LATE PLEISTOCENE GLACIOFLUVIAL AND GLACIOMARINE SEDIMENTS ON THE NORTH SIDE OF THE MINAS BASIN, NOVA SCOTIA

by

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Submitted in partial fulfilment of the requirements for the degree of Doctor of Philosophy at Dalhousie University, Halifax, Nova Scotia, May, 1980.

DALHOUSIE UNIVERSITY

FACULTY OF GRADUATE STUDIES

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TO MY PARENTS

Dos est magna parentium virtus

TO DIXON AND GERRY

Amicum perdere est damnorum maximum

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ABSTRACT

A discontinuous glaciomarine outwash terrace on the north shore of the Minas Basin and ice contact stratified drift north of Parrsboro were mapped. Detailed sedimentological studies on some of the deposits provide insight into deglacial depositional processes.

The outwash terrace consists predominantly of marine Gilbert type deltas with molds and casts of *Portlandia arctica* in the bottomset beds. The topset beds and glaciofluvial gravel inland of the deltas were deposited by braided meltwater streams with sustained velocities (inferred) of 1 to 4 m/sec and with flow depths of 1 to 2 m. The predominant crude bedding of the gravel is a result of rapid downstream sediment movement. Shifting of the streams produced a planar, erosional unconformity between the topsets and finer grained foresets. Bed load sediment avalanched down the foreset slope while finer sediment was carried in suspension onto the bottomset slope. In the winter, the supply of meltwater ceased and clay and fine silt settled out of suspension, producing bottomset varves where coarse silt and sand had been deposited. Infrequent large slumps of the foresets produced turbidity currents which deposited coarse sand in the bottomsets. Foreset progradation caused instability and failure of the bottomsets which occasionally submerged parts of the delta and caused secondary foreset deposition.

Pebble counts show that as the receding valley ice crossed bedrock boundaries, different rock types were supplied to the delta. During and after deltaic deposition, postglacial rebound exceeded sea level rise and the deltas became emergent, with emergence increasing steadily to the west. Meltwater then eroded terraces in the glaciomarine and inland

glaciofluvial sediments. The northward receding ice in the Parrsboro Gap built a small recessional moraine that forms the present drainage divide between Parrsboro River and River Hebert at Gilbert Lake. Holocene sea level rise and subsidence of the land have resulted in submergence of the east end of the terrace and substantial erosion of the deposits. Eroded sediments are subaerially preserved only at Advocate Harbour.

Radiocarbon dating of organic sediment in kettles and lakes is inconclusive. The pollen assemblage and stratigraphy of the Leak Lake core suggest correlation with outwash deposits in southwestern New Brunswick, dated at 13,000 to 13,325 yr BP. Glaciomarine sediment, which appears to be an offshore equivalent to the outwash terrace, in the Minas Basin is about 14,000 years old, suggesting that the terrace is slightly older than was previously thought.

ACKNOWLEDGEMENTS

It is not possible to thank all of the people that have helped me during the course of this study and it is probable but unintentional that some significant contributors may have been omitted in this section.

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Many people kindly provided nonscientific assistance and I am as grateful to them as to the aforementioned. My roommates, especially those from '1470,' made my graduate student days enjoyable and memorable and my last roommate, Ric Bezanson, kindly housed me during the difficult 'thesis writing' stage. I thank the people of Advocate Harbour and Parrsboro for the many friendships that were struck and the many kindnesses supplied along the way. Peter, Winnie, Sam and Brigit McCaig provided hospitality and the finest of cuisine at the Ottawa House, my favourite home away from home. Thanks are also extended to my many opponents and partners on the Parrsboro tennis court and unknowingly you have made my days in the field well known at Dalhousie. My officemates, especially those of the 'Penthouse

Gangue,' instigated many interesting discussions, only some of which concerned my thesis. I express my heartfelt thanks to Art Atkinson for his assistance on many different aspects of the thesis but mostly for his friendship. The coring of some of the lakes was done under severe conditions and most people could not have been paid enough to help.

Lastly, I thank Dianne Crouse for typing the first draft of the thesis and Joan Checholik for typing the final version as well as doing an excellent job of editing.

CHAPTER 1. INTRODUCTION

THE THESIS TOPIC

In the winter of 1974, Dr. Gerard Middleton of McMaster University suggested the study of the glaciomarine outwash terrace on the north shore of the Minas Basin as a potential thesis topic. The terrace consists predominantly of Gilbert type deltas that are characteristic of lakes rather than seas (Ashley, 1972, p. 35). The Bay of Fundy is macrotidal with tidal ranges reaching 16.3 m, and Swift and Borns (1967, p. 698) and Grant (1970, p. 685) use the deltas as evidence that the early postglacial tides were minimal. But even if the Bay of Fundy was nontidal or microtidal in early postglacial times, the structure of the deltas was thought to be atypical of marine conditions although few, if any, marine Gilbert type deltas have been studied in detail. The only beaches described in the terrace complex occur at Advocate Harbour, and the reason implied for this is that the western end of the terrace is more open to wave attack (Swift and Borns, 1967, p. 697). However, other parts of the terrace, such as Port Greville, are also open to waves travelling up the Bay of Fundy, so there seem to be other controlling factors as to why the beaches occur only at Advocate Harbour. These were the problems that piqued interest in the outwash terrace and led to an initial investigation.

A field trip to the north shore of the Minas Basin in May of 1974 revealed that the Nova Scotia Department of Highways was excavating a pit in the raised beach at Advocate Harbour. Continuous excavation provided a unique three dimensional exposure of the beach. It was decided to study the beach in detail and to compare the beach to the nearest delta, which is at Spencers Island. Although field work was carried out on both the raised beach at Advocate Harbour and the raised delta at Spencers Island, the beach study was fruitful beyond expectation and provided sufficient material for an M.Sc. thesis (Wightman, 1976).

The original problems that remained unanswered by the M.Sc. thesis and additional problems that were raised during the course of the work at Advocate Harbour made it worthwhile to study the whole of the raised terrace area, which is the subject of this thesis.

PHYSIOGRAPHY

The peninsula that lies between Chignecto Bay and Minas Basin, hereafter called Chignecto Peninsula, comprises igneous and metamorphic highlands and sedimentary lowlands. The dominant geomorphic feature on the southern part of the peninsula is the Cobequid Highlands (Fig. 1.1). The Cobequids trend east-west and extend from Cape Chignecto in the west to a little beyond the eastern end of the Minas Basin. In general, they are low (125 to 225 m), narrow (2 to 10 km) and consist of hills in the west, of which the higher ones are plutons. To the east, the Cobequids broaden to approximately 17 km and form a plateau at an elevation of approximately 300 m, although elevations of almost 375 m are attained; this surface is the remnant of a broad Cretaceous peneplain (Goldthwait, 1924). The physiography of the Cobequids reflects the increasing proportion of igneous rocks to the east. Two major wind gaps, the Parrsboro

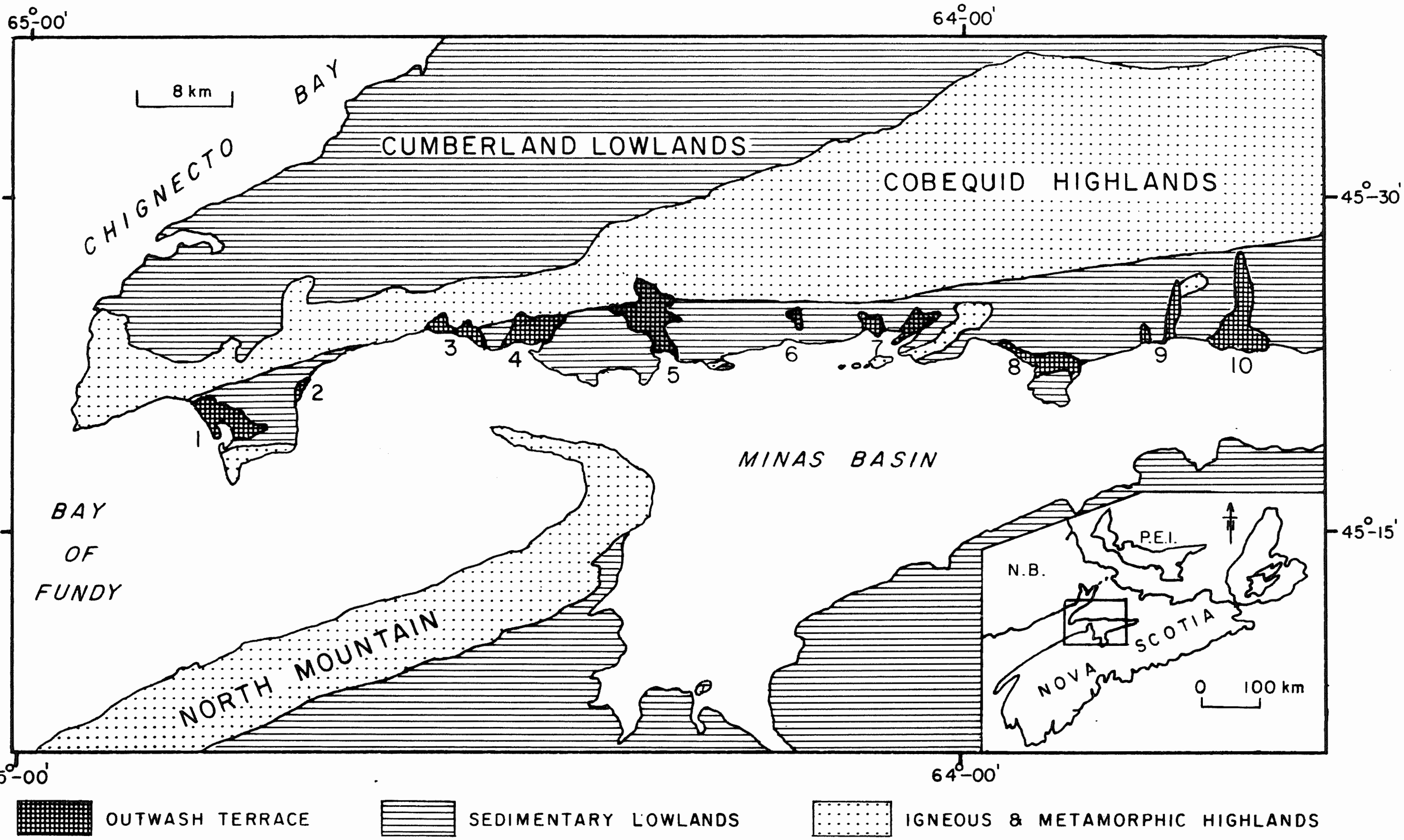


FIGURE 1.1. Geomorphic provinces of the Minas Basin area and outwash terrace in thesis study area. Locations of outwash deposits are: 1. Advocate Harbour, 2. Spencers Island, 3. Port Greville area, 4. Diligent River, 5. Parrsboro, 6. Moose River, 7. Five Islands, 8. Economy, 9. Bass River and 10. Portapique. (Modified after Swift and Borns, 1967.)

Gap and the Folly Lake Gap, provide breaks through the Cobequids (Goldthwait, 1924, p. 25).

The Cobequid Complex (Weeks, 1948, p. 7) is composed mainly of low grade metamorphic Silurian to Middle Carboniferous stratified rocks that are intruded by Devonian to Carboniferous plutonic rocks. From Parrsboro eastwards, there are lesser amounts of higher grade metamorphic rocks, including gneisses and schists, of Paleozoic and Proterozoic age.

The southern edge of the Cobequids is bounded by the Cobequid Fault, which is remarkably straight and is part of the larger Glooscap Fault system (King *et al.* 1975). Erosion of the softer sedimentary rock south of the fault has created a precipitous fault line scarp.

South of the Cobequid Fault there are several occurrences of Triassic basalt, most notably at Cap D'Or near Advocate Harbour and at Five Islands. The basalts also form highlands, the occurrence at Cap D'Or reaching an elevation of 170 m while at Five Islands, Economy Mountain reaches an elevation of 215 m.

The Cobequids are bordered by a narrow belt of sedimentary lowlands to the south and a broader area of sedimentary lowlands to the north. The "lowlands" are lower than the Cobequids but have substantial elevations and relief near the Cobequids. On the southern side, the Cobequid Fault separates the highlands from Carboniferous sedimentary rocks, which are in turn separated by faults from Triassic sedimentary and basaltic rocks. The Carboniferous and Triassic sedimentary rocks form a discontinuous strip of lowlands with the softer Triassic rocks at lower elevations than the Carboniferous. Thus, the lowlands generally slope towards the Minas Basin.

West of Port Greville the lowlands are present only between Spencers Island and Advocate Harbour where they are roughly 6 km wide and reach elevations of 150 m. East of Port Greville, the lowlands generally widen, due mostly to a broadening of the Triassic sedimentary rocks, towards the head of the Minas Basin where they are approximately 14 km wide. The Triassic sedimentary rocks are generally lower than 100 m while the Carboniferous rocks are generally lower than 150 m.

The discontinuous outwash terrace on the north shore of the Minas Basin lies on this sedimentary lowland, occurring in river valleys and extending varying distances inland from the mouths of the valleys. The distance is inversely related to the width and steepness of the valleys and, as the valleys steepen and narrow in the Cobequid Highlands, the deposits occur predominantly south of the Cobequid Fault (Fig. 1.1).

Carboniferous sedimentary rocks of the Cumberland Basin unconformably overlie the older Cobequid rocks on the north side of the highlands and the area is termed the Cumberland Lowlands. In general, the lowlands are highest near the Cobequids (up to 225 m) and slope northward to Chignecto Bay. This increase in elevation towards the Cobequids causes the shoreline to steepen as the lowlands narrow to the west.

PREVIOUS WORK

Chalmers (1894, p. 24M) referred to parts of the outwash terrace on the north shore of the Minas Basin and recognized the importance of the deposits in defining the extent of postglacial marine onlap. Goldthwait (1924, pp. 100, 152) discussed part of the terrace at Parrsboro and suggested that, because of postemergence erosion, its shoreward elevation of 18 m might not represent the maximum extent of marine onlap. He also noted the "step like series of broad terraces" in the delta and recognized that they were cut by the Parrsboro River, and not by the sea, as the delta was uplifted. Goldthwait (1924, p. 152) also gives elevations for other parts of the terrace at Diligent River, Spencers Island and West Advocate that Chalmers (1894) had listed.

Borns (1965) described ice wedge casts in the terrace but also considered the general features of the deposits and came to some conclusions on the genesis of the terrace. He suggested that "the last Pleistocene ice sheet to cover northern Nova Scotia (Chignecto Peninsula) dissipated primarily by downwasting, probably by downmelting." Borns interpreted the deposition of the terrace as follows. The Cobequid Mountains cut the ice sheet into two parts and meltwater from the ice south of the mountains deposited valley trains that merged into deltas at the sea. The meltwater and outwash were closely associated with masses of stagnating ice and some of the ice masses melted after the deposition of outwash had ceased, as shown by a large kettle (≃60 m x 9 m) at Port Greville that developed after the cutting of an 8 m fluvial terrace. Borns also stated that many of the smaller kettles were not infilled because of the shifting positions of the meltwater streams.

Borns (1965) correlated the stagnating ice sheet on the Chignecto Peninsula with the ice sheet that constructed the end moraine system extending from Cherryfield, Maine to St. John, New Brunswick. Marine shells from the moraine were dated at 13,325 yr BP (I-GSC-7). As corroborating evidence, Borns mentions that Livingstone and Livingstone (1958,

p. 356) also concluded that the last ice on Cape Breton was of Cary age (11,700 to 13,600 yr BP; Flint, 1955). The ice wedge casts, which post-date the accumulation of the outwash, were thus assigned a Valders age. In addition, Borns (1965, p. 1224) concluded that "nothing was found to suggest the presence in the area of outwash of more than one age."

Borns (1966, p. 51) reaffirms his interpretation of deglaciation in northern Nova Scotia as one of "thinning, separation and stagnation." He also records discovering casts of the marine pelecypod *Portlandia glacialis* in the deposits, which provide substantive evidence that the terrace is glaciomarine. He elaborated on his previous pattern of deglaciation as follows. During deglaciation, the marine areas like the Bay of Fundy and its end basins, the Minas Basin and Chignecto Bay, became ice free sooner than the land areas because of more rapid melting and calving of the ice. This separated the ice on the Chignecto Peninsula from ice over New Brunswick and ice on mainland Nova Scotia. Ice contact deltas were deposited along the northern shore of the Minas Basin as the sea level rose against the wasting ice mass.

Borns (1966, p. 52) reiterates his age correlations for the outwash terrace (Port Huron; 11,850 to 13,400 yr BP) and ice wedge casts (Valders; 10,300 to 11,400 yr BP) which are partly supported by an average age of 10,585 ± 47 yr BP on 13 samples of charcoal at the Debert paleo-Indian site. However, there are no frozen ground features at the site to suggest the presence of permafrost during occupation (Borns, 1966, p. 53).

Borns' (1965, 1966) work was capped by a more detailed study by Swift and Borns (1967) and their paper may be summarized as follows. Swift and

Borns (1967) named the deposits of the terrace the Five Islands Formation and divided it into two members: 1. the upper, glaciofluvial Saints Rest Member, that disconformably overlies 2. the lower, marine Advocate Harbour Member. The Advocate Harbour Member is composed of two marine lithosomes, glaciodeltaic and glaciolittoral. The glaciodeltaic lithosome extends from Spencers Island to Five Islands, while the glaciolittorial lithosome occurs only at Advocate Harbour.

The glaciodeltaic lithosome is made up of deltas with a tripartite structure (topset, foreset and bottomset beds). The topset beds are coarse, imbricated fluvial gravels that unconformably overlie the foreset beds, which are finer grained and dip from 20° to 34°. The bottomset beds are rhythmites, coarsening upwards from clay/silt through clay/sand to clay/gravel interbeds.

The glaciolittoral lithosome at Advocate Harbour consists of several raised spits enclosing a lagoon, similar to the modern Advocate Harbour shoreline (Fig. 1.2). The internal structure of the spits is typical beach stratification (McKee, 1957), with foreshore beds dipping 5° to 14°S (seaward), and the sediment is generally gravel with sand matrix. Wave activity and longshore drift must have been similar to that of today, as the fossil and modern spits have a similar orientation.

The glaciofluvial lithosome (Saints Rest Member) of sandy gravel disconformably overlies the marine lithosomes and has sedimentary structures and textures typical of a shallow braided stream (Eynon and Walker, 1974). Kettles are numerous. Swift and Borns (1967) imply that this lithosome is present at all outcrops along the terrace (their Fig. 15, p. 709), and they envisage it is a separate, later event.

