation took place in the basin. At the time of the marine incursion, isostatic uplift was comparatively rapid (see below), resulting in the initial emergence of the more elevated regions following a brief period of submergence during which little sedimentation occurred over a short period of time. Conversely, the low-lying areas were submerged for a longer period of time and are characterized by thicker accumulations of marine sediments, as seen in Figure 4.

The marine regression is also responsible for the coarsening upward sequence preserved in many sites. Sections 22, 25, 27, 29, 31, 35, 36, 37 and 38 show a typical off-lap sequence in which deep-water silts grade upward into littoral or deltaic sands. Falling sea level results in waves reworking older marine deposits exposed on topographic highs, with sediments redeposited in the basin, a process observed today in Arctic fjords (Piper et al., 1983). Sand is retained in coastal zones as regression proceeds, resulting in coarser-grained sediments being deposited in the littoral zone. Occasional severe storms may concentrate gravel and cobbles, and leave them stranded as raised beaches.

Most commonly, deep-water silts which settled out of suspension in a relatively tranquil environment, grade into silty sand, and then either massive sand or rhythmically bedded strata (Facies F). Deposited in a shallow water environment, the alternating layers of sandy silt and silty sand may represent cycles of discharge (related to drainage of the basin) and sedimentation occurring in response to annual climatic cycles or to short-term weather changes within the annual cycle. Kerr (1984b) described a lenticular, wavy and flaser bedded sequence overlying these rhythmites, which are characteristically found in tidal environments (Reineck and Singh, 1980). However, poorly laminated to massive sand, interpreted as littoral deposits (Facies F), are more widespread. Their high degree of sorting could be representative of a constant energy level over a relatively long period, common to the littoral environment (Folk, 1980). In opposition to this, a diverging view is presented by the absence of beaches or wave-cut benches which suggest that wave activity was probably minimal in the postglacial sea. Although the degree of sorting is diagnostic of the energy level of waves (Conybeare, 1979), it is possible that beaches were not developed as they are frequently associated with periods of relative stillstand (Conybeare, 1979).

Where rivers enter the sea, coarse-grained sediment is deposited near their mouths, and fine suspended sediments are carried out away from the mouth and deposited relatively slowly from suspension. In this environment, deltas are formed as seen in Sections 20 and 23 (Appendices D-20 and D-23). Unfortunately, both stratigraphic sequences are poorly exposed, making the following interpretation only tentative. In Section 20, the gradational contact between Facies I and Unit 1 is interpreted as bottomset beds, although stratification is not discernible due to the fine-grained nature of the sediment. The foresets dip at 22° towards the northeast. They are unconformably overlain by trough cross-bedding (topsets) which may represent sand-dominated braided stream deposits (Rust, 1978), or small scour and channel-fill features (Reineck and Singh, 1980), where reverse grading may occur. In Section 23, the distinction between bottomset and foreset bedding is not clear. Topset beds may represent a fluvial dominated tidal environment which has been overlain by prograding fluvial deposits and subsequently overlain and stabilized by peat.

During and following ice retreat, the coastal region rose with respect to sea level in response to isostatic adjustments. A number of stages in the postglacial marine regression have been identified on the basis of deltas and sedimentary structures characteristic of littoral environments. Although regression was a continuous event, five major phases are particularly well represented, corresponding to elevations of 110 m, 90 m, 70 m, 45 m and 25 m a.s.l. (Figure 10).

The first stage at 110 m a.s.l. (Section 23) is characterized by a large delta which is actually a series of poorly defined overlapping fans at 120 m, 130 m and 140 m a.s.l., which are believed to be a single deltaic complex. However, the 110 m stage is clearly marked by an arcuate front measuring 3 km wide, while the delta attains 6 km in length, from its front to the apex at 140 m a.s.l. This delta consists of sand and silt, and was found to be barren of micro and macrofossils. Its marine origin can only be established by its altitude in the regional context. Approximately 50 km to the east, Bruneau (1984) mapped a comparable delta at 100 m a.s.l. in the Coppermine River valley (Figure 10).

The second phase is marked by two deltas situated along tributaries of the Richardson and Rae Rivers, as well as a raised beach ridge, all occurring at 90 m a.s.l. Both deltas have their surfaces scarred by abandoned braided stream channels and neither

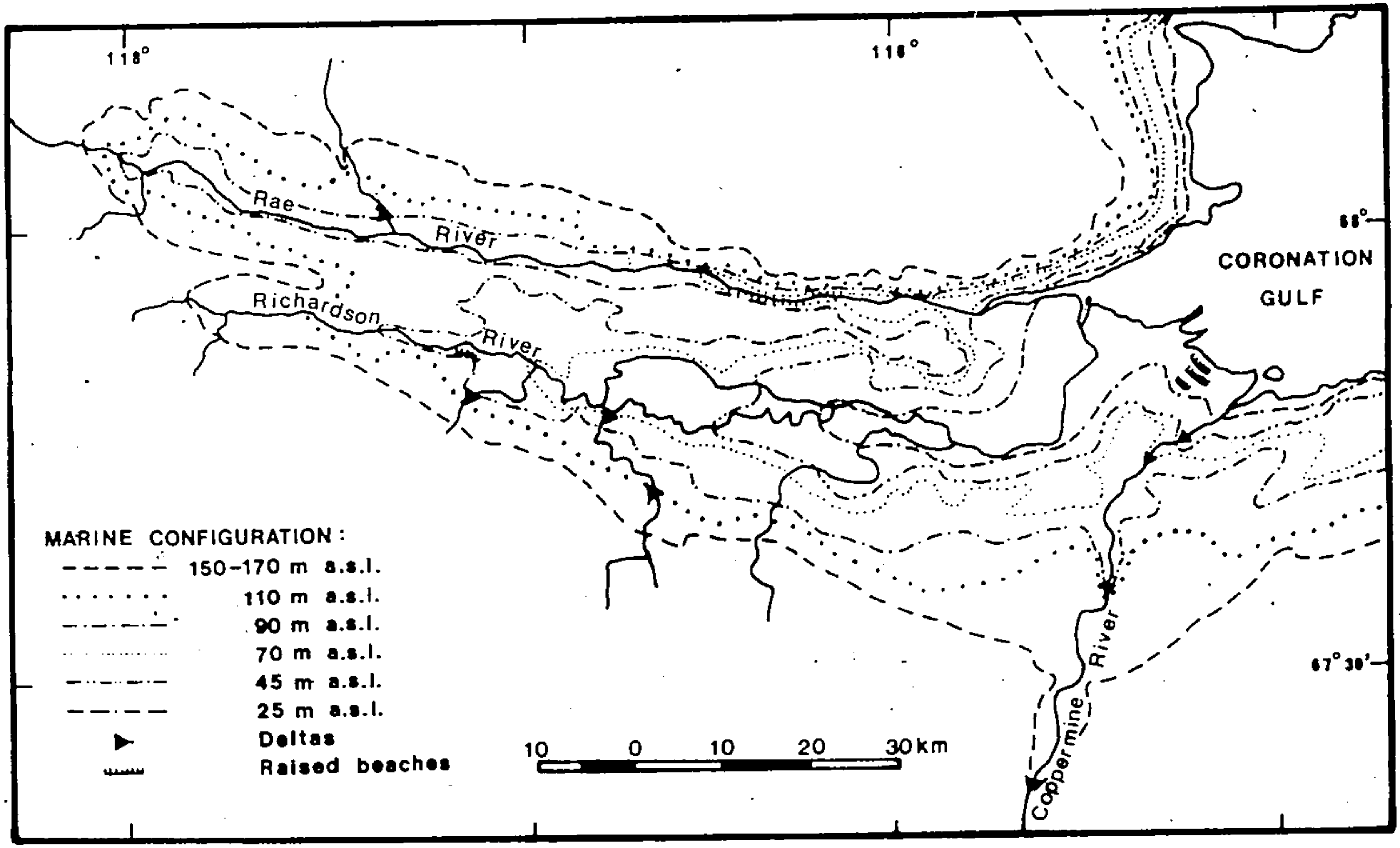


FIGURE 10. Palaeogeographic map showing the marine limit and successive stages of marine regression.

appears to have any marine fossils. However, one of the beach segments mapped by Mercier (1986) was found to be fossiliferous (Macoma calcarea). These bivalves provided an age of 10,300 ± 240 years B.P. (GSC - 3663), and since some bivalves were found in life position, it is believed that this date can be associated with a water plane at 90 m a.s.l. (Mercier, 1986). The beaches consist of glaciofluvial material which has been reworked, forming long narrow strandlines. They are the only beaches reported from the study area. The rarity of raised beaches could result in part from extensive solifluction and cryoturbation of the surficial deposits which may have obliterated such features elsewhere.

The third marine stage is associated with two deltas at 70 m a.s.l., one in the study area (Section 20), the other in the Coppermine River valley (Figure 10). Samples taken from Section 20 (Appendix D-20) were found to contain both marine bivalves and microfossils. Apart from this relatively small delta, no other evidence for this regressive phase was observed in the study area. However, in Section 25 (Appendix D-25), fossils of Mya truncata situated at an altitude of 73 m a.s.l. yielded an age of 9,430 ± 110 years B.P. (GSC - 3941). Unfortunately, it is not possible to state with certainty that these specimens are at or near the sea level of their age, although faunal assemblages and stratigraphic evidence suggest that they represent a relatively shallow water environment.

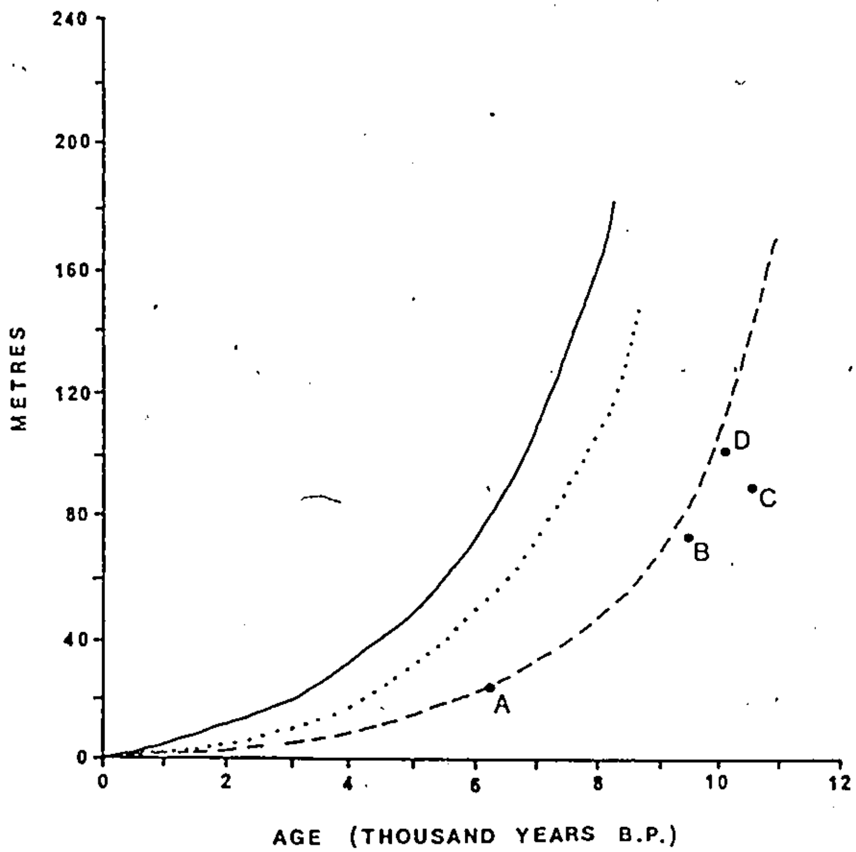
A fourth phase has been identified at 45 m a.s.l. based on the surficial distribution of a widespread littoral sand facies and stratigraphic evidence (Section 31). Faunal evidence is indicative of shallow sublittoral and littoral environments (cf. Palaeoenvironmental Reconstruction). This phase, or one closely associated to it, may be related to a delta in the Coppermine River valley located at 40 m a.s.l. (Bruneau, 1984).

Finally, a fifth phase established at 25 m a.s.l. is characterized by a succession of sedimentary structures indicative of intertidal conditions in an estuarine environment (Kerr, 1984b). Driftwood (Salix sp.) recovered from these sediments was dated at 6,100 ± 80 years B.P. (GSC - 4009) and is believed to be at sea level of this age.

From the features described above, it is not possible to determine whether the retreat of the sea from the study area was a gradual decline in sea level, or one or a series of relatively rapid falls followed by periods of greater stability. However, postglacial uplift curves can be used to approximate the rate of emergence, and thus permit a more accurate interpretation of the marine regression. Postglacial uplift represents the vertical displacement of the crust following deglaciation and can be estimated by adding a correction for eustatic changes in sea level to the altitude of the marine limit (Jelgersmae, 1966). Eustatic variations are simultaneous worldwide changes in sea level (Walcott, 1972, p. 855). In the Canadian Arctic, corrections for eustatic sea-level rise are between 0.15 m and 0.3 m/ 100 years (Andrews, 1970). Possible sources of errors in this calculation include the measurements of the elevations of the samples dated, environmental interpretations of sites and errors in radiocarbon dates themselves.

In the Coronation Gulf region, there are few absolute postglacial uplift values. Walcott (1972) established an emergence curve for the Bathurst Inlet area (Figure 1), which indicates an average rate of glacioisostatic uplift of 2.5 m/100 years (Figure 11). Andrews (1970), based on a mathematical model, calculated a mean rate of rebound of 2 m/100 years to 2.5 m/100 years for this same area. He also estimated uplift rates of 3 m to 3.5 m/100 years at 8,000 years B.P., and 1.5 m/100 years at 6,000 years B.P.,

FIGURE 11. Postglacial uplift curves for the Coronation Gulf area, N.W.T. (modified from Walcott, 1972).



Bathurst Inlet area:

——— Probable maximum sea level position

········ Probable minimum sea level position

Present study area:

- - - - - Estimated sea level position

Sample	Sample Number	Age	Elevation (m a.s.l.)
A	GSC-4009	6100 ± 80 B.P.	23
B	GSC-3941	9430 ± 110 B.P.	73
C	GSC-3663	10300 ± 240 B.P.	90
D	GSC-3327	9880 ± 90 B.P.	100

which are comparable to Walcott's values of 2.5 m/100 years at 8,000 years B.P. and 1.3 m/100 years at 6,000 years B.P.

Figure 11 illustrates the maximum and minimum probable sea level positions for the Bathurst Inlet area. The curve proposed by the present author for the study area differs significantly from the uplift curves in the Bathurst Inlet area because of differential isostatic rebound. The marine limit near Bathurst Inlet approaches 200 m to 250 m a.s.l., whereas in the western part of the Richardson and Rae River basin, it attains only 150-170 m a.s.l. The dashed curve (Figure 11) is based on four radiocarbon dates of marine bivalves. It is extrapolated to the left of B and D as these two dates represent bivalves which probably lived in shallow waters and do not represent sea level at their age. Similarly, site C is unlikely to have been at sea level 10,300 years ago. The present author believes that these fossils (GSC - 3663) once lived in waters ten's of metres deep, as Macoma calcarea is a relatively deep water species, not associated with littoral environments (cf. Palaeoenvironmental Reconstruction). In addition, if the curve were to pass through this point, the marine incursion would be pushed further back into time to approximately 12,000 B.P., a date which is not compatible with the present model of deglaciation for the

Coronation Gulf area. The average rate of uplift for the study area is 1.4 m-1.6 m/100 years, which is in accordance with the less important isostatic rebound in the western part of Coronation Gulf (Blake, 1970).

c) Palaeoenvironmental Reconstruction

A total of 15 samples representing 6 stratigraphic sections (Appendices D-25, 20, 40, 31, 37, 38) were found to contain both marine micro and macrofossils. These sections, located at elevations of 76 m, 70 m, 65 m, 42 m and 20 m a.s.l. respectively, are believed to be representative of the postglacial marine sedimentary record.

No species were recorded from the marine shoreline during maximum inundation (150-170 m a.s.l.). Although macrofossils have been noted up to 90 m a.s.l., microfossils appear to be absent above 76 m a.s.l.

A rapid emergence during the Late Pleistocene could be responsible for both the absence of fauna at higher elevations, as well as for the lack of shoreline features indicating that long periods of time with a constant sea level seem unlikely. The initial marine environments following the incursion must have been characterized by extreme environmental conditions developed in the vicinity of the glacier front.

Reduced salinities, ice movement, rapid sedimentation and other adverse conditions are likely to have impeded the colonization of these waters, thus rendering the sediments barren of life (Mode *et al*., 1983).

It is hoped that the following analysis of the marine fossils, i.e., foraminifers, ostracodes and mollusks, will contribute to a better understanding of Late Pleistocene and Holocene Arctic fauna, and that this information may be used in future studies to estimate ancient Arctic environmental conditions.

i) Foraminifera

A total of 33 foraminifera species have been recorded in postglacial marine sediments from the study area (Appendix E). Apart from Craig (1960) who listed 7 foraminifera species outside of the study area, no other investigation dealing with Pleistocene fossils has been carried out in the region.

Based primarily on the distribution and relative abundance of different species, three biofacies have been established (Table 1). This interpretation is also supported by lithological data since distinct granulometric trends exist in all of the sections studied.

TABLE 1. Distribution and abundance (%) of Foraminifera.

	BIOFACIES														
	A			B			C								
	SAMPLE NUMBER														
SPECIES	F15	F13	F7	F14	F12	F6	F11	F10	F9	F8	F5	F4	F3	F2	F1
ELPHIDIUM ORBICULARE	60	19	25		94	86	57	86	76	77	93	82	100	55	
E. CLAVATUM	<1	1		80		11	28	14	20	17	4	18		31	100
E. WILLIAMSONI							<1		<1						
E. BARTLETTI		<1												<1	
ELPHIDIUM SP.	<1														
BUCCELLA FRIGIDA		14					14		2		3			5	
PSEUDOPOLYMORPHINA NOVANGLIAE	2					2				3				8	
QUINQUELOCULINA STALKERI		1							<1					<1	
SCUTULORIS TEGMINIS									<1						
CORNUSPIRA INVOLVENS		<1				<1									
REOPHAX CURTUS				10											
R. SCORPIURUS		<1													
ALVEOPHRAGMIUM CRASSIMARGO		1		10											
A. cf. JEFFREYSI		<1													
HAPLOFRAGMOIDES SP.		1													
HYPERAMMINA SP.		<1													
CASSIDULINA RENIFORME	13	29	50												
ISLANDIELLA HELENAE	10	7	12												
I. NORCROSSI	6	2													
VIRGULINA LOEBLICHI	<1	22			3										
ELPHIDIELLA GROENLANDICA	4														
ELPHIDIELLA SP.	<1														
DENTALINA FROBISHERENSIS		<1													
D. ITTAI	<1														
PYRGO WILLIAMSONI	<1	1													
OOLINA MELO					3										
OOLINA SP.			<1												
FISSURINA SP.		<1													
PARAFISSURINA SP.	<1														
POLYMORPHINA SP.		<1													
LAGENA GRACILLIMA	<1								<1						
L. SEMILINEATA			<1												
TRILOCULINA TRIHEDRA		1	12												
NUMBER OF SPECIES	14	20	6	3	3	4	4	2	7	4	3	2	1	6	1
FORAMINIFERAL NUMBER	148	500	16	10	38	105	117	183	167	30	41	22	2	192	2

Biofacies A

Biofacies A, represented by samples F7, F13 and F15, is characterized by a relatively high species diversity (29 species) and generally, a greater number of specimens. The sediments associated with this biofacies consist of clayey-silt representing an off-shore lithofacies of the post-glacial sea. Foraminifera and stratigraphic evidence suggest that deposition took place in comparatively deep water (approximately 50 to 150 m in depth) with near-normal marine conditions in the central parts of the basin which reflect environments far from ancient shorelines at their time of deposition. This high diversity assemblage is probably related to rapidly improving environmental conditions in Early Holocene times following the invasion of the sea. No transition from an earlier low diversity highly glacial fauna was observed. No fossils were recorded in marine clay 1 m above the glaciolacustrine-marine contact, so it is possible that no such early assemblage existed or it may have been overlooked during the sampling process.

The characteristic species of this assemblage are Cassidulina reniforme, Islandiella helenae, I. norcrossi, Virgulina loeblichi and the presence of 15 other different species (Table 1) which are restricted to Biofacies A. Cassidulina reniforme has been reported as a typical species of a Pleistocene glaciomarine fauna from North Norway (Hald et al., 1984), as well as from Holocene glaciomarine deposits of Svalbard archipelago (Nagy, 1984). Elverhøi et al., (1980) noted its presence in Spitsbergen from an area lying just outside the front of a grounded glacier in water depth of about 50 m. Although C. reniforme is restricted to deeper water environments in the present study, it was found at all depths in the Gulf of St. Lawrence (Rodrigues and Hooper, 1982a) and is associated with silty clay of variable depth in Frobisher Bay, where it reflects present marine conditions (Dowdeswell et al., 1985). The latter also noted a decrease in its abundance in ice-proximal glacial conditions. Nagy (1984) noted that C. reniforme was a dominant species where bottom salinity was around 33 ppt; it is also indicative of a bottom-water temperature of 3°C (Rodrigues, personal comm., 1985).

A second important constituent of Biofacies A is Islandiella helenae. This species is considered to be essentially an arctic species (Feyling-Hanssen et al., 1976) and was noted in Pleistocene marine deposits from Baffin Island (Feyling-Hanssen, 1976).

It is essentially a deeper-water species which has been reported from a number of various localities: it is present in Hudson Bay from 50 m to 230 m depth (Leslie, 1965), in the Bering Sea in waters 100 m to 200 m deep (Anderson, 1963), and from depths of 13 m to 223 m off Alaska, Canada and Greenland (Loeblich and Tappan, 1953). I. helenae is also associated with silty clay of variable depths in Frobisher Bay, reflecting present marine conditions (Dowdeswell et al., 1985). Rodrigues and Hooper (1982a) noted that it was a dominant species at 50 m to 200 m depth in the Gulf of St. Lawrence, and was associated with salinities of 31 ppt to 34 ppt and water temperatures of - 2° to 4° C.

Islandiella norcrossi, also restricted to Biofacies A, is present in the Bering Sea from depths of 100 m to 200 m (Anderson, 1963), and is also an abundant species in Hudson Bay at 100 m (Leslie, 1965) and a predominant species between 140 m and 150 m (Wagner, 1968). Rodrigues and Hooper (1982a) found I. norcrossi to dominate at approximately 250 m depth in the Gulf of St. Lawrence, although its depth range was variable. This species was also recorded in Pleistocene marine deposits from Baffin Island (Feyling-Hanssen, 1976).

A third foraminifera species which is most abundant in Biofacies A is _Virgulina loeblichi_. It is considered to be a typical species of a Pleistocene glaciomarine fauna in North Norway (Hald _et al._, 1984) and reflects present marine conditions in Frobisher Bay (Dowdeswell _et al._, 1985) where it is associated with silty clay of variable depth, and which does not tolerate ice-proximal glacial conditions of lower salinity. Unlike the present study, _V. loeblichi_ was reported to have a shallow depth range by Loeblich and Tappan (1953) of about 21 m to 64 m in Alaska. However, Rodrigues and Hooper (1982a) noted that although its depth range varied from 125 m to 520 m, it was a dominant species at 390 m depth, associated with salinities of 33 to 35 ppt and water temperatures of 1° to 5°C.

A number of other deep-water foraminifera species are found only in Biofacies A (Table 1). One of these is _Triloculina trihedra_, found associated with cold water high latitude environments (Cronin, 1979) but according to Loeblich and Tappan (1953), restricted to the Arctic. It was reported by Vilks (1969) to occur below 200 m in the Canadian arctic archipelago, and also between 200 m and 400 m in waters of 4° to 5°C and 34 ppt to 35 ppt in the Gulf of St. Lawrence (Rodrigues and Hooper, 1982a).

Its presence has also been recorded by Markussen *et al.*, (1985) in cores at depths of 3,000 m from the eastern Arctic Basin.

Pyrgo williamsoni, found between 400 m and 520 m depth in waters of 34 ppt to 35 ppt and 4° to 5°C in the Gulf of St. Lawrence (Rodrigues and Hooper, 1982a), has also been noted by Dowdeswell *et al.*, (1985) from Frobisher Bay where it can tolerate ice-proximal glacial conditions of low salinity.

Another species, *Buccella frigida*, is referred to by Todd and Low (1967) as typical of cold water areas and has been recorded in marine Pleistocene deposits from Baffin Island (Feyling-Hanssen, 1976). It is found today in the Gulf of St. Lawrence between 55 m to 400 m depth, associated with salinities of less than 33 ppt to 35 ppt, and temperatures of - 2° to 5°C. (Rodrigues and Hooper, 1982a). Wagner (1968) found the range of *B. figida* variable, from 43 m to 276 m, but was most abundant between 117 m and 126 m depth in Hudson Bay.

The genus *Lagena* is represented by *L. semilineata* and *L. gracillima*. Both are distributed throughout the Arctic and are found in Alaskan waters of less than 50 m (Loeblich and Tappan, 1953). Leslie (1965) found *L. gracillima* to be a dominant species around 100 m in Hudson Bay, whereas Vilks (1969) noted its occurrence below 200 m depth in the Canadian arctic archipelago. It was also reported from the Bering Sea (Anderson, 1963) with a depth range of 100 m to 200 m. L. semilineata was found at variable depth, from 55 m to 200 m in waters of the Gulf of St. Lawrence with salinities of less than 33 ppt to 34 ppt and temperatures ranging from - 2° to 4°C (Rodrigues and Hooper, 1982a).