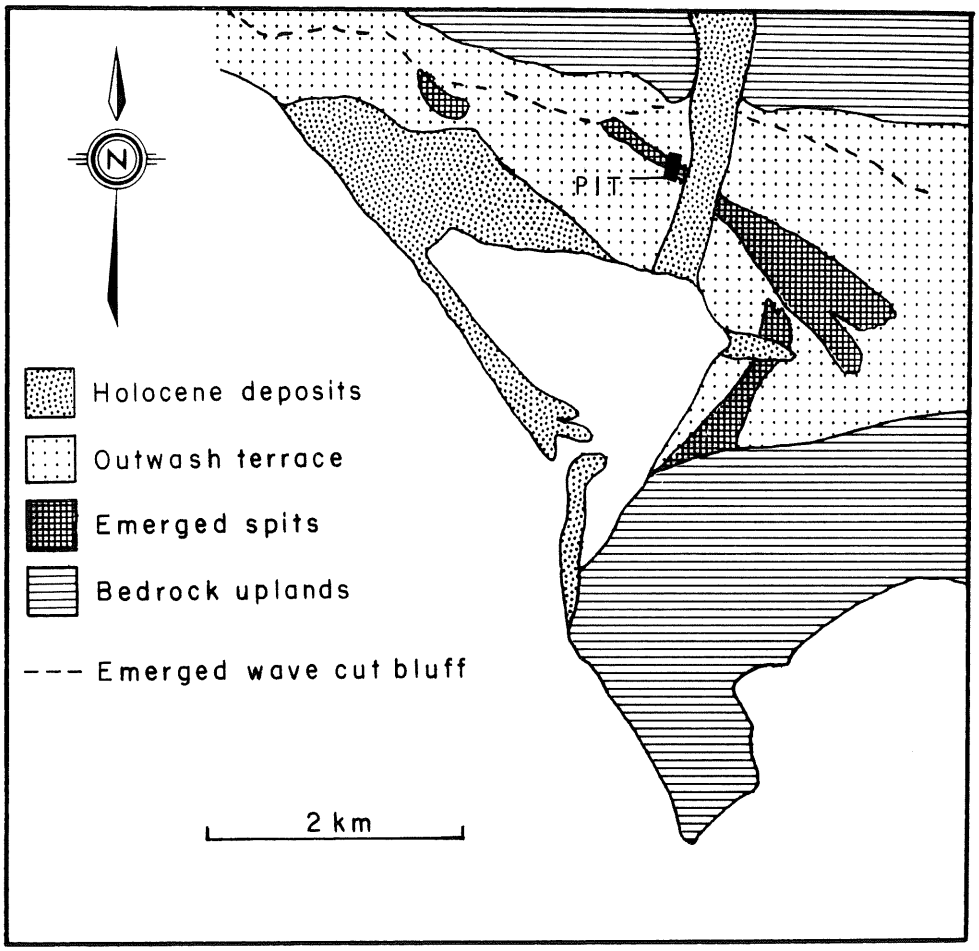


FIGURE 1.2. Surficial geology of Advocate Harbour area with location of gravel pit studied by Wightman (1976). (Modified from Swift and Borns, 1967.)

Although they had no direct evidence of age, Swift and Borns (1967) follow Borns' (1965, 1966) and place the deposition of the terrace between Port Huron time and Valders time of the classical sequence. They suggested the following sequence of events for the formation of the terrace. As the ice dissipated in the Minas Basin, it was followed by a rising sea level. Ice receded in the valleys on the north shore of the Minas Basin and was replaced by prograding deltas as far north as the Cobequid Fault (Fig. 1.3, top). The upper surfaces of the deltas rose with the rising sea level. The zone of rapid isostatic uplift, following the receding ice front, reached the northern shore of the Minas Basin and the deltas emerged. Dissection of the upper surfaces of the deltas produced a minimum of 6 m of relief. Subaerial alluvial fans then prograded across the dissected delta surfaces, producing the glaciofluvial lithosome (middle, Fig. 1.3). As the supply of outwash material diminished, the terrace continued to emerge, and underwent a second dissection, forming the present drainage system. When emergence slowed to a negligible rate, the sea advanced to its present position (Fig. 1.3, bottom).

Goldthwait's (1924, p. 150) marine limits for the north shore of the Minas Basin show that emergence increases towards the west (Fig. 1.4) Borns (1966, p. 54) and Swift and Borns (1967, p. 707) agreed with this in general but had more complete data and placed the zero isobase farther to the west at Bass River (Fig. 1.5). This means that the eastern end of the Minas Basin is submerged and has never had a higher sea level than at present. Swift and Borns (1967) used the contact between the Advocate Harbour Member and the Saints Rest Member to define the marine/nonmarine limit and measured the elevation of the contact from Saints Rest (0 m,

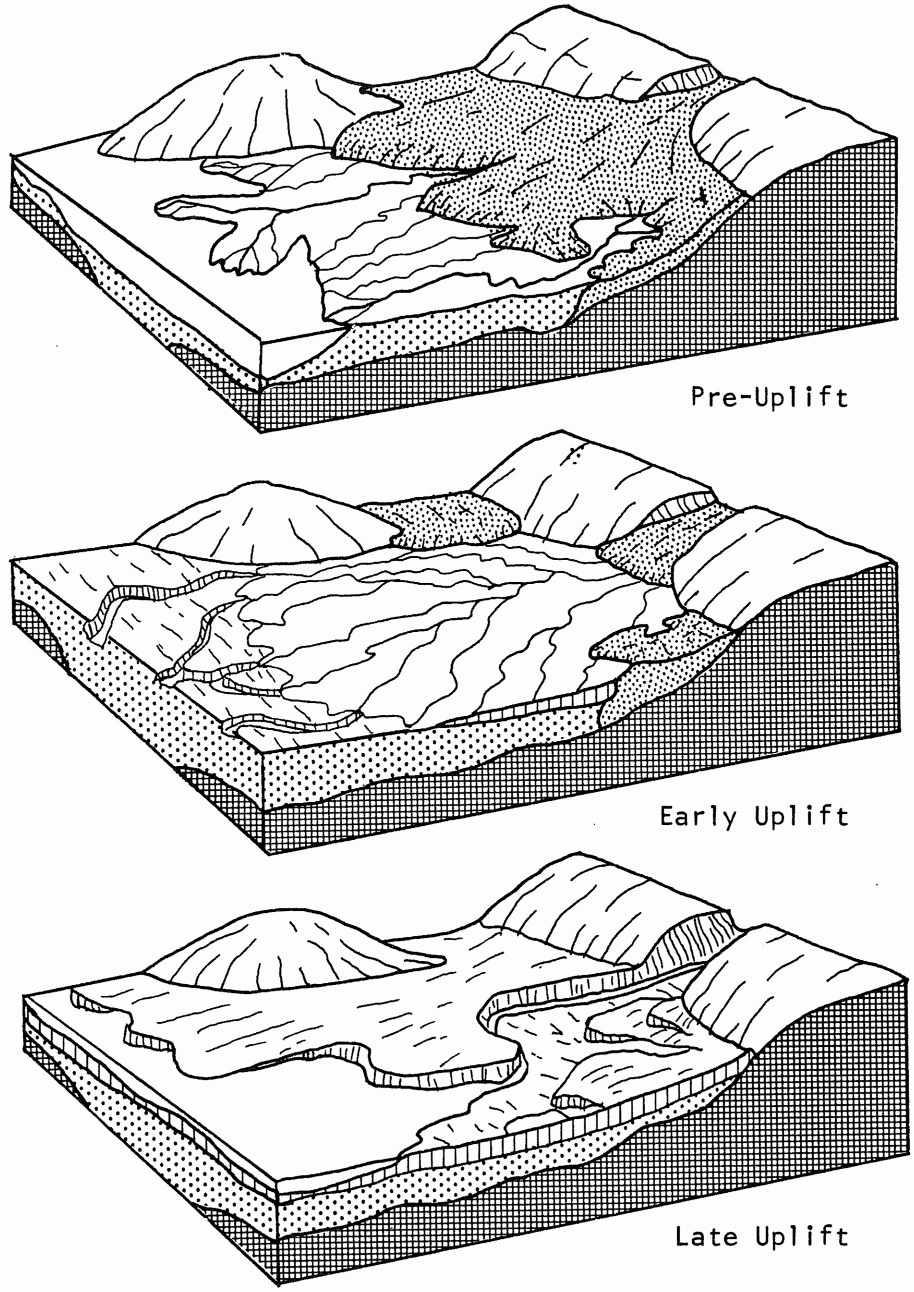


FIGURE 1.3. Evolution of the Minas north shore outwash terrace. Top, growth of marine deltas; Middle, uplift and erosion of marine plain — growth of subaerial fans based on Cobequid scarp; Bottom, modern terrace after uplift, dissection and sea level rise. (After Swift and Borns, 1967.)

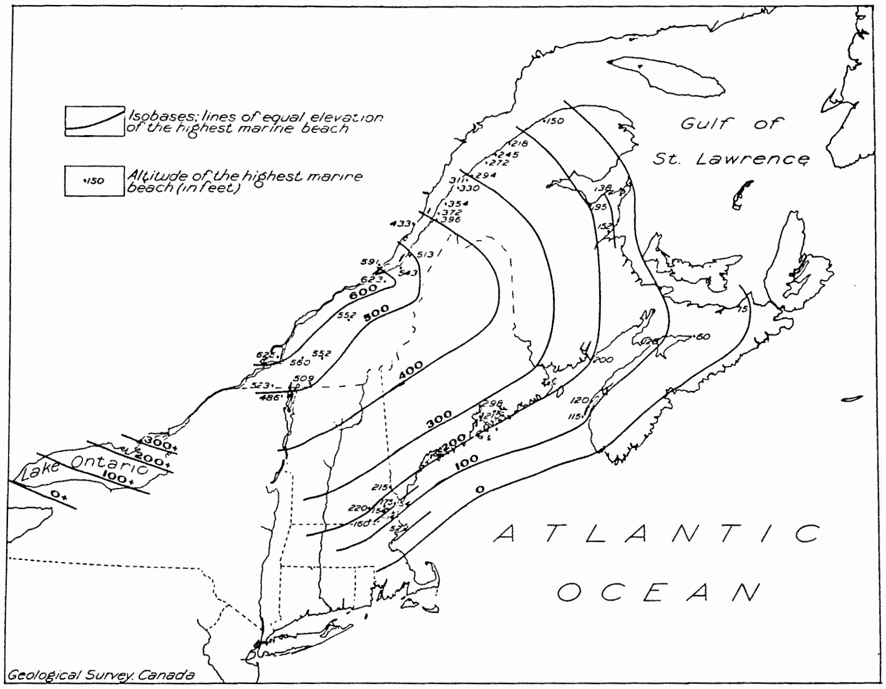


FIGURE 1.4. Postglacial upwarping shown by isobases. (From Goldthwait, 1924.)

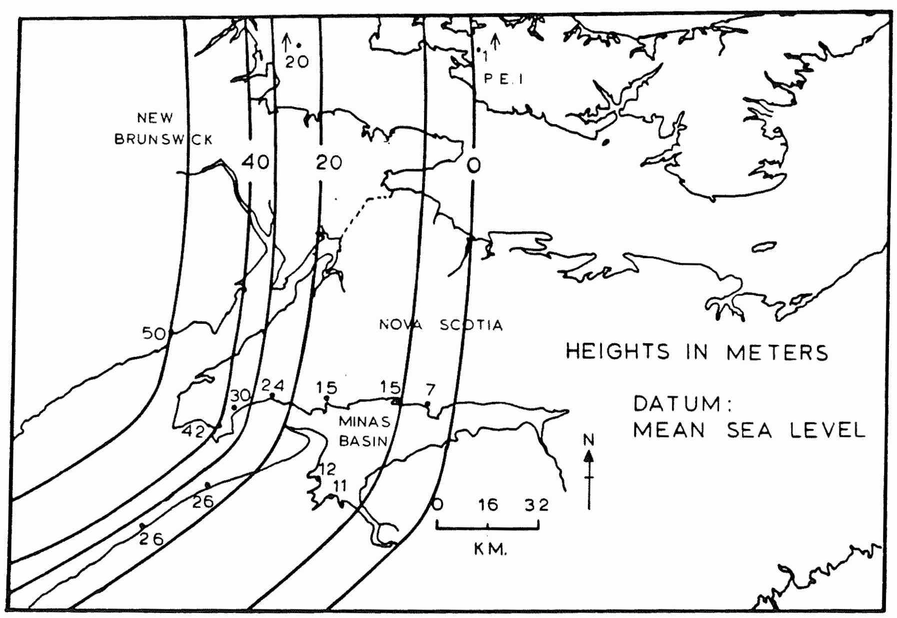


FIGURE 1.5. Isobases of postglacial emergence. Prince Edward Island data from Prest (1962, 1964); data for the south shore, Bay of Fundy from Hickox, 1962. (From Swift and Borns, 1967.)

mean sea level) to Spencers Island (33 m). At Advocate Harbour, the height of the emerged wave cut bluff was used to define postglacial emergence.

Swift and Borns (1967) made three important observations about their isobases of emergence: 1. the isobases give only a minimum amount for postglacial uplift as the sea has also been rising during the period since the deposition of the terrace; 2. the zero isobase is not the eastward limit of postglacial uplift as an undefined amount of uplift occurred before the formation of the terrace; and 3. postglacial tectonic movements, including those associated with deglaciation, affect the emergence data.

The raised beach at Advocate Harbour was the subject of a study that was largely sedimentological and dealt with a problem within, rather than on, the terrace (Wightman, 1976). Some of the results will be mentioned in later chapters. The outwash terrace is mentioned in several studies of psotglacial emergence (Welsted, 1976; Wightman and Cooke, 1978) but Swift and Borns' (1967) paper remains the most detailed study of the terrace prior to this thesis.

SPECIFIC PROBLEMS

Thesis field work was designed to concentrate on several problems that were outlined at the start of the project. As stated previously, the limited occurrence of the beaches in the terrace (only at Advocate Harbour) and especially the sedimentology of marine Gilbert type deltas interested the author. It was also decided to map the terrace deposits in hopes that the delineation of their exact extent and location, as well as the different types of deposits, would make possible a better understanding of the deglaciation of the Chignecto Peninsula. The relationship between the deposits and the drainage systems responsible for them was poorly understood and the maps would help to establish this. The delta at Parrsboro is a striking example, as the southward flowing Parrsboro River is relatively short (≈7.5 km) although the delta is quite large. This is a consequence of the drainage divide at Gilbert Lake in the Parrsboro Gap, which also creates a relatively long (≈25 km), northward flowing River Hebert. The bedrock source for the gravel in the deposits had not been delineated and this bears upon the length of the drainage system as well as on the direction of ice movement on the Chignecto Peninsula.

More detailed elevation data on the deposits were needed to improve understanding of the postglacial emergence of the terrace as well as the fluvial erosion to which Goldthwait (1924, p. 100) refers. Elevation data on the deltaic topset/foreset contact would make possible a better approximation of the marine limit along the north shore of the Minas Basin.

Finally, if no pelecypod shell material could be found in the deposits, coring of selected kettle ponds and lakes would be carried out so that pollen analysis and radiocarbon dating of the sediment would hopefully give a minimum age for the terrace.

GILBERT TYPE DELTAS

Coarse grained Gilbert type deltas comprise the bulk of the outwash deposits in the Minas terrace and an understanding of this type of delta is essential to the proper interpretation of the deglaciation of the north shore of the Minas Basin. In a model delta there is a tripartite structure, with topset, foreset and bottomset beds and it is the steep angle (10° to 35°) of the foreset beds that is characteristic (Fig. 1.6).

Braided streams deposit the topset beds, which form a layer of glaciofluvial outwash on the foreset beds. The main bedding planes are nearly horizontal, although the sediment within the beds can be cross-stratified. The topsets are the coarsest part of the delta and they thicken landward to maintain the necessary stream gradient. At the delta front, the sediment in the lower part of the channels is deposited below the level of the body of water into which the delta is building. The depth to which the stream sediment is deposited below the water level is dependent upon the depth of the channels but because the braided streams are relatively shallow, this depth is not great. Inland from the delta front where the bulk of the topset beds are deposited, the braided streams are above the water level so the topsets can be regarded as effectively subaerial deposits.

The topset/foreset contact marks the change from stream deposition to submarine deposition on the foreset slope and, as the braided streams are rapidly shifting, the topset/foreset contact often becomes a flat unconformity. Because the stream channels are shallow, the position of the topset/foreset contact is very close to that of the water level. For this reason, the topset/foreset contacts of the deltas in the Minas terrace are used to determine the marine limit (elevation of the maximum extent of postglacial marine submergence; Farrand and Gajda, 1962, p. 6), as suggested by Wightman and Cooke (1978, p. 62).

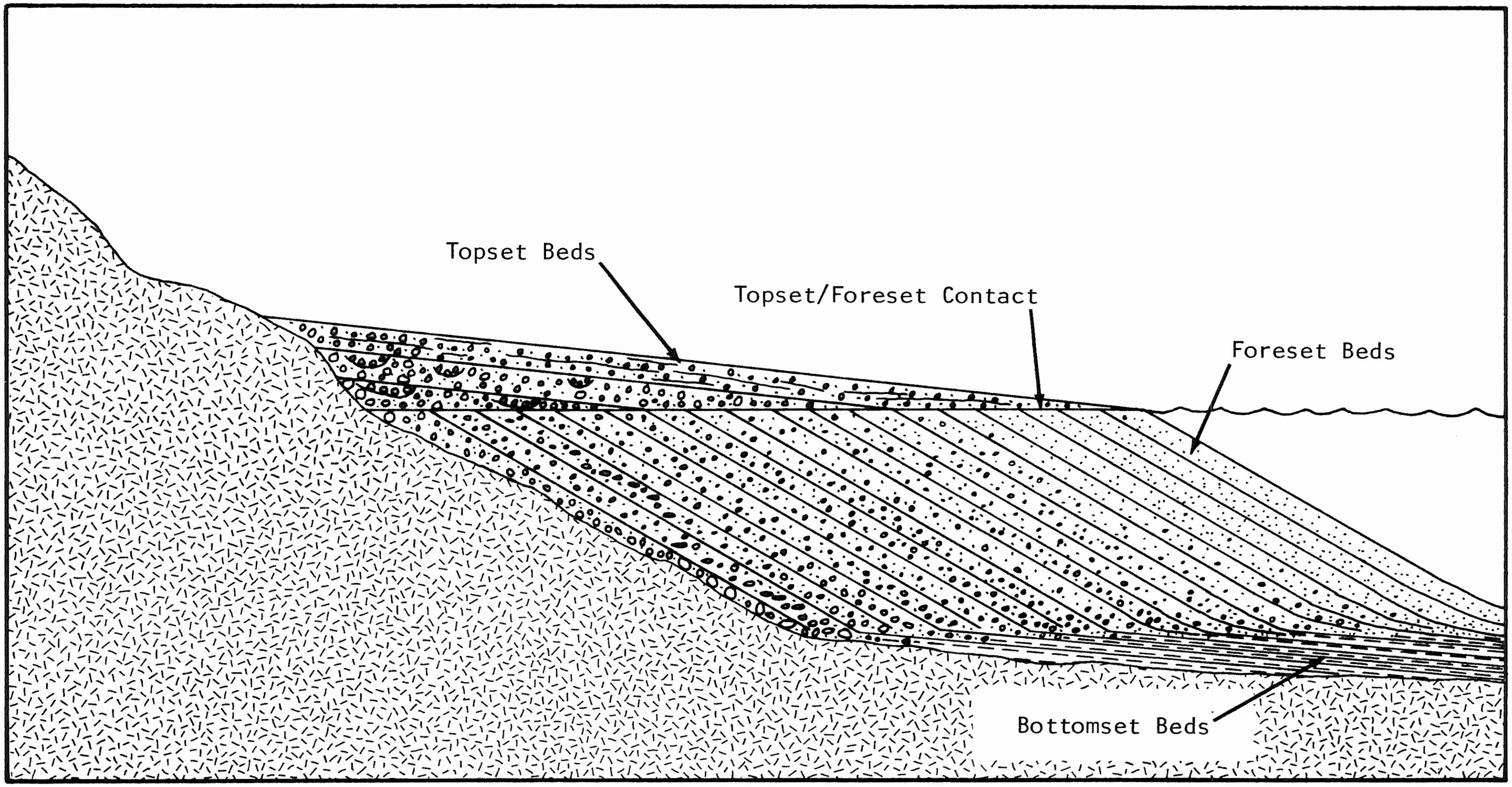


FIGURE 1.6. Schematic diagram of a model proglacial Gilbert type delta. (Modified after Gilbert, 1890).

The bulk of the deltaic sediment is deposited in the foreset beds, the dip of which represents the former angle of the delta front upon which bed load sediment was deposited. The thickness of the foreset facies is approximately equal to the depth of the body of water. The bottomset beds are fine grained, suspended load sediments that are deposited from the toe of the foreset slope seaward. They interfinger with the foresets and as the delta continues to build, the topsets and foresets prograde over the bottomsets.