Two forms of the genus Dentalina are associated with Biofacies A. Although both species are rare (Table 1), their appearance should be noted. D. ittai has been reported from different regions and at variable depths: 25 m in the Bering Sea (Anderson, 1963), from Alaskan waters of less than 50 m (Loeblich and Tappan, 1953), 52 m to 100 m in Hudson Bay (Leslie, 1963), and from 55 m to 400 m depth in the Gulf of St. Lawrence, where salinities range from less than 33 ppt to 35 ppt, and temperatures vary between - 2° and 5°C (Rodrigues and Hooper, 1982a). D. frobisherensis was also reported from a number of arctic and subarctic regions: 40 m to 152 m depth in Hudson Bay (Leslie, 1965; Wagner, 1968), greater than 200 m in the Canadian arctic archipelago (Vilks, 1969), and from between

125 m and 520 m depth from the Gulf of St. Lawrence, associated with salinities of 33 ppt to 35 ppt and water temperatures of 1° to 5°C (Rodrigues and Hooper, 1982a).

A species which is common in one sample from this assemblage (Table 1) is Elphidiella groenlandica. In the Gulf of Alaska, Todd and Low (1967) found it associated with a muddy substrate which is a comparable environment to the present study. However, Loeblich and Tappan (1953) noted its presence in Alaskan waters less than 50 m deep. The occurrence of E. groenlandica in Biofacies A suggests that this species is not necessarily restricted to shallow waters and can be found at greater depths.

Biofacies A is characterized by the occurrence of five agglutinated foraminifera. Although rare, their presence in this assemblage is of some importance as they are associated with a clayey-silt substrate as opposed to a muddy sand or sand substrate of a more shallow environment. This aspect of arenaceous forms will be discussed further under Biofacies C.

Two species of the genus Alveophragmium have been found in the marine sediments of the study area. This genus is more characteristic of deeper-water environments as A. crassimargo is present at depths of 100 to 200 m in both the Bering Sea (Anderson, 1963) and Hudson Bay (Leslie, 1965). It has been reported by Vilks (1969) from the Canadian arctic archipelago in waters less than 200 m deep and as a predominant species in samples from 145 m (Wagner, 1968), but with a range of 43 m to 276 m in Hudson Bay. Hooper (1975) noted that in the St. Lawrence Estuary, mean depth for this species was only 40 m, and that the water temperature and salinity varied from - 1° to 8°C and 21 ppt to 32 ppt respectively. *A. crassimargo* was also associated with a silty sand substrate as opposed to a muddy substrate as in the present study. *A. jeffreysi* is found in the upper 200 m from the Canadian arctic archipelago (Vilks, 1969) and is a dominant species in samples from 199 m depth in Hudson Bay (Wagner, 1968), although it occurs in waters 43 m to 276 m deep.

The species *Reophax scorpiurus* appears to have a variable range in depth since it is found in Alaskan waters of less than 50 m deep (Loeblich and Tappan, 1953) and in Hudson Bay (Wagner, 1968) where it varies from 43 m to 276 m, becoming a predominant species at 117 m depth. Its presence in Biofacies A, therefore, is not totally unexpected.

Finally, Elphidium species contribute only moderately to this biofacies in comparison with Biofacies B and C (Table 1). Their relatively low numbers are significant in Biofacies A as they commonly comprise 90% of the total population in the littoral zone. Thus, lower numbers of these species can be expected in deeper-water environments where conditions are more favorable to a diverse assemblage.

Biofacies B

This biofacies is poorly represented in the stratigraphic record, so it is difficult to give a detailed description of its characteristics. It is comprised of samples F6, F12 and F14 (Table 1) and is marked by a decrease in both species diversity and the number of specimens present with respect to Biofacies A. The sediments associated with Biofacies B consist of sandy clayey silt representing a sublittoral facies. The fauna and stratigraphic relationship of the samples with other sediments suggest that this biofacies is a transitional zone between the deeper, off-shore environment of Biofacies A and the shallow, near-shore conditions of Biofacies C. In this context, its bathymetric range could possibly represent a depth of 30 m to

50 m. This low diversity assemblage is probably related to a minor deterioration in ecological conditions such as lowering of the water temperature, lowering of salinity, an increase in freshwater supply, and a change in the substrate. The above factors or a combination of them may be used to explain the general decrease in fauna brought upon by the shallowing of the environment caused by the lowering of relative sea level during the marine regression.

Biofacies B is dominated by Elphidium clavatum and E. orbiculare (Table 1) which are usually indicative of cold, shallow environments. However, the presence of Alveophragmium crassimargo, Reophax curtus and Virgulina loeblichi which may occur at greater depths, indicates a possible mixing of deep and shallow-water forms resulting in an intermediate depth assemblage representing a transition between species adapted to more saline waters, a muddy substrate and deeper water, and those which are better suited and more tolerant of brackish water, sandy substrate and overall harsher conditions of the littoral zone in the Arctic.

Biofacies C

This biofacies, represented by samples F1, F2, F3, F4, F5, F8, F9, F10 and F11, is characterized by a slight increase in species diversity and a greater number of specimens over Biofacies B (Table 1). The substrate associated with Biofacies C consists of sand or clayey sand interpreted as being a littoral and sublittoral lithofacies, deposited in waters of up to approximately 30 m depth.

The dominance of the genus *Elphidium* in this assemblage is of particular importance as it comprises ubiquitous species indicative of cold, shallow waters of low-salinity (Cronin, 1976). As noted in Champlain Sea studies (Cronin, 1979), nearshore assemblages are dominated by various *Elphidium* species, suggesting that water was brackish and probably less than 30 m in depth. Similar observations were made by Bergen and O'Neil (1979), who stated that Alaskan fauna dominated by *Elphidium* species are found at depths of less than 30 m. The relatively shallow arctic assemblage of the Chuckchi Sea living at depths of 6 m to 71 m (Cooper, 1964) is also dominated by species of the genus *Elphidium*. The shallow-water inshore fauna described by Cooper (1964) is similar to that of the present study as indicated by the presence of *Elphidium orbiculare* and *Buccella frigida* (Table 1). The shallow-water fauna along the Siberian coast described by Todd and Low (1966) is also comparable to Biofacies C.

As noted in Table 1, *Elphidium* species comprise from 86% to 100% of the total fauna. This type of dominance by various *Elphidium* species is not uncommon and is known to occur in low diversity assemblages from shallow Alaskan waters where the genus *Elphidium* encompasses 92% of the fauna (Lagoe, 1979b). More specifically, *E. clavatum* is typically restricted to relatively shallow, brackish waters such as estuarine and deltaic environments (Loeblich and Tappan, 1953; Anderson, 1963; Walton, 1964). It was found to be characteristic of marginal marine conditions, from shallow Alaskan waters (Lagoe, 1979b) to the Gulf of St. Lawrence (Bartlett and Molinsky, 1972), where it usually inhabits water less than 10 m to 20 m deep. However, Rodrigues and Hooper (1979a) have noted its presence as a dominant species as deep as 375 m, in waters of 34.5 ppt and temperatures of 4° to 5°C.

E. orbiculare, recorded in Holocene glaciomarine deposits of Baffin Island (Feyling-Hanssen, 1976) and the Svalbard archipelago (Nagy, 1984), has commonly been found at depths of less than 50 m, both in the Chuckchi Sea (Cooper, 1964) and in

Hudson Bay (Leslie, 1965). This species is also typical of shallow-water inshore fauna from the Canadian arctic archipelago (Vilks, 1969), as well as low diversity assemblages from shallow Alaskan waters (Lagoe, 1979b).

Not as common as other species, *E. bartletti* and *E. williamsoni* (Table 1) have been reported in waters of 55 m to 200 m depth (Rodrigues and Hooper, 1979a), but also in Pleistocene sediments of shallower waters. The abundance of *Elphidium* species which can tolerate salinities ranging from less than 18 ppt to greater than 28 ppt (Fillon and Hunt, 1974), seems to be connected to shallow water brackish conditions.

Buccella frigida is another typical shallow water species (Cooper, 1964; Vilks, 1969), and is thought to signify depths ranging from 20 m to 40 m (Cronin, 1976). Although this species is not restricted to Biofacies C (Table 1), it is here that it is found in a greater number of samples than in other biofacies, reflecting its wider spatial distribution in Biofacies C.

An important characteristic of Biofacies C is the apparent absence of agglutinated foraminifera. Several recent studies in arctic and subarctic regions have shown that arenaceous forms may contribute significantly to the foraminifera population.

In the Canadian Arctic, Vilks (1969) and Lagoe (1977) found that shallow water faunas are dominated by agglutinated species. This was also noted from the Eastern Siberian Coast (Todd and Low, 1966), where the fauna consists chiefly of arenaceous forms, associated with waters 10 m to 55 m deep. Agglutinated species are relatively common in Hudson Bay (Leslie, 1965), as well as in the Bering Sea (Anderson, 1963) in which arenaceous forms comprise 75% to 100% of the fauna between approximately 20 m to 100 m depth. Hooper (1975) noted that agglutinated foraminifera are characteristic of shallow waters with temperatures of - 1° to 8°C and salinities of 21 ppt to 32 ppt in the St. Lawrence Estuary.

From the above, the absence of arenaceous species from Biofacies C and their presence in Biofacies A could therefore be interpreted as anomalous. However, as noted earlier, the species occurring in this study are not restricted to shallow water environments. Additional detailed studies, which are beyond the scope of this investigation, would be required to more fully understand the palaeoenvironmental aspects and implications of these foraminiferal assemblages.

Although the possibility of transport and redeposition of foraminifera is present, it is difficult to estimate the importance of this process. Mixing of the fauna could have occurred, notably in the latter stages of sedimentation when erosion of previously deposited marine sediments could take place by downcutting of rivers. However, reworking of the sediments seems to have played a minor role as resedimentation should lead to random distribution of the microfauna, but this does not occur. Furthermore, very little abrasion of larger foraminifera was observed. Palaeogeographic conditions would also have favored comparatively minor displacement of faunas, as suggested by the relatively gentle slope of the sea floor.

The nature of the substrate has been considered as probably the most important factor in the distribution of foraminifera (Loeblich and Tappan, 1953; Wagner, 1968). Inasmuch as the foraminifera appear to be related to bottom sediment, i.e., shallow-water fauna of the sandy Biofacies C, and the deeper-water assemblage in the clayey silt of Biofacies A, the writer believes that salinity is the limiting factor of distribution in this study. It is the ability of the genus Elphidium to tolerate brackish water and harsh, rapidly fluctuating environmental conditions which permits it to dominate a zone which is generally inhospitable. Species requiring stable, more saline environments are more likely to be found at greater depths where more favorable near-normal marine conditions exist. It is believed that the relationship between fauna and sediments simply reflects the stratigraphic position of the microfauna within a typical regressive marine sequence in which shallow marginal marine conditions are commonly associated with a high energy coarse-grained sediment and where normal marine conditions are related to a deeper low energy zone with a fine-grained substrate. Thus, based mainly on comparisons with studies of recent ecology, high diversity assemblages such as Biofacies A reflect average marine salinities whereas low diversity associations as noted in Biofacies C, frequently point to brackish water environments, regardless of the nature of the substrate.

ii) Ostracoda

Temperature and salinity sensitive ostracode species provide a useful tool for inferring temporal environmental changes during the history of marine inundation in the basin of the Richardson and Rae Rivers. Although ostracodes are not as common as foraminifers (Table 2), they generally can provide a greater potential for indicating palaeoenvironmental conditions because of distinct freshwater, brackish water and marine species. The number of microfossils recovered from each sample is highly variable, and consequently the number of species also varies. A total of 16 different ostracode species were recorded (Appendix F) and are believed to be fairly representative of the ecological conditions.

Ostracodes have been used to define two environmentally distinct phases of the sea (Table 2); Assemblage A corresponds to Biofacies A of the foraminiferal study, whereas Assemblage B represents Biofacies B and C which cannot be individually distinguished with the use of ostracodes.

Assemblage A

Assemblage A, represented by samples F13 and F15, is associated with clayey silt deposits considered to be a deep-water lithofacies of the early postglacial sea. It is characterized by the presence of _Cytheropteron simplex_, _C. paralatissimum_, _Roundstonia globulifera_ and _Palmenella limicola_ which are restricted to this assemblage. _Cytheropteron simplex_ has been recorded in deep-water silts

TABLE 2. Distribution and abundance (%) of Ostracoda.

	ASSEMBLAGES									
	A		B							
	SAMPLE NUMBER									
SPECIES	F13	F15	F6	F12	F5	F4	F9	F8	F11	F2
EUCYTHERIDEA BRADII	17	40	21	18	34	100		14	34	4
E. MACROLAMINATA	17	20	8	26	33		26		66	2
E. PUNCTILLATA			5							2
CYTHEROPTERON PARALATISSIMUM	8		3							
C. SIMPLEX	34			4						
PALMENELLA LIMICOLA	16	40								
ROUNDSTONIA GLOBULIFERA	8									
HETEROCYPRIDEIS SORBYANA				48			6			53
PSEUDOCYTHEREIS SIMPSONENSIS				4				78		32
PARACYPRIDEIS PSEUDOPUNCTILLATA			63				31			
LOXOCONCHA VENEPIDERMOIDEA					33		13			
LIMNOCYTHERE cf. *HERRICKI*							6			
ILYOCYPRIS GIBBA							12			
I. BRADYI								8		
ACANTHOCYTHEREIS DUNELMENSIS							6			
RABILIMIS cf. *SEPTENTRIONALIS*										7
NUMBER OF SPECIES	6	3	5	5	3	1	7	3	2	6
NUMBER OF SPECIMENS	12	5	39	23	3	1	16	14	3	105

and clays of the Champlain Sea (Cronin, 1977b), and according to Whatley and Masson (1979), its distribution is restricted to the Pleistocene. It is considered to be a marine species (S > 30 ppt) indicative of salinities of 29 ppt to 35 ppt (Cronin, 1981) and temperatures of 0° to 12°C. Brasier (1983) also stated that Cytheropteron species are deeper water ostracodes found particularly on fine-grained substrates. C. paralatissimum, occurring widely in Arctic seas, is characteristic of sublittoral Arctic faunas (Neale and Howe, 1975).

Palmenella limicola has been recorded from Pleistocene marine deposits of northern Alaska (Swain, 1963), and is reported living in waters of the Gulf of Alaska, the fjords of Norway and off eastern Greenland (Hazel, 1970). The temperature tolerance of this species varies from 0° to about 12°C, and it has a depth range of 5 m to 200 m. P. limicola is considered to be a marine species as it is associated with salinities of 18 ppt to 35 ppt (Neale, 1963) and was found to represent a relatively deep water facies typified by glaciomarine silty clays of the Champlain Sea (Cronin, 1977a, 1981).

Assemblage B

This assemblage is represented by samples F6, F12, F5, F4, F9, F8, F11 and F2 (Table 2). It is dominated by species representative of brackish water environments (0.5 ppt to 30 ppt), as well as three freshwater (S > 0.5 ppt) ostracode species. The genus Eucytheridea is a common constituent of all samples in both assemblages A and B. E. bradii and E. punctillata are generally considered to be brackish water species which can tolerate salinities from 10 ppt to 35 ppt, and both species are abundant in sandy Champlain Sea deposits (Cronin, 1977b). Their annual temperature range is from 0° to 15°-20° C, but their depth range varies slightly: 5 m to 435 m for E. punctillata and 5 m to 750 m for E. bradii (Hazel, 1970). A third species, E. macrolaminata, generally lives in frigid climatic zones, and is known to live off the coast of Norway where temperatures range from - 2° to 5°-10° C (Elofson, 1941).

Heterocyprideis sorbyana is also relatively abundant in Assemblage B (Table 2). Cronin (1977a) considered this species to be the most common ostracode species of the Champlain Sea because of its broad temperature and salinity tolerances. This shallow, brackish water ostracode lives in waters of 2 ppt to 35 ppt, with temperatures ranging from - 2° to 18°C (Neale, 1963). H. sorbyana is associated with fine and coarse-grained substrates in the present study, suggesting that its distribution is unaffected by changes in the substrate (Benson et al., 1983). Although this species is most commonly found in shallow water environments, it can occur as deep as 435 m (Hazel, 1970).

The extreme rarity of Acanthocythereis dunelmensis in Assemblage B attests to the predominant brackish water nature of the fauna. A. dunelmensis is a marine species indicative of salinities of approximately 32 ppt to 35 ppt and temperatures of 0° to 12°C (Elofson, 1941; Cronin, 1977a).

The importance of Pseudocythereis simpsonensis, Paracyprideis pseudopunctillata and Loxoconcha venepidermoidea is difficult to ascertain as these three species have only been previously reported from Pleistocene marine deposits of northern Alaska (Swain, 1963), where they have been considered representative of the marine facies as a whole. Based on their distribution in the present study (Table 2), it would appear that they are characteristic of relatively shallow, marginal marine environments, since they contribute significantly in some cases to the ostracode fauna.

Rabilimis septentrionalis, characteristic of Arctic faunas, has also been reported from sublittoral deposits in Alaska, Greenland, and from Russian Harbour in waters 2 m to 30 m deep (Neale and Howe, 1975).

The presence of three freshwater ostracode species found in Assemblage B is also of interest. Two species of the genus Ilyocypris found in the marine sediments of this study have been reported as common constituents of freshwater Pleistocene deposits by Cronin (1977a) and by Swain (1963). The habitats of these species include a wide range of environments, from ephemeral ponds and intermittent streams to permanent running waters. I. bradyi has been found in shallow-water, stagnant environments (Robinson, 1978) and lakes of low salinity (De Deckker et al., 1979); Limnocythere herricki is associated with shallow lakes (Staplin, 1963). However, the introduction of non-marine forms into the marine environment should be expected. Transport and mixing of freshwater, brackish water and marine faunas is particularly evident in estuaries and deltas (Neale, 1969). The freshwater ostracodes are likely to drift out in a layer of less dense water, overlying a salt wedge in which tidal currents bring marine forms landwards. The influx of freshwater from the ancestral Richardson and Rae Rivers may have resulted in the low diversity ostracode assemblages noted in this study, as low diversity faunas from parts of the Beaufort Sea were attributed to reduced salinities caused by the discharge of the Mackenzie River (Briggs, 1980).

Assuming that some life forms were present in the proglacial lake prior to the marine incursion, no fauna from the lacustrine environment survived the initial period of mixing between the freshwater of Glacial Lake Richardson and the incoming marine water. There is also no record of an early glaciomarine ostracode fauna, so that the first and oldest assemblage encountered (Assemblage A) represents saline conditions already existing at greater depth (50 m to 150 m), as indicated by the presence of marine ostracode species.

With decreasing depth, the stability of the environment generally decreases as does salinity. This is normally accompanied by an increase in the energy level of the environment, as well as an increase in the grain size of the sediments. The mixed associations of marine, brackish and freshwater ostracodes suggest periods of fluctuating salinities, affected by an increasing influence of freshwater run-off in a progressively shallowing estuarine environment. Salinity appears to be the fundamental factor determining the distribution of ostracodes. Low diversity assemblages dominated by Eucytheridea and Heterocyprideis are characteristic of brackish waters of estuaries and near-shore conditions, reflecting the instability of the environment, notably oscillations in salinity caused by run-off from the drainage basin. This is not totally unexpected if one considers the geometry of the sea with the basin (Figure 10). As the sea gradually regressed to lower elevations due to isostatic rebound, an estuarine environment persisted throughout this period, resulting in the dilution of the marine waters.

The nature of the substrate seems to have had little effect on ostracode distribution, since species appear to be found equally in all environments. However, Pokorný (1978) noted that coarse-grained sediments contained small ostracode populations, whereas heterogeneous sediments supported a larger, more diversified fauna. Despite the large depth range of certain ostracode species, the fauna present in Assemblage A includes species usually associated with shallow waters whose temperature does not exceed 10°-12°C. Although some species listed in Table 2 have the ability to live in warmer waters, this upper temperature limit is defined by

Cytheropteron simplex and Palmenella limicola. Whether similar conditions occurred throughout this entire region in areas affected by the postglacial marine inundation is not yet known, and will be the subject of a future study by the writer.

iii) Macrofossils

The macrofossil assemblage recorded in postglacial marine sediments is dominated by mollusc species (Table 3). In addition to the species listed in Appendix G, Craig (1960) noted the presence of Mya pseudoarenaria from Holocene marine deposits north of the study area. Several other different species of marine shells collected by O'Neill (1924) along the coast of Dolphin and Union Strait were not observed in the study area. Fossils listed in Table 3 represent an impoverished bivalve fauna, suggesting that environmental conditions were fairly rigorous. Although there is no clear faunal succession apparent in Table 3, it is possible to define some noticeable differences between the various samples studied.

The first of these is the occurrence of the genus Portlandia in samples F13 and F15, which have previously been assigned to a deep-water facies. P. arctica has been reported from Hudson Bay (Wagner, 1968), where it was found to live at depths of

TABLE 3. Distribution of macrofossils.

	SAMPLE NUMBER							
SPECIES	F13	F15	F12	F6	F11	F2	F8	F3
GASTROPOD SP.		X						
PORTLANDIA ARCTICA		X						
P. LENTICULA	X							
MACOMA BALTHICA	X				X	X		X
M. CALCAREA				X		X		
HIATELLA ARCTICA				X		X		
MYTILUS EDULIS			X		X			X
MYA TRUNCATA						X		X
SAXICAVA ARCTICA					X			
BALANUS SP.							X	

100 m to 210 m, occurring most commonly between 117 m and 196 m. Ockelman (1958) noted its presence at depths of 10 m to 50 m near melt-water streams off the fronts of glaciers in Greenland, where cold, freshwater influx occurs. This species was associated with clay sediments of the Champlain Sea, representing a relatively deep offshore facies in which salinities ranged from 20 ppt to 34 ppt (Rodrigues and Richard, 1983, 1985). *P. arctica* has also been associated with muddy glaciomarine environments in waters which do not exceed 0°C (Hillaire-Marcel, 1980; Mode *et al.*, 1983). The occurrence of *P. lenticula* is an additional indication of a deep water environment as this species has been found in waters 101 m to 276 m deep from Hudson Bay, where bottom-water temperatures did not exceed - 1°C (Wagner, 1968).

The remaining samples, F12, F6, F11, F2, F8 and F3, contain six other types of molluscs which are generally associated with slightly shallower waters. *Macoma calcarea*, a typically arctic species, has been recorded living in waters 76 m to 276 m deep from Hudson Bay (Wagner, 1968), but was found to be most abundant at the shallower depth range. This species lives in waters which do not exceed 0°C and whose salinities are greater than 30 ppt (Hillaire-Marcel, 1980). It is adapted to both sand and clay substrates (Rodrigues and Richard, 1983, 1985).

Hiatella arctica has also been reported from a range of water depths: 35 m to 165 m in Hudson Bay (Wagner, 1968) and 50 m to 60 m in Greenland (Ockelman, 1958), although Wagner (1968) found it most abundant between 35 m and 60 m depth. This species is associated with sand and muddy sand from shallow water of the Champlain Sea and tolerates salinities of 20 ppt to 32 ppt (Rodrigues and Richard, 1983, 1985). Its occurrence in F6 and F2 is in accordance with the interpretation of waters approximately 30 m to 60 m deep..

Another typically arctic species, *Mya truncata*, was found in a shallow water deltaic environment (Sample F3, Appendix D-20) and a sublittoral zone (Sample F2, Appendix D-25). Wagner (1968) reported this species present at depths of 35 m to 97 m in Hudson Bay, but was most abundant from waters 35 m to 50 m deep. Two other species associated with a near-shore environment are *Mytilus edulis* and *Saxicava arctica* (Table 3). The former is largely restricted to the intertidal environment, and although it prefers rocky surfaces, it can survive anchored on fine to coarse-grained sediments or other objects on the substrate (Stanley, 1970). However, it is not as well adapted to brackish water as are other molluscs. S. arctica, known from only one site (Table 3), is a burrower which inhabits sandy littoral deposits (Sowerby, 1859).

The most common bivalve is Macoma balthica. This is a ubiquitous species which can tolerate a wide range of salinities and temperatures. This ability provides it with a considerable advantage over other species. It is common to intertidal and shallow subtidal environments but has also been found living in waters as deep as 276 m (Wagner, 1968; Stanley, 1970). M. balthica is also adapted to both coarse-grained sediments of the littoral zone, as well as muddy deposits of offshore environments, a fact which explains its presence in a wide range of samples (Table 3). This variable depth range is accompanied by a high tolerance of fluctuating salinities, from [illegible] ppt to 35 ppt (Hillaire-Marcel, 1980).

The pelagic larvae of M. balthica require a minimum temperature of 10°C to survive (Lammens, 1967), which indicates that the shallow marine waters attained temperatures of 10°-12°C at least for a short period of time during the summer.

From their distribution, it would appear that certain molluscs are controlled primarily by depth, and that texture of the bottom sediments is of secondary importance. Hiatella arctica, Mya truncata, Macoma balthica and M. calcarea were the only species recorded above 73 m a.s.l. The highest elevation at which fossils (M. calcarea) were found in the basin is 90 m a.s.l., situated more than 60 km from the present coastline. Species found below 15 m a.s.l. in the eastern part of the basin (Portlandia arctica and P. lenticula) are characteristic of more truly marine conditions as indicated by the foraminiferal Biofacies A and the ostracode Assemblage A. However, the first species to enter the inundated basin were probably tolerant of lower salinity conditions, but they are not represented here, unless P. arctica is a relict species of an earlier glaciomarine facies which has not yet been observed in the stratigraphic record. The molluscs found at higher elevations are generally related to shallow water environments with coarse-grained sediments deposited at successively lower elevations as the sea regressed. Therefore, the waters of the postglacial sea may have been density stratified with respect to salinity; water with salinities of 30 ppt to 34 ppt occupied the central and eastern deeper parts of the basin and were overlain by waters with salinities of less than 30 ppt present at shallower depths. The arctic affinities of the fauna and the apparent absence of a more diversified bivalve assemblage can best be explained by a combination of controlling factors such as the relatively cold temperature of the water, an intermittent and/or insufficient supply of nutrients, the detrimental effects of coastal and littoral ice, and more importantly, considerable fluctuations in salinity.