Bates (1953, p. 2125) concluded that Gilbert type deltas are deposited where the inflowing water is equal in density to the water into which the stream is flowing and he termed this homopycnal inflow. However, Axelsson (1967, p. 17) points out that Gilbert type deltas also form in situations of unequal density inflow. It will be shown in later chapters that the deltas in the Minas terrace were formed by overflow (inflow less dense) conditions. The most important factor in forming steep foreset beds is an abundance of bed load sediment (Axelsson, 1967, p. 18; Friedman and Sanders, 1978, p. 501) and not the type of inflow. Therefore, Gilbert type deltas tend to be coarse grained and most were deposited by rapidly shifting, shallow braided streams in water significantly deeper than the stream channels (Axelsson, 1967, p. 26; Friedman and Sanders, 1978, p. 504).

AREA AND METHODS OF STUDY

The outwash terrace along the north shore of the Minas basin was studied from Portapique in the east to Advocate Harbour in the west, and is usually referred to in the thesis as the Minas terrace. At Parrsboro, the study continued inland to the north through the Parrsboro Gap to Newville Lake, then east to West Brook and, in less detail, north from Newville Lake to just beyond the Chignecto Game Sanctuary. The inland part of the study deals mainly with ice contact stratified features while the coastal part deals with glaciofluvial-glaciomarine outwash.

Field work was carried out in the fall of 1974 and in the summers of 1976, 1977 and part of 1978. The main part of the field work involved identification of the deglacial features, followed in many instances by detailed sedimentological descriptions. Samples were taken for grain size analyses and the data are listed in Appendix 1. Single measurements of clasts given in the thesis are of the a-axis unless stated otherwise.

MAPPING

After the deglacial deposits were identified, they were mapped using 1975, 1:10,000 colour air photographs and an Old Delft scanning stereoscope II for interpolation from field data. There are 10 maps (Fig. 1.7) 8 of which are included in the text of the thesis and the other 2 (D, E) are in the map pocket at the back of the thesis. Elevation and various other field data are plotted on the maps, including bedrock boundaries that have been taken from Donohoe and Wallace (1978). The limits of the deltas were established using the limits of the delta plain although the proximal parts of the plain may be composed only of glaciofluvial gravel and thus may not be deltaic in the strict sense. Continuous exposure is required to place the landward limit of the foreset beds and this is not available.

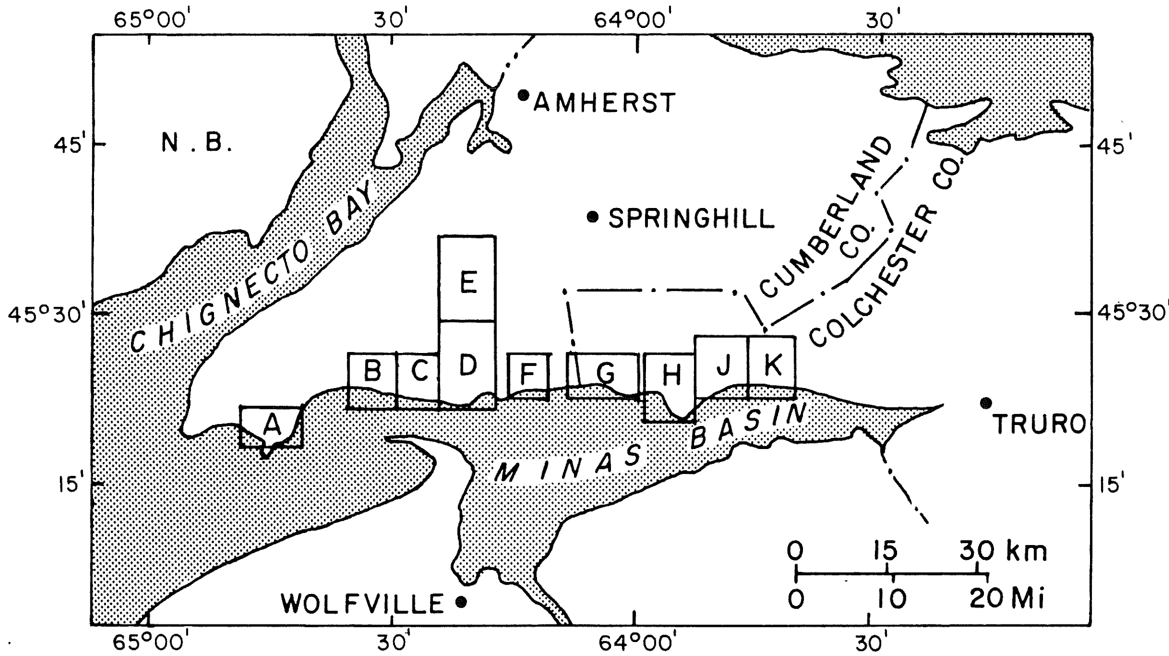


FIGURE 1.7. Location of surficial geology maps. A. Advocate Harbour, B. Port Greville, C. Diligent River, D. Parrsboro, E. West Brook, F. Moose River, G. Five Islands, H. Economy, J. Bass River, K. Portapique.

SPOT HEIGHTS

Elevations of features were taken using a Paulin micro altimeter, model M-1, incremented in 1 foot intervals. Corrections were made by making traverses in loops starting and ending at a bench mark of known elevation. Both Geodetic Survey of Canada and Nova Scotia Land Survey control monuments were used for bench marks and each traverse lasted approximately 45 minutes. After one area had been covered, a reading would be taken at a more appropriately located bench mark for a new area and the same pattern would be repeated. From the changes in the altimeter readings at the bench marks, correction curves were drawn for each day and corrections were made to the field readings by interpolation from the curves. The readings were converted to metres for plotting on the maps.

It is difficult to define the error margin on the individual readings as on certain days the air mass was very stable and there were only very small corrections. On other days an air mass would move in and the readings would change rapidly in a short time (up to 15 m in one hour). However, a fair degree of accuracy was attained and in some areas where readings happen to have been taken in close proximity to Nova Scotia Land Survey control monuments that were only discovered subsequently, the elevations differed by no more than 2/3 metre (2 feet). In general, accuracy to within ± 1.5 m is probably attained.

When elevations were taken of the topset/foreset contact of the deltas, the base readings were taken at the Higher High Water mark on the beaches. Elevations of the contact and the H.H.W. mark were taken within 5 minutes of each other and the agreement of the readings was within

± 1/3 metre. There may be some error in the elevation of the H.W.W. mark, derived from the Canadian Hydrographic Service tide tables, but it is probably quite accurate, to within ± 0.5 m. As the readings were double or triple checked, the accuracy of the topset/foreset contact elevations (marine limits) for the various areas is probably within ± 1.0 m. A transit and level were used to survey the marine limit at Spencers Island and various features at Advocate Harbour.

PEBBLE COUNTS

Pebble counts were done in the field at each of the deposits, so microscope work was not involved. Representative pebbles were collected by the author from various deposits along the north shore of the Minas Basin and were identified by Howard Donohoe and Peter Wallace of the Nova Scotia Department of Mines as to lithology and possible bedrock source. These pebbles were used as a reference suite to which other pebbles could be compared for identification. Pebble counts were then taken at each of the outwash deposits, as the grain size of the sediment varied from deposit to deposit. All pebbles with a sieve diameter greater than -2.5ϕ were counted. In some areas, the sediment was fine enough that pebbles coarser than -4ϕ were almost nonexistent but in most deposits the sediment was a coarse pebble gravel and the counts were predominantly on pebbles larger than -4ϕ.

Usually, 110 pebbles were counted at each sample site. There were 1 or 2 sample sites in each genetic unit of sand or gravel at each sample station (e.g. several sample sites in both foreset and topset beds). There were from 1 to 4 sample stations at each location depending upon the size of the exposure, with larger exposures generally along the shoreline and smaller exposures in borrow pits. For most deposits pebble counts were done at several locations (e.g. pebbles counted at shoreline exposure and at inland pit). As there are 9 deposits in the Minas terrace, a total of almost 8,500 pebbles were counted, excluding the deposits north of Parrsboro. The pebble counts, and how and where they were done, are listed in Appendix 2.

LAKE BATHYMETRY

A Kelvin Hughes MS26B 14 kHz echo sounder was used to chart the bathymetry and indicate the type of bottom sediment in Gilbert and Newville Lakes.

CORING

A 1 1/2" (3.8 cm) Livingstone corer modified after a model obtained from J.H. McAndrews of the Royal Ontario Museum, Toronto, was used to core several lakes and kettle ponds.

^{14}C DATING

Organic samples were sent for dating to the radiocarbon laboratories at the Geological Survey of Canada (Dr. Wes Blake Jr.), the Ministry of Natural Resources, Quebec (Dr. Pierre Lasalle) and Dalhousie University (Dr. J.G. Ogden III).

CHAPTER 2. PORTAPIQUE

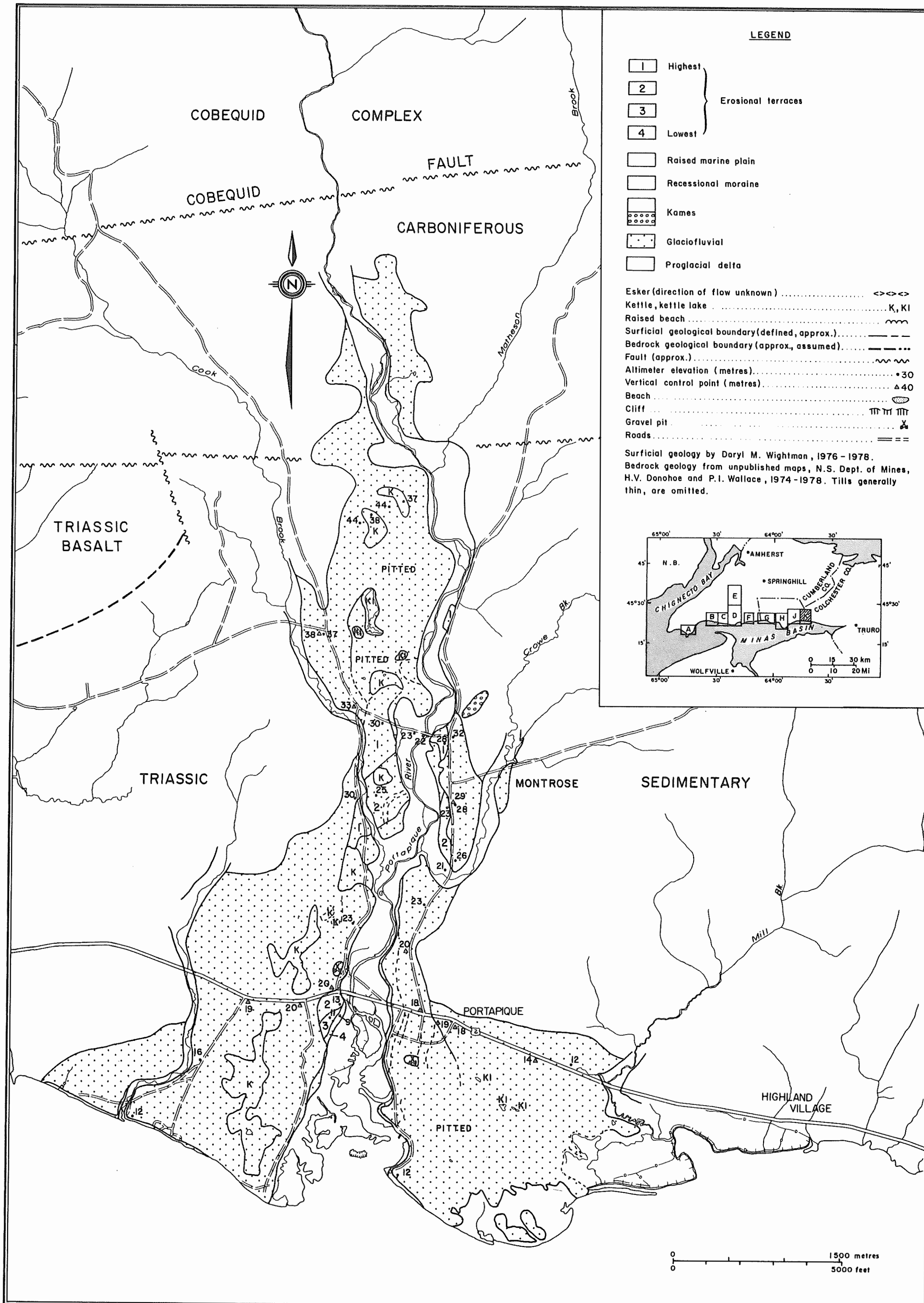
SURFICIAL GEOLOGY

Portapique is roughly 50 km east of Parrsboro and 35 km west of Truro. It is the most easterly of the deposits in the thesis study area and is the only deposit not reported upon by Swift and Borns (1967).

The deposit at Portapique, shown on map K, figure 2.1, is very large and extends inland farther than any of the other deposits, almost 9 km along the Portapique River valley. The deposit is approximately 5.5 km wide at the distal (southern) end and narrows inland to about 0.5 km at the proximal end. Triassic Blomidon and Wolfville Formations (undifferentiated) underlie all but the northern 1.5 km, which is underlain by the Upper Carboniferous Parrsboro Formation.

The deposit reaches at least 44 m in elevation to the north and slopes down to 12 m or less at the Minas shoreline. The flat upper surface of the deposit contrasts with the topography of the surrounding bedrock to define fairly sharply the limits of the outwash. In the south the surrounding bedrock has a low, rolling topography with hills 20 to 35 m high. Elevations increase to the north to 50 m but in the Carboniferous bedrock, they increase to 75 to 100 m, with rugged topography, while at the Cobequid Fault elevations rise sharply to a plateau at 250 m or more.

There are several incised terraces on the Portapique deposit, but they are difficult to map and correlate because of the relatively small differences in elevation between the terraces and the upper surface and between the terraces themselves. The terraces are also not extensive



Base map modified from N.S. Dept. of Lands & Forests 1=15,840 Map Series.

FIGURE 2.1. Map K. Surficial geology of the Portapique area, Colchester County, Nova Scotia

and the distance between them creates difficulties in correlation. If they have been correlated correctly, the terraces at and south of the road that crosses the valley at Montrose have an eastward as well as a southward slope.

Although there is no designated terrace on the east side of the Portapique River south of the highway, there does appear to be some erosion and several traces of paleochannels. There is also the trace of a paleochannel on the second order terrace west of the Portapique River at Montrose.

The upper surface of the deposit slopes seaward at a gradient of roughly 5.5 m/km (Fig. 2.2). At the southern end, where the deposit widens, there are also secondary slopes away from the higher central part along the Portapique River estuary. Although the deposit reaches an elevation of at least 44 m on its proximal end, the relief between the upper surface and the Portapique River is relatively constant at about 10 m. South of the highway the relief between the deposit and the estuary drops below 10 m as the deposit slopes down to a bluff of variable height (1 to 4 m) at the shoreline.

The most striking feature of the outwash at Portapique is the number of kettles, kettle ponds and pitted areas. Two large kettles occur on the west side of the Portapique River, one ($\simeq$1.4 km x 0.4 km) south of the highway and the other ($\simeq$0.8 km x 0.3 km) just north of the highway. Both kettles are very irregularly shaped but are elongated and aligned in a north-south direction. One of the smaller kettles occurs on a second order terrace west of Montrose. Even though the southern part of the deposit has numerous kettles, the upper surface is fairly

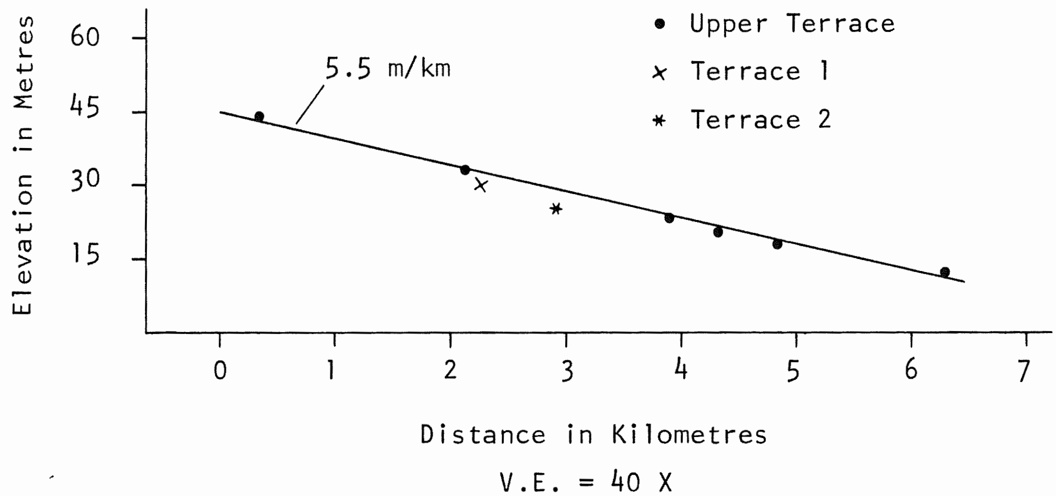


FIGURE 2.2. Gradient of the upper surface of the deposit at Portapique. Profile essentially north-south.

flat, characteristic of the deposits in the terrace. However, north of Montrose the upper surface is very rough and kettled. The kettles on the deposit at Portapique are different from those elsewhere in that they are preferentially elongated north-south, as compared to east-west at other deposits.

SEDIMENTOLOGY

The gravel at Portapique is poorly exposed along the shore as the bluff is low, with a large portion covered by talus, or is nonexistent. The best exposure is in DuPaul's gravel pit just north of the highway on the west side of the river. The pit is about 10 m deep and almost that thickness is exposed when the pit is being actively excavated (Fig. 2.3). The sediment is mainly gravel with a sand matrix. In the middle of the pit face, tabular cross beds are common, often graded (the effect of gravity [Reineck and Singh, 1975, p. 19]) and can be openwork. The cross beds are up to 1.5 m thick and are generally less than 10 m long. Dip of the beds os 20° to 25°S, indicating a paleoflow in that direction. There are also channel shaped depressions up to 10 m wide and 1 m deep, interpreted as paleochannels. The paleochannels are infilled by channel fill cross bedding (Reineck and Singh, 1975, p. 92) and/or tabular cross beds. There are a few sand beds (Fig. 2.4; sample 78-1, Fig. 2.5) but they are generally thin and discontinuous. The gravel at the top of the pit has a crude horizontal stratification. The gravel (sample 78-3, Fig. 2.5) has a sand matrix and is coarser than the gravel lower in the pit (sample 78-2, Fig. 2.5) that is frequently cross bedded. Sand beds are even less abundant in the coarse gravel at the top of the pit.

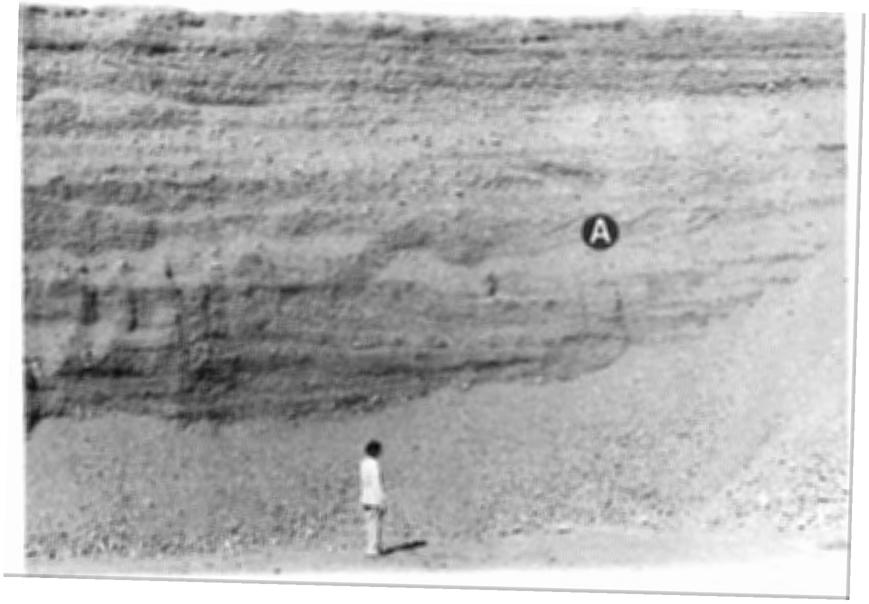


FIGURE 2.3. West face of DuPaul's pit, Portapique. Tabular cross beds (A) in central part of face. Gravel coarser and horizontally bedded at top.