CHAPTER 6

Conclusion

The purpose of this thesis was to investigate the nature of Late Quaternary stratigraphy in the basin of the Richardson and Rae Rivers, N.W.T. This objective was attained by identifying a succession of depositional environments on the basis of sedimentary structures, grain size, facies associations and stratigraphic relationships.

The methods employed aided in the interpretation of sedimentary sequences which record a series of distinct sedimentary facies characteristic of particular environments in a well defined succession. Beginning with the retreat of the Laurentide Ice Sheet, these facies resulted from a wide variety of rapidly changing environmental conditions. Ice retreat is initially represented by basal till and/or flow till lying directly on bedrock. These multi-till sequences are the result of changes in depositional style at the ice margin, i.e., sediments deposited directly by the ice and those which have resulted from flow processes.

Subaqueous outwash deposits issuing from subglacial conduits form a distinctive facies sequence: proximal cross-bedded gravels deposited near the mouth of the ice tunnel, massive poorly sorted sand, ripple-drift cross-laminated fine sand and silt, and distal glaciolacustrine silty clay rhythmites. Some of these successions are thought to represent an esker to varve transition in which bedding structures reflect a progressive decrease in current velocity in the downstream direction in relatively deep water. It is possible to have lateral variations between this type of sequence and till due to fluctuations between deposition of outwash and glacial drift which is not a continuous process.

Sedimentation of coarse and fine-grained facies was dominated by underflows, both near the ice margin and at a distance, and the generally conformable contacts between the finer-grained sediments indicate that deposition was a continuous event. Strong seasonal variations in run-off and winter ice cover on lakes led to deposition of varved couplets, with the silt layer deposited during the summer and the clay-rich layer resulting from suspension during the winter. Palaeocurrent directions and lithofacies associations are compatible with the configuration of the reconstructed ice fronts during deglaciation.

The continuous eastward retreat of the ice front permitted lake levels to fall as lower outlets were freed of ice. The glacier, shortly after having liberated the highlands to the northeast of the basin, permitted the marine incursion which marked the end of the last phase of the glacial lake. The sudden passage from a lacustrine environment to a marine environment is marked by the cessation of the deposition of glaciolacustrine rhythmites by turbidity currents and the beginning of a marine sedimentation characterized by massive silty-clay deposits resulting from the flocculation of fine-grained particles. Following an initial maximum invasion, the sea gradually regressed to lower elevations due to crustal rebound. During this stage, numerous deltas were formed, and a coarsening upward marine succession was deposited as a result of emergence.

The marine micro and macrofauna also serve as alternate sources of evidence used to corroborate this interpretation. Fossils from the oldest marine sediments (clayey-silts) record cold, near-normal marine conditions in deep water environments. These are overlain by younger coarser-grained marine sediments characterized by an Elphidium dominated foraminiferal fauna, ubiquitous ostracode species and an impoverished bivalve assemblage. Seasonal changes in temperature, salinity fluctuations, extensive ice cover and turbulence in the littoral zone were major factors limiting the fauna to only species tolerant of dramatic changes in their habitat.

In summary, this study has contributed to a better understanding of Late Quaternary stages of deglaciation and postglacial history in this region of the Northwest Territories. Stratigraphic and sedimentological data were presented in order to describe the succession of sedimentary environments resulting primarily from the retreat of the ice front. The study of such sequences provides more detailed information that may serve to support and complete the reconstruction of the Late Quaternary history of an area which cannot be inferred by the mapping of surficial deposits alone.

Other detailed investigations would be required to further define the depositional environments presented in this thesis. These could include geochemical analyses of tills to compare and contrast the mode of deposition of these deposits, as well as stratigraphic and sedimentological studies on a smaller scale of particular sites.

A greater number of radiocarbon dates are also needed to complete an absolute temporal framework of the marine regression and to help establish a more accurate postglacial uplift curve. Furthermore, the writer intends to continue his investigation of glaciomarine and marine deposits exposed along the coast of Dolphin and Union Strait, Coronation Gulf and Queen Maud Gulf, areas which have received little attention to date.

REFERENCES

AGTERBERG, F., and BANERJEE, I., 1969: Stochastic model for the deposition of varves in glacial Lake Barlow-Ojibway, Ont., Canada; Can. Jour. Earth Sci., v. 6, pp. 625-652.

ALLAN, R., LYNCH, J., and LUND, N., 1972: Regional geochemical exploration in the Coppermine River area, District of Mackenzie: a feasibility study in permafrost terrain; Geol. Surv. Can., Paper 71-33, 52 p.

ANDERSON, G., 1963: Distribution patterns of Recent foraminifera of the Bering Sea; Micropaleontology, v. 9, pp. 305-317.

ANDERSON, J., DOMACK, E., and KURTZ, D., 1980: Observations of sediment-laden icebergs in Antarctic waters: implications to glacial erosion and transportation; Jour. Glaciol., v. 25, pp. 387-397.

ANDREWS, J., 1970: Present and postglacial rates of uplift for glaciated northern and eastern North America derived from postglacial uplift curves; Can. Jour. Earth Sci., v. 7, pp. 703-715.

REFERENCES - Continued

ARNBORG, L., 1955: Hydrology of the Glacier River Austurfljot (Hoffellssandur, Part II): Geografiska Annaler, v. 37, pp. 185-201.

ASHLEY, G., 1975: Rhythmic sedimentation in Glacial Lake Hitchcock, Massachusetts-Connecticut. In Glaciofluvial and Glaciolacustrine Sedimentation, A. Jopling and B. McDonald (Eds), Soc. Econ. Paleon. Mineral., Spec. Publ. n. 23, pp. 304-320.

ASHLEY, G., SOUTHARD, J., and BOOTHROYD, J., 1982: Deposition of climbing-ripple beds: a flume simulation: Sedimentology, v. 29, pp. 67-79.

BANERJEE, I., 1973: Sedimentology of Pleistocene glacial varves in Ontario, Canada; Geol. Surv. Can., Bull. 226, Part A, pp. 1-44.

BANERJEE, I., and MCDONALD, B., 1975: Nature of esker sedimentation. In Glaciofluvial and Glaciolacustrine Sedimentation, A. Jopling and B. McDonald (Eds), Soc. Econ. Paleon. Mineral., Spec. Publ. n. 23, pp. 132-154.

REFERENCES - Continued

BARAGAR, W., and DONALDSON, J., 1973: Coppermine and Dismal Lakes Map Areas; Geol. Surv. Can., Paper 71-39, 20 p., Maps 1,337a, 1,338a.

BARTLETT, G., and MOLINSKY, L., 1972: Foraminifera and the Holocene history of the Gulf of St. Lawrence; Can. Jour. Earth Sci., v. 9, pp. 1,204-1,215.

BENSON, R., DELGROSSO, R., and STEINECK, P., 1983: Ostracode distribution and biofacies, Newfoundland continental slope and rise; Micropaleontology, v. 29, n. 4, pp. 430-453.

BERGEN, F., and O'NEIL , P., 1979: Distribution of Holocene Foraminifera in the Gulf of Alaska; Jour. of Paleontology, v. 53, n. 6, pp. 1,267-1,292.

BLAKE, W., Jr., 1970: Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates and differential postglacial uplift in the Eastern Queen Elizabeth Islands; Can. Jour. Earth Sci., v. 7, pp. 634-664.

BLATT, H., MIDDLETON, G., and MURRAY, R., 1980: Origin of Sedimentary Rocks; Prentice-Hall, Inc., 782 p.

REFERENCES - Continued

BOOTHROYD, J., 1984: Glaciolacustrine and glaciomarine fans: a review; Programs with Abstracts, GSA Northeastern Section, v. 16, p. 4.

BOULTON, G., 1968: Flow tills and related deposits on some Vestspitsbergen glaciers; Jour. Glaciol., v. 7, pp. 391-412.

__________, 1972: Modern glaciers as depositional models for former ice sheets; Jour. Geol. Soc. London, v. 128, Part 4, pp. 361-393.

BRASIER, M., 1983: Microfossils; George Allan and Unwin, London, pp. 122-146.

BRIGGS, W., 1980: The modern ostracode fauna of the continental shelf, slope and rise of the southeastern Beaufort Sea; American Quat. Assoc. Abstracts with Programs, 6th Mtg, Orono, Maine, p. 49.

BRUNEAU, H., 1984: L'évolution postglaciaire du secteur aval de la rivière Coppermine, T.N.-O.; Thèse de maîtrise, Département de géographie, Université d'Ottawa, Ottawa, 201 p.

REFERENCES - Continued

CAMPBELL, F., 1983: Stratigraphy of the Rae Group, Coronation Gulf Area, Districts of Mackenzie and Franklin; In Current Research, Part A, Geol. Surv. Can., Paper 83-1A, pp. 43-52.

CHEEL, R., and RUST, B., 1982: Coarse-grained facies of glacio-marine deposits near Ottawa, Canada; In Research in Gla~~cial,~~ Glacio-Fluvial, and Glacio-Lacustrine Systems; R. Davidson-Arnott, W. Nickling, B. Fahey (Eds), 6th Guelph Symp. on Geomorph., 1980, pp. 279-295.

CLAYTON, K., 1973: Glacial and fluvioglacial landforms; Hafner Publ. Co., 242 p.

COHEN, J., 1979: Deltaic sedimentation in glacial Lake Blessington, County Wicklow, Ireland; In Moraines and Varves, Ch. Schuchter and A. Balkema (Eds), Rotterdam, pp. 357-368.

COLEMAN, J., and PRIOR, D., 1981: Deltaic sand bodies; Americ. Assoc. Petrol. Geol., pp. 5-65.

CONYBEARE, C., 1979: Lithostratigraphic Analysis of Sedimentary Basins; Academic Press, 555 p.

COOPER, S., 1964: Benthonic foraminifera of the Chuckchi Sea; Cushman Found. Foram. Res. Contrib. 15, pp. 79-104.

REFERENCES - Continued

CRAIG, B., 1960: Surficial geology of the North-Central District of Mackenzie, Northwest Territories; Geol. Surv. Can., Paper 60-18, Map 24-1960.

CRONIN, T., 1976: Preliminary report on paleoecology of the southern Champlain Sea as indicated by benthonic Foraminifera; Maritime Sediments, Spec. Publ. 1, Part B.

__________, 1977a: Late-Wisconsin marine environments of the Champlain valley (New York, Quebec); Quaternary Research, 7, pp. 238-253.

__________, 1977b: Champlain Sea Foraminifera and Ostracoda: a systematic and paleoecological synthesis; Géogr. phys. Quat., v. XXXXI, n. 1-2, pp. 107-122.

__________, 1979: Late Pleistocene benthic foraminifers from the St. Lawrence Lowlands; Jour. Paleontology, v. 53, n. 4, pp. 781-814.

__________, 1981: Paleoclimatic implications of Late Pleistocene marine ostracodes from the St. Lawrence Lowlands; Micropaleontology, v. 27, n. 4, pp. 384-418.

DAMUTH, J., and EMBLEY, R. 1979: Upslope flow of turbidity currents on the northwest flank of the Ceará Rise: western Equatorial Atlantic; Sedimentology, v. 26, pp. 825-834.

REFERENCES - Continued

DE DECKKER, P., GEURTS, M.-A., and JULIA, R., 1979: Seasonal rhythmites from a Lower Pleistocene lake in northeastern Spain; Palaeogr., Palaeoclimatol., Palaeoecol., v. 26, pp. 43-71.

DE GEER, G., 1912: A geochronology of the last 12,000 years; Int. Geol. Cong., XI Stockholm, 1910, Compte Rendu, v. 1, pp. 241-258.

__________, 1940: Geochronologia Suecica principles; Kungl. Svenska Vetenskapsakademiens Handlingar, Series 3, 18, pp. 1-367.

DEPARTMENT OF ENERGY, MINES AND RESOURCES, 1980: Principal mineral areas of Canada; Map 900A.

DOMACK, E., 1984: Rhythmically bedded glaciomarine sediments on Whidbey Island, Washington; Jour. Sed. Petrol., v. 54, n. 2, pp. 589-602.

DOMACK, E., and LAWSON, D., 1985: Pebble fabric in an ice-rafted diamicton; Jour. of Geology, v. 93, pp. 577-591.

REFERENCES - Continued

DOWDESWELL, J., OSTERMAN, L., and ANDREWS, J., 1985: Quartz sand grain shape and other criteria used to distinguish glacial and non-glacial events in a marine core from Frobisher Bay, Baffin Island, N.W.T., Canada; Sedimentology, v. 32, pp. 119-132.

DREIMANIS, A., 1976: Tills, their origin and properties; In Glacial Till, R. Legget (Ed), R. Soc. Canada, Spec. Publ. 12, pp. 11-49.

____________, 1982: Work Gp. 1, INQUA Commission on genesis and lithology of Quaternary deposits; In Report of Activities 1977-1982, C. Schluchter (Ed), ETH, Zurich, pp. 12-31.

DREWRY, D., and COOPER, A., 1981: Processes and models of Antarctic glaciomarine sedimentation; Annals of Glaciology, v. 2, pp. 117-122.

EDWARDS, M., 1978: Glacial Environments; In Sedimentary Environments, H. Reading (Ed), Elsevier, New York, 557 p.

ELOFSON, O., 1941: Zur Kenntnis der marinen Ostracoden Schwedens mit besonderer Berücksichtigung des Skageraks; Zool. Bidr. Uppsala, v. 19, pp. 215-534.

REFERENCES - Continued

ELVERHØI, A., LIESTØL, O., and NAGY, J., 1980: Glacial erosion, sedimentation and microfauna in the inner part of Kongfjorden, Spitsbergen; Norsk Polarinst. Skrifter 172, pp. 33-61.

EYLES, N., EYLES, C., and MIALL, A., 1983: Lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences, Sedimentology, v. 30, pp. 393-410.

EYLES, N., and MIALL, A., 1984: Glacial Facies; In Facies Models, R. Walker (Ed), Geoscience Canada, Reprint Series 1, 2nd edition, pp. 15-38.

FEYLING-HANSSEN, R., 1976: The stratigraphy of the Quaternary Clyde Foreland Formation, Baffin Island, illustrated by the distribution of benthic foraminifera; Boreas, v. 5, pp. 77-94.

FEYLING-HANSSEN, R., and BUZAS, M., 1976: Emendation of Cassidulina and Islandiella helenae new species; Jour. Foram. Research, v. 6, pp. 154-158.

REFERENCES - Continued

FILLON, R., and HUNT, A., 1974: Late Pleistocene benthic Foraminifera of the southern Champlain Sea: Paleotemperature and paleosalinity indications; Maritime Sediments, v. 10, n. 1, pp. 14-18.

FOLK, R., 1980: Petrology of Sedimentary Rocks; Hemphill, 61 p.

FOLK, R., and WARD, W., 1957: Brazos River bar: a study in the significance of grain size parameters; Jour. Sed. Petrol., v. 27, pp. 3-26.

FRANCIS, E., 1974: Glacial sediments: a selective review; In Ice Ages: Ancient and Modern; A. Wright and F. Moseley (Eds), Seel House Press, pp. 43-68.

GILBERT, R., 1982: Contemporary sedimentary environments on Baffin Island, N.W.T., Canada: Glaciomarine processes in fjords of eastern Cumberland Peninsula; Arctic and Alpine Research, v. 14, pp. 1-12.

__________, 1983: Sedimentary processes in Canadian Arctic fjords; Sedimentary Geology, v. 36, pp. 147-175.

REFERENCES - Continued

GUSTAVSON, T., 1975: Sedimentation and physical limmology in proglacial Malaspina Lake, southeastern Alaska; In Glaciofluvial and Glaciolacustrine Sedimentation, A. Jopling and B. McDonald (Eds), Soc. Econ. Paleon. Mineral., Spec. Publ. n. 23, pp. 249-263.

GUSTAVSON, T., ASHLEY, G., and BOOTHROYD, J., 1975: Depositional sequences in glaciolacustrine deltas; In Glaciofluvial and Glaciolacustrine Sedimentation, A. Jopling and B. McDonald (Eds), Soc. Econ. Paleon. Mineral., Spec. Publ. n. 23, pp. 264-281.

HALD, M., and VORREN, T., 1984: Modern and Holocene foraminifera and sediments on the continental shelf off Troms, North Norway; Boreas, v. 13, pp. 133-154.

HALDORSEN, S., and SHAW, J., 1982: The problem of recognizing melt-out till; Boreas, v. 11, pp. 261-277.

HARLAND, W., HEROD, K., and KRINSLEY, D., 1966: The definition and identification of tills and tillites; Earth Sci. Rev., v. 2, pp. 225-256.

REFERENCES - Continued

HARRISON, S., 1975: Turbidite origin of glaciolacustrine sediments, Woodcock Lake, Pennsylvania; Journ. Sed. Petrol., v. 45, pp. 738-744.

HAZEL, J., 1970: Ostracode zoogeography in the southern Nova Scotian and northern Virginian faunal provinces, U.S. Geol. Surv., Prof. Paper 529-E, 21 p.

HILLAIRE-MARCEL, C., 1980: Les faunes des mers postglaciaires du Québec: quelques considérations paléoécologiques; Géogr. phys. Quat., v. XXXIV, n. 1, pp. 3-59.

HOOPER, K., 1975: Foraminiferal ecology and associated sediments of the lower St. Lawrence Estuary; Jour. Foram. Research, v. 5, pp. 218-238.

INQUA Commission on Genesis and Lithology of Quaternary Deposits, 1982: In Report of Activities, 1977-1982, C. Schluchter (Ed), ETH, Zurich.

JELGERSMAE, S., 1966: Sea-level changes during the last 10,000 years; In Proceedings of the International Symposium on World Climate 8,000 to 0 B.C., J. Sawyer (Ed), pp. 54-71.

REFERENCES - Continued

JOPLING, A., and WALKER, R., 1968: Morphology and origin of ripple-drift cross-laminations, with examples from the Pleistocene of Massachusetts; Jour. Sed. Petrol., v. 38, pp. 971-984.

KARROW, P., 1984: Discussion on lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences; Sedimentology, v. 31, pp. 883-898.

KENNEY, T., and CHAN, H., 1972: Use of radiographs in a geological and geotechnical investigation of varved soil; Can. Geotech. Jour., v. 9, pp. 195-205.

KERR, D., 1984a: Stratigraphy, sedimentology and palaeoecology of Late Quaternary sediments, Richardson River Basin, N.W.T., B. Sc. Thesis, Department of Geology, University of Ottawa, Ottawa, 92 p.

________, 1984b: Marine regression and palaeoenvironments during the Holocene in the Richardson River basin, District of Mackenzie, N.W.T.; Géogr. phys. Quat., v. XXXVIII, n. 2, pp. 185-192.

REFERENCES - Continued

KERR, D., and MERCIER, A., 1985: Stratigraphic evidence for the existence of a glacial lake in the Richardson and Rae River basins, N.W.T., Contributions to the Geology of the Northwest Territories, INAC, v. 2, pp. 55-62.

KEYS, J., and WILLIAMS, K., 1984: Rates and mechanisms of iceberg ablation in the d'Urville Sea, Southern Ocean; Jour. of Glaciology, v. 30, n. 105, pp. 218-222.

KRANCK, K., 1975: Sediment deposition from flocculated suspensions; Sedimentology, v. 22, pp. 111-123.

_________, 1981: Particulate matter grain-size characteristics and flocculation in a partially mixed estuary; Sedimentology, v. 28, pp. 107-114.

KRUGER, J., and MARCUSSEN, I., 1976: Lodgement till and flow till: a discussion; Boreas, v. 5, pp. 61-64.

KUENEN, Ph. H., 1951: Mechanics of varve formation and the action of turbidity currents; Geol. Foren. Forhandl. Bd 73, H. 1, pp. 69-84.

LAGOE, M., 1977: Recent benthic Foraminifera from the central Arctic Ocean; Jour. of Foram. Research, v. 7, pp. 100-130.

REFERENCES - Continued

LAGOE, M., 1979a: Recent benthonic foraminiferal biofacies in the Arctic Ocean; Micropaleontology, v. 25, pp. 214-224.

________, 1979b: Modern benthic Foraminifera from Prudhoe Bay, Alaska; Jour. of Paleontology, v. 53, pp. 258-262.

LAMBERT, A., and HSÜ, K., 1979: Non-annual cycles of varve-like sedimentation in Walensee, Switzerland; Sedimentology, v. 26, pp. 453-461.

LAMMENS, J., 1967: Growth and reproduction in a tidal flat population of Macoma balthica; Netherlands Jour. Sea Research, v. 3, pp. 315-382.

LECKIE, D., and MCCANN, S., 1982: Glacio-lacustrine sedimentation on low slope prograding delta; In Research in Glacial, Glacio-fluvial, and Glacio-lacustrine systems; R. Davidson-Arnott, W. Nicling and B. Fahey (Eds), 6th Guelph Symp. on Geomorphology, 1980, pp. 261-278.

LESLIE, R., 1965: Ecology and paleoecology of Hudson Bay Foraminifera; Bedford Inst. of Oceanogr., Dartmouth, Nova Scotia, Rept. 65-6, 192 p.

REFERENCES - Continued

LOCAT, J., 1982: Contribution à l'étude de l'origine de la structuration des argiles sensibles du Canada; Thèse de Doctorat, Université de Sherbrooke, Sherbrooke, Québec.

LOEBLICH, A., and TAPPAN, H., 1953: Studies in Arctic foraminifera; Smithsonian Miscel. Coll., v. 121, 150 p.

LOWE, D., 1979: Sediment gravity flows: their classification and some problems of application to natural flows and deposits; Soc. Econ. Paleont. Mineral., Spec. Publ. n. 27, pp. 75-82.

MARKUSSEN, B., ZAHN, R., and THIEDE, J., 1985: Late Quaternary sedimentation in the eastern Arctic Basin: stratigraphy and depositional environments; Palaeogeography, Palaeoclimatology, Palaeoecology, v. 50, pp. 271-284.

MAY, R., 1977: Facies model for sedimentation in the glaciolacustrine environment; Boreas, v. 6, pp. 175-180.

MERCIER, A., 1984: Glacial lake in the Richardson and Rae River basins, N.W.T., Géogr. phys. Quat., v. XXXVIII, n. 1, pp. 75-80.

REFERENCES - Continued

MERCIER, A., 1986: Historique du Quaternaire Supérieur du bassin de la rivière Richardson, T.N.-O., Thèse de maîtrise, Département de géographie, Université d'Ottawa, Ottawa, 242 p.

MIALL, A., 1984: Deltas; In Facies Models; R. Walker (Ed), Geoscience Canada, Reprint Series 1, 2nd edition, pp. 105-118.

MIDDLETON, G., HAMPTON, M., 1973: Sediment gravity flows: mechanics of flow and deposition; In Turbidites and Deep Water Sedimentation; Soc. Econ. Paleont. Mineral., Pacific Sec. Short Course Lecture Notes, pp. 1-38.

MODE, W., NELSON, A., and BRIGHAM, J., 1983: A facies model of Quaternary glacial-marine cyclic sedimentation along eastern Baffin Island, Canada; In Glacial-Marine Sedimentation, B. Molnia (Ed), Plenum Press, pp. 495-533.

MORGENSTERN, N., 1967: Submarine slumping and the initiation of turbidity currents; In Marine Geotechnique, A. Richards (Ed), University of Illinois Press, pp. 189-220.

NAGY, J., 1984: Quaternary glaciomarine deposits and foraminifera from Edgeøya, Svalbard; Boreas, v. 13, pp. 319-232.

REFERENCES - Continued

NEALE, J., 1963: Factors influencing distribution of Recent British Ostracoda; In Ostracoda as ecological and palaeo-ecological indicators, H. Puri (Ed), Publ. Staz. Zool. Napoli, suppl. n. 33.

________, 1969: The taxonomy, morphology and ecology of Recent Ostracoda; Edinburgh, Oliver and Boyd, pp. 251-267.

NEALE, J., and HOWE, H., 1975: The marine Ostracoda of Russian Harbour, Novaya Zemlya and other high latitude faunas; In Biology and Paleobiology of Ostracoda, F. Swain (Ed), Bull. Am. Paleont., v. 65, pp. 381-431.

OCKELMAN, W., 1958: The zoology of East Greenland: marine Lamellibranchiata; Meddelelser om Grønland, 122.

O'NEILL, J., 1924: Report of the Canadian Arctic Expedition, 1913-1918, v. XI, The Geology of the Arctic Coast of Canada; West of the Kent Peninsula, 107 p.

OVENSHINE, A., 1970: Observations on iceberg rafting in Glacier Bay, Alaska and the identification of ice-rafted deposits; Geol. Soc. Am. Bull., v. 81, pp. 891-894.

PETTIJOHN, F., and POTTER, P., 1964: Atlas and glossary of primary sedimentary structures; Berlin, Springer-Verlag.

REFERENCES - Continued

PICKRILL, R., and IRWIN, J., 1983: Sedimentation in a deep glacier-fed lake - Lake Tekapo, New Zealand; Sedimentology, v. 30, pp. 63-75.

PIPER, D., LETSON, J., DE IURE, A., and BARRIE, C., 1983: Sediment accumulations in low-sedimentation, wave dominated, glaciated inlets; Sed. Geol., v. 36, pp. 195-215.

POKORNÝ, V., 1978: Ostracoda; In Introduction to Marine Micropaleontology, B. Haq and A. Boersma (Eds), Elsevier, New York, pp. 109-150.

POWELL, R., 1981: A model for sedimentation by tidewater glaciers; Annals of Glaciology, v. 2, pp. 129-134.

QUIGLEY, R., 1980: Geology, mineralogy, and geochemistry of Canadian soft soils: a geotechnical perspective; Can. Geotech. Jour., v. 17, pp. 261-285.

QUIGLEY, R., and OGUNBADEJO, T., 1972: Clay layer fabric and oedometer consolidation of a soft varved clay; Can. Geotech. Jour., v. 9, pp. 165-175.

REINECK, H., and SINGH, I., 1980: Depositional Sedimentary Environments; Springer-Verlag, 2nd edition, 549 p.

REFERENCES - Continued

ROBINSON, E., 1978: The Pleistocene, In A stratigraphical index of British Ostracoda; R. Bate and E. Robinson (Eds), Seel House Press, Liverpool, p. 466.

RODRIGUES, C., and HOOPER, K., 1982a: Recent benthonic foraminiferal associations from offshore environments in the Gulf of St. Lawrence; Jour. Foram. Research, v. 12, pp. 327-352.

_________, 1982b: The ecological significance of Elphidium clavatum in the Gulf of St. Lawrence, Canada; Jour. Paleontology, v. 56, pp. 410-422.

RODRIGUES, C., and RICHARD, S., 1983: Late glacial and postglacial macrofossils from the Ottawa-St. Lawrence Lowlands, Ontario and Quebec; In Current Research, Part A, Geol. Surv. Can., Paper 83-1A, pp. 371-379.

_________, 1985: Temporal distribution and significance of Late Pleistocene fossils in the western Champlain Sea basin, Ontario and Quebec; In Current Research, Part B, Geol. Surv. Can., Paper 85-1B, pp. 401-414.

RUST, B., 1977: Mass flow deposits in a Quaternary succession near Ottawa, Canada; diagnostic criteria for subaqueous outwash; Can. Jour. Earth Sci., v. 14, pp. 175-184.

REFERENCES - Continued

RUST, B., 1978: Depositional models for braided alluvium; In Fluvial Sedimentology, Can. Soc. Petrol. Geol., Mem. 5, A. Miall (Ed), pp. 605-625.

RUST, B., and ROMANELLI, R., 1975: Late Quaternary subaqueous outwash deposits near Ottawa, Canada; In Glaciofluvial and Glaciolacustrine Sedimentation, A. Jopling and B. McDonald (Eds), Soc. Econ. Paleon. Mineral., Spec. Publ. n. 23, pp. 177-192.

SAUNDERSON, H., 1975: Sedimentology of the Brampton Esker and its associated deposits: an empirical test of theory; In Glaciofluvial and Glaciolacustrine Sedimentation, A. Jopling and B. McDonald (Eds), Soc. Econ. Paleon. Mineral., Spec. Publ. n. 23, pp. 155-176.

SELLEY, R., 1970: Ancient Sedimentary Environments; Chapman and Hall Ltd., London, 237 p.

SHAW, J., 1975: Sedimentary successions in Pleistocene ice-marginal lakes; In Glaciofluvial and Glaciolacustrine Sedimentation, A. Jopling and B. McDonald (Eds), Soc. Econ. Paleon. Mineral., Spec. Publ. n. 23, pp. 281-303.

REFERENCES - Continued

SHAW, J., 1977: Sedimentation in an alpine lake during deglaciation, Okanagan Valley, British Columbia, Canada; Geografiska Annaler, 59A, pp. 221-240.

________, 1982: Melt-out till in the Edmonton area, Alberta, Canada; Can. Jour. Earth Sci., v. 19, pp. 1,548-1,569.

SHEPARD, F., 1954: Nomenclature based upon sand-silt-clay ratios; Jour. Sed. Petrol., v. 24, pp. 151-158.

SHILTS, W., 1978: Detailed sedimentological study of till sheets in a stratigraphic section, Samson River, Quebec; Geol. Surv. Can., Bull. 285, 30 p.

________, 1981: Surficial geology of the Lac Mégantic area, Québec; Geol. Surv. Can., Memoir 397, 102 p.

SHOLKOVITZ, E., 1976: Flocculation of dissolved organic and inorganic matter during the mixing of river water and seawater; Geochim. Cosmochim. Acta, 40, pp. 831-845.

SMITH, A., 1959: Structures in the stratified late glacial clays of Windermere, England; Jour. Sed. Petrol., v. 29, pp. 447-453.

REFERENCES - Continued

SMITH, N., and ASHLEY, G., 1985: Proglacial lacustrine environments; In Glacial Sedimentary Environments, G. Ashley, J. Shaw, N. Smith (Eds), Soc. Econ. Paleon. Mineral., ~~Short Course~~ No. 16, pp. 135-216.

SMITH, N., and SYVITSKI, J., 1982: Sedimentation in a glacier-fed lake: the role of pelletization on deposition of fine-grained suspensates; Jour. Sed. Petrol., v. 52, pp. 503-513.

SOUTHARD, J., 1971: Representation of bed configurations in depth-velocity-size diagrams; Jour. Sed. Petrol., v. 41, pp. 903-915.

SOWERBY, G., 1859: Illustrated index of British shells; p. vi.

STANLEY, S., 1970: Relation of shell form to life habits of the Biva~~lvia (Mollusca)~~; Geol. Soc. of America, Memoir 125, 296 p.

STAPLIN, F., 1963: Pleistocene Ostracoda of Illinois; Jour. of Paleont., v. 37, pp. 1,164-1,203.

REFERENCES - Continued

STEVENS, R., 1985: Glaciomarine varves in late Pleistocene clays near Göteborg, southwestern Sweden; Boreas, v. 14, pp. 127-132.

ST-ONGE, D., 1980: Glacial Lake Coppermine, north-central District of Mackenzie, Northwest Territories; Can. Jour. Earth Sci., v. 17, pp. 1,310-1,315.

__________, (in press): Quaternary Geology and Geomorphology of the Lower Coppermine River Valley, District of Mackenzie N.W.T.; Field Excursion A22, INQUA 1987, 42 p.

ST-ONGE, D., and BRUNEAU, H., 1982: Dépôts meubles du secteur aval de la rivière Coppermine, T.N.-O.; In Recherches en cours, partie B, Com. géo. Canada, Etude 82-1B, pp. 51-55.

ST-ONGE, D., and GEURTS, M.A., 1982: Les formes d'effondrement et le mode de déglaciation de la région du lac Quingaluk, T.N.-O., Canada; Géogr. phy. Quat., v. XXXVI, n. 1-2, pp. 233-240.

SWAIN, F., 1963: Pleistocene Ostracoda from the Gubik Formation, Arctic Coastal Plain, Alaska; Jour. Paleont., v. 37, pp. 798-834.

REFERENCES - Continued

SYVITSKI, J., 1978: Sedimentological advances concerning the flocculation and zooplankton pelletization of suspended sediments in Howe Sound, British Columbia: a fjord receiving glacial meltwater; Ph. D. Thesis, University of British Columbia, Vancouver, B.C., 291 p.

THOMAS, G., and CONNELL, R., 1985: Iceberg drop, dump, and grounding structures from Pleistocene glacio-lacustrine sediments, Scotland; Jour. Sed. Petrol., v. 55, pp. 243-249.

TODD, R., and LOW, D., 1966: Foraminifera from the Arctic Ocean off the eastern Siberian Coast; U.S. Geol. Survey, Prof. Paper n. 550-C, pp. C79-C85.

________ , 1967: Recent Foraminifera from the Gulf of Alaska and southeastern Alaska; U.S. Geol. Surv., Prof. Paper n. 573-A, 46 p.

TORRANCE, K., 1983: Towards a general model of quick clay development; Sedimentology, v. 30, pp. 547-555.

VILKS, G., 1969: Recent Foraminifera in the Canadian Arctic; Micropaleontology, v. 15, pp. 35-60.

REFERENCES - Concluded

WAGNER, F., 1968: Faunal study, Hudson Bay and Tyrrell Sea; In Earth Science Symposium on Hudson Bay, Geol. Surv. Can., Paper 68-53, pp. 7-29.

WALCOTT, R., 1972: Late Quaternary vertical movements in eastern North America: quantitative evidence of glacio-isostatic rebound; Reviews of Geophysics and Space Physics, v. 10, pp. 849-884.

WALTON, W., 1964: Recent foraminiferal ecology and Paleoecology; In Approaches to Paleoecology, J. Imbrie and N. Newell (Eds), John Wiley and Sons, pp. 151-232.

WHATLEY, R., and MASSON, D., 1979: The ostracode genus Cytheropteron from the Quaternary and Recent of Great Britain; Rev. Espanola Micropal, v. 11, pp. 223-277.

WHITEHOUSE, V., JEFFREY, L., and DEBRECHT, J., 1960: Differential settling tendencies of clay minerals in saline waters; In Clays and Clay Minerals, Proc. 7th Nat. Conf. on Clays and Clay Minerals, Nat. Acad. Sci., Nat. Res. Coun., pp. 1-76.

APPENDIX A

Grain Size Parameters

(after Folk and Ward, 1957)

Sedimentary Facies	Mean (Ø)	Standard Deviation (Ø)	Skewness
A	3.60	2.70	-0.29
B	5.08	2.10	0.18
C	1.13	0.74	0.21
D	1.53	0.56	0.01
E	3.79	0.81	0.38
F	2.02	0.49	0.07
G	5.35	0.90	0.33
Ha	9.30*	0.72*	- 0.49*
Hb	8.20*	1.02*	0.12*
I	6.74	1.75*	0.33*

* Extrapolated from graph.

Résumé of grain size analysis procedure

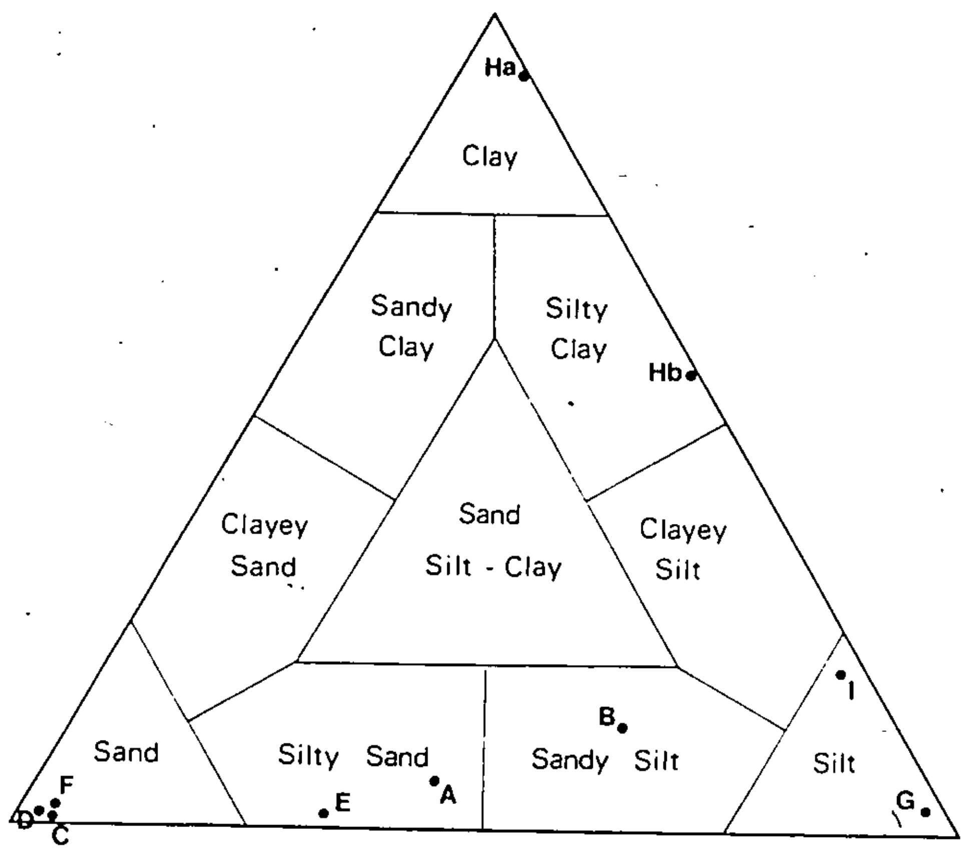
For sand-sized particles, the sample is air-dried and split to approximately 50 g. Sieves are stacked in order, from - 2 Ø (at top) to 4 Ø, using ½ Ø intervals. Samples are added, and sieves are shaken on Ro-top for 15 minutes. Afterwards, sieves are inverted to remove residual grains and each fraction is weighed.

APPENDIX A

Silt and clay size distribution is determined by a technique involving particle settling (pipette analysis). This method consists of preparing a suspension of mud of low concentration. Approximately 15 g of sediment and dispersant are added to a 1 litre graduated cylinder. Cylinder is stirred vigorously. 20 seconds after the stir rod emerges, the pipette is inserted to a depth of 20 cm and exactly 25 ml of the suspension is withdrawn. These withdrawals are continued at specified time intervals. Suspensions are transferred to 50 ml weighed beakers and permitted to evaporate. Residues are weighed to 0.0001 g when dry. At each of these times, all particles of a determined diameter (Stokes' Law) will have settled below the chosen level. From the weight of sediment recovered in samples of known volume, the grain size distribution can be calculated.

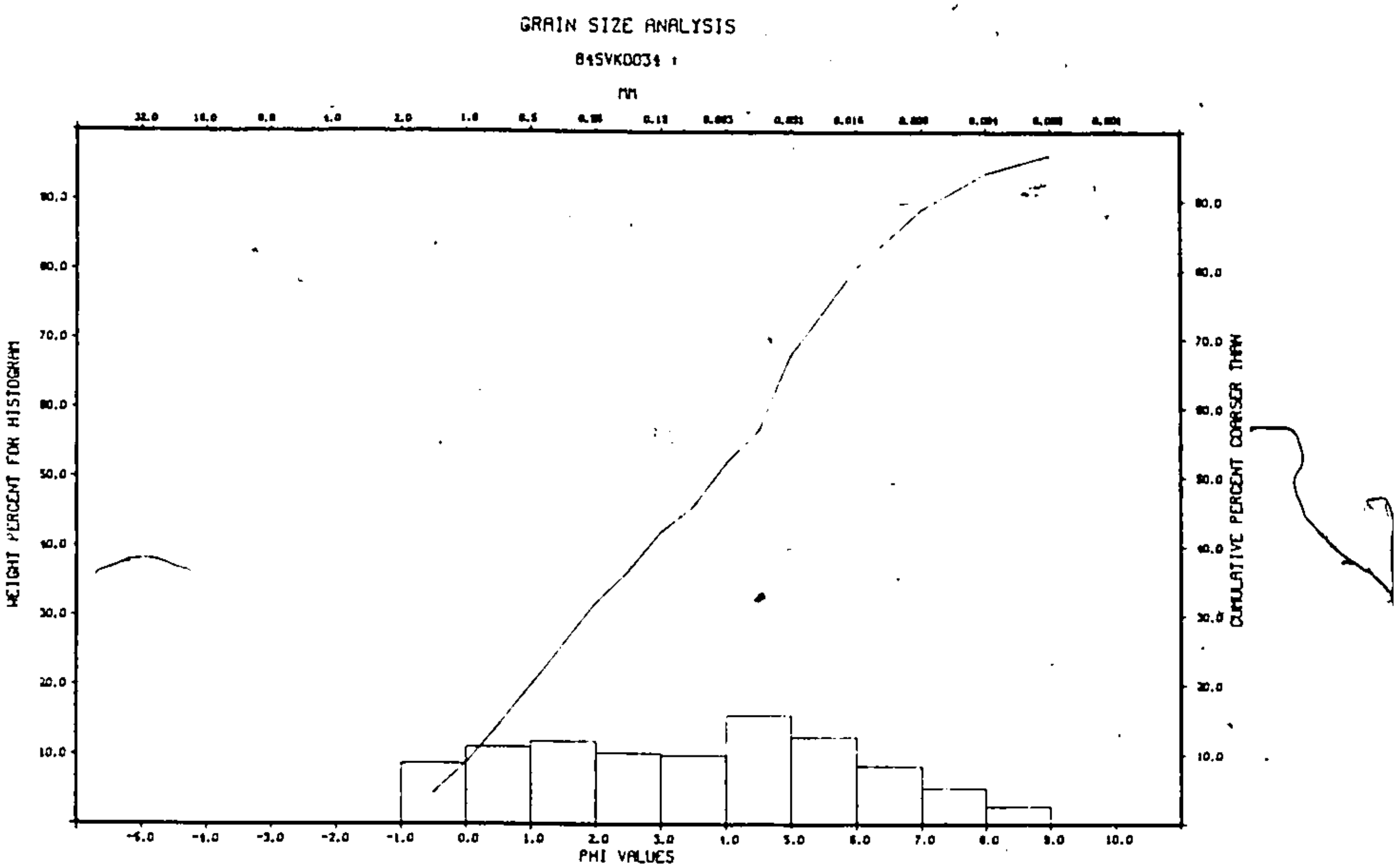
APPENDIX B

Nomenclature classification of sedimentary facies



APPENDIX C-1

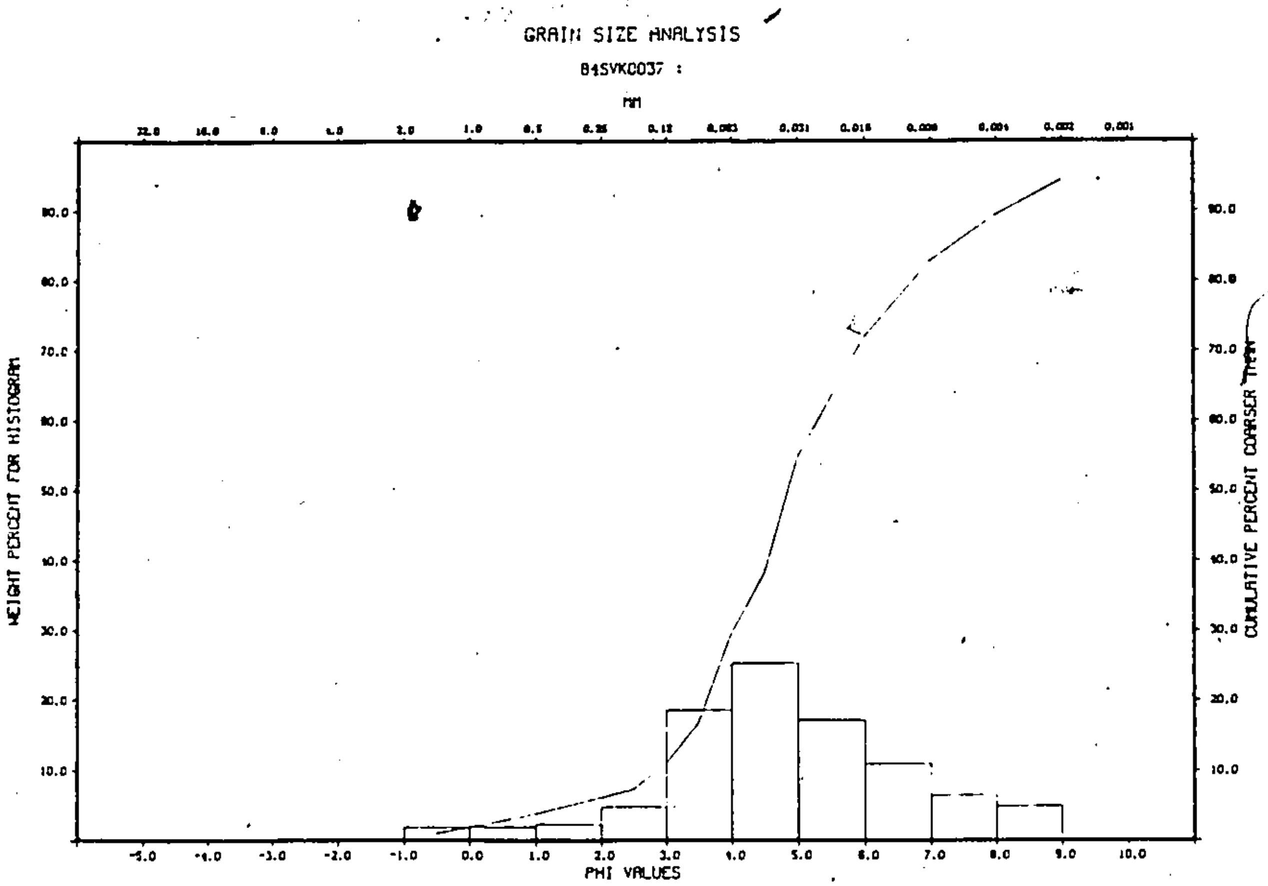
Facies A



Class(Ø)	Weight(g)	Weight Percent	Cumulative Percent
- 1.0-0.0	4.40	8.78	8.78
0.0-1.0	5.58	11.13	19.92
1.0-2.0	5.98	11.94	31.86
2.0-3.0	5.15	10.27	42.14
3.0-4.0	4.98	9.93	52.08
4.0-5.0	3.08	15.72	67.80
5.0-6.0	0.54	12.59	80.39
6.0-7.0	0.38	8.44	88.84
7.0-8.0	0.27	5.23	94.08
8.0-9.0	0.21	2.58	96.65
>9.0	0.17	3.34	100.00

APPENDIX C-2

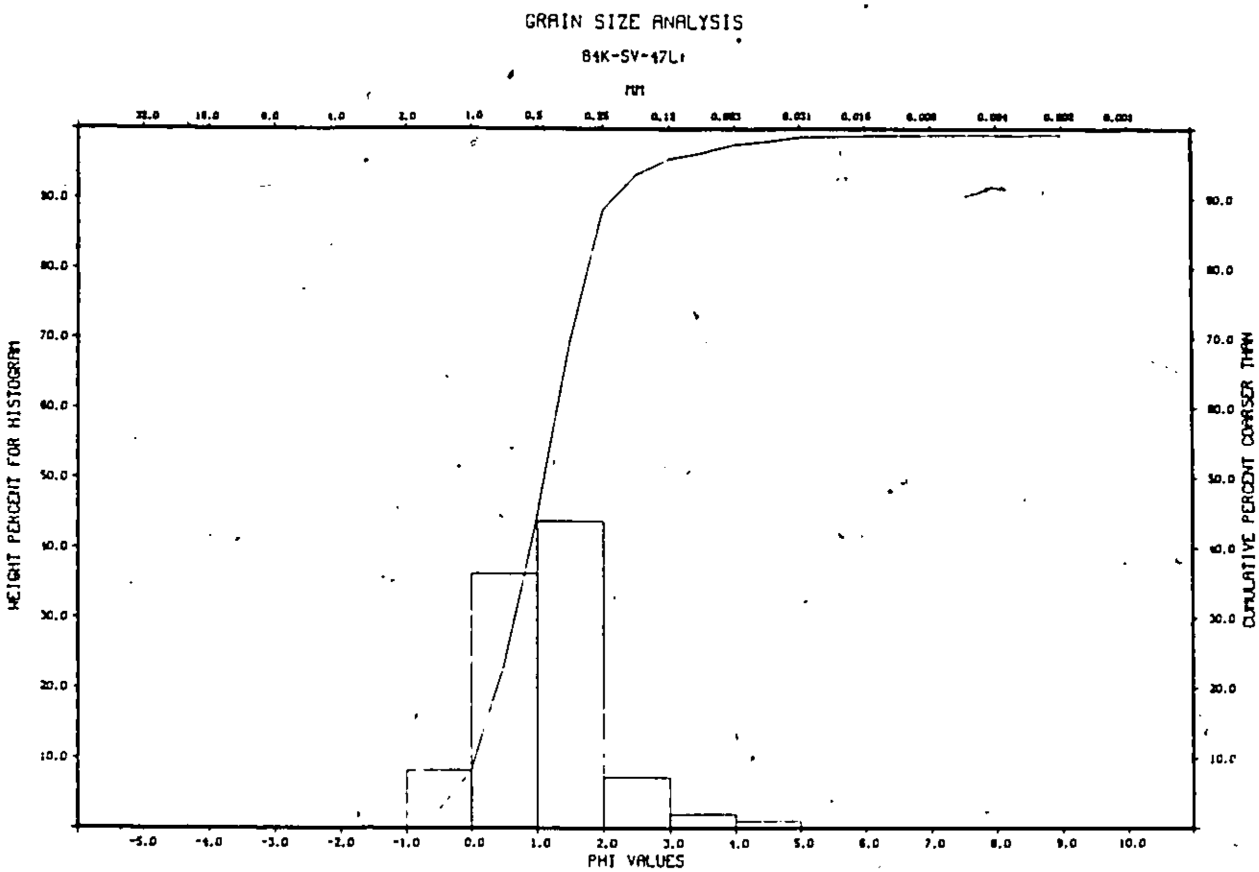
Facies B



Class (Ø)	Weight (g)	Weight Percent	Cumulative Percent
- 1.0-0.0	0.98	1.89	1.89
0.0-1.0	1.01	1.94	3.85
1.0-2.0	1.20	2.32	6.18
2.0-3.0	2.50	4.83	11.01
3.0-4.0	9.50	18.58	29.60
4.0-5.0	5.40	25.29	54.89
5.0-6.0	0.72	17.19	72.09
6.0-7.0	0.49	10.94	83.03
7.0-8.0	0.35	6.47	89.50
8.0-9.0	0.27	4.89	94.40
>9.0	0.21	5.59	100.00