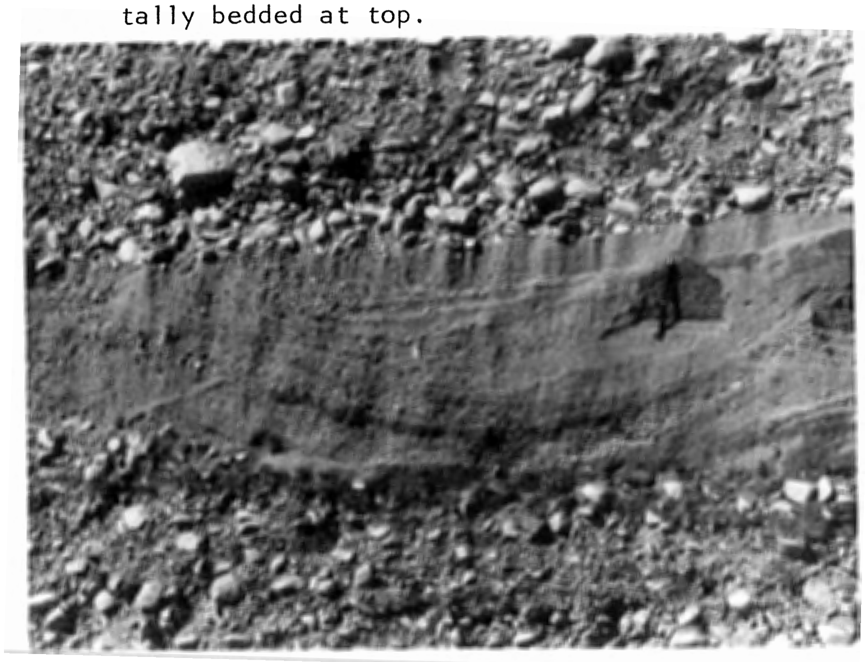


FIGURE 2.4. Sand lens in middle of west pit face, DuPaul's pit, Portapique. Sand has channel fill cross bedding. Sample 78-1 from sand by pen.

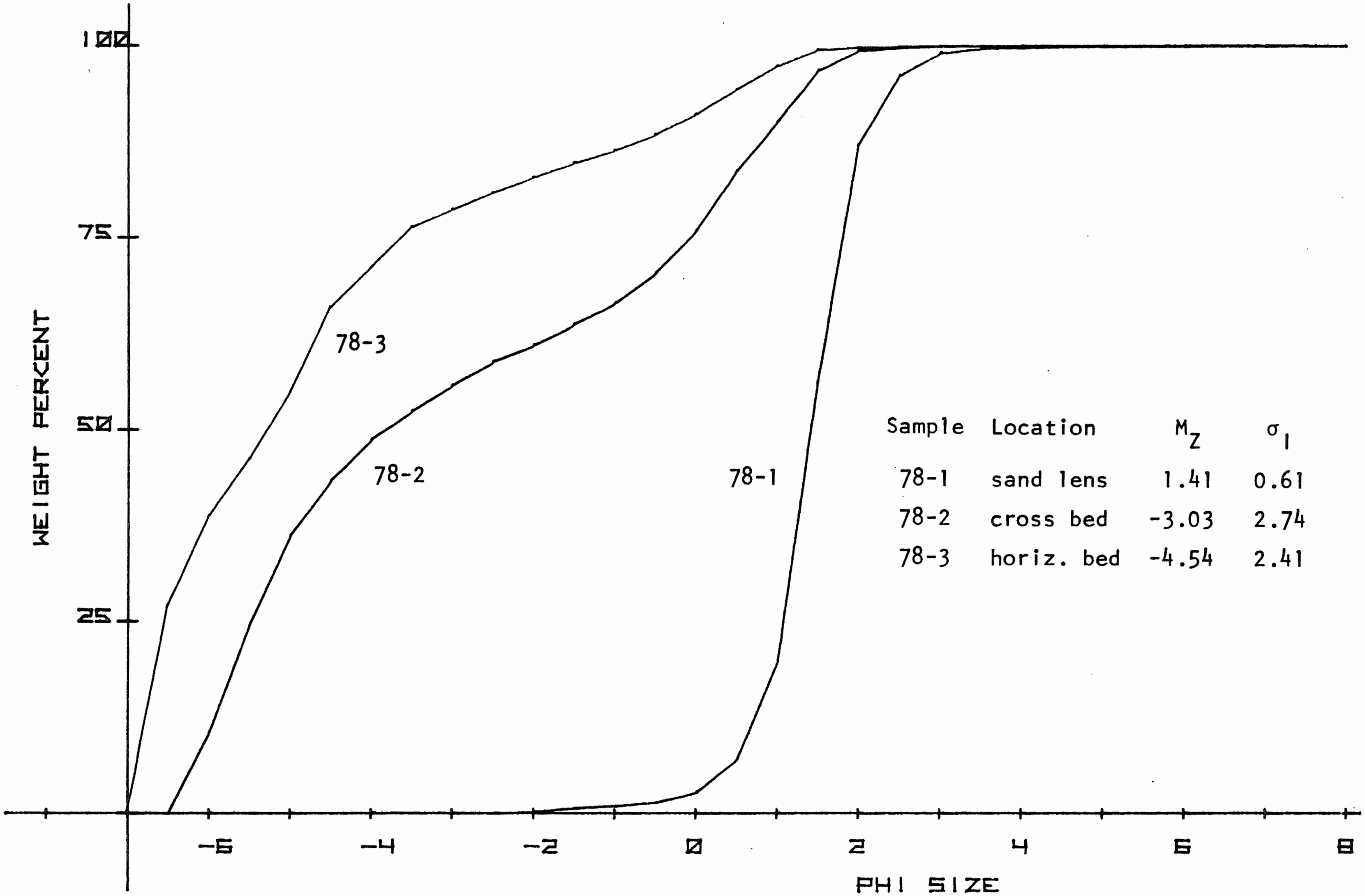


FIGURE 2.5. Grain size analyses, Portapique. M_Z = Graphic Mean, σ_I = Inclusive Graphic Standard Deviation for this and all subsequent grain size analyses.

INTERPRETATION

The gravel exposed in DuPaul's pit and along the shoreline bluffs is interpreted as glaciofluvial in origin, which is consistent with Swift and Borns' emergence data. The gravel has several features that Swift and Borns (1967, p. 697) list as criteria for recognizing the glaciofluvial Saints Rest Member: crude horizontal stratification, cut and fill structures (especially in the sand) and an abundance of kettles. These authors also imply that fluvial terraces (p. 698, 703) are characteristic of this lithosome and, although they are present on the Portapique deposit, they are not restricted to glaciofluvial deposits as the discussions of the deltas will show.

Swift and Borns (1967, p. 698) interpret the glaciofluvial lithosome as a braided stream deposit because of a relict pattern of channels on the surface. However, the coarseness of the sediment and lack of silt and clay (less than 1/2%, Fig. 2.5) are more diagnostic of this environment. Miall (1977, p. 7) states that "Glacial outwash streams are almost invariably braided," and many recent studies of modern outwash systems (Jopling and McDonald, 1975) confirm this statement. The conditions necessary for braiding, abundant coarse, non-cohesive bed load, strongly fluctuating and high discharge and steep slopes (Miall, 1977, p. 7) are characteristic of many glacial meltwater systems.

HORIZONTAL BEDS

The crude horizontal stratification of the gravel at the top of the pit is the most consistent type of bedding in outwash sediments (Church and Gilbert, 1975, p. 61). The glaciofluvial sediments in the Minas terrace corroborate this statement. The initial deposition of very coarse sediment and later entrapment of fines at lower flows leads to discontinuous horizontal bedding (Fahnestock, 1963, p. A22). This type of stratification indicates deposition in high discharge streams too shallow, even at flood stage, for slip faces to form (Rust, 1975, p. 246; Church and Gilbert, 1975, p. 61). Longitudinal bars that lack slip faces or a more immature form of 'diffuse gravel sheet' (Hein and Walker, 1977, p. 569) are bedforms that develop in this environment. Hein and Walker (1977, p. 569) suggest that horizontal stratification is related more to high discharge than to water depth. At high discharge, sediment is moved rapidly downstream and there is little time for vertical aggradation. At lower discharges, the gravel accretes vertically as well as moving downstream and slip faces form.

The thickness of the tabular cross beds below the rudimentary bedding (Fig. 2.3) indicates that streams were at least 1.5 m deep during the deposition of the cross beds. As ice was presumably melting back as it released sediment and meltwater, the overlying crudely bedded gravel was deposited in a slightly more distal position than the tabular cross beds. The streams might be expected to be somewhat shallower in a more distal position but studies on modern outwash systems (Boothroyd and Ashley, 1975, p. 218; Hein and Walker, 1977, p. 565) show little change in channel depth from the lower part of the upper fan to the lower part of the mid fan. It will also be shown by hydrodynamic calculations that stream depths were probably in the 1 to 2 m range during the deposition of the crudely bedded gravel. Thus, the streams should have been deep enough for slip faces to form and the horizontal stratification is probably related to more rapid downstream sediment or to the slightly coarser grain size. The crudely bedded gravel at the top of the pit is probably an aggraded sequence of diffuse gravel sheets (Hein and Walker's 1977 model).

TABULAR CROSS BEDS

Most tabular cross beds are deposited on the slip faces of bars. Miall (1977, p. 14) restricts the development of slip faces to linguoid (transverse) bars. Others (Boothroyd and Ashley, 1975, p. 220; Gustavson *et al.* 1975, p. 279; Collinson, 1978, p. 21) do not restrict the development of slip faces or tabular cross beds to linguoid bars and state that longitudinal bars can also have slip faces. Miall (1977, p. 14) further states that linguoid bars (and hence tabular cross beds) are most typical of sandy braided rivers. However, Hein and Walker (1977, p. 565) observed foreset slopes (slip faces) in the midstream and downstream reaches of the Kicking Horse River. The gravel was finer (mean clast size = -4.6ϕ, midstream; -3.7ϕ, downstream) in these reaches than in the upstream reach (mean clast size = 5.6ϕ) where slip faces were absent. Discharge was lower in the midstream and downstream reaches and there was sufficient time for the gravel to aggrade and develop slip faces. Their model predicts crude horizontal bedding in the coarse, proximal areas and both cross and horizontal beds in the finer, more distal areas (Hein and Walker, 1977, p. 569).

Studies of post-Late Wisconsinan outwash reveal that tabular cross beds of gravel are not uncommon in outwash of this age, and various depositional mechanisms have been proposed for the cross bedding.

McDonald and Banerjee (1971, p. 1298) attribute tabular cross beds of pebble gravel to the downstream migration of transverse bars. Eynon and Walker (1974, p. 58) suggest that gravel foreset beds (up to 4 m thick) were deposited on the downstream side of a bar core formed by previous erosion of the outwash. Costello and Walker (1972, p. 397) suggest that large tabular cross beds of gravel (up to 3 m thick) were deposited on the slip faces of mid channel bars. Rust (1975, p. 247) argues that the large foreset beds described by Costello and Walker (1972) could not have formed on an outwash plain as even at flood stage, the streams would not be deep enough. He suggests that the foreset beds were deposited in ice melt out depressions (kettles) but the sedimentological relationships described by Costello and Walker (1972) do not fit that interpretation very well.

The depositional mechanism for the tabular cross beds at Portapique was not evident as talus covered the beds on the upstream end. However, the beds are present and they are composed of gravel. While tabular cross beds or foreset slopes may not be common in the gravel parts of modern outwash systems, it appears that they are more common in the outwash deposited at the end of the last glaciation. This implies that there are differences in processes between the modern and paleo-outwash systems that have been studied.

VERTICAL SEQUENCE

The textural relationship between the cross beds and the crude horizontal beds is consistent with Walker and Hein's (1977) model; the cross beds are composed of finer gravel ($M_z = -3.03\phi$) than the horizontal beds ($M_z = -4.54\phi$). However, their position in the vertical sequence is the reverse of what is predicted. Most studies (Boothroyd and Ashley, 1975, p. 218; Church and Gilbert, 1975, p. 66) show that grain size decreases downstream and therefore, in a retreating ice front situation, the gravel higher in the vertical sequence would represent a more distal depositional environment. Three possible reasons for the inverse relationship at Portapique are: 1. the ice advanced after the deposition of the cross beds and the crudely bedded gravel was deposited in a more proximal environment than the cross beds; 2. as the ice retreated it released coarser sediment in the fluvial transport range and, although the downstream fining continued, coarser sediment was supplied to each location along the system; or 3. the upper gravel was deposited during higher discharges than the lower gravel. Few data exist to support any of these possibilities but there is no evidence of an ice advance anywhere along the terrace. As for differences in discharge, hydrodynamic calculations (discussed below) show that the difference in the maximum stream velocity for the cross beds and for the horizontal beds is small. Deposition of the sand matrix necessitates fluctuations in stream velocities both for the middle and upper parts of the sequence and fluctuations in discharge are also a requirement for braiding. Thus, a difference in the size of sediment supplied to the system is favoured by the process of elimination of the other alternatives but there are no positive data to support this conclusion.

HYDRODYNAMIC CONSIDERATIONS

GRAVEL

The shape of the cumulative curves and poor sorting of samples 78-2 and 78-3 (Fig. 2.5) imply that the samples are composed of two better sorted end members. As has been suggested, the sand matrix is probably the result of infiltration at low flow. The samples were split at -1ϕ and replotted as separate samples (Fig. 2.6) to confirm this. The gravel fractions are very coarse ($M_z = -4.71\phi$, $M_z = 5.44\phi$) and the sand fractions are coarse grained. Hydrodynamic calculations were made to estimate various stream parameters and to test the theory of sand infiltration at lower flows. The results are given in table 1.

The equations used to calculate U, the average stream velocity, and Fr, the Froude number listed in table 1 are as follows. Blatt *et al.* (1972, p. 87) related τ_o, the critical shear stress, to U as

$$\tau_o = \frac{f\rho U^2}{8} \qquad (1)$$

where f is the friction factor and ρ is the fluid density. Rearranging the equation to solve for U gives

$$U = \left[\frac{8\tau_o}{f\rho}\right]^{1/2} \qquad (2)$$

where the flow depth, D, is known, the Froude number is related to stream velocity (Blatt *et al.*, 1972, p. 84) as

$$Fr = \frac{U}{(gD)^{1/2}} \qquad (3)$$

where g is the acceleration due to gravity (980 cm/sec^2).

Samples 78-2 and 78-3 have sharp cut offs on the coarse end of the distribution, at approximately -5.5ϕ and -6.5ϕ respectively as seen in

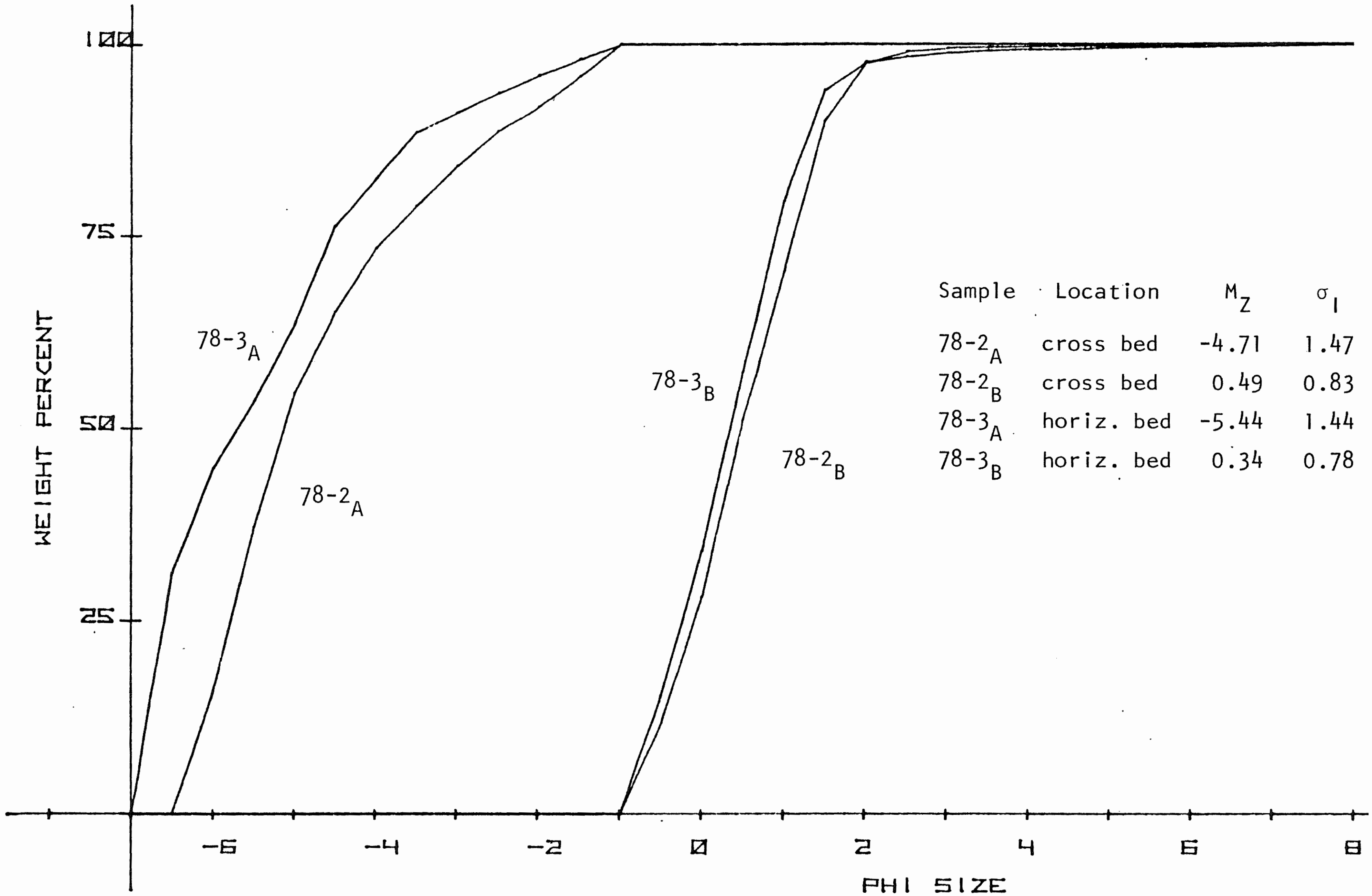


FIGURE 2.6. Grain size analyses, glaciofluvial gravel, Portapique. Replot of samples 78-2 and 78-3 split at -1ϕ.

TABLE I. Hydrodynamic calculations for the glaciofluvial sediment at Portapique.