APPENDIX C-3

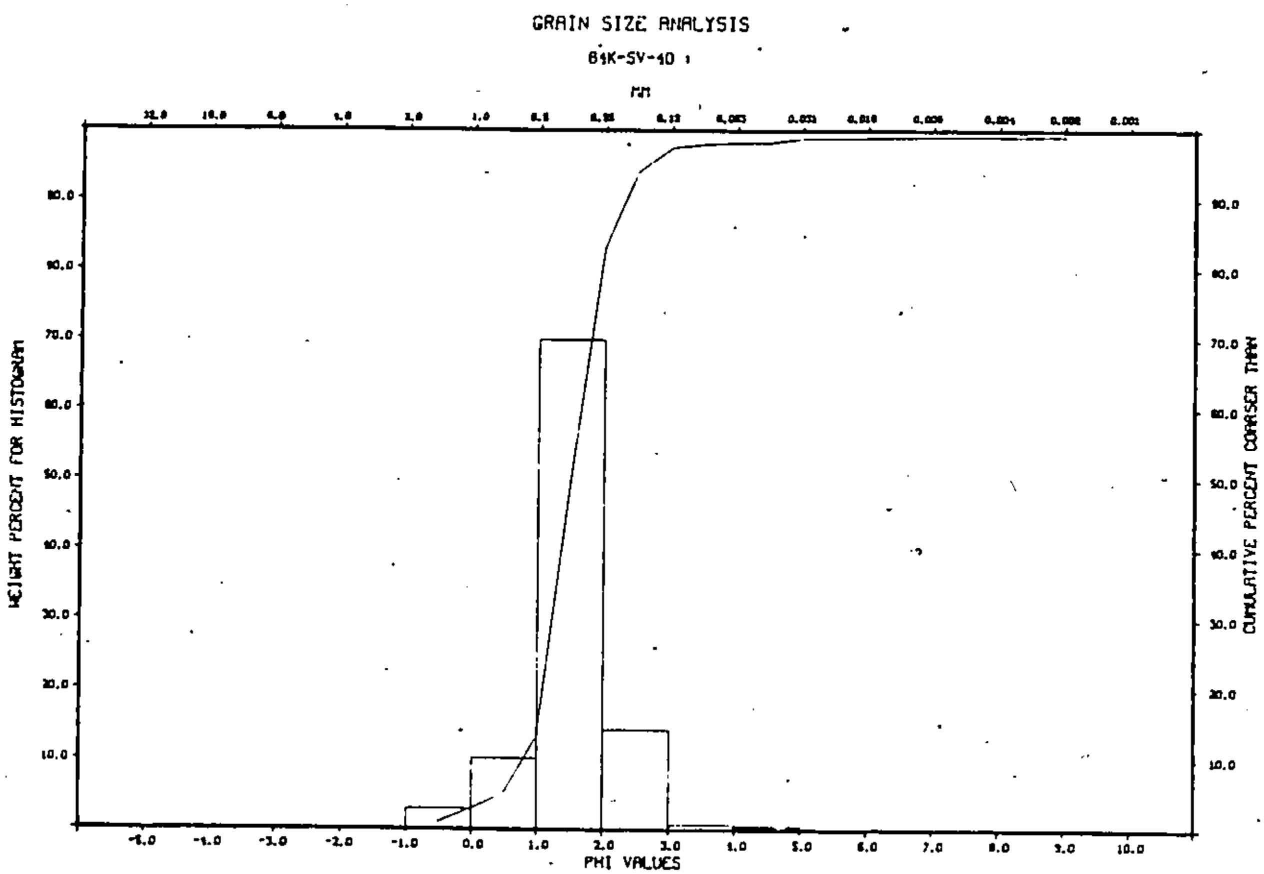
Facies C



Class (Ø)	Weight (g)	Weight Percent	Cumulative Percent
- 1.0-0.0	4.05	8.26	8.26
0.0-1.0	17.81	36.33	44.59
1.0-2.0	21.49	43.84	88.44
2.0-3.0	3.57	7.27	95.72
3.0-4.0	0.99	2.01	97.74
4.0-5.0	0.38	1.01	98.76
5.0-6.0	0.15	0.26	99.02
6.0-7.0	0.15	0.09	99.12
7.0-8.0	0.14	0.01	99.14
8.0-9.0	0.14	0.02	99.16
>9.0	0.14	0.83	100.00

APPENDIX C-4

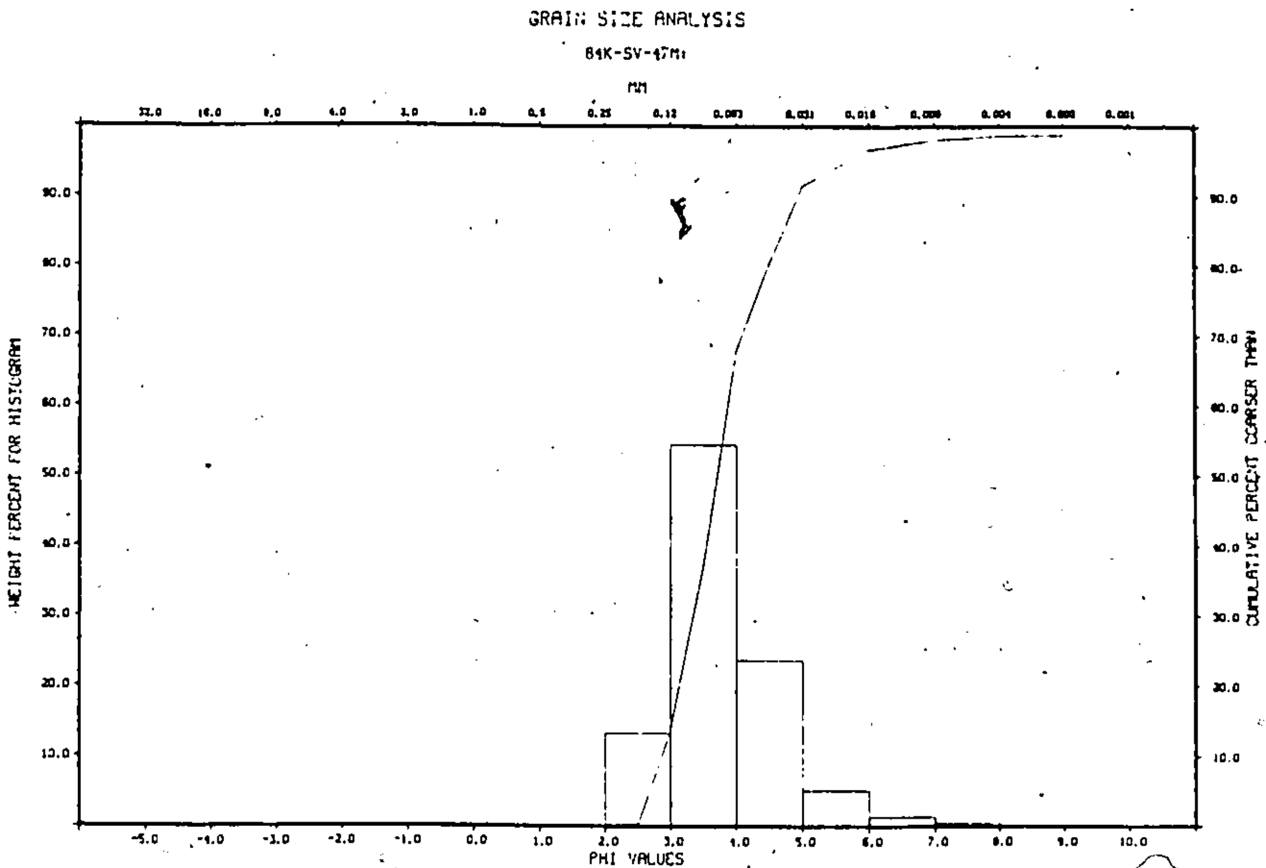
Facies D



Class (Ø)	Weight(g)	Weight Percent	Cumulative Percent
- 1.0-0.0	1.48	2.97	2.97
0.0-1.0	5.09	10.24	13.22
1.0-2.0	34.78	70.00	83.23
2.0-3.0	7.08	14.24	97.48
3.0-4.0	0.37	0.74	98.22
4.0-5.0	0.20	0.63	98.86
5.0-6.0	0.15	0.04	98.90
6.0-7.0	0.15	0.23	99.13
7.0-8.0	0.14	0.02	99.24
>9.0	0.14	0.75	100.00

APPENDIX C-5

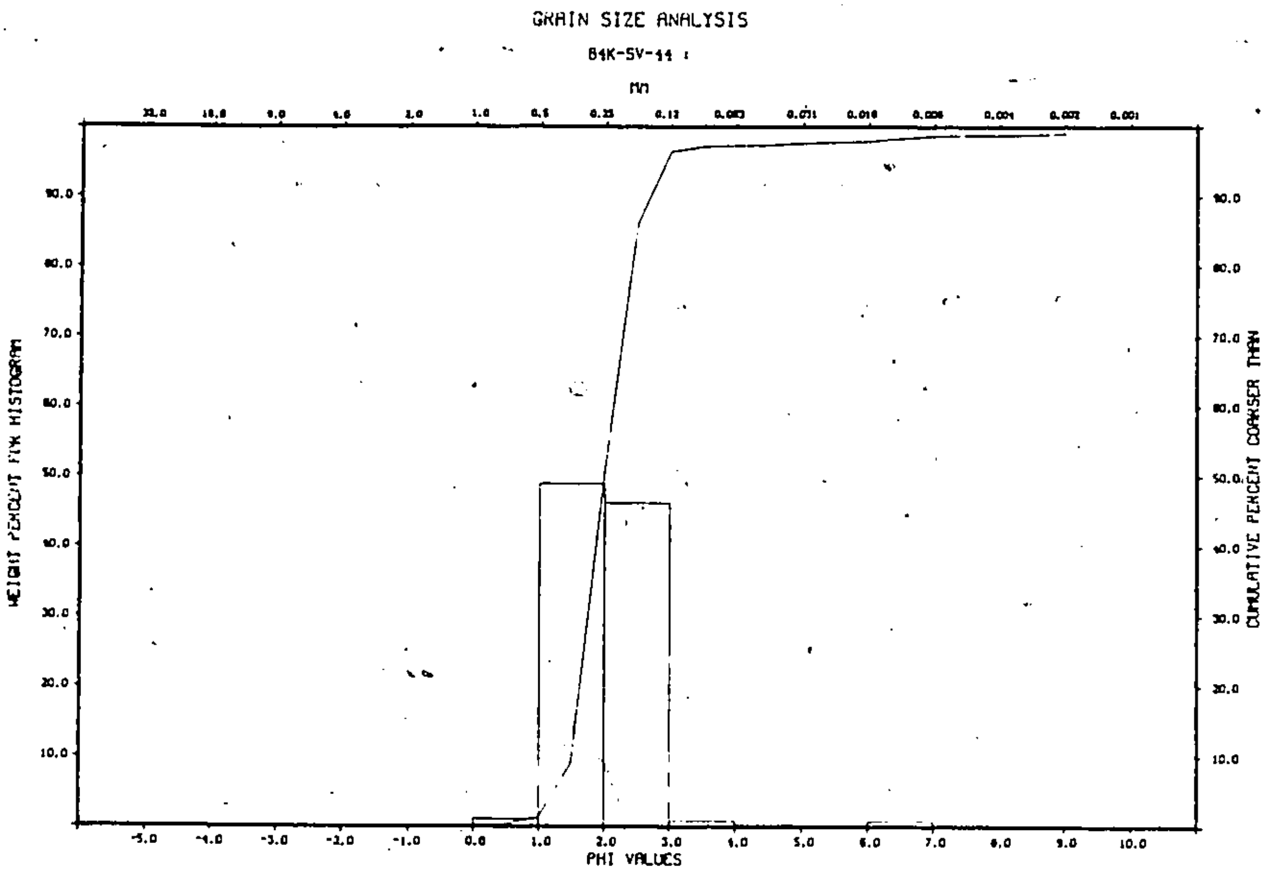
Facies E



Class (Ø)	Weight(g)	Weight Percent	Cumulative Percent
-1.0-0.0	0.01	0.02	0.02
0.0-1.0	0.02	0.04	0.06
1.0-2.0	0.02	0.04	0.11
2.0-3.0	5.97	13.29	13.84
3.0-4.0	24.45	54.46	67.87
4.0-5.0	6.03	23.67	91.54
5.0-6.0	0.23	5.14	96.68
6.0-7.0	0.17	1.38	98.06
7.0-8.0	0.16	0.57	98.64
8.0-9.0	0.15	0.10	98.75
>9.0	0.15	1.24	100.00

APPENDIX C-6

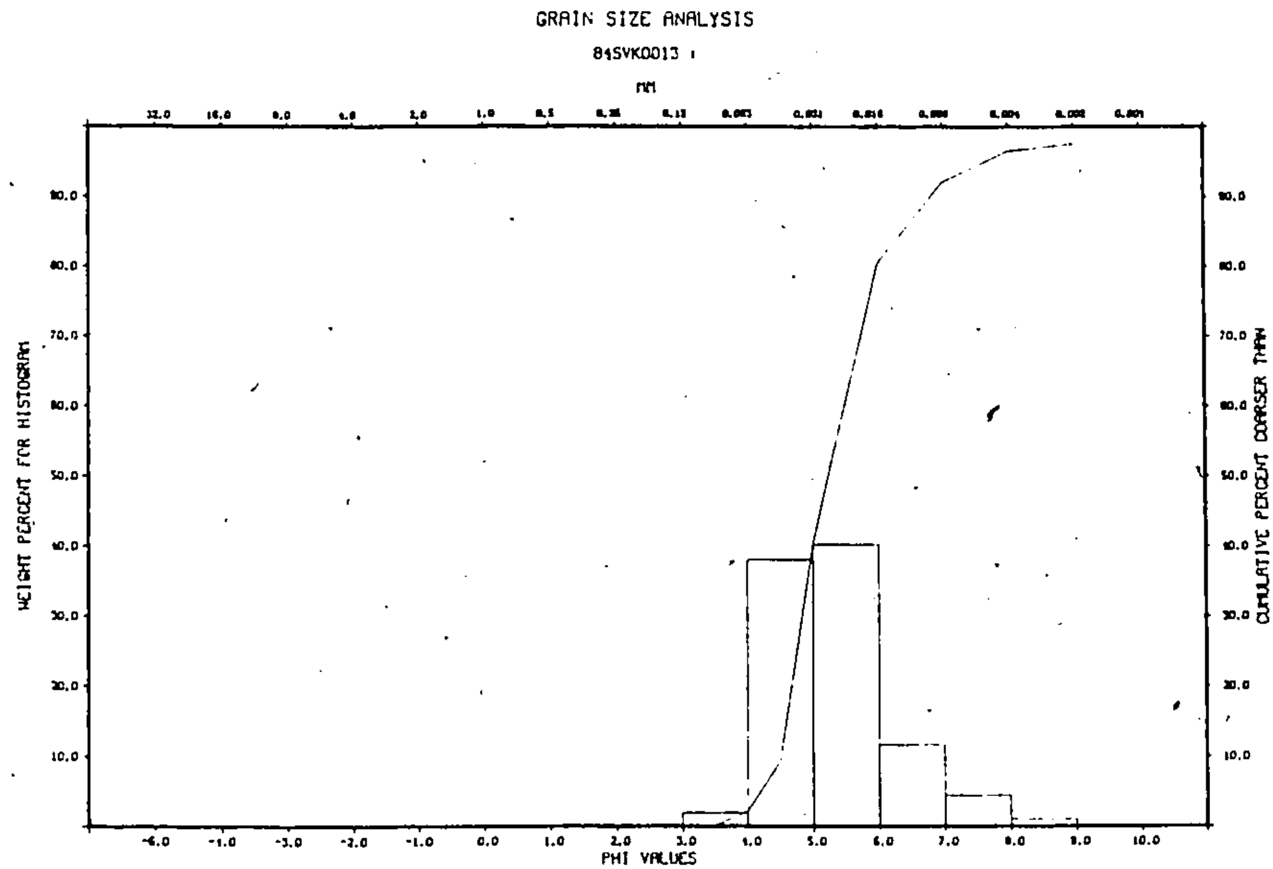
Facies F



Class (Ø)	Weight (g)	Weight Percent	Cumulative Percent
- 1.0-0.0	0.09	0.19	0.19
0.0-1.0	0.53	1.13	1.76
1.0-2.0	22.76	48.92	50.27
2.0-3.0	21.49	46.19	96.47
3.0-4.0	0.39	0.83	97.31
4.0-5.0	0.18	0.30	97.62
5.0-6.0	0.16	0.28	97.91
6.0-7.0	0.16	0.78	98.69
7.0-8.0	0.15	0.08	98.77
8.0-9.0	0.15	0.24	99.01
>9.0	0.15	0.98	100.00

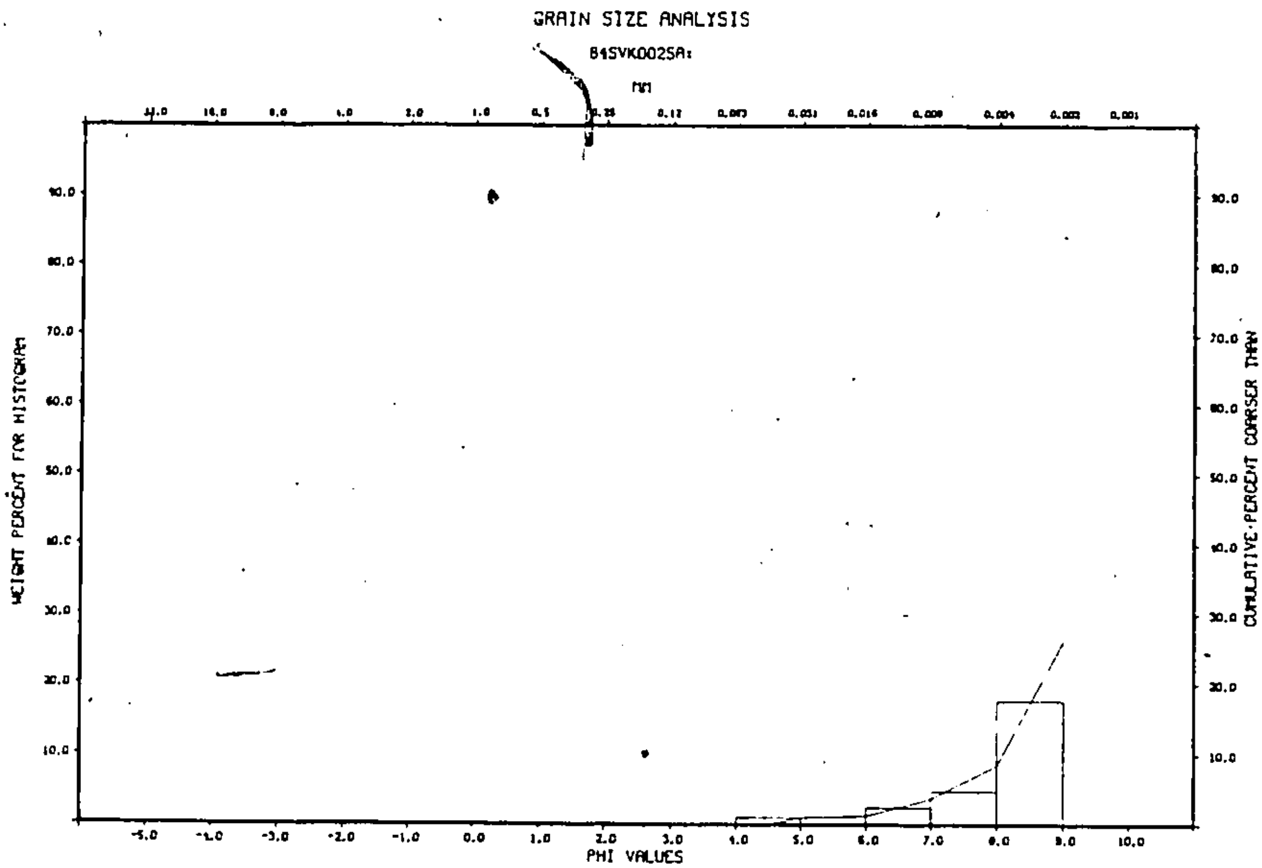
APPENDIX C-7

Facies G



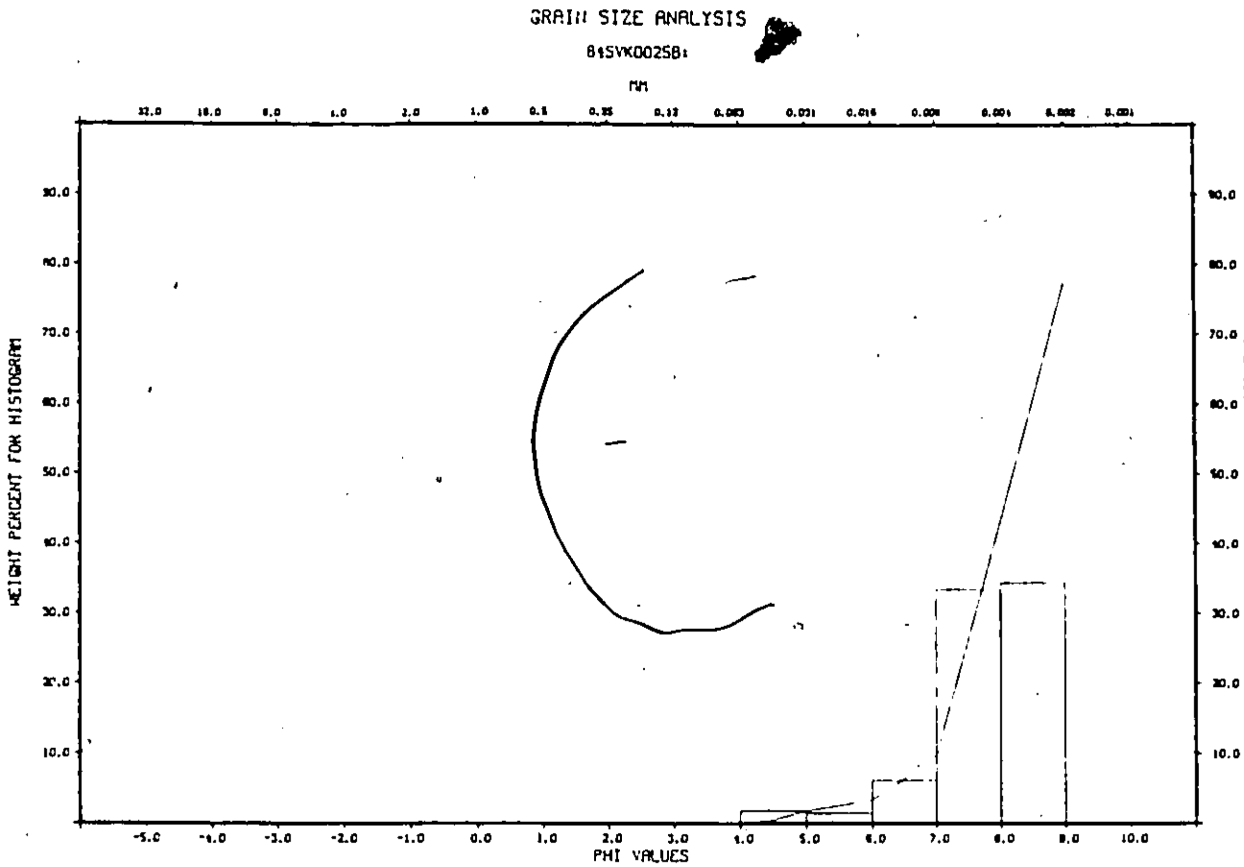
Class (Ø)	Weight(g)	Weight Percent	Cumulative Percent
- 1.0-0.0	0.00	0.00	0.00
0.0-1.0	0.01	0.02	0.02
1.0-2.0	0.02	0.04	0.07
2.0-3.0	0.03	0.06	0.14
3.0-4.0	0.83	1.95	2.09
4.0-5.0	5.18	38.07	40.18
5.0-6.0	0.77	40.23	80.42
6.0-7.0	0.34	11.62	92.04
7.0-8.0	0.22	4.43	96.48
8.0-9.0	0.17	1.05	97.53
>9.0	0.16	2.46	100.00

Facies Ha



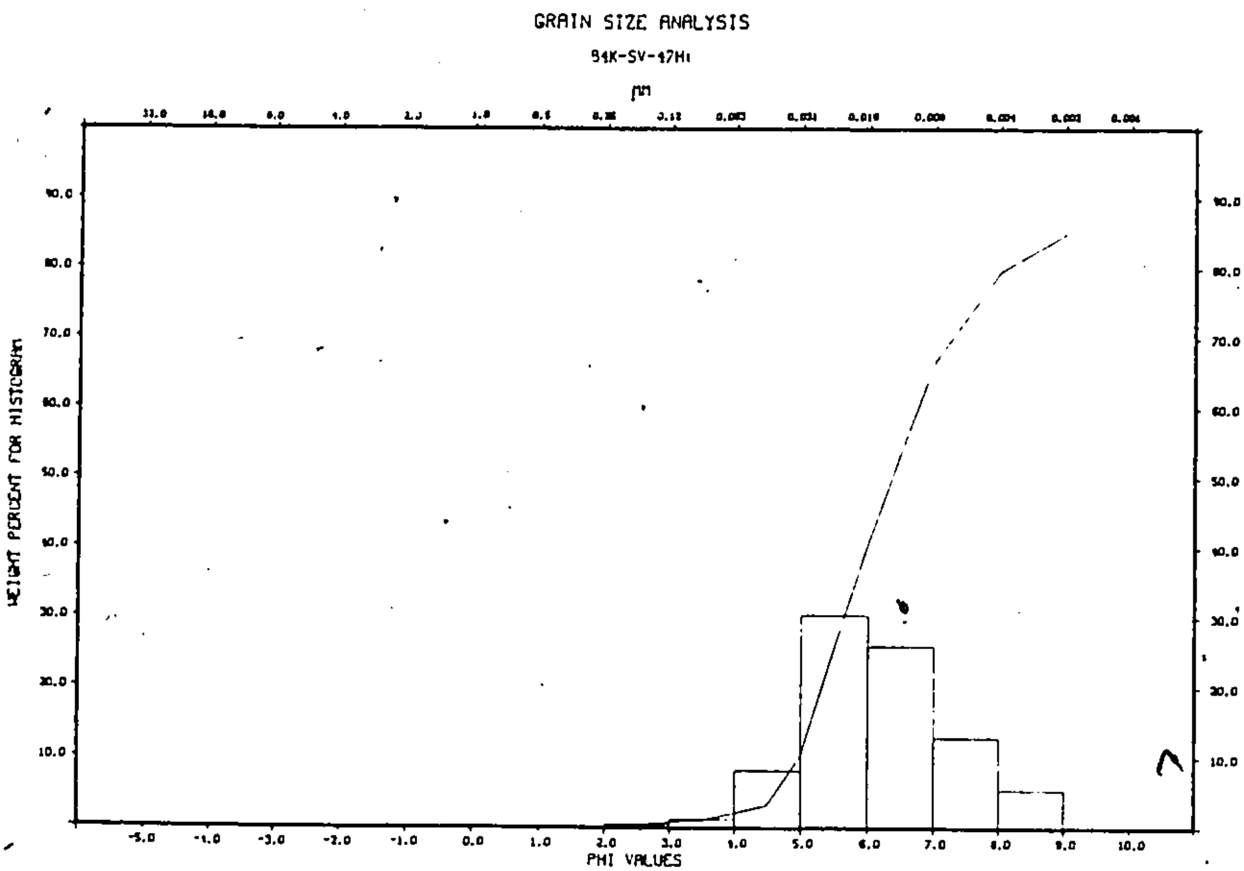
Class (Ø)	Weight(g)	Weight Percent	Cumulative Percent
- 1.0-0.0	0.00	0.00	0.00
0.0-1.0	0.00	0.00	0.00
1.0-2.0	0.00	0.00	0.00
2.0-3.0	0.00	0.00	0.00
3.0-4.0	0.00	0.00	0.00
4.0-5.0	1.31	1.04	1.04
5.0-6.0	1.25	0.25	1.30
6.0-7.0	1.25	2.45	3.76
7.0-8.0	1.22	4.78	8.55
8.0-9.0	1.16	17.68	26.23
>9.0	0.97	73.76	100.00

Facies Hb



Class (Ø)	Weight(g)	Weight Percent	Cumulative Percent
- 1.0-0.0	0.00	0.00	0.00
0.0-1.0	0.00	0.00	0.00
1.0-2.0	0.00	0.00	0.00
2.0-3.0	0.00	0.00	0.00
3.0-4.0	0.00	0.00	0.00
4.0-5.0	1.27	1.88	1.88
5.0-6.0	1.05	1.53	3.41
6.0-7.0	1.03	6.21	9.63
7.0-8.0	0.97	33.45	43.08
8.0-9.0	0.66	34.39	77.47
>9.0	0.34	22.52	100.00

Facies I

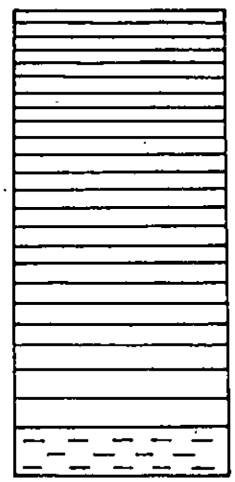


Class (Ø)	Weight(g)	Weight Percent	Cumulative Percent
- 1.0-0.0	0.02	0.03	0.03
0.0-1.0	0.02	0.03	0.07
1.0-2.0	0.10	0.18	0.26
2.0-3.0	0.31	0.59	0.86
3.0-4.0	0.67	1.27	2.14
4.0-5.0	1.99	8.25	13.67
5.0-6.0	1.31	30.41	40.81
6.0-7.0	0.91	25.96	66.78
7.0-8.0	0.57	12.92	79.70
8.0-9.0	0.40	5.44	85.14
>9.0	0.33	14.85	100.00

APPENDIX D-1

Section 1

Elevation of top: 205 m a.s.l.



Facies

H Silty clay rhythmites, 5 m thick; couplets range in thickness from 20 cm at base and decrease to 3 cm at top.

G Very fine sandy silt, 1 m thick; massive to poorly horizontally laminated; grades sharply into Facies H.

LEGEND FOR APPENDIX D

Lithofacies

I

H

G

F

E

D

C

B

A

Symbols

climbing ripples

▲ dropstones

soft sediment deformation

● fossils

peat

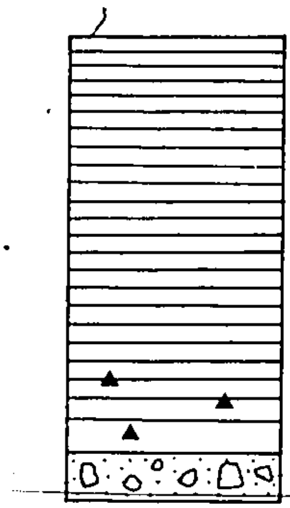
F3 fossil sample location

palaeocurrent direction

APPENDIX D-2

Section 2

Elevation of top: 180 m a.s.l.



Facies

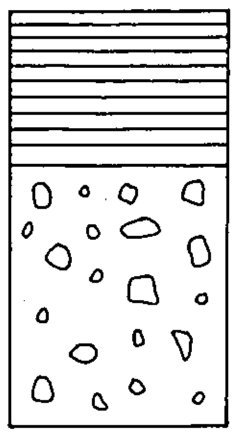
H Silty clay rhythmites, 5 m thick; couplets range in thickness from 10 cm at base to 1 cm at top (Plate 7); rare dropstones, up to 20 cm diameter occur at base.

A Massive, poorly sorted diamicton, 0.5 m in thickness; angular to well rounded clasts, up to 0.75 m diameter; sharply overlain by Facies H.

APPENDIX D-3

Section 3

Elevation of top: 152 m a.s.l.



Facies

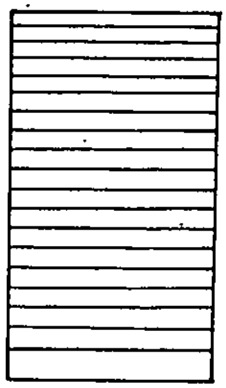
H Silty clay rhythmites, 4 m thick; couplets vary in thickness from 10 cm at base to 2 cm at top.

B Massive poorly sorted diamicton 7 m in thickness; clasts variable in composition, angular to rounded, ranging from 2 cm to 50 cm in diameter; sharply overlain by Facies H.

APPENDIX D-4

Section 4

Elevation of top: 142 m a.s.l.



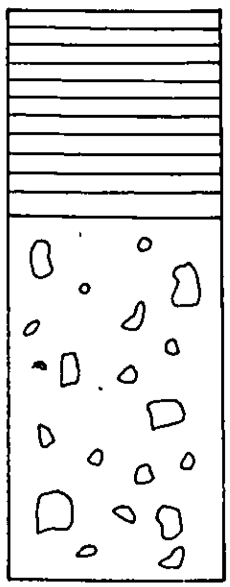
Facies

II Silty clay rhythmites, 7 m in thickness; couplets range from 10 cm thick at base, to 1 cm at top; minor gravel and cobble lag on surface.

APPENDIX D-5

Section 5

Elevation of top: 122 m a.s.l.



Facies

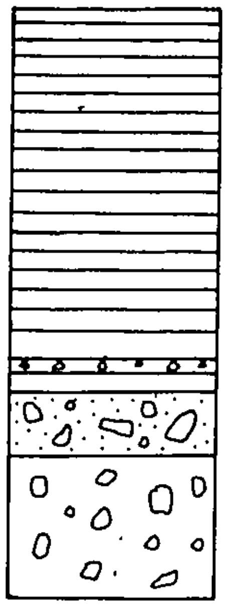
H Silty clay rhythmites, 6 m thick; couplets range in thickness from 10 cm at base and decrease to 2.5 cm at top.

B Poorly sorted, massive silty diamicton, 9 m in thickness; numerous erratics of variable size and composition, but predominance of well rounded green sandstone cobbles; gradational contact over 30 cm with Facies H (Plate 13).

APPENDIX D-6

Section 6

Elevation of top: 100 m a.s.l.



Facies

H Silty clay rhythmites, 10 m in thickness; thickness of couplets ranges from 10 cm at base to 2 cm at top.

B Poorly sorted silty diamicton, 10 cm thick; sharp upper and lower contacts with Facies H.

H Silty clay rhythmites, 50 cm thick, forming draped layers over irregular upper surface of Facies A.

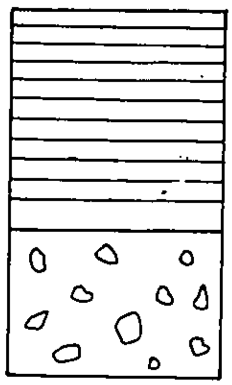
A Massive poorly sorted sandy diamicton with abundant gravel and cobble, 2 m in thickness; sharp lower boundary with Facies B.

B Poorly sorted silty diamicton with high concentration of angular gray sandstone and dolomite, clasts up to 2 m diameter, 4 m thick.

APPENDIX D-7

Section 7

Elevation of top: 80 m a.s.l.



Facies

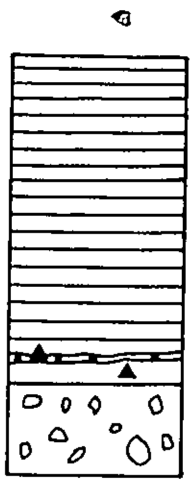
H Silty clay rhythmites, 6 m thick; couplets range in thickness from 10 cm at base, to 2.5 cm at top of section.

B Poorly sorted, massive silty diamicton, 4 m thick; sharply overlain by thinly bedded alternating layers of Facies H and B, 2 to 6 cm thick, over a distance of 30 cm; diamicton is rich in gravel-sized angular clasts of green sandstone (Plate 14).

APPENDIX D-8

Section 8

Elevation of top: 120 m a.s.l.



Facies

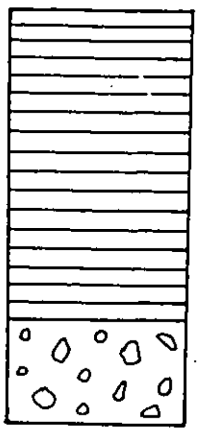
H Silty clay rhythmites, 5.5 m thick; couplets range in thickness from 2.5 cm at base to 1 cm at top; lower 25 cm is interbedded with thin diamicton layers; occasional dropstones present.

B Massive, poorly sorted silty diamicton, 2 m in thickness; clasts up to 0.75 m diameter, consisting of angular bedrock slabs (green sandstone) to well rounded cobbles of various lithologies; sharply overlain by Facies H.

APPENDIX D-9

Section 9

Elevation of top: 100 m a.s.l.



Facies

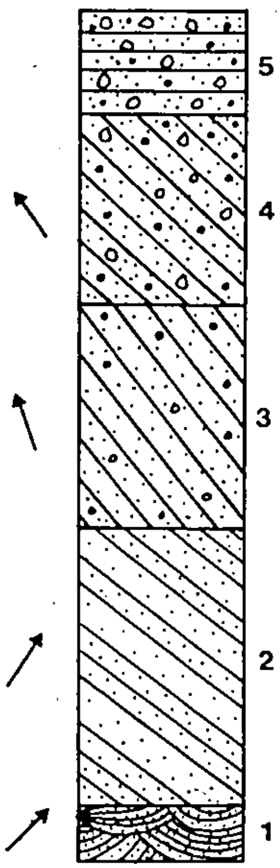
H Silty clay rhythmites, 5 m in thickness; couplets vary in thickness from 3 cm at base to 1 cm at top of section.

B Poorly sorted massive silty diamicton, 1.5 m thick; angular to rounded clasts up to 0.75 m diameter, various lithologies; sharply overlain by Facies H.

APPENDIX D-10

Section 10

Elevation of top: 260 m a.s.l.



Units

5. Planar bedded medium to coarse sand, with granules, gravel and rare cobbles, 3 m thick.

4. Medium to coarse cross-bedded sand with granules and minor gravel, 5 m thick; bed thickness ranges from 30 cm to 1 m; beds dip 25° towards 325°; abrupt erosional contact with Unit 5.

3. Medium to coarse sand with minor granules, 6 m in thickness; tabular and rare trough cross-bedding are present; beds are 30 to 90 cm thick, and dip 30° towards 340°; sharp upper contact with Unit 4.

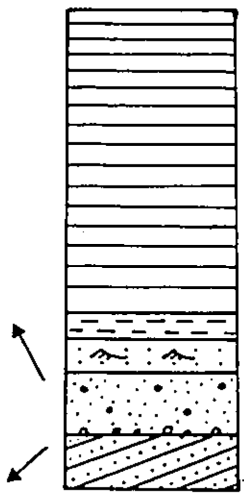
2. Tabular cross-bedded medium to coarse sand, 7.3 m thick; beds range in thickness from 10 to 35 cm, and dip 23° towards 035°; erosional sharp contact with Unit 3.

1. Medium to coarse sand, 1.5 m in thickness; trough cross-stratification towards 045°; sharp erosional contact with Unit 2.

APPENDIX D-11

Section 11

Elevation of top: 80 m a.s.l.



Facies

H Silty clay rhythmites, 8 m thick; couplets range in thickness from 30 cm at base, and decrease to 4 cm at top; gravel and cobble lag on surface.

G Finely laminated silt, 75 cm thick; upper and lower gradational contacts.

E Silty sand, 90 cm thick; cosets of climbing ripples (type B) overlain and underlain by horizontal laminae, 4 cm to 5 cm thick.

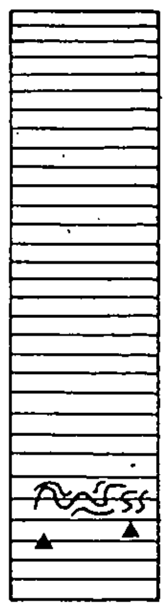
D Medium to coarse structureless sand, 1.6 m thick; rare granules present, sharp erosional lower boundary with Facies C.

C Coarse sand with rare gravel, 1.5 m thick; cross-bedded sand dipping 21° to 34° towards 230°; beds range in thickness from 20 to 50 cm; gravel lag concentrated at upper contact with Facies D.

APPENDIX D-12

Section 12

Elevation of top: 80 m a.s.l.



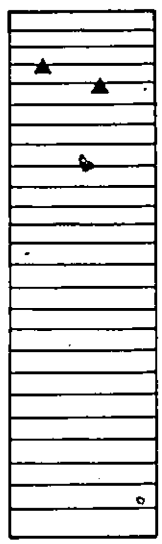
Facies

H Silty clay rhythmites, 24 m thick; couplets range in thickness from 30 cm at base, to 1 cm at top of section; intraformational contorted bedding restricted to 1 m thick band; rare dropstones occur near base of unit, rounded clasts up to 20 cm diameter.

APPENDIX D-13

Section 13

Elevation of top: 61 m a.s.l.



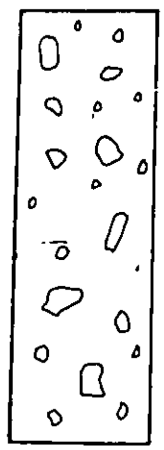
Facies

H Silty clay rhythmites, 24 m in thickness; couplets vary from 25 cm thick at base, to 4 cm at top of section; rhythmites thin and thicken vertically in a random distribution; dropstones are rare, but attain 1.5 m diameter, are well rounded, and are restricted to upper zone.

APPENDIX D-14

Section 14

Elevation of top: 80 m a.s.l.



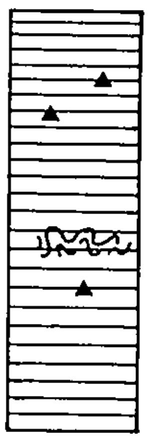
Facies

B Silty diamicton, massive and poorly sorted, 15 m in thickness; angular to rounded clasts of various lithologies and size, up to 1 m diameter; slabs of green sandstone are abundant.

APPENDIX D-15

Section 15

Elevation of top: 53 m a.s.l.



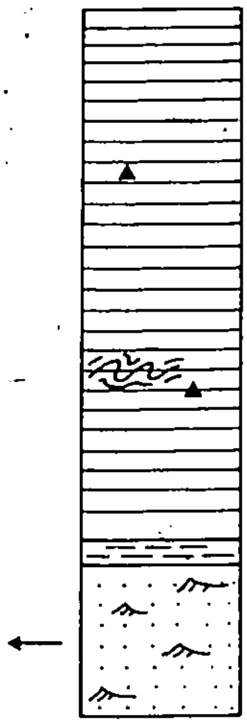
Facies

H Silty clay rhythmites, 15 m thick; couplets range in thickness from 15 cm at base and decrease to 3 cm at top of section; soft sediment deformation is rare and is restricted to mid-section; small dropstones and associated sand and gravel concentrations found at different levels; surface may exhibit cobbles and gravel, and rare marine bivalves forming lag.

APPENDIX D-16

Section 16

Elevation of top: 61 m a.s.l.



Facies

H Silty clay rhythmites, 14.5 m thick; couplets range from 25 cm at base to 3 cm at top of section; minor convolute bedding between undisturbed rhythmites; rare dropstones less than 10 cm diameter present.

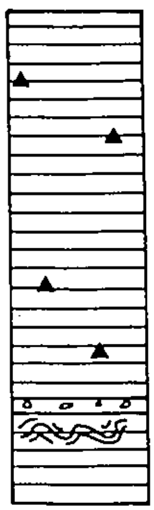
G Silty layer, 50 cm in thickness, finely laminated with two thin clay layers; upper gradational contact with Facies H.

E Silty sand, 4 m thick; composed of climbing ripples and parallel laminations: type A, B, and sinusoidal laminae (Plate 3); palaeocurrent directions towards 270°; upper gradational boundary with Facies G.

APPENDIX D-17

Section 17

Elevation of top: 68 m a.s.l.



Facies

H Silty clay rhythmites, 21 m thick; couplets range in thickness from 30 cm at base, and thin upward to 0.5 cm at top, but there is considerable fluctuation in this trend; dropstones up to 30 cm diameter are present.

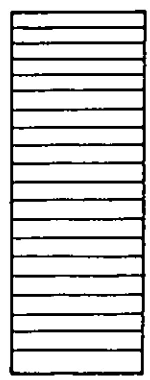
B Poorly sorted silty diamicton, 10 cm thick; rounded to angular clasts up to 7 cm diameter; sharp upper and lower contacts with Facies H.

H Silty clay rhythmites, 5 m in thickness; couplets vary from 30 cm to 20 cm thick; convolute bedding (folds and diapir) are restricted to 1.5 m to 2 m thick bands, underlain and overlain by planar rhythmites.

APPENDIX D-18

Section 18

Elevation of top: 90 m a.s.l.



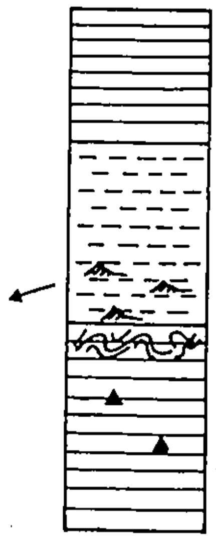
Facies

H Silty clay rhythmites, 8 m thick; couplets range from 10 cm at base to 2.5 cm at top of section; no apparent deformation or dropstones.

APPENDIX D-19

Section 19

Elevation of top: 70 m a.s.l.



Facies

H Silty clay rhythmites, 7.5 m thick; couplets range in thickness from 15 cm at base to 1 cm at top; clay layer in couplets is reddish brown in colour.

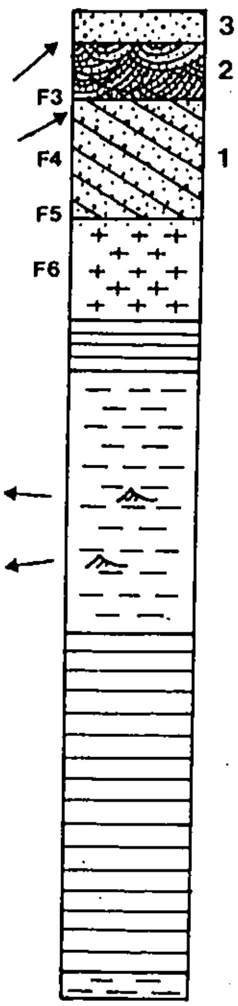
G Silty unit, 6.5 m thick; structures range from massive to horizontally laminated to cross-laminated; alternating sequences of type B climbing ripples, sinusoidal laminae and planar laminae; palaeocurrent direction towards 250°; upper and lower gradational contacts with Facies H.

H Silty clay rhythmites, 11 m in thickness; couplets range from 30 cm to 4 cm thick; contorted bedding 1 m thick underlies Facies G; dropstones do not exceed 10 cm diameter.

APPENDIX D-20

Section 20

Elevation of top: 70 m a.s.l.



Facies and Units

3. Medium to coarse-grained sand, I m thick; horizontally laminated.

2. Fine to coarse-grained sand, 1.5 m thick; trough cross-stratification with sets 35 cm thick, which show a coarsening-upward trend; palaeocurrents towards 048°; upper gradational contact with Unit 3.

1. Medium to coarse sand, 4.5 m thick; tabular cross-stratified beds with fining upward cycles, dipping 22° towards 060°; sharp erosional contact with Unit 2.

I Massive silt with greater concentration of sand at upper gradational contact with Unit 1, 4.5 m thick.

H Silty clay rhythmites, 2 m thick; couplets vary in thickness from 20 cm at base to 3 cm at top; salty taste; sharp contact with Facies I.

G Silty unit, 10 m in thickness; composed of beds 30 cm to 2 m thick of climbing ripples (Plate 4), undulatory laminae, and rarely type A ripples; gradational contacts with Facies H.

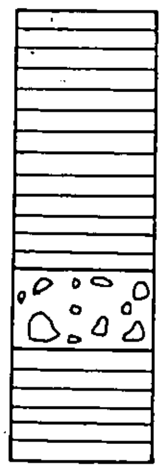
H Silty clay rhythmites, 13 m thick; couplets vary from 35 cm at base to 3 cm at top; gradational contacts with Facies G.

G Silty, laminated deposit, 1 m thick; upper gradational contact with Facies H.

APPENDIX D-21

Section 21

Elevation of top: 70 m a.s.l.



Facies

H Silty clay rhythmites, 7 m thick; couplets range in thickness from 40 cm at base to 2 cm at top of section.

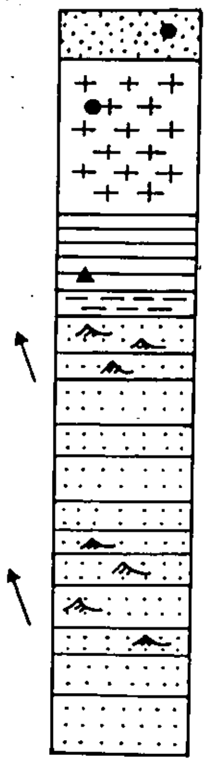
B Poorly sorted massive silty diamicton, 2 m thick; clast size and lithology variable; sharply overlain and underlain by Facies H.

H Silty clay rhythmites, 3 m thick; couplets vary from 10 cm at base to 4 cm at top of bed.

APPENDIX D-22

Section 22

Elevation of top: 80 m a.s.l.



Facies

F Fine to medium sand, 3 m thick; massive to laminated, fossiliferous.

I Massive silt with minor sand concentration at upper boundary with Facies F, 6.35 m thick.

H Silty clay rhythmites , 4 m in thickness; couplets range from 25 cm to 10 cm thick; rare dropstones present; upper boundary with Facies I is sharp.

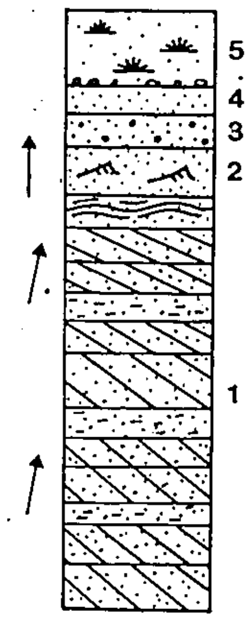
G Silty layer, 50 cm thick; faintly laminated to massive; upper and lower gradational contacts with Facies H and E respectively.

E Silty sand, 16 m thick; composed of cosets of type A, B cross-laminations, sinusoidal laminations, parallel laminations, and may also be massive; palaeocurrents towards 340°.

APPENDIX D-23

Section 23

Elevation of top: 110 m a.s.l.



Units

5. Sandy peat, 1.5 m thick.

4. Silt and fine sand, 1.2 m thick; planar and cross-laminated structure; overlain by 25 cm thick gravel bed.

3. Gravelly sand, 1.25 m thick; massive to poorly stratified; upper and lower gradational contacts.

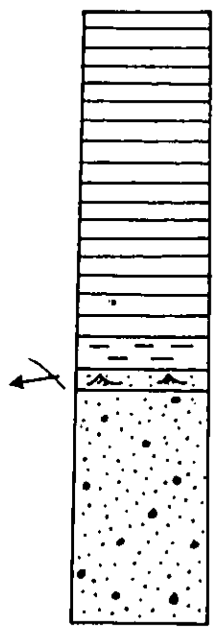
2. Silty sand, 2 m thick; planar and cross-laminations; palaeocurrent directions towards 000°.

1. Alternating strata of massive to laminated fine to medium sand, and tabular cross-bedded sand, separated by thin silty clay layers 3 cm to 7 cm thick; beds range from 10 cm to 1 m thick; cross-beds dip 39° towards 010°; sinusoidal laminations of silty sand near upper contact with Unit 2.

APPENDIX D-24

Section 24

Elevation of top: 65 m a.s.l.



Facies

H Silty clay rhythmites, 18 m thick; couplets range from 30 cm at base to 2.5 cm at top of section.

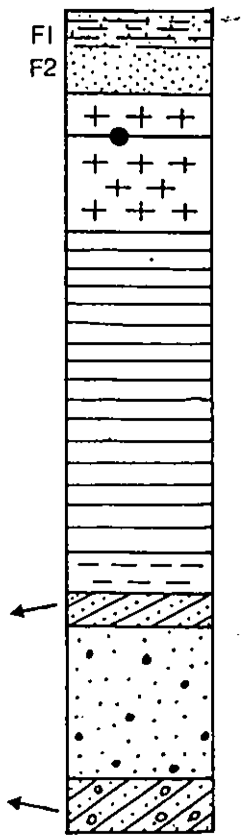
G Silty layer, 2 m thick; massive to finely laminated; upper and lower gradational contacts with Facies H and E respectively.