Bedding type	Gravel size (ϕ)	τ_o (dynes/cm2)	K		Flow depth (m)	f	U (m/sec)	Fr	U_{slope} (m/sec)	Fr_{slope}	U* (cm/sec)	W		$U\tau_o$ (ergs/cm^2/sec)	U_{power} (m/sec)	Fr_{power}	$\tau_{o(power)}$	
			ϕ	mm								(mm)	(ϕ)				(dynes/cm^2)	(mm)
horiz.	-3.5	110	-5.4	42.2	2	.05	1.3	.29	-	-	10.5	0.81	0.30	-	-	-	-	-
horiz.	-6.5	830	-5.4	42.2	1	.067	3.1	.99	2.5	.80	28.8	2.3	-1.20	-	-	-	-	-
horiz.	-6.5	830	-5.4	42.2	2	.05	3.6	.81	4.2	.95	28.8	2.3	-1.20	-	-	-	-	-
x-bed	-3.0	74	-4.7	26	2	.042	1.2	.27	-	-	8.6	0.70	0.51	-	-	-	-	-
x-bed	-5.5	430	-4.7	26	2	.042	2.9	.66	4.5	1.02	20.7	1.5	-0.58	-	-	-	-	-
sand	1.0	2.7	1.41	0.38	0.2	.0235	0.3	.21	1.9	1.36	-	-	-	4000	1.1	.79	36.4	4.2
sand	1.0	2.7	1.41	0.38	1	.0159	0.4	.13	5.2	1.66	-	-	-	4000	1.3	.42	31.7	3.8

figure 2.6. The fine ends of the gravel components dropped off at about -3.0ϕ and -3.5ϕ, respectively. For comparison, calculations were made for each sample at both the coarse and the fine ends of the gravel distribution. The critical shear stress necessary to initiate bed load movement, τ_o, is given for different grain sizes in Blatt *et al.* (1972, p. 91). To obtain f, the friction factor, the relative bed roughness K/D must be calculated. The height of the projections on the bed, K, can be estimated as the mean grain size of the gravel fraction. The flow depth, D, was taken as 2 m for the cross beds. The beds were up to 1.5 m thick so the streams had to be deeper than that to deposit the beds. The flow depth was taken as 2 m for the horizontal beds as well, but for the coarse fraction a calculation was also made for D = 1 m to show what the effect of shallow depth would be on moving the coarse fraction. A Moody diagram (Streeter, 1966, p. 260) can then be used to obtain f.

A second equation independent of grain size was used to calculate stream velocity and serves as a check on the previous calculations. The average stream velocity is related to the stream slope by the equation (Blatt *et al.*, 1972, p. 87)

$$U_{slope} = \left[\frac{8g}{f}\right]^{1/2} (DS)^{1/2} \qquad (4)$$

where S is the stream slope expressed as a dimensionless number. The stream slope can be approximated by the gradient of the upper surface, 5.5 m/km (Fig. 2.2).

To compare the size of the clasts to the size of the matrix, the shear velocity U* must be calculated. For a given value of τ_o (Blatt *et al.*, 1972, p. 87)

$$U^* = \left[\frac{\tau_o}{\rho}\right]^{1/2} \qquad (5)$$

The shear velocity must be equal to, or greater than, the settling velocity (w) of the sediment in order to keep it in suspension (Harms *et al.*, 1975, p. 138). Thus, the diameter of a quartz sphere with a settling velocity, w, equal to U* gives an estimate of the maximum diameter of the sediment that will be kept in suspension. Blatt *et al.* (1972, p. 54) give settling velocities for quartz spheres and in table 1, w is the grain size of the sediment that has a settling velocity equal to U*.

As shown in table 1, the stream velocity required to initiate bed movement in -6.5ϕ gravel (horizontal bed) at a flow depth of 2 m is 3.6 m/sec. Under those conditions, the suspended load is very coarse as granule gravel can be maintained in suspension. Changing the flow depth to 1 m only drops the velocity to 3.1 m/sec. Calculations for the -3.5ϕ gravel over the same bed give a stream velocity of 1.3 m/sec and sediment up to coarse grained sand will be kept in suspension. Therefore, stream velocities had to drop to 1.3 m/sec or lower for the coarse grained sand to be deposited as matrix in the horizontal beds. This is still a very high flow and the good sorting of the coarse grained sand (Fig. 2.6) reflects this as finer sediment was kept in suspension. The stream velocity necessary to initiate bed load movement for the coarse gravel (-5.5ϕ) in the cross beds is calculated to be 2.9 m/sec, which is 0.7 m/sec lower than the velocity for the horizontal beds.

The velocities calculated using the stream slope are in fairly close agreement with the velocities calculated using the grain size. Where the flow depth was assumed to be 1 m for the horizontal beds, the slope velocity (2.5 m/sec) drops below the velocity calculated using grain size. The slope velocity is too low to move sediment of the observed size and therefore the flow depth (1 m) is probably incorrect. Where U_{slope} is larger than U there is no problem because U is the minimum velocity required to move bed load of that grain size. As is shown in table 1, changing the depth has a marked effect on U_{slope}.

The Froude numbers are quite high and even exceed unity for U_{slope} for the tabular cross beds. Flows are supercritical at Froude numbers greater than 1 and subcritical at Froude numbers less than 1 (Blatt *et al.*, 1972, p. 84). Supercritical flows would not be expected over long distances because of their instability. The stream velocity of 3.6 m/sec for the horizontal gravel gives a Froude number of 0.81 which is high enough that antidunes may have formed (Boothroyd and Ashley, 1975, p. 220).

It is interesting to compare the flow parameters estimated in table 1 to measured flow parameters for the Scott outwash fan (Boothroyd and Ashley, 1975, p. 197). Over the lower part of the upper fan and mid fan (a distance of 10 km), velocities ranged from 0.40 to 1.55 m/sec at flow depths of 0.10 to 0.70 m, giving Froude numbers from 0.27 to 0.95. On the upper fan (over a distance of 2.5 km) the velocities ranged from 1.75 to 3.0 m/sec at flow depths of 0.70 to 1.0 m. The accompanying Froude numbers varied from 0.75 to 1.24. Taking flow depth into consideration, there is quite good agreement between the estimates in table 1 and the flow parameters for the upper mid fan and lower part of the upper fan on the Scott outwash system.

SAND

The same calculations discussed in the gravel section were made for sample 78-1 (Fig. 2.5), taken from the finer part of the sand lens (Fig. 2.4). The results are listed in table 1. The sand has a mean grain size of 1.41ϕ, which was used for obtaining bed roughness, and a coarser limit of about 1ϕ. The flow depth (0.2 m) was obtained from the depth of the sand lens.

The calculated values for U and U_{slope} are widely discrepant. The stream velocity calculated from the grain size (0.3 m/sec) is probably too low as the finer side of the lens was sampled. The stream velocity calculated from the slope (1.9 m/sec) appears to be unrealistically large. Conspicuously absent from the sand are ripples, even in the overlying horizontal sand (Fig. 2.4). The grain size of the sand is finer than 0.6 mm so ripples would form in the lower flow regime (Blatt *et al.*, 1972, p. 119, 121). Because stream power and flow regime are related, the stream velocity required for upper flow regime conditions can be calculated. Stream power can be defined as (Blatt *et al.*, 1972, p. 121)

$$\text{stream power} = U\tau_o \qquad (6)$$

For sand 0.38 mm in diameter, the transition to upper flow regime conditions occurs at about 4000 ergs/cm^2/sec (Allen, 1970a, p. 79). If (1) is multiplied by U

$$U\tau_o = \frac{f\rho U^3}{8} \qquad (7)$$

and $U\tau_o$ is the stream power (6). Rearranging the equation to solve for U gives

$$U = \left[\frac{8}{f\rho}\right]^{1/3} \quad (U\tau_o)^{1/3} \tag{8}$$

Setting the stream power equal to 4000 ergs/cm^2/sec gives a velocity U_{power} of 1.1 m/sec and a Froude number, Fr_{power}, of 0.79 (Table 1). The Froude number is close to the value for antidune formation. Thus, the sedimentary structure of the sand implies that velocities were in the 1.1 m/sec range, which is between the values calculated for U and U_{slope}. The lack of sediment finer than medium grained attests to the winnowing strength of the current. This velocity can be checked by solving for τ_o using (6). The value ($\tau_{o\ power}$ = 36.4 dynes/cm^2) is the critical shear stress for sediment 4.2 mm in diameter (Blatt *et al.*, 1972, p. 91). The small pebbles and granules on the south side of the lens (Fig. 2.4) are consistent with this value.

For comparison, the same calculations were made for the sand using a flow depth of 1 m as some of the flat bedded sand may have been deposited in deeper water. The discrepancy between U and U_{slope} increases because U_{slope} has a very high value of 5.2 m/sec. The value of U_{power} (1.3 m/sec, using the lower to upper flow regime boundary) once again appears to be more reasonable and is only 0.2 m/sec more than that calculated for a flow depth of 0.2 m. Thus, changing flow depth greatly affects U_{slope} but has less effect on velocities calculated using grain size or stream power. However, the Froude number drops to 0.42. The corresponding shear stress ($\tau_{o\ power}$) is sufficient to move granule gravel.

In summary, although the sand was deposited at low discharge when stream power was too low to move the coarse gravel, the flow velocities were still probably quite high ($\simeq$1 m/sec). The overall range of estimated velocities is about 1 to 4 m/sec. If lower velocities existed, the preservation potential of their deposits must have been low.

OVERVIEW

The abundance of kettles indicates that the outwash was deposited in close proximity to dissipating ice. The kettles show that some ice persisted until after deposition had ceased on the upper surface and, in one instance, on a terraced surface. The particularly rough surface north of Montrose is unlike the surfaces of other deposits in the Minas terrace and this outwash must have been deposited just before the supply of sediment ceased. By the time the ice had melted, there was little sediment to infill or smooth the surface.

The slope of the upper surface, dip of the tabular cross beds and location of the deposit indicate that meltwater was discharging to the south along the Portapique River valley. The sedimentological data suggest a braided outwash environment with coarse gravel (up to -7ϕ) and flows from 1 to 2 m deep reaching velocities of 4 m/sec. There is no indication of delta foresets or bottomsets in the exposed sediments. The deposit is graded towards the Minas Basin and postdepositional marine erosion has undoubtedly removed much of the distal end of the deposit. One section of the present shoreline bluff has been observed to recede in the order of a metre per year over the period from 1975 to 1978.

CHAPTER 3. BASS RIVER AREA

SURFICIAL GEOLOGY

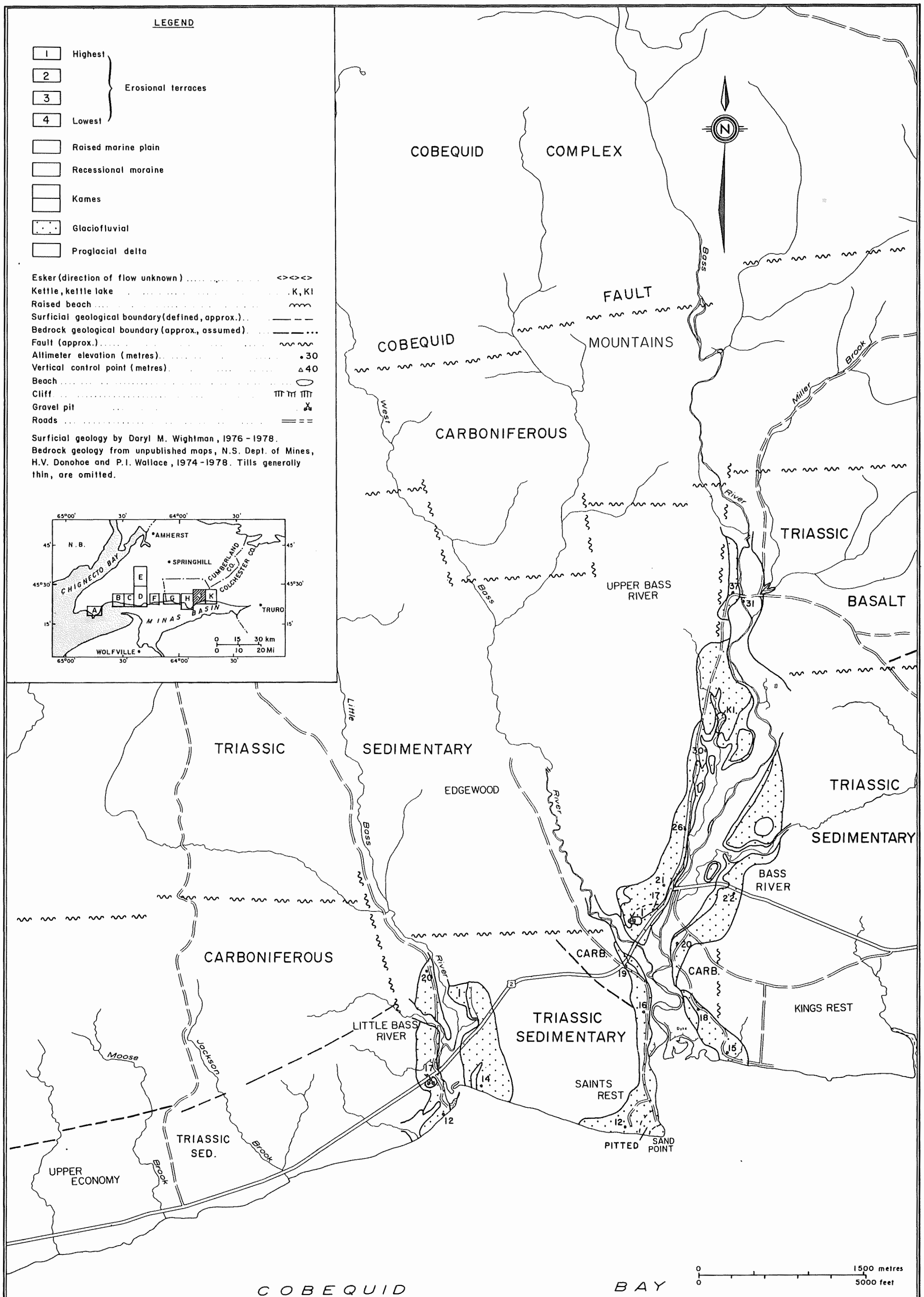
Bass River is almost equidistant from Parrsboro, 43 km to the west and Truro, 40 km to the east at the head of the Minas Basin. It is approximately 10 km west of Portapique. There are two deposits of glaciofluvial gravel in the Bass River area: the main deposit at Bass River that extends well inland, north of the confluence of Miller Brook and Bass River and a small deposit at Little Bass River (Fig. 3.1).

BASS RIVER

At Bass River, the deposit is narrow but it extends farther inland than the deposits to the west, although not as far as the deposit at Portapique. Along the Minas shoreline it occurs on both sides of the mouth of the Bass River but the combined width is less than 1 km. However it extends inland (north) along the Bass River valley for a distance of almost 5.5 km, keeping a relatively constant width (less than 1 km) except near the northern limit where it narrows to less than 0.25 km. In several places it is broken into isolated 'buttes' by stream erosion.

The gravel overlies the Triassic Blomidon and Wolfville Formations (undifferentiated) for most of its length. The exceptions are at its northern extremity where it may overlie Triassic North Mountain Basalt and for roughly 1 km centred on the confluence of the West Bass River and Bass River where it overlies the Upper Carboniferous Parrsboro Formation.

The deposit reaches an elevation of 37 m on its proximal (north) end and slopes down to 12 m at its distal end. The deposit is confined



Base map modified from N.S. Dept. of Lands & Forests 1:15,840 Map Series.

FIGURE 3.1. Map J. Surficial geology of the Bass River area, Colchester County, Nova Scotia.

to the Bass River valley and its limits are fairly easily recognized because of the flat nature of the upper surface of the gravel and the rolling topography of the surrounding hills. The hills near the shoreline generally increase in elevation to the west as the belt of soft Triassic red beds narrows and at Bass River they are slightly higher (over 45 m) than at Portapique. North of the deposit the elevations rise sharply in the rugged Carboniferous hills to 150 m or more. At the Cobequid Fault the land rises sharply again to a plateau north of the fault at elevations of 275 m or more.

The upper surface of the glaciofluvial gravel slopes seaward with little east-west component. The seaward gradient is 5.0 m/km (Fig. 3.2) and there is relatively little scatter amongst the data. Except for the eroded central portion over which the Bass River flows, there is only one small terrace, 3 to 4 m below the upper surface, in central Bass River just north of a gravel pit. The pitted surface of the outwash west of Sand Point corresponds with the kettles that Swift and Borns (1967) mention at Saints Rest.

Although the deposit reaches at least 37 m in elevation to the north, there is only a 6 m drop to the Bass River, which flows over bedrock. South of this where the upper surface is approximately 30 m in elevation, there is roughly a 15 m drop to the Bass River. However the relief between the Bass River and the upper surface is roughly 10 m or less for much of the deposit.

LITTLE BASS RIVER

The deposit at Little Bass River is exposed along the Minas shoreline on both sides of that river, is less than 0.5 km wide at each

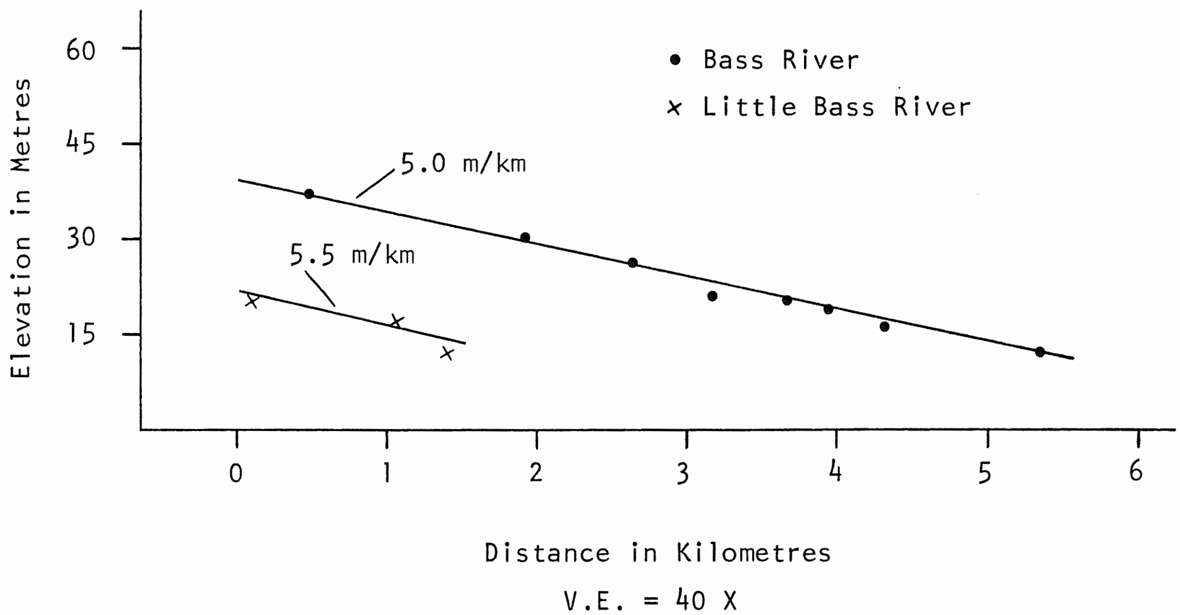


FIGURE 3.2. Gradients of the upper surfaces of the glaciofluvial gravel at Bass River and Little Bass River. Profiles essentially north-south.

exposure and extends inland for approximately 1.5 km. It overlies the Triassic red beds as at Bass River. The deposit reaches 20 m in elevation at its proximal (north) end and slopes down to 12 m at its distal end. The surrounding bedrock reaches elevations of 30 m or more and the bedrock/gravel contact is easy to place. Hills in the Carboniferous immediately to the west reach heights of 90 m or more.

The seaward gradient of the upper surface of the gravel is 5.5 m/km (Fig. 3.2) but this is not a reliable figure as the deposit is small, data are sparse and the data are relatively scattered. A terrace, apparently of the first order, occurs on the east side of the river north of the highway but no elevation was obtained for it.