E Silty sand, 1 m thick; consists of planar and cross-laminated cosets.

D Massive to poorly laminated coarse sand, 12.3 m thick; grades rapidly into Facies E.

APPENDIX D-25

Section 25
Elevation of top: 76 m a.s.l.



Facies

F Medium to fine sand, 4 m thick; massive at base, grading into sand and sandy silt rhythmites, 2 cm to 10 cm thick with gradational contacts.

I Massive silt, 6 m thick; grades abruptly into Facies F.

H Silty clay rhythmites, 17 m in thickness; couplets range from 25 cm to 2.5 cm at the top; no dropstones or deformations observed; sharply overlain by Facies I.

G Silty layers, 2 m thick; massive at base, becoming laminated at top of unit; gradational contact with Facies C.

C Medium to coarse sand, 2 m in thickness; cross-bedded towards 260°; lower gradational boundary with Facies D.

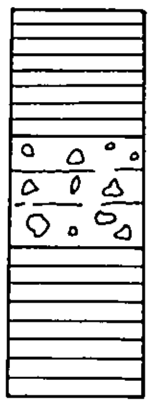
D Massive to laminated medium to coarse sand, 8 m thick; contact with lower Facies C is gradational.

C Coarse sand with minor gravel and some cobbles up to 15 cm diameter; 3 m thick; tabular cross-bedded strata dip 24° towards 280°, and average 30 cm thick.

APPENDIX D-26

Section 26

Elevation of top: 55 m a.s.l.



Facies

H Silty clay rhythmites, 3.5 m thick; couplets range from 8 cm to 2.5 cm at top of section.

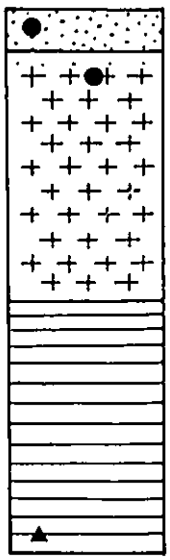
B Poorly sorted silty diamicton, 3 m thick; comprised of strata 15 cm to 40 cm thick; clasts up to .75 cm diameter, angular to rounded and of various lithologies; sharp upper and lower contacts with Facies H.

H Silty clay rhythmites, 4 m in thickness; couplets range in thickness from 25 cm at base to 4 cm in upper part of unit.

APPENDIX D-27

Section 27

Elevation of top: 50 m a.s.l.



Facies

F Fine sand, 1.5 m thick; fossiliferous and structureless.

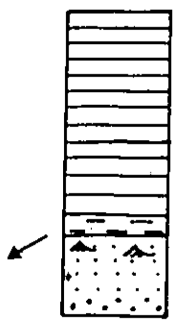
I Massive silt with minor sand at upper gradational contact with Facies F, 9.5 m in thickness.

H Silty clay rhythmites, 10 m in thickness; couplets range from 35 cm at base to 1 cm at top of section; clay layers at base contain sandy gravel layers and lenses; contact with Facies I is sharp.

APPENDIX D-28

Section 28

Elevation of top: 80 m a.s.l.



Facies

H Silty clay rhythmites, 5.5 m thick; couplets range in thickness from 20 cm at base to 3 cm at top of section; upper part of this unit has rhythmites which taste salty.

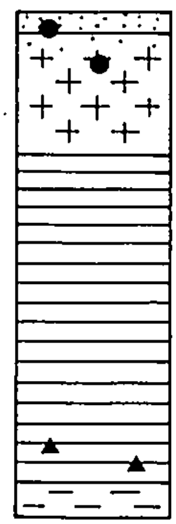
G Silty layer, 50 cm thick; planar laminated at top and sinusoidal laminations at base, with gradational contacts with both Facies H and E.

E Medium to fine sand with greater proportion of silt at top of unit, 2 m in thickness; sand grades upward into ripple-drift cross-laminations (type A, B); palaeocurrent direction towards 240°.

APPENDIX D-29

Section 29

Elevation of top: 40 m a.s.l.



Facies

F Fine to medium sand with rare granules, 1 m in thickness, fossiliferous.

I Massive silt, 8 m thick; fossiliferous, grades into Facies F.

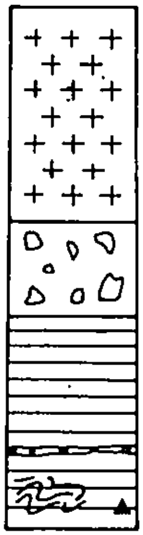
H Silty clay rhythmites, 20 m thick; couplets decrease in thickness from 25 cm to 2 cm thick at top of section; rare dropstones at base with diameters inferior to 30 cm; sharply overlain by Facies I.

G Laminated silt unit, 2 m in thickness; occasional sand partings in silt layers; grades upward into Facies H.

APPENDIX D-30

Section 30

Elevation of top: 40 m a.s.l.



Facies

I Massive clayey silt, 8.25 m in thickness.

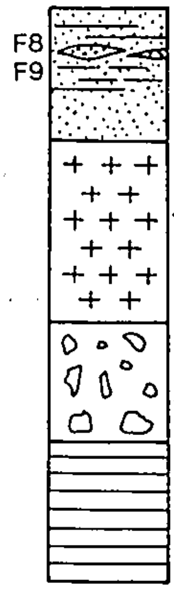
B Structureless poorly sorted silty diamicton, 3.75 m thick; angular to rounded clasts of various lithologies, up to 1 m diameter; very rare sand and gravel lenses; grades sharply into Facies I.

H Silty clay rhythmites, 8 m in thickness; couplets range from 20 cm at base to 2.5 cm at top of unit; thin 10 cm thick diamicton bed (Plate 6) rich in small angular green sandstone clasts, continuous over 30 m, but pinches and swells locally; small-scale deformations in form of diapirs, and rare dropstones present near base of unit.

APPENDIX D-31

Section 31

Elevation of top: 42 m a.s.l.



Facies

F Massive, poorly laminated sand, 5.25 m in thickness; rare pebbly lenses near upper surface; fossiliferous.

I Structureless clayey silt, 7 m thick; grades rapidly into Facies F; very rare clayey stringers suggest a poorly defined bedding structure.

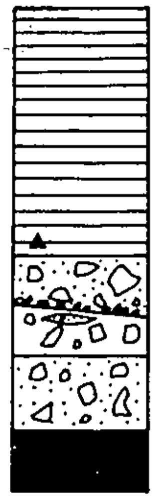
B Poorly sorted, massive silty diamicton, 4.25 m thick; clasts range up to 1 m diameter; rapid gradational contact with Facies I.

H Silty clay rhythmites, 5.5 m thick; couplets vary from 15 cm to 3 cm thick; sharp contact with Facies B showing depressed rhythmites.

APPENDIX D-32

Section 32

Elevation of top: 26 m a.s.l.



Facies

H Silty clay rhythmites, 7 m thick; couplets range from 15 cm at base to 2.5 cm at top of section.

A Sandy, massive diamicton, poorly sorted, 1.5 m thick; sharply overlain by Facies H.

B Structureless silty diamicton, 0.25 m to 1.25 m thick; rare gravel and sand lenses and discontinuous layers (Plate 2), as well as thin clay lenses in med.-fine sand layers, 2 cm to 5 cm thick near upper contact with Facies A; this contact is marked by a granule and gravel lag deposit.

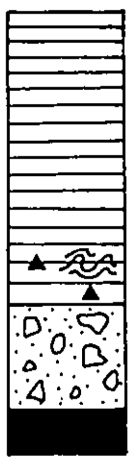
A Sandy, non-sorted diamicton, 2 m in thickness; sharply overlain by Facies B; numerous angular clasts of green sandstone.

R Bedrock: smooth, locally irregular green sandstone, 1.75 m thick.

APPENDIX D-33

Section 33

Elevation of top: 30 m a.s.l.



Facies

H Silty clay rhythmites, 11 m in thickness; couplets range from 15 cm at base to 3 cm at top of section; rare dropstones (30 cm diameter) and soft sediment deformation near base of unit.

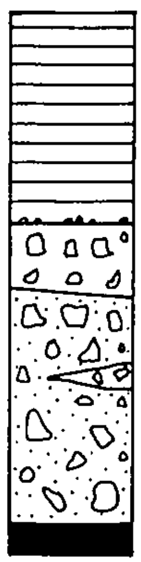
A Poorly sorted, structureless sandy diamicton, 4 m thick; clasts of variable size, composition and degree of roundness; sharply overlain by Facies H.

R Bedrock, 1.75 m thick, green sandstone.

APPENDIX D-34

Section 34

Elevation of top: 27 m a.s.l.



Facies

H Silty clay rhythmites, 5.75 m thick; couplets range in thickness from 15 cm at base to 2.5 cm at top of section; no apparent dropstones or deformation.

B Poorly sorted silty diamicton, 2 m thick; structureless, clasts of different lithologies, rounded to angular; separated from Facies H by stony clay layer.

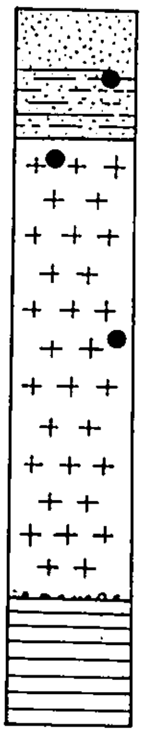
A Massive sandy diamicton, 6 m in thickness; rounded to angular clasts of green sandstone, up to 1 m diameter; contains lenses of diamicton (Facies B), .25 m to .75 m thick; sharp contacts with silty diamicton.

R Bedrock, green sandstone, 1 m thick.

APPENDIX D-35

Section 35

Elevation of top: 28 m a.s.l.



Facies

F Medium to fine sand, 4.5 m thick; upper 2 m are massive; lower 2.5 m composed of silty sand layers (1 cm) and silt layers (5 cm) which alternate.

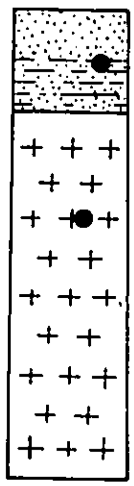
I Massive clayey silt, 17 m in thickness; fossiliferous, upper gradational contact with Facies F; at base of unit, there is a discontinuous sandy horizon with rare granules and gravel, 15 cm thick.

H Silty clay rhythmites, 4.5 m thick; couplets range in thickness from 15 cm at base, to 2.5 cm near top of unit; rhythmites have a salty taste; sharply overlain by Facies I.

APPENDIX D-36

Section 36

Elevation of top: 23 m a.s.l.



Facies

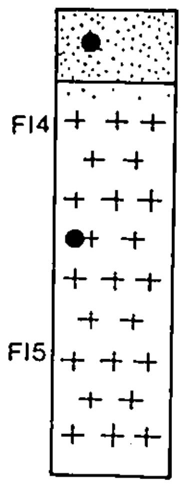
F Medium to fine fossiliferous sand, 4 m in thickness; consists of upper 2 m of massive to poorly bedded sand, underlain by sandy silt and silt alternating layers, 1 to 5 cm thick.

I Massive clayey silt, 14 m thick; fossiliferous, grading upward into Facies F.

APPENDIX D-37

Section 37

Elevation of top: 20 m a.s.l.



Facies

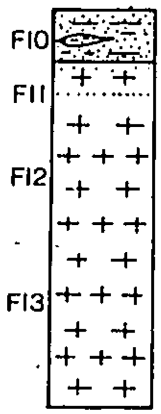
F Medium to fine fossiliferous sand, 2 m in thickness; massive to poorly laminated.

I Structureless silt, 11 m thick; fossiliferous, grading upward into Facies F, with minor sand increase near upper boundary.

APPENDIX D-38

Section 38

Elevation at top: 20 m a.s.l.



Facies

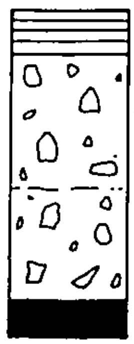
F Medium to fine fossiliferous sand, 2 m thick; massive to faintly laminated with occasional silt lenses, 10 cm thick by 50 cm long.

I Massive clayey silt, 13 m thick; fossiliferous; blue-gray in colour at base grading to brown-gray near top; upper contact with Facies F gradational.

APPENDIX D-39

Section 39

Elevation of top: 40 m a.s.l.



Facies

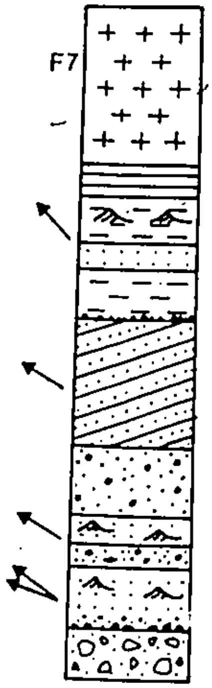
H Silty clay rhythmites, 1 m in thickness; couplets average 10 cm thick.

B Massive, poorly sorted silty diamicton, 6.5 m thick; poorly defined bedding plane at mid-section; clasts of various lithologies, up to 1 m diameter, angular to rounded; sharply overlain by Facies H.

R Bedrock, red sandstone, 1 m thick.

APPENDIX D-40

Section 40
Elevation of top: 65 m a.s.l.



Facies

I Massive clayey silt, fossiliferous near top, 6 m thick.

H Silty clay rhythmites, 2 m in thickness; couplets have gradational contacts, taste salty and vary in thickness from 40 cm to 15 cm near top; rapid gradational contact with Facies I.

G Silty unit, 2 m thick; type B climbing ripples at base, overlain by sinusoidal laminae and planar laminae; palaeocurrents towards 315°; gradational upper and lower surfaces.

E Laminated sand and silt, 1 m thick, with minor sinusoidal laminae.

G Massive to laminated silt, 2 m thick.

C Coarse sand, 4.5 m thick, 15-30 cm thick strata dip 18° to 21° towards 300°; sharply overlain by 15 cm gravel lag.

D Medium to coarse sand with rare gravel, 2.25 m in thickness; cross-laminated at top, grading downward to planar laminae; upper erosional contact.

E Normal graded medium to fine sand, 1.25 m thick; laminated at base grading to type B ripples.

D Massive to poorly laminated coarse sand with rare gravel, 1 m thick; rare red clay rip-up clasts.

E Normal graded bed, laminated coarse sand at base to rippled fine sand at top, 2 m thick; erosional upper contact.

A Poorly sorted sandy diamicton, 2 m thick; reddish brown in colour with abundant angular clasts of red sandstone.

APPENDIX E

Foraminifera of the Postglacial Sea in the Basin of the Richardson and Rae Rivers, N.W.T.

Alveophragmium crassimargo (Norman)

A. cf. jeffreysi (Williamson)

Buccella frigida (Cushman)

Cassidulina reniforme Nørvang

Cornuspira involvens (Reuss)

Dentalina frobisherensis Loeblich and Tappan

D. ittai Loeblich and Tappan

Elphidiella groenlandica Cushman

Elphidiella sp.

Elphidium bartletti Cushman

E. clavatum Cushman

E. orbiculare (Brady)

E. williamsoni Haynes

Elphidium sp.

Fissurina sp.

Haplofragmoides sp.

Hyperammina sp.

Islandiella helenae Feyling-Hanssen and Buzas

I. norcrossi (Cushman)

Lagena gracillima (Seguenza)

L. semilineata Wright

APPENDIX E - (Concluded)

Oolina melo d'Orbigny

Oolina sp.

Parafissurina sp.

Polymorphina sp.

Pseudopolymorphina novangliae (Cushman)

Pyrgo williamsoni (Silvestri)

Quinqueloculina stalkeri Loeblich and Tappan

Reophax curtus Cushman

R. scorpiurus Montfort

Scutuloris tegminis Loeblich and Tappan

Triloculina trihedra Loeblich and Tappan

Virgulina loeblichi Feyling-Hanssen

APPENDIX F

Ostracoda of the Postglacial Sea in the Basin of the Richardson and Rae Rivers, N.W.T.

Acanthocythereis dunelmensis (Norman)

Cytheropteron paralatissimum Swain

C. simplex Whatley and Masson

Eucytheridea bradii (Norman)

E. macrolaminata (Elofson)

E. punctillata (Brady)

Heterocyprideis sorbyana (Jones)

Ilyocypris bradyi Sars

I. gibba (Ramdohr)

Limnocythere cf. *herricki* Staplin

Loxoconcha venepidermoidea Swain

Palmenella limicola (Norman)

Paracyprideis pseudopunctillata Swain

Pseudocythereis simpsonensis Swain

Rabilimis cf. *septentrionalis* (Brady)

Roundstonia globulifera (Brady)

APPENDIX G

Macrofossils of the Postglacial Sea in the Basin of the Richardson and Rae Rivers, N.W.T.

Hiatella arctica (Linné)

Macoma balthica (Linné)

M. calcarea (Gmelin)

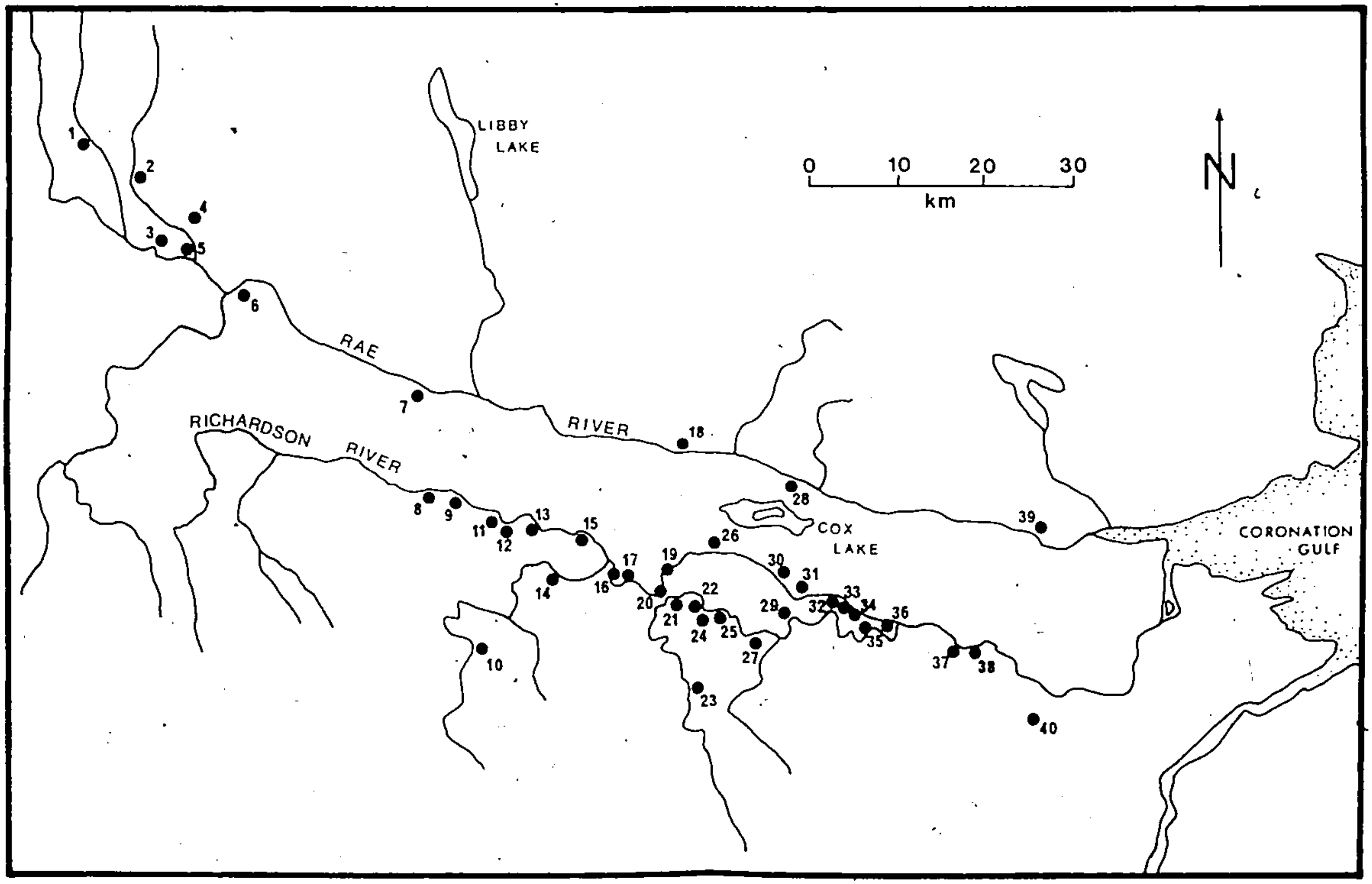
Mya truncata (Linné)

Mytilus edulis (Linné)

Portlandia arctica (Gray)

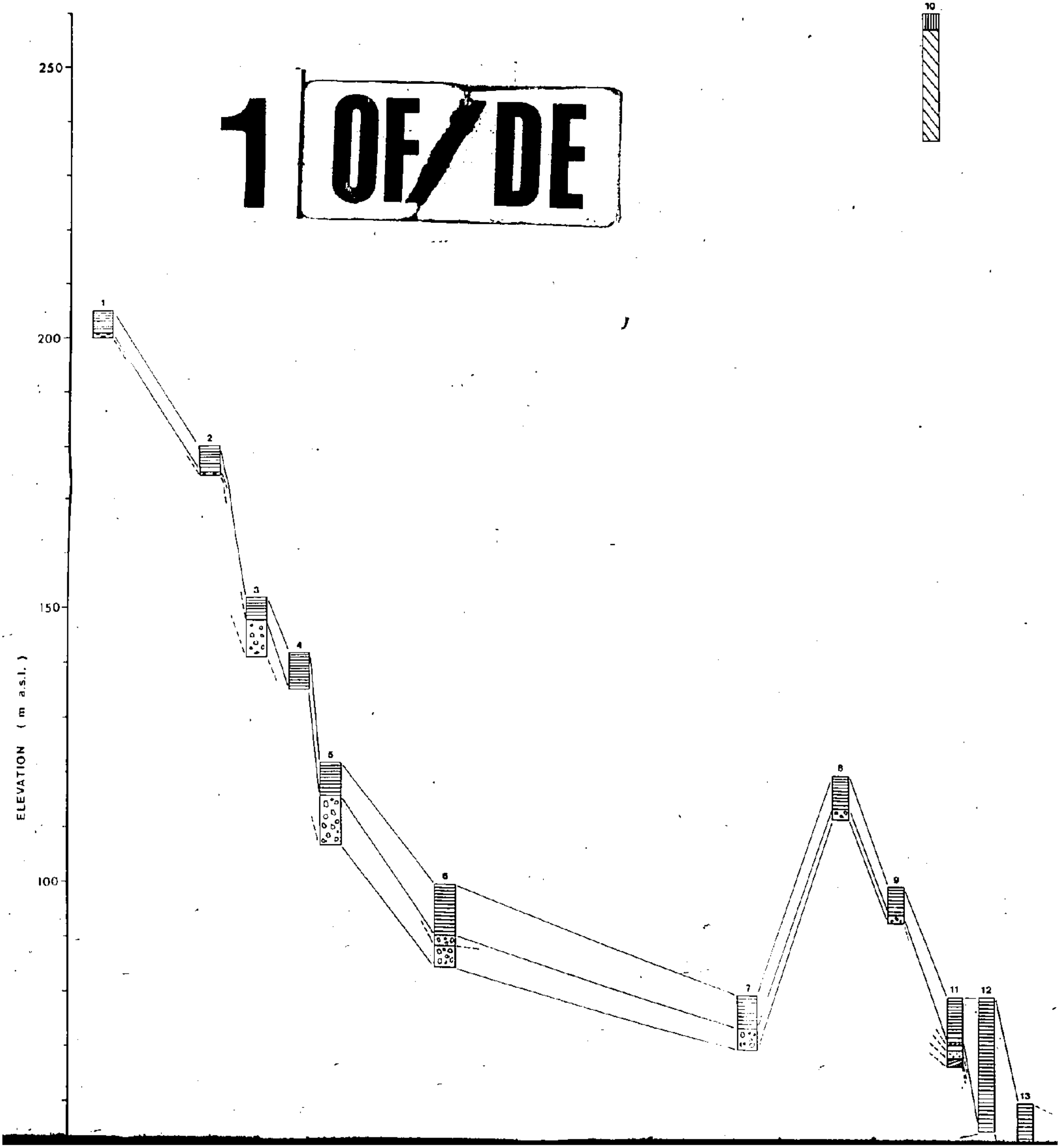
P. lenticula (Möller)

Saxicava arctica Linné

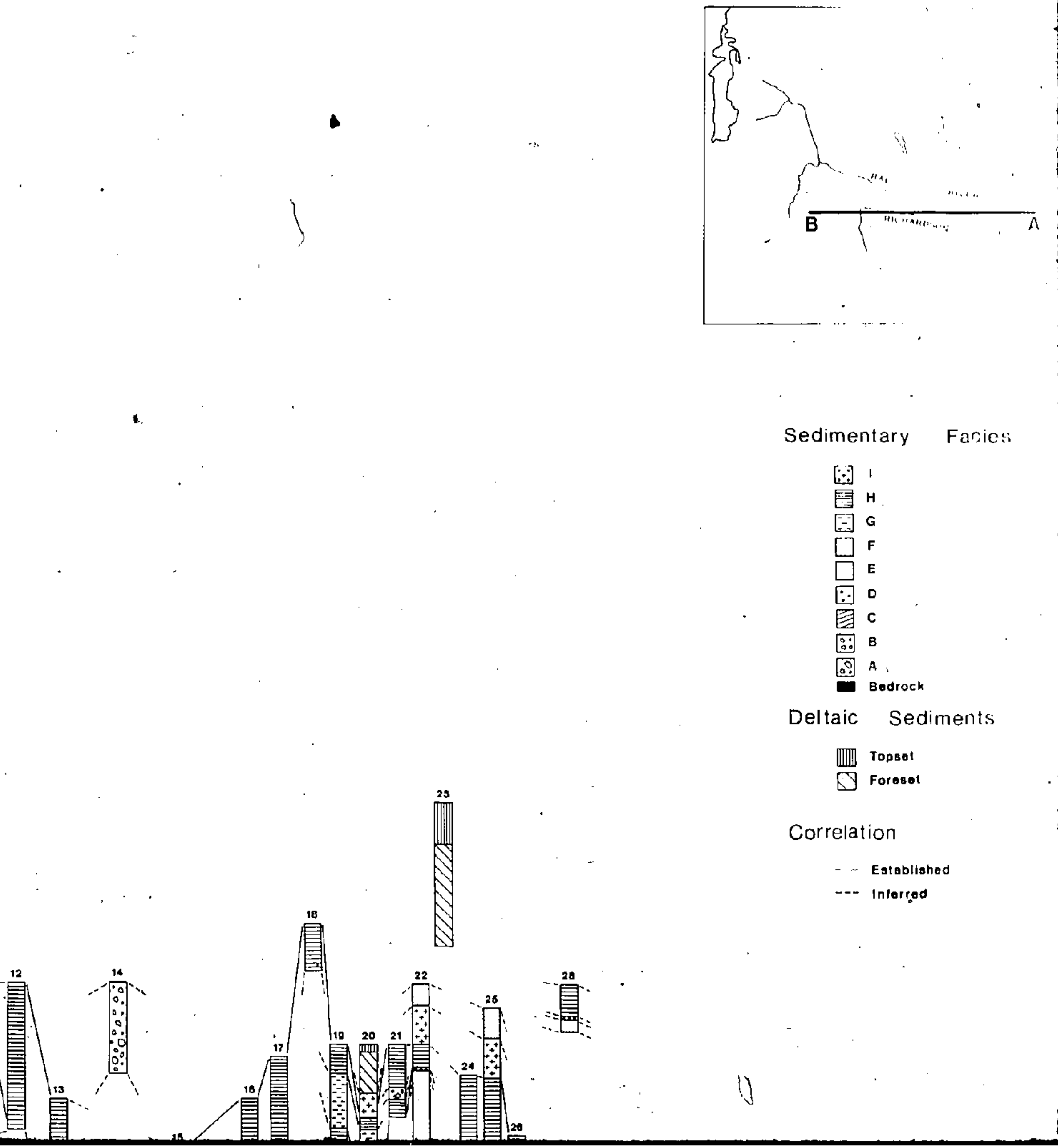


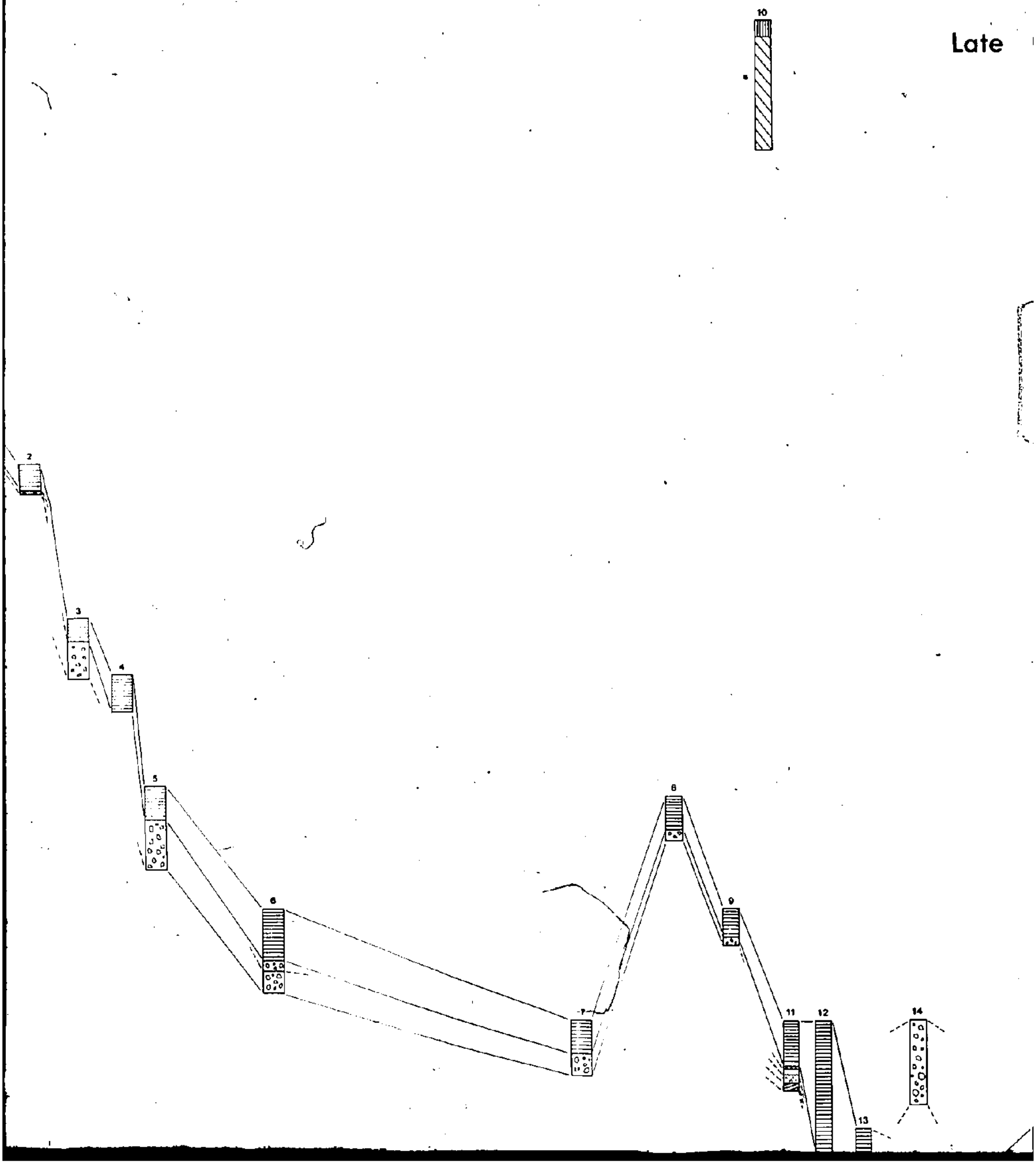
Location of stratigraphic sections.
(To accompany Figure 4)

1 OF/DE



Late Quaternary Lithostratigraphic Correlation



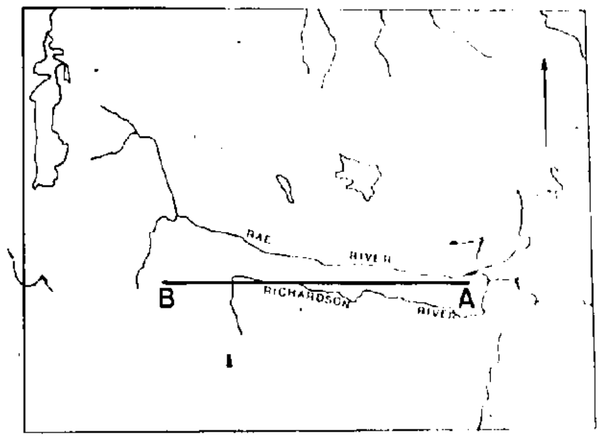


Late
10
2
3
4
5
6
7
8
9
11
12
13
14

Late Quaternary Lithostratigraphic Correlation

2

OF/DE



Sedimentary Facies

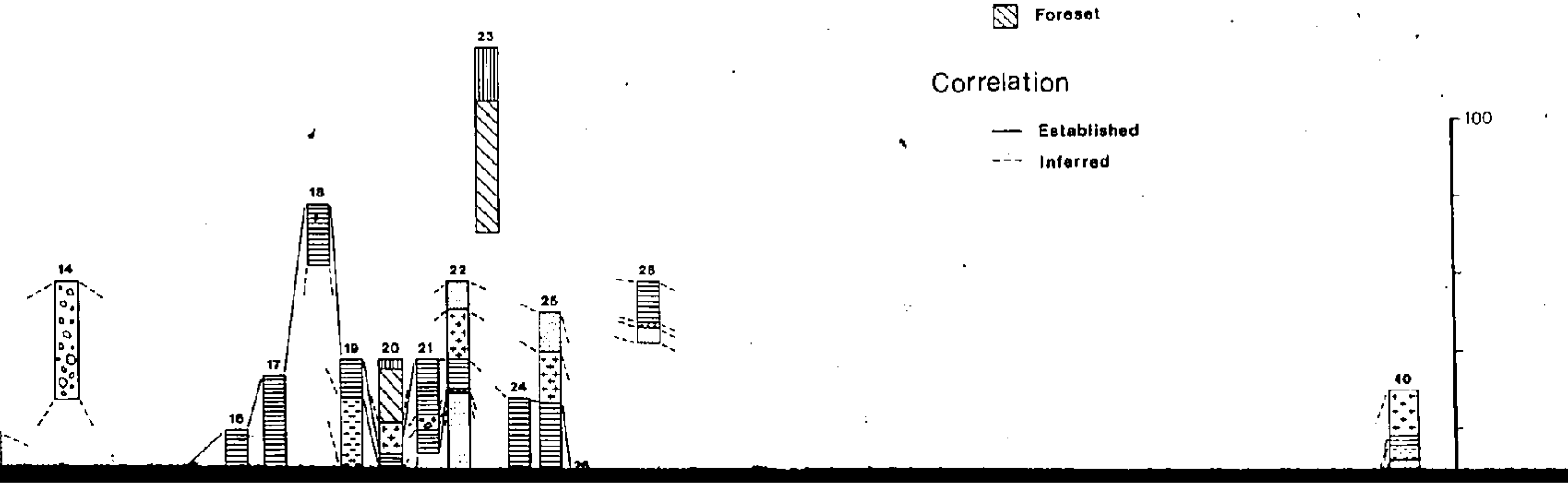
- I
- H
- G
- F
- E
- D
- C
- B
- A
- Bedrock

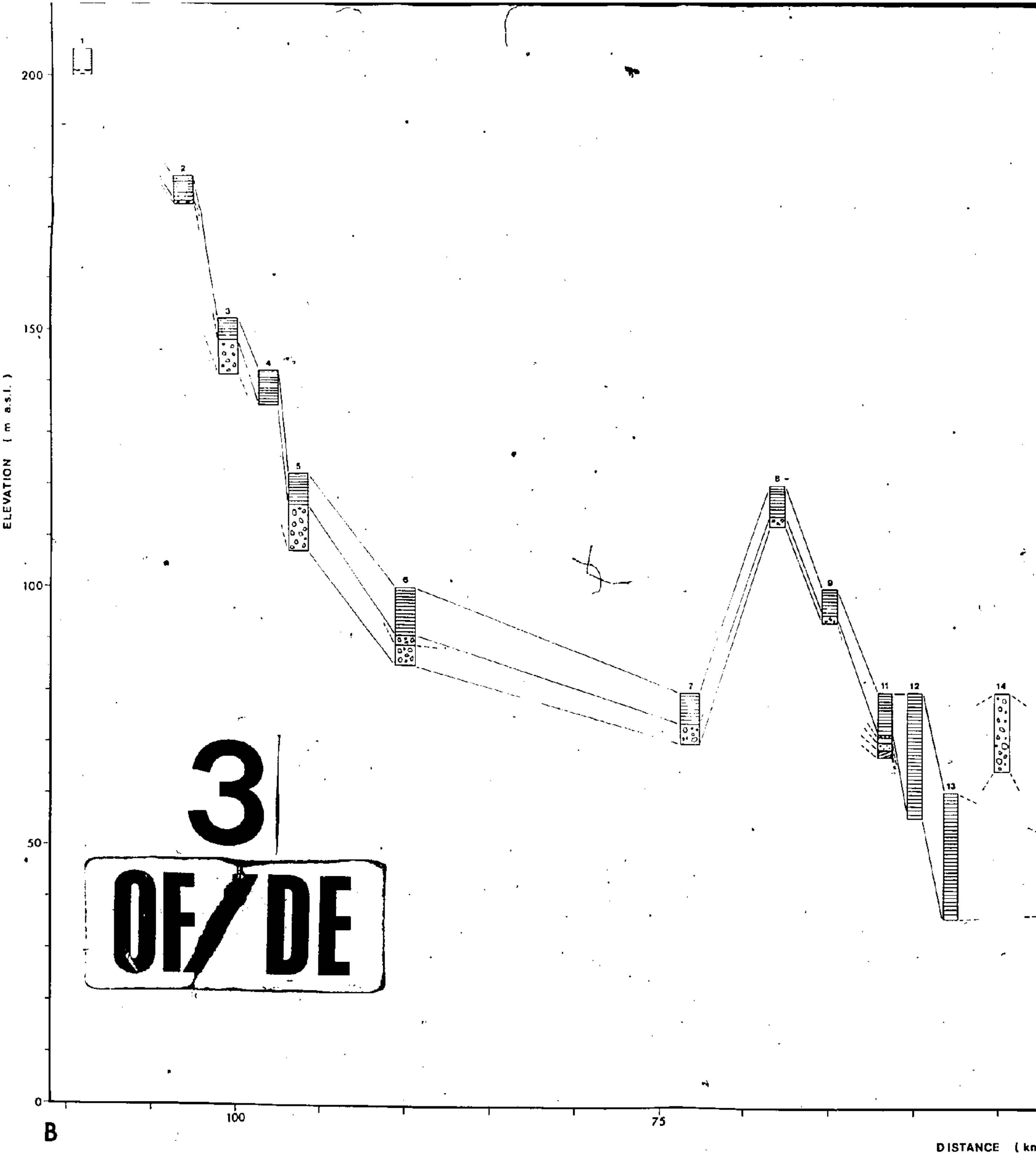
Deltaic Sediments

- Topset
- Foreset

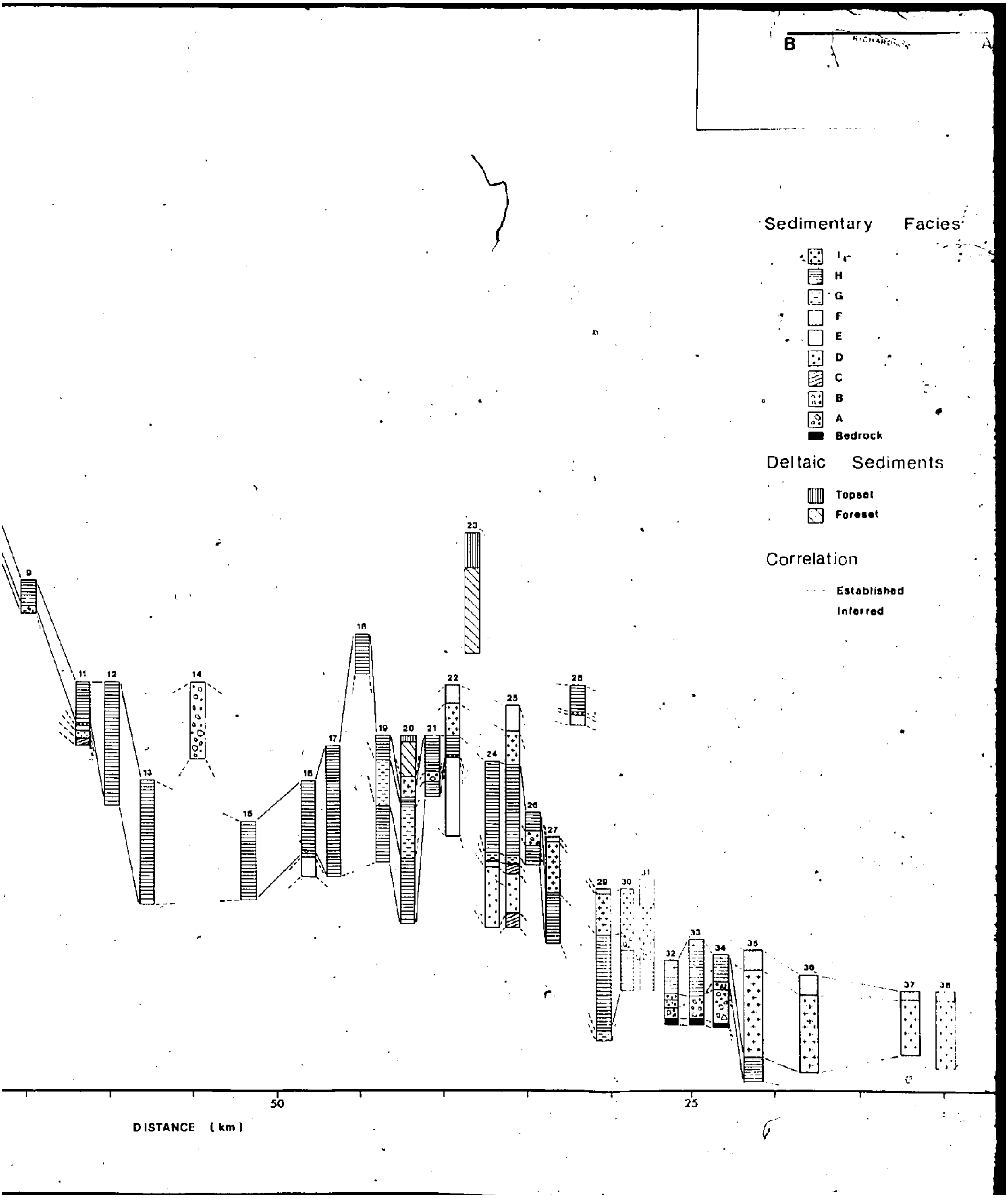
Correlation

- Established
- Inferred

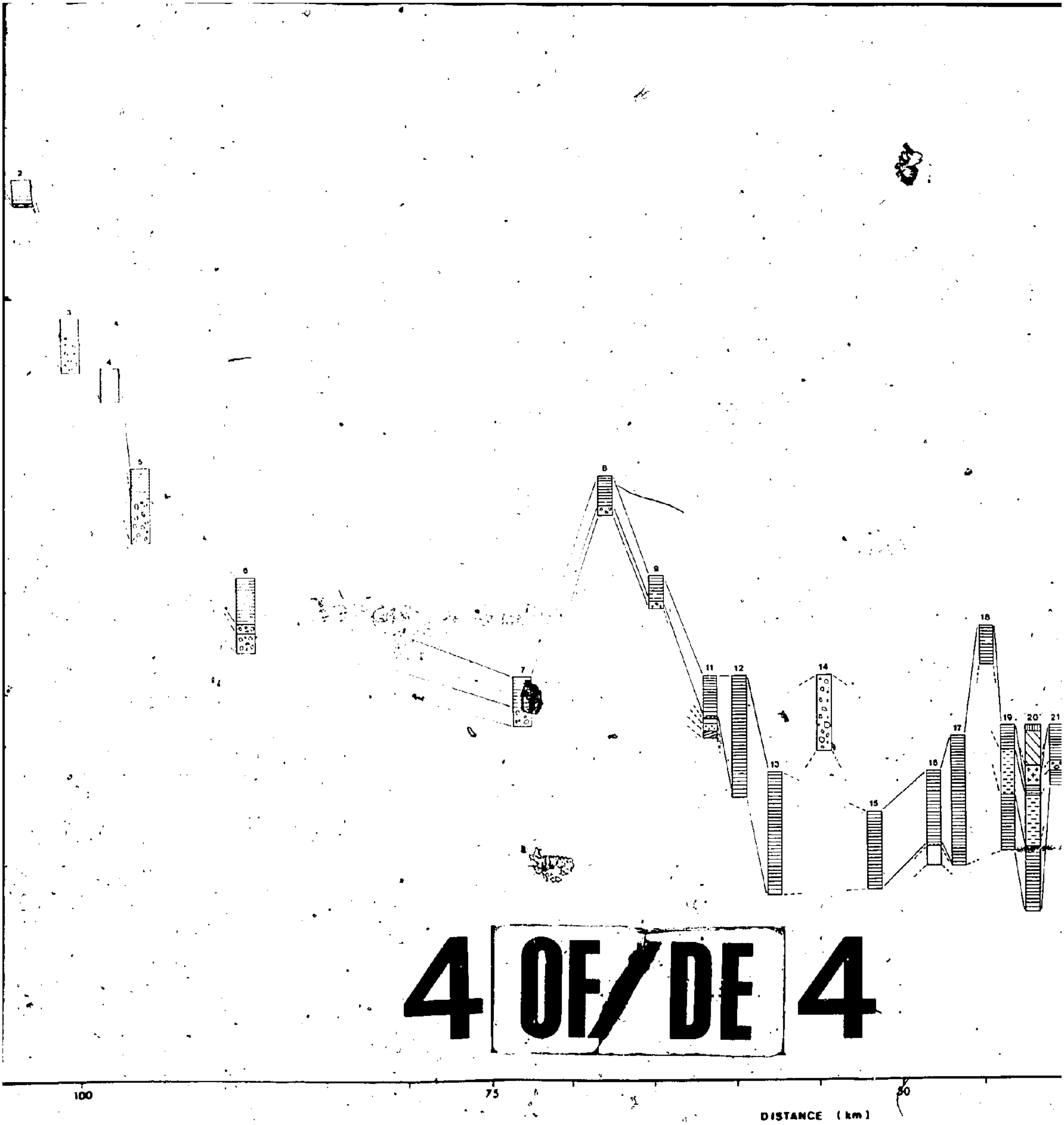



ELEVATION (m a.s.l.)
200
150
100
50
0
1
2
3
4
5
6
7
8
9
11
12
13
14
100
75
B
DISTANCE (km

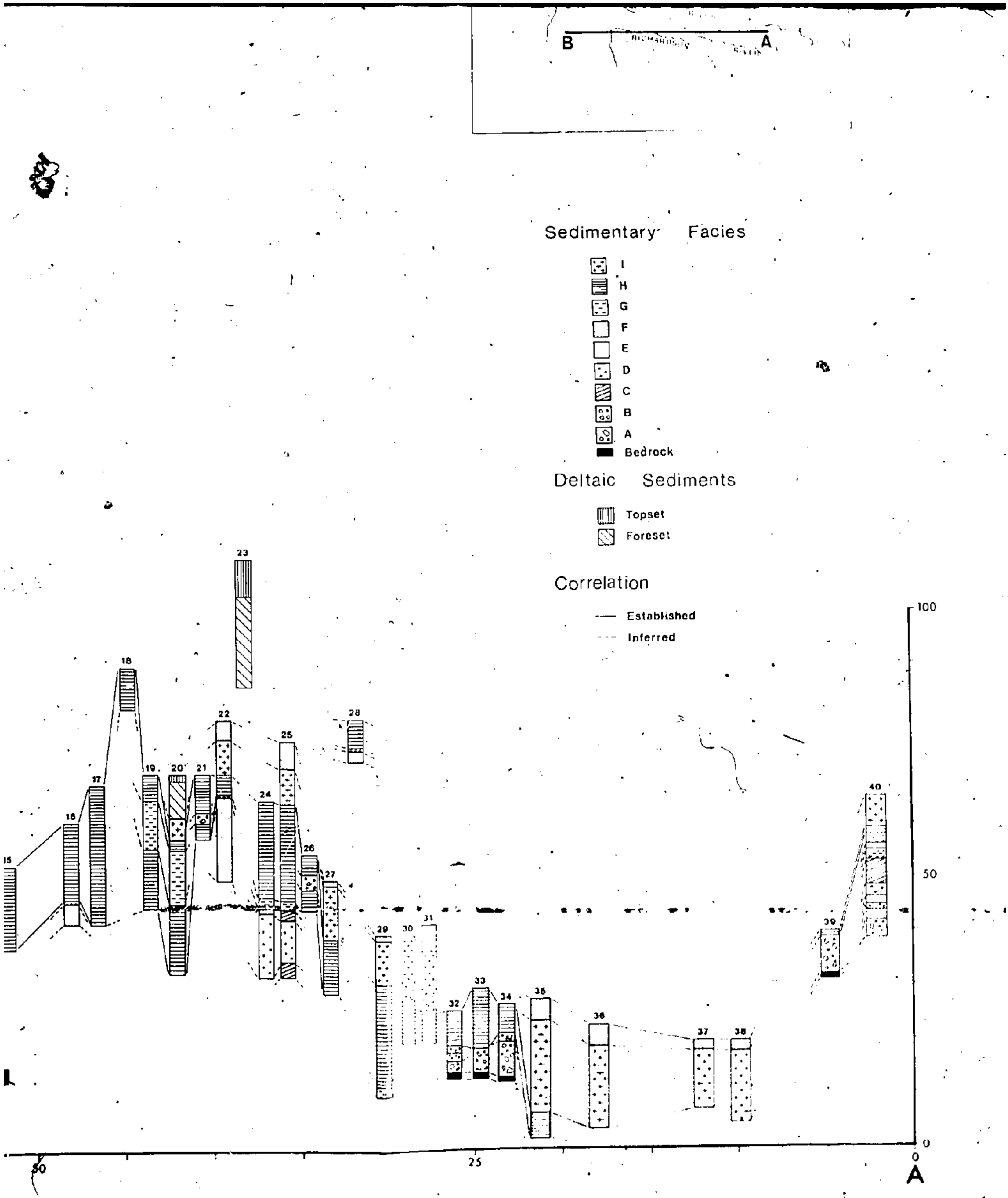
3
OF/DE



B
Sedimentary Facies
I
H
G
F
E
D
C
B
A
Bedrock
Deltaic Sediments
Topset
Foreset
Correlation
Established
Inferred
9
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
50
25
DISTANCE (km)



4 OF/DE 4



B
A
Sedimentary Facies
I
H
G
F
E
D
C
B
A
Bedrock
Deltaic Sediments
Topset
Foreset
Correlation
Established
Inferred
100
50
0
50
25
0
A