SEDIMENTOLOGY

BASS RIVER

At Bass River, the main exposures are along the shore west and north of Sand Point, Saints Rest. This is the type area for the Saints Rest Member (glaciofluvial) of the Five Islands Formation (Swift and Borns, 1967, p. 703). North of Sand Point the bluff is 3 to 5 m high and the exposure is better than west of Sand Point where the bluff is a little lower (1 to 4 m high). In the central part of the northern exposure the gravel, from 1.5 m to 4 m thick, erosionally overlies soft Triassic red beds, up to 2 m thick (Fig. 3.3). The red beds thicken to the north but end abruptly with a sharp erosional contact. At the truncation, the upper metre of the red beds is in contact with gravel but the lower metre abuts a partly washed till which thins out to the north over several metres. Just north of the bedrock truncation, a large clast of Triassic red bed (Fig. 3.4), which was really a clast of sand rather than sandstone, occurred in the gravel.

The gravel is similar in texture and structure to the crude horizontal beds at Portapique. It is closedwork (sand matrix) and the crude bedding, imparted by changes in grain size and by discontinuous sand beds (Fig. 3.4), ranges in thickness from 10 to 50 cm. The horizontally stratified sand beds, up to 30 cm thick and tens of metres long, are more abundant in the upper part of the gravel. There is one occurrence of tabular cross beds, 0.5 m thick and several metres long, that dip 20°S. The overall grain size is similar to that of the horizontal beds at Portapique but the gravel is slightly coarser at the bedrock contact.

LITTLE BASS RIVER

One of the best exposures of fluvial structures in the entire terrace is along the shoreline west of the Little Bass River. The exposure is 4 to 5 m high but in the central section there is a red till, up to 1 m thick, at the base. The gravel is closedwork and generally finer grained than at Saints Rest, although some boulders up to about 25 cm in diameter are present, particularly just above the till where they are probably an erosional lag. Unlike the outwash at Portapique and Bass River, sand beds are common.

The striking difference between the gravel here and elsewhere along the terrace is the abundance of cut and fill structures. The structures are interpreted as paleochannels and are infilled by channel fill cross bedding, tabular cross beds, massive sand or gravel or some combination of these. The cross stratification dips south to southwest, indicating paleoflow in that direction. The sand beds (5 to 30 cm thick) are more

FIGURE 3.3. Eastward facing exposure of glaciofluvial gravel overlying Triassic red beds, Saints Rest. Red beds are truncated to north. Fence pickets on side of bank attest to rapid shoreline erosion.

FIGURE 3.4. Clast of Triassic red bed (at trowel) in outwash at Saints Rest. Trowel easily removes sand. Discontinuous sand beds (coarse) in middle to upper part of gravel.

abundant in the lower part of the exposure (Fig. 3.5) and form sandy intervals over 1 m thick. Most of the sand is parallel laminated or bedded but some is cross stratified.

The gravel in the lower two thirds of the exposure is mainly cross stratified with tabular cross beds. The cross beds infill channel shaped depressions with the coarsest gravel at the bottom of the cross bed, but can have sandy toesets (Pettijohn *et al.*, 1973, p. 351) that continue up the opposite side of the paleochannel. Some paleochannels are cut in sand and infilled with gravel (Fig. 3.5). One set of tabular cross beds is not exposed at the bottom but is over 1 m thick and appears to infill a paleochannel. The largest exposed paleochannel, approximately 5 m across and 1 m deep, has a very coarse, massive gravel infill (Fig. 3.6). The gravel along the top of the exposure is commonly coarse and massive to crudely horizontally stratified (Fig. 3.6), similar to the glaciofluvial gravel at Saints Rest and Portapique.

INTERPRETATION

BASS RIVER

As has been mentioned previously, the gravel exposed along the bluff at Saints Rest is the 'type' section for the glaciofluvial Saints Rest Member. The structure and texture of the gravel are similar to those at Portapique and the depositional environment is interpreted to be essentially the same - shallow (1 to 2 m), braided streams with high flow velocities. The predominance of massive to crude horizontal bedding implies rapid downstream sediment movement.

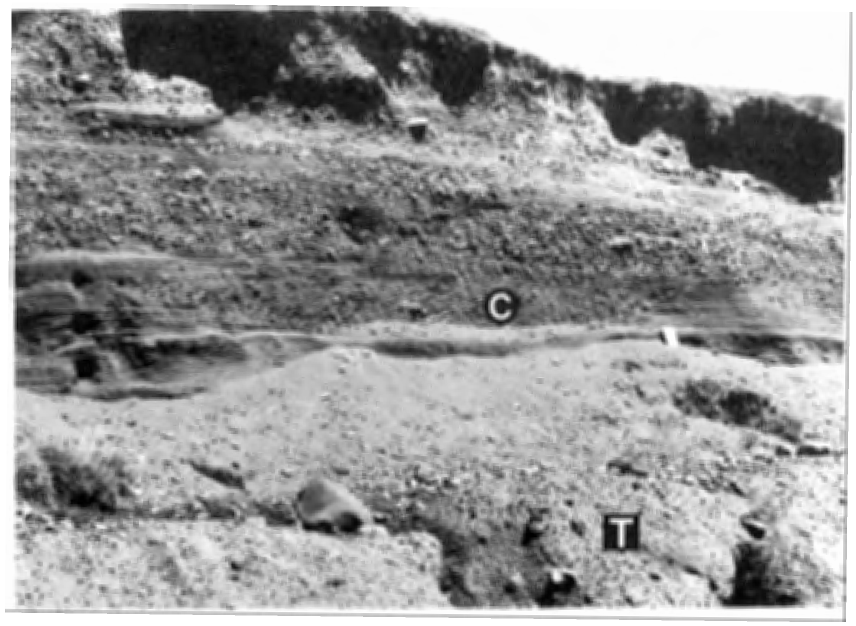


FIGURE 3.5. Gravel channel deposit (C) cut in sand. Channel infill consists of tabular cross beds. Sand, abundant in bottom half of exposure, is both cross and horizontally stratified. Till (T) in lower part of photograph.

FIGURE 3.6. Large channel deposit (C) composed of coarse, massive gravel.

The clast of Triassic red bed (Fig. 3.4) in its present state could not withstand transportation and deposition with the surrounding gravel. There appears to be two probable explanations for its emplacement: 1. the clast was frozen during transportation and deposition, or 2. the clast has been delithified by postdepositional processes. As the lithification of the clast varies little from the lithification of the Triassic red beds at Saints Rest and elsewhere (they are friable to unlithified), it seems most likely that the clast was frozen during transportation. This interpretation is also consistent with the rarity of Triassic clasts or pebbles. If the red beds had been competent enough to withstand erosion and deposition, Triassic pebbles would be common in the gravel as it is the underlying bedrock.

LITTLE BASS RIVER

The gravel exposed at Little Bass River is also glaciofluvial in origin but is different from glaciofluvial gravel at Saints Rest and Portapique, or elsewhere in the Minas terrace. The abundance of sand in the lower part of the sequence, the fineness of the gravel and the numerous cut and fill structures are all uncommon.

The predominance of sand could be the result either of stream power being too low to move coarse gravel or of little gravel being supplied to the system. Most of the sand is parallel laminated so upper flow regimes were common but the presence of dune cross bedding and absence of ripple cross stratification means that stream power did drop into the upper part of lower flow regime conditions (Blatt *et al.*, 1972, p. 121).

Although the gravel in the middle of the sequence is finer than the gravel at Portapique and Saints Rest, there are large pebbles scattered throughout the gravel. This suggests that the critical shear stress, τ_o, was sufficient to move gravel of that size but the supply of pebble gravel was low. The one channel deposit of very coarse gravel (Fig. 3.6) shows that at least some streams were strong enough to transport coarse sediment. The coarser gravel at the top of the sequence is similar in structure and texture to that at Saints Rest and Portapique and similar processes are interpreted for its transport and deposition.

The cut and fill structures are composed of tabular cross beds, channel fill cross beds, or both. Channel fill cross beds are the result of sediment accretion along the sides and bottom of the channel and do not necessarily represent lower flow regime sedimentation. The tabular cross beds in the sand were probably deposited by the downstream migration of linguoid or transverse bars (Miall, 1977, p. 14). Tabular cross beds composed of gravel are supposed to be rare so their abundance here is of particular interest. For this reason, the gravel cross beds are discussed in the following, separate section.

ORIGIN OF GRAVEL CROSS BEDS, LITTLE BASS RIVER

As stated previously, gravel cross beds are more common in Late Pleistocene outwash sequences than in modern braided streams. Rust (1975, p. 246) and Miall (1977, p. 34) invoke special circumstances, such as a confined valley, to increase flow depth sufficiently to develop gravel slip faces. Their discussions, as well as that of Hein and Walker (1977, p. 569) involve cross beds deposited from the downstream migration of inchannel bars. The outwash at Little Bass River shows that another mechanism was important in depositing gravel cross beds.

All of the gravel cross beds at Little Bass River infill paleochannels, the shape of which show that the interchannel/channel margins were relatively steep. Gravel was transported across the top of, and deposited on the downstream side of, the interchannel bars. The steep slope of the downstream side of the bars (upstream side of the channels) acted as a depositional slip face and the cross beds prograded across the channels until the channels were infilled. This obviously required flow depths that exceeded the channel depths. Moreover, the flow depth over the interchannel areas had to be sufficient to transport gravel. The very coarse gravel (Fig. 3.6) was deposited as a massive channel infill so that stream power was probably insufficient to transport gravel of that size over the interchannel bars. Similarly, Boothroyd and Ashley (1975, p. 219) observed that the coarsest gravel was moved in the deeper parts of the system. Thus, large areas of the outwash must have been flooded at high discharge when the cross beds were deposited. At lower discharges, the flows would occupy the unfilled channels.

This process of gravel cross bed deposition depends upon the existence of relatively steep sided, well defined channels and large fluctuations in discharge. Many channels in the gravel sections of modern outwash systems appear to be shallower and broader with less concavity than those at Little Bass River (Fahnestock, 1963, p. A17; McDonald and Banerjee, 1971, p. 1287; Rust, 1975, p. 246). The paleochannels at Bass River are cut into sand or fine gravel which might be a factor in their shape. Large fluctuations in discharge are necessary because it seems difficult to imagine the channels being cut at high flow, when flooding and channel infilling were prevalent. The channels were probably cut at lower, but still strong, flows.

This process of interchannel bars serving as nuclei for the deposition of cross beds is similar to that proposed by Eynon and Walker (1974). They observed foresets up to 4 m high deposited on the downstream side of a bar core. The bar core was an eroded remnant caused by the incision and erosion of streams on the outwash plain (Eynon and Walker, 1974, p. 57). At Little Bass River, the amount of time for incision and erosion appears to have been less, so only incision, with little lateral migration, of the channels was accomplished.

Because the cut and fill structures are more numerous and better developed at Little Bass River than elsewhere in the terrace, the finer grain size may have been important in this process. The sand and fine gravel may have permitted channel incision at lower flows that would not have been accomplished, or at least achieved with greater difficulty, in coarser gravel. The fineness of the gravel allowed sediment transport over the interchannel bars while gravel as coarse as that at Saints Rest or Portapique was probably limited to the deeper parts of the system.

OVERVIEW

BASS RIVER

There are fewer kettles in the gravel at Bass River than at Portapique but the kettle pond and pitted area on the upper surface indicate that the outwash was still deposited in close proximity to dissipating ice.

The slope of the upper surface, the dip of the tabular cross beds and the location of the deposit show that meltwater was discharging to the south along the Bass River valley. The similarities in structure and texture of the sediment to that at Portapique imply a similar depositional environment of braided streams. The glaciofluvial gravel north of Sand Point directly overlies bedrock so it is not topset gravel there and also is probably not topset gravel to the north of Saints Rest. Swift and Borns (1967, p. 708) state that the contact between the Advocate Harbour and Saints Rest Members lies halfway down the beach (approximately mean sea level) at Saints Rest. It is possible that west of, and at, Sand Point the glaciofluvial gravel is underlain by deltaic sediments but this was not observed.

LITTLE BASS RIVER

There are no kettles or clasts of unlithified sediment to indicate proximity to ice or freezing conditions, but it is assumed that the outwash was deposited in conditions similar to those of Portapique and Bass River.

Sedimentological data imply deposition by braided streams that probably flooded the surface of the deposit at high discharge. The dip of the tabular cross beds and slope of the upper surface of the deposit indicate a southward meltwater discharge out of the Little Bass River valley, although the deposit does not extend as far inland along the valley as do the deposits at Bass River and Portapique. The western part of the deposit overlies till or bedrock along the shoreline so the gravel does not represent the topset facies of a delta.

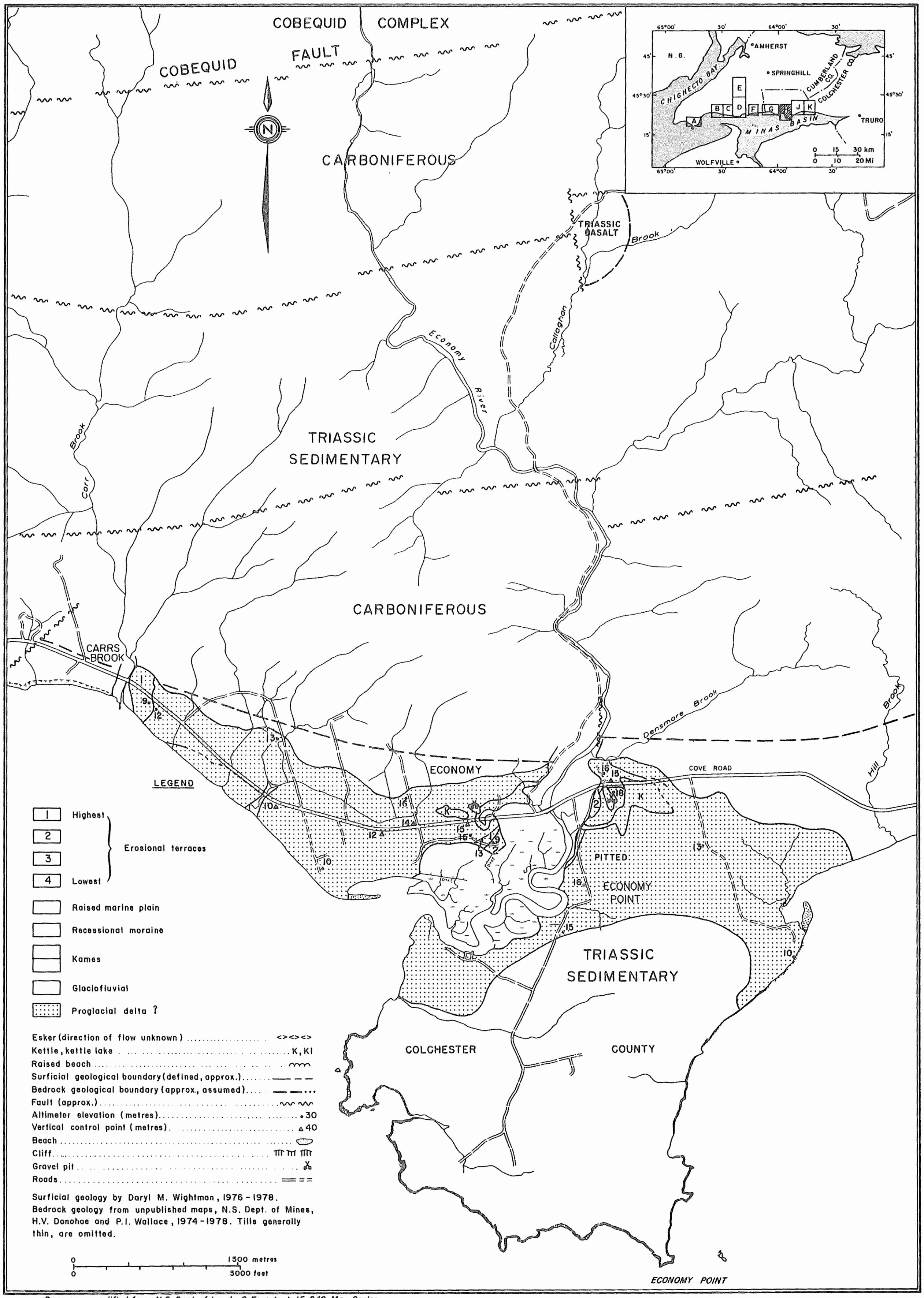
CHAPTER 4. ECONOMY

SURFICIAL GEOLOGY

Economy is located about 10 km west of Bass River and 35 km east of Parrsboro. The surficial geology at Economy is somewhat enigmatic (Fig. 4.1). Gravel pits and the low shoreline bluffs, generally 1 to 3 m high, expose glaciofluvial gravel but circumstantial evidence, dealt with later, points to a deltaic origin for the deposit.

The deposit overlies, and is bounded by, low lying Triassic Blomidon and Wolfville Formations (undifferentiated). North of the deposit for all but the extreme western part, the gravel thins out over the gently sloping red beds and the contact is difficult to place. The gravel reaches 16 m above sea level in several places along its northern limit and the contact is at about this elevation. The boundary between the soft Triassic red beds and the Carboniferous Parrsboro Formation is at about 30 m and the gradient steepens abruptly as the Carboniferous hills reach elevations of over 135 m. At the extreme western end of the deposit, the gravel abuts the Carboniferous to the north (Fig. 4.1). Economy Peninsula to the south of the deposit is comprised of low (below 30 m) hills of Triassic bedrock and the outwash/bedrock contact (at ≃15 m) is also difficult to place.

The upper surface of the deposit generally slopes seaward, but on Economy Peninsula there is an eastward slope to the surface. The low elevation of the surface, the somewhat conjectural boundaries and the narrow, elongated (along the shoreline) shape of the deposit make the calculation of exact gradients meaningless. The gradients (from 2 to



Base map modified from N.S. Dept. of Lands & Forests 1:15,840 Map Series.

FIGURE 4.1. Map H. Surficial geology of the Economy area, Colchester County, Nova Scotia.

4 m/km) generally steepen near the bedrock boundaries (e.g. north of the highway west of Economy River) and are shallow near the shoreline. Several small terraces are incised below the surface (Fig. 4.1).

The surface of the deposit east of Economy Point Road is quite irregular and pitted. North of this pitted surface is a large irregularly shaped kettle (longest dimension ≈500 m). The kettle has been breached on the southwest corner by stream erosion leading to the Economy River estuary. A smaller (300 m x 75 m) kettle occurs just north of the highway in Central Economy and an irregular indentation just east of this in the Economy River valley may be the result of a partially eroded kettle.

SEDIMENTOLOGY

The best shoreline exposures are on the east side of Economy Point at the end of Cove Road (2 to 3 m bluff) and along the shore from the Economy River estuary to Carrs Brook (1 to 2 m bluff). Some gravel is exposed in a small pit (owned by G. Morrisson) in Economy and better exposure exists in a larger pit just east of Economy Point Road.

The sediment exposed in the pits and along the shoreline is gravel with a sand matrix, with few sand beds and lenses. The gravel is mainly massive but in places there is a rough horizontal stratification. Southward dipping tabular cross beds up to 1 m thick are present in the gravel pit east of Economy Point Road. Although gravel is exposed in the walls of Morrisson's pit, Mr. G. Morrison (1977, personal communication) stated that the whole pit is underlain by sand which is at least 4 m deep in one part (depth of a water well). A small shovel dug hole in the pit floor exposed sand beds that dipped 25°SW.

GLACIOFLUVIAL OR GLACIODELTAIC ORIGIN(?)

The gravel exposed along the bluffs and in the pits at Economy is similar in structure and texture to the gravel exposed at Saints Rest and Portapique and is therefore interpreted as glaciofluvial in origin. Swift and Borns (1967, p. 708) state that the contact between the Saints Rest and Advocate Harbour Members is exposed at the base of the bluff (top of the beach, ≈8 m above sea level) along the shoreline between Economy and Carrs Brook. Vic Prest (1978, personal communication) also stated that foreset beds were visible along this beach. From 1976 to 1978, the berm of the beach was well developed and the underlying sediment, as well as most of the low bluff, was completely covered.

The sand at the bottom of Morrisson's pit was stratified similar to foreset sand (dip of 25°SW), but the exposure was not large enough to permit any firm interpretation. The gravel/sand transition at the base of the pit is similar to gravel/sand transitions at the topset/foreset contacts of known deltas in the terrace. Moreover, the base of the pit has an elevation of 10 m (altimeter) which is very similar to the elevation of Swift and Borns' marine limit along the shore.

Thus, it is probable that the glaciofluvial gravel exposed along the shoreline is underlain by foreset beds, in which case the glaciofluvial gravel is topset gravel and the deposit is a delta. However, the interpretation is not based on direct observation by the author and relies on reports by others. This is the reason for the question mark behind the term 'proglacial delta' on the map, figure 4.1.

OVERVIEW

The pitted surface east of Economy Point Road and the two kettles give evidence of ice blocks, surrounded by outwash, that melted after deposition on the upper surface had ceased. A possible third kettle, mentioned in the surficial geology section, gives a rough east-west alignment to the kettles which might represent a former ice front position. Thus, some of the deposit could have been laid down while the ice front was within the present limits of the deposit.

The glaciofluvial gravel that is exposed in the pits and shoreline bluffs is interpreted as a braided stream deposit that probably represents the topset facies of a delta. The main source of meltwater discharge appears to have been the Economy River valley as the upper surface of the delta decreases in elevation away from it. Drainage from some of the smaller brooks to the west of Economy River may have influenced the delta to some degree.

CHAPTER 5. FIVE ISLANDS AREA

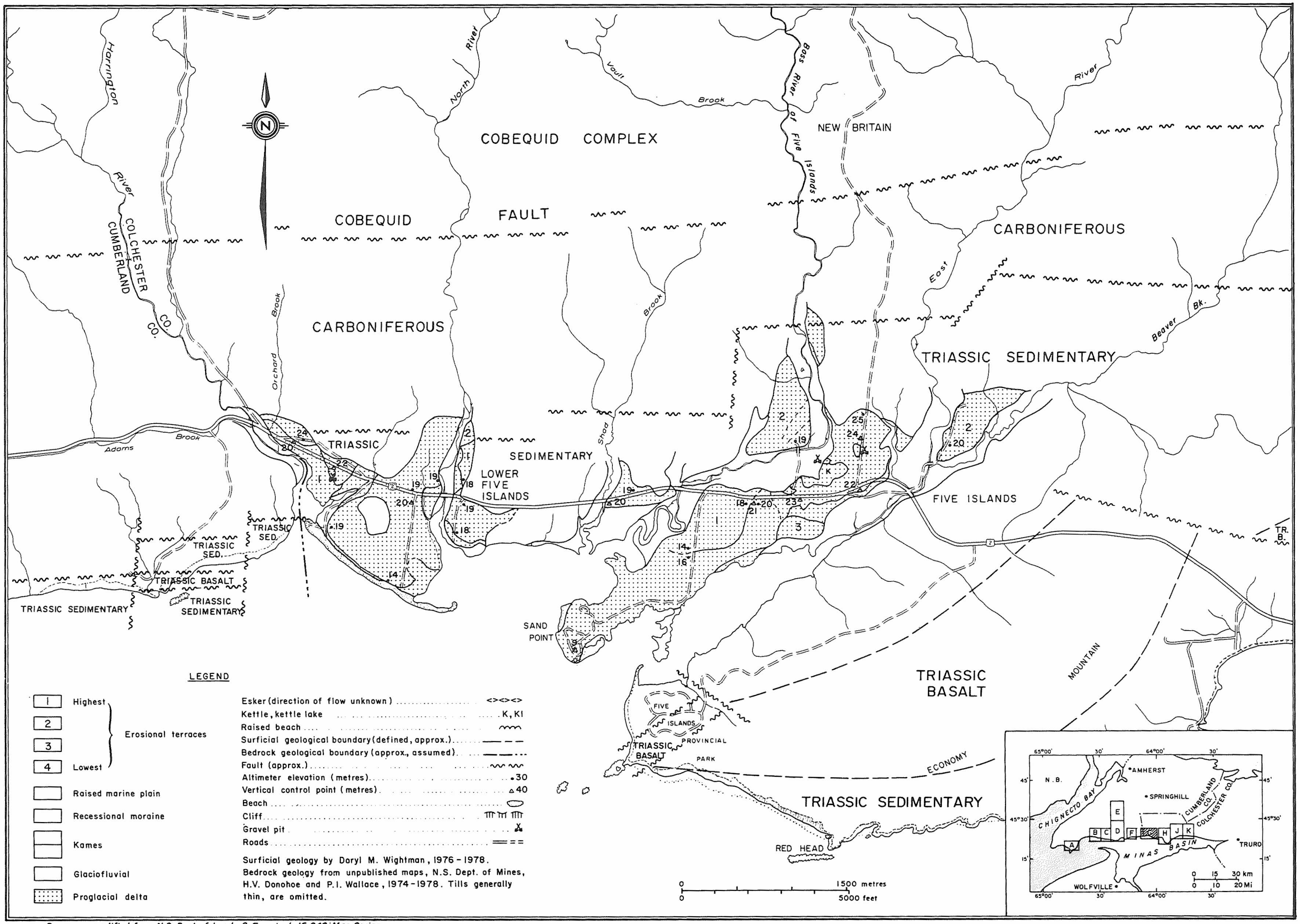
SURFICIAL GEOLOGY

Five Islands is located approximately 10 km west of Economy and 20 km east of Parrsboro. Swift and Borns (1967, p. 695) show one continuous deposit from Lower Five Islands to Five Islands and, although they only mention an exposure at Lower Five Islands, they imply that the whole deposit is deltaic. The author has mapped two separate deposits (Fig. 5.1) comprising a delta at Lower Five Islands and a larger deposit at Five Islands proper that is tentatively mapped as a delta but lacks the exposure to prove it.

LOWER FIVE ISLANDS

At Lower Five Islands, the main part of the delta lies between the Harrington and North Rivers, with a smaller part just east of the North River. The most proximal portions of the delta overlie the Upper Carboniferous Parrsboro Formation, while the bulk of the delta overlies the Triassic Scots Bay Formation.

At the proximal (north) end, the delta reaches 24 m in elevation and slopes down to 14 m at the distal end. North of the delta the bedrock hills rise sharply to elevations over 100 m but across the Harrington River from the delta, the Carboniferous is lower (≃50 m in elevation) but rises to the west. A knoll of Triassic protrudes through the central part of the delta but it does not reach 30 m in height. On the east side south of highway 2 the sediments thin out over low (below 30 m) Triassic bedrock.



Base map modified from N.S Dept. of Lands & Forests 1:15,840 Map Series.

FIGURE 5.1. Map G. Surficial geology of the Five Islands area, Colchester County, Nova Scotia.

The upper surface of the delta at Lower Five Islands slopes seaward at approximately 6.5 m/km (Fig. 5.2), calculated in a southeastern direction from the proximal part of the delta at the Harrington River. Data are more sparse along the North River but the gradient appears as steep as, or steeper than, south of highway 2, but north of the highway it appears to be shallower. There are several terraces eroded in the delta along the Harrington and North Rivers.

FIVE ISLANDS

At Five Islands proper, the deposit extends discontinuously from Shad Brook east to Beaver Brook and is broken into several parts by the Bass River of Five Islands and the East River. The deposit overlies the Triassic Scots Bay Formation except for a small northern part that extends on to the Upper Carboniferous Parrsboro Formation along the Bass River of Five Islands. The maximum elevation of the outwash is 25 m on the proximal part between the Bass River of Five Islands and East River. From here it slopes seaward to an elevation of 9 m on the distal portion at Sand Point. The bedrock surrounding the deposit rises to elevations of 75 m or more, but the slopes are not quite as steep as those found in the bedrock surrounding the deltas to the west because of the soft Triassic bedrock. The small part of the deposit just east of Shad Brook is probably very thin.

The gradient of the upper surface of the deposit at Five Islands proper is unusual. In the lower part south of highway 2, the gradient is approximately 6.0 m/km (Fig. 5.3). North of the highway the gradient is much lower, roughly 2.0 m/km which is less than half of the gradient to

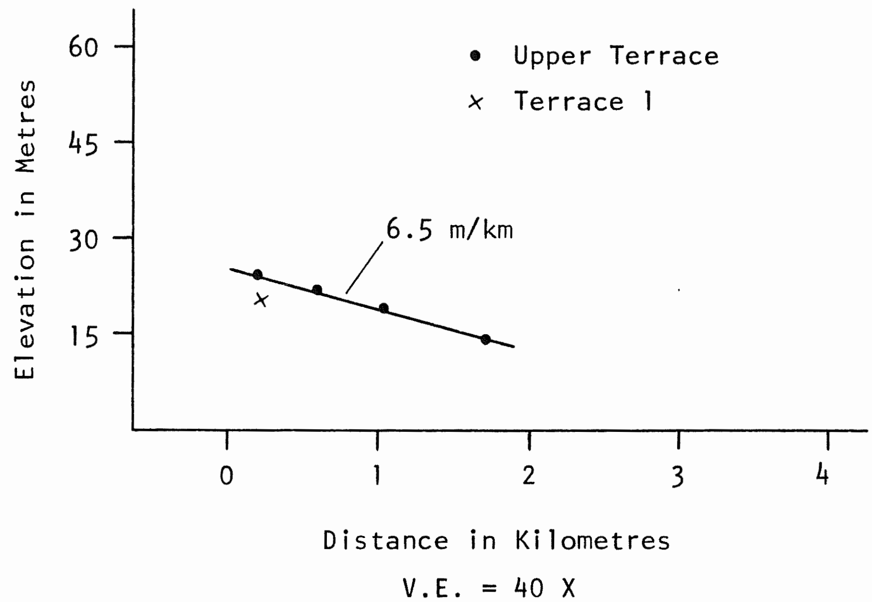


FIGURE 5.2. Gradient of the upper surface of the delta at Lower Five Islands. Profile northwest-southeast along the east side of Harrington River.

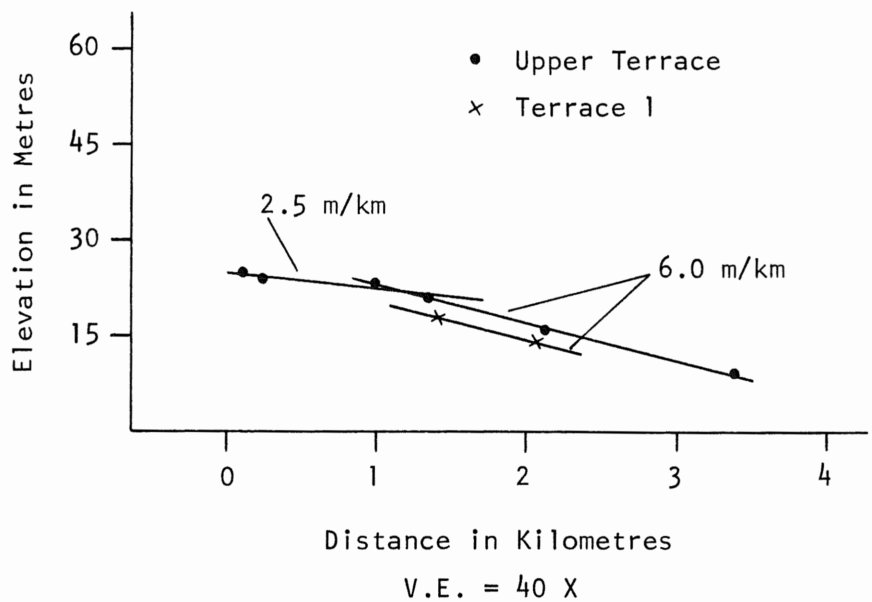


FIGURE 5.3. Gradients of the upper and terraced surfaces of the deposit at Five Islands. Profile northeast-southwest from proximal part of deposit along road to New Britain, to Sand Point.

the south. A fairly large terrace occurs south and east of the Bass River of Five Islands and its gradient appears similar to the gradient of the upper surface, which is 6.0 m/km (Fig. 5.3). Two terraces have been correlated as second order, although they may be first order, and there is no elevation data for the small terrace, tentatively correlated as third order, near the estuary of the East River.

SEDIMENTOLOGY

The main exposure is at the Lower Five Islands delta along the shore between the Harrington and North Rivers. This is one of the most important exposures in the Minas terrace as it is the 'type' section for the Five Islands Formation (Swift and Borns, 1967, p. 703). It is 19 m above sea level at the proximal (W) end and slopes down to 14 m at the distal end. One other exposure at Lower Five Islands will be discussed briefly at the end of this section. Several small exposures occur at Five Islands but most are too thin (Sand Point) or small to be of any value.

LOWER FIVE ISLANDS

Topset Facies

Description

At the proximal end of the delta, the topset beds are about 3.5 m thick. Three units, based on lithology and colour, are discernible within the topset beds (Fig. 5.4), although the lowest unit is not extensive and the grain size distribution of a sample taken from this unit poses questions about its origin. The division of the topset facies into 3

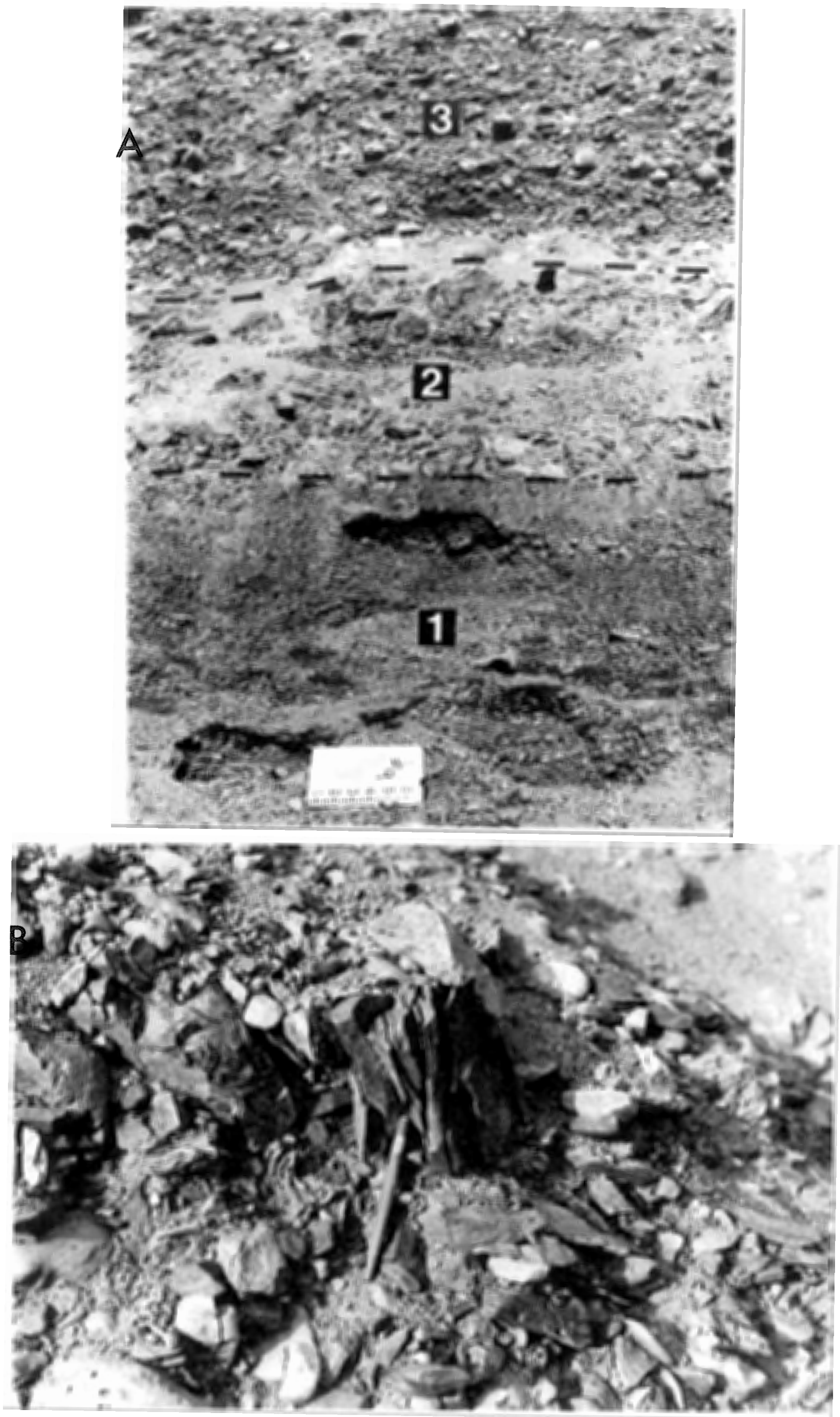


FIGURE 5.4. A. Three units of the topset facies at proximal end of Lower Five Islands delta exposure. Field book at topset/foreset contact. B. Sandstone clasts in unit 2. Sandstone has tendency to split along bedding planes.

units is dealt with again in Chapter 13 where the detailed lithologic compositions of the units are presented.

Unit 1.

The lowest unit has a red tinge to its basic grey colour and about 27 percent of the pebbles are of sedimentary (non Cobequid) origin. The unit is finer grained (sample 78-11, Fig. 5.5) than the other topset units. The red colour is imparted by the matrix of which silt and clay compose only 0.16 percent in sample 77-11.

The grain size distribution of sample 77-11 is markedly different from that of the other topset samples and in fact looks similar to that of sample 77-16 from the distal foreset beds. They are not identical because sample 77-11 is better sorted (steeper curve) than sample 77-16. In figure 5.6 sample 77-11 is plotted with two fine grained (gravel) topset samples from Spencers Island (74-199, 74-203) and the cumulative curves are very similar. It is possible that, due either to sampling technique or differences in sedimentary processes, finer grained fluvial gravel has a more sigmoidal shaped cumulative curve than coarser fluvial gravel in this setting.

The structure of the gravel in unit 1 is also that of topset, not foreset, gravel. Unit 1 occurs above a definite topset/foreset contact and there appears to be no disruption of the underlying foresets or bottomsets. The gravel is horizontally stratified, similar to other topset beds.

Unit 1 occurs only on the proximal end of the exposure and is the least extensive of the units as it is traceable with certainty over a

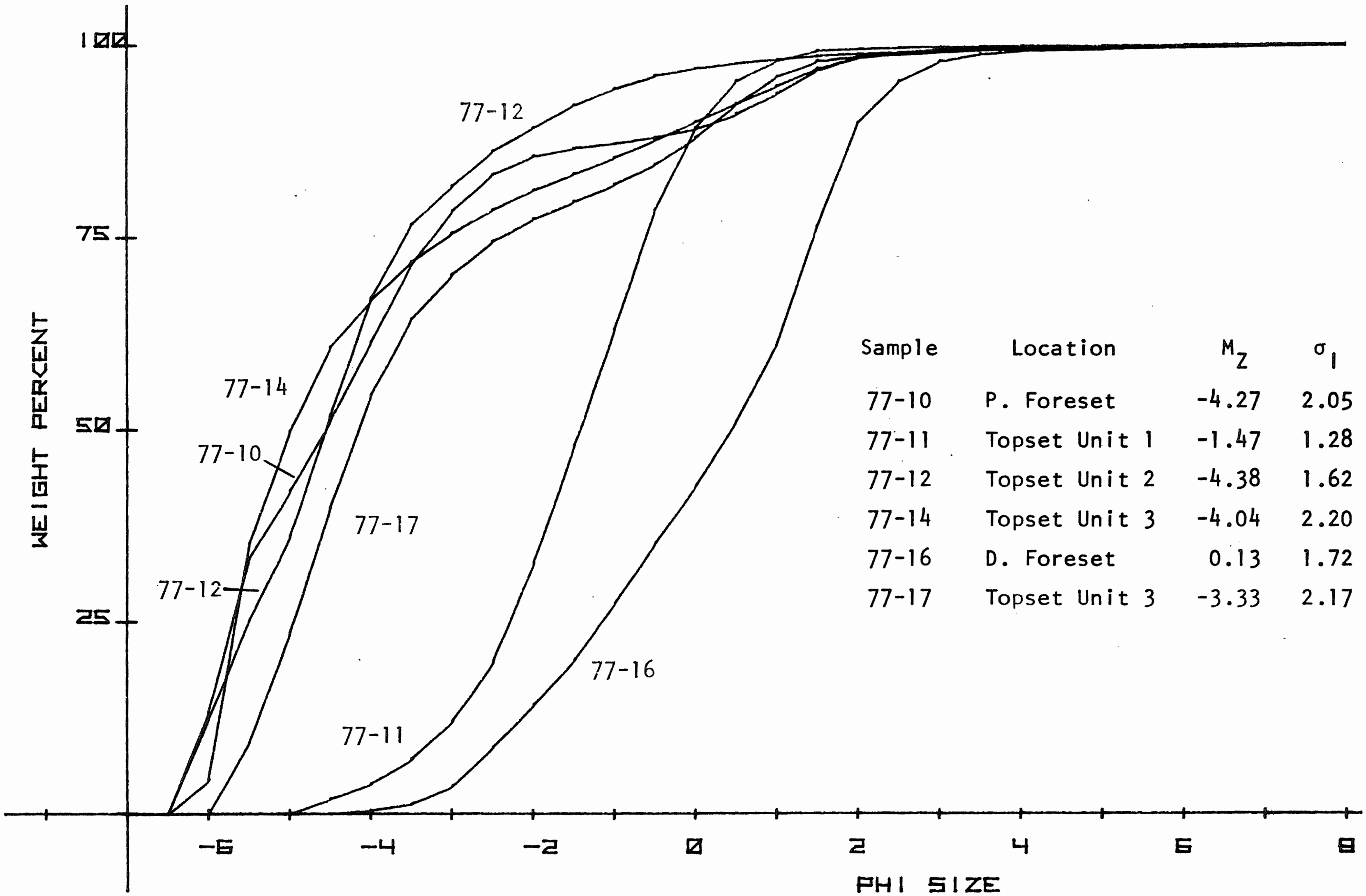


FIGURE 5.5. Grain Size analyses, foreset and topset facies, Lower Five Islands.

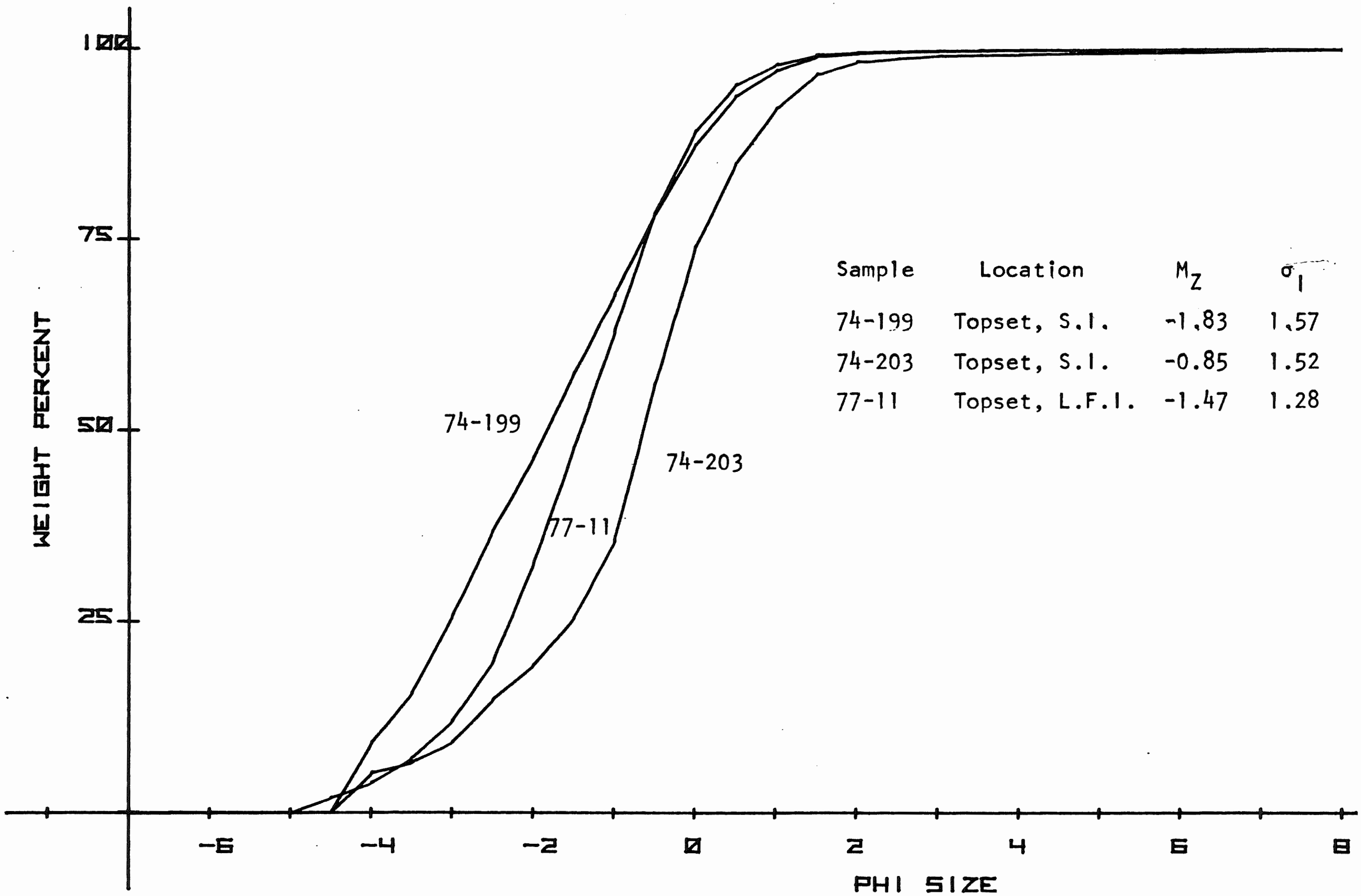


FIGURE 5.6. Grain size analyses, topset facies, Lower Five Islands (77-11) and Spencers Island.

distance of only 50 m from the western end of the exposure. It is approximately 0.75 m thick but the thickness varies. This unit is truncated to the east by a large channel shaped deposit (A, Fig. 5.7) that is 35 m wide with a maximum depth of 1.1 m. The topset/foreset contact beneath unit 1 is an erosional unconformity that is generally flat but it is indented by several channel shaped depressions up to 0.5 m deep. East of the large channel deposit, there are several questionable occurrences of unit 1, less than 10 m long and 1 m thick, that either thin out or are truncated.

Unit 2

Unit 2 lacks the red tinge of unit 1 and is grey in colour. It is similar in composition to unit 1, with 29 percent of the pebbles being sedimentary in origin but unit 2 is much coarser as is shown in figure 5.5 by sample 77-12, taken from the base of the unit. The coarseness of unit 2 is one of its distinguishing characteristics (Fig. 5.4). Cobbles and boulders of sandstone or tectonized sandstone, up to 70 cm x 40 cm x 30 cm, are common to abundant and rapid lateral changes in grain size are frequent.

Unit 2 is generally massive but in places a rough horizontal stratification, 30 to 50 cm thick, is imparted by changes in grain size. Coarser gravel tends to outline the bottoms of channel deposits (A, Fig. 5.7) and some channel deposits are very coarse throughout. In one coarse channel deposit, the b-axes of the ten largest cobbles was measured and the mean is 19.3 cm. This is a very representative measurement as the cobbles were in two adjacent groups (4 and 6) of clast to clast contact.

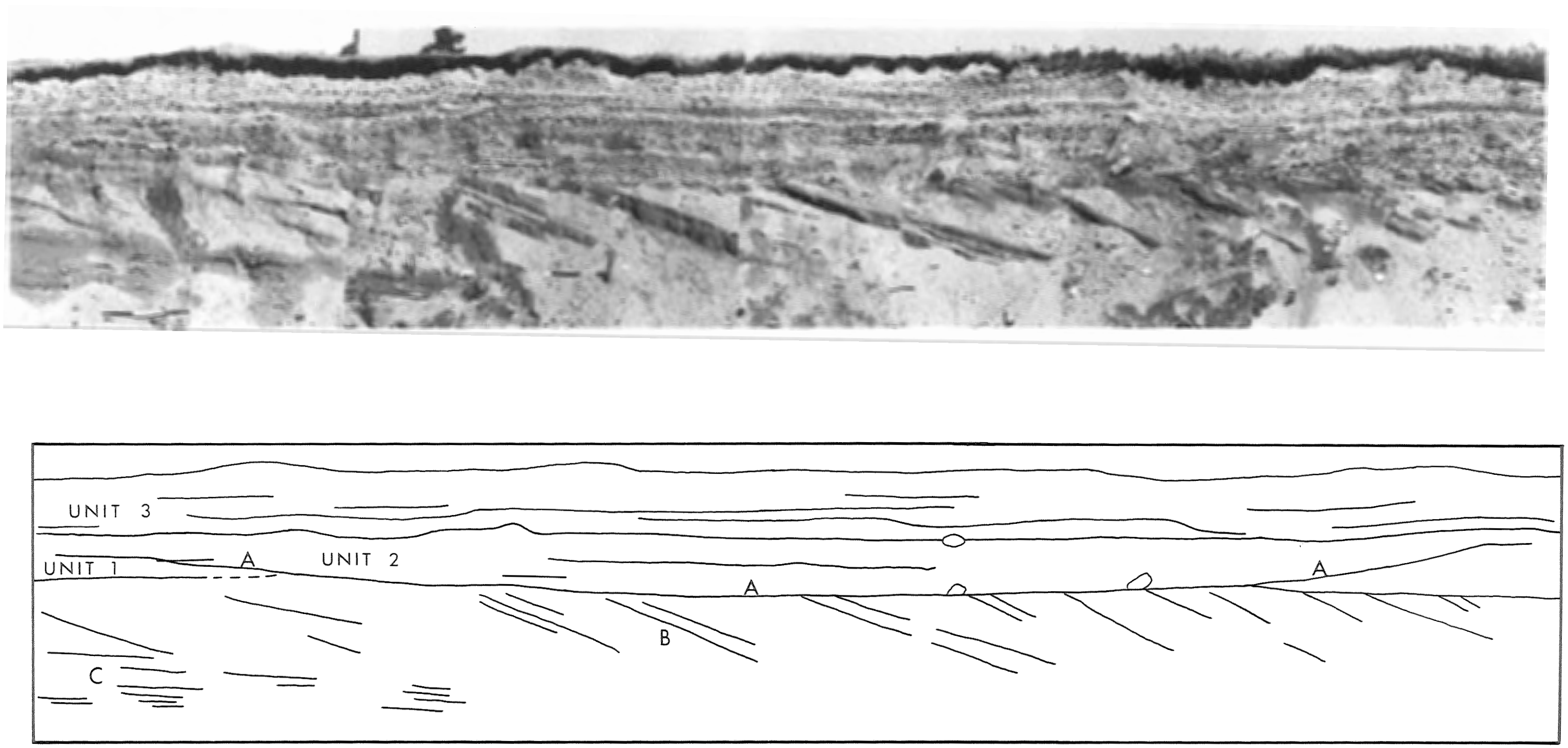


FIGURE 5.7. Photolog of proximal part of exposure, Lower Five Islands. Divisions on scales are 0.5 m, as in all other photologs.

At the proximal end of the delta, unit 2 erosionally overlies unit 1 and relief on the contact is less than 0.5 m. In places, there is some gradation between the units and the contact is difficult to place. Here, unit 2 is approximately 0.85 m thick and it thickens to the east as unit 1 is truncated (Fig. 5.7). The channel deposit that truncates unit 1 is part of unit 2 and in the middle of the channel deposit, unit 2 attains a maximum thickness of 1.7 m.

From the point where unit 1 is erosionally truncated in the proximal part of the exposure to the mid part of the exposure, unit 2 overlies the foreset beds with an angular unconformity. In the proximal part of the exposure, the topset/foreset contact is shallowly undulating with relief in the order of 1 m. The lows are always caused by channel deposits. One particularly coarse channel deposit, composed of pebble to cobble gravel, is 12 m wide and extends 1 m into the foreset beds.

In the mid delta area, unit 2 is approximately 1.2 m thick but in places where unit 3 thickens, unit 2 is thin and possibly even absent. The topset/foreset contact is more planar than at the proximal part of the delta, but it is difficult to place in certain areas where the underlying foreset beds are similar in grain size to unit 2.

Unit 3

Unit 3 is the most conspicuous of the units as it has a distinctive yellow brown colour, an abundance (88%) of Cobequid (igneous and metamorphic) pebbles and an almost vertical weathering profile. The colour is derived partly from the sand matrix (only 0.28% silt and clay in sample 77-14, Fig. 5.5) and partly from the abundance (51%) of granite pebbles.

Samples 77-14 (proximal) and 77-17 (distal) in figure 5.5 show that the unit is fairly coarse and becomes slightly finer from proximal to distal. The gravel fraction in unit 3 has a more even grain size, with fewer cobbles and boulders, than that of unit 2. The gravel has a crude horizontal stratification that is imparted by discontinuous sand beds, most numerous in the distal section, that are usually 5 to 20 cm thick. The sand is generally coarse to very coarse grained and horizontally stratified, but there is some cross stratified sand and the grain size does get as fine as medium grained.

From the proximal to the mid parts of the exposure, unit 3 erosionally overlies unit 2, but the contact is not always sharp and in some places there is a mixing of rock types between the two units. Relief on the contact is less than 1 m. Unit 3 varies in thickness but is usually from 1.5 to 1.8 m thick.

Eastwards from the proximal end of the delta, the lowest 2 units successively thin out so that at the distal end of the delta, unit 3 constitutes the entire topset facies. As unit 2 thins out from the mid to the distal end, the contact between the two units is difficult to establish in some places. Even in the distal part of the delta, patches of unit 2 or concentrations of sandstone in unit 3 occur near the topset/foreset contact. Thus, the contacts between the topset units are generally sharp but they can be gradational and subjective. At the distal end of the delta, unit 3 attains its maximum thickness of 2.3 m.

Interpretation

The topset beds can be considered the equivalent of the glaciofluvial gravel that has been described at Portapique and Bass River as the topset beds are deposited by streams which cross the delta, transporting sediment to the delta front (Gilbert, 1890, p. 68; Church and Gilbert, 1975, p. 86). Gustavson *et al.* (1975, p. 264) state:

> "It is the fluctuations in discharge, velocity, depth, sediment load and other hydrologic parameters of the meltwater streams that account for the variations in the continuum of sedimentary structures linking deposits of the glaciofluvial, glaciodeltaic and glaciolacustrine environments."

Units 1, 2 and 3

Although unit 2 is coarser than unit 1 and unit 1 has a red tinge to its colour, the pebble components of the two units are basically the same. Unit 3 is distinctly different with a much higher content of Cobequid rock types in the pebble fraction. Swift and Borns' (1967, p. 706) model of northward receding ice margins during the deposition of the deposits is considered the key to understanding the genesis of the three topset units.

As glacial ice advances, it erodes and incorporates bedrock into the ice. Therefore, at any given location, the ice will contain sediment representative of the bedrock over which it has passed. The erosion of the clasts during transport, and hence the distance downstream that the clasts will survive depends, at least partly, upon the mode of transport of the clasts within the ice (Andrews, 1975, p. 123) and the competency of the clasts.

If the last ice advance over the Cobequid Highlands was from the north (Nielsen, 1976, p. 98), ice that sat over a particular bedrock formation would contain pebbles from that formation and from bedrock to the north. Outwash and till at Portapique and Bass River show that clasts of the Triassic red beds were too incompetent to survive glacial erosion, transportation and subsequent deposition by meltwater. However, the closest and most likely source for the red component in the colour of unit 1 is the Triassic red beds over which most of the delta at Lower Five Islands lies. The lithology of the pebbles in unit 1 is consistent with an ice front position over the red beds. Carboniferous sedimentary rocks and the Cobequid complex lie immediately north of the Triassic (Fig. 5.1) and both have typical representatives in the pebble fraction of unit 1. About 27 percent of the pebbles are of sedimentary origin. The predominance of Cobequid rock types is probably partly due to the width of the Cobequids compared to the width of the Carboniferous Parrsboro Formation (≈13.5 km versus 2.25 km) and partly due to the resistance of Cobequid pebbles to comminution during glacial and glaciofluvial transportation.

As the ice front receded to a position on the Carboniferous rocks, clasts representative of the Carboniferous and of rock types to the north would be released. Red matrix would no longer be supplied to the sediment as ice in that position had not crossed the Triassic south of the Cobequid Fault. The pebble lithology of unit 2 is consistent with an ice front position over the Carboniferous. Although the percentage of sedimentary rock (29%) is only slightly greater than in unit 1 (27%), the occurrence of large cobbles and boulders of sandstone is conspicuous in the unit.

Retreat of the ice front northward across the Cobequid Fault would result in the discharge of pebbles representative of the Cobequid Complex and farther travelled sediment from the north. The low percentage of sedimentary rock in unit 3 (12%) and the abundance of granite (51%) agrees with an ice front position in the Cobequids. The low percentage of sedimentary rock suggests that clasts derived from bedrock north of the Cobequids were largely comminuted by the time ice had crossed the Cobequids and/or the erosion of Cobequid bedrock was very high and the other rock types were diluted. The former concept appears more plausible. The sedimentary rock that is present was probably picked up as the streams reworked the previously deposited outwash.

Texture and Structure

The topset beds are similar in texture and structure to the glaciofluvial sediments of the deposits east of Five Islands and the depositional processes are interpreted to be the same. Especially striking is the similarity between unit 3 and the topset(?) gravel at Economy, the glaciofluvial gravel at Saints Rest and the upper part of the glaciofluvial sequence at Portapique. All are rich in igneous and metamorphic rock types, relatively uniform in clast size (upper limit generally -6.5 to -7ϕ), closedwork with a medium to coarse grained sand matrix, massive to crudely bedded and contain infrequent, discontinuous lenses of unrippled sand. They are interpreted as an aggraded sequence of diffuse gravel sheets deposited by braided streams. In unit 3, the slight downstream fining of the gravel fraction (samples 77-14, 77-17, Fig. 5.5) and downstream increase in sand beds are typical of braided streams (Boothroyd and Ashley, 1975, p. 218; Church and Gilbert, 1975, p. 66; Ballantyne, 1978, p. 146).

Unit 1 is slightly finer grained and distinct beds are better developed, but the overall process is probably the same. Unit 2 is very coarse in places and some of the channel deposits are composed of massive coarse gravel such as that at Little Bass River (Fig. 3.6). Flow velocities must have been very high (>4 m/sec) to transport this gravel. Even though the maximum paleochannel depth is only 1.1 m, the flow depth was probably greater as the coarse channel deposits are almost as coarse at the top as at the bottom. With flow depth greater than channel depth, large areas on the upper surface of the delta must have been flooded at times.

There is a noticeable difference in the number of recognizable channel deposits within the topsets from the proximal to distal parts of the exposure. At the proximal end, channel deposits are common and the topset/foreset contact is frequently indented by the deposits. The apparent extraordinary width (35 m) of one channel deposit is probably the result of an oblique cut through a paleochannel as the deposit is only 1.1 m thick. Channel deposits are less recognizable in the mid delta, and in the distal part of the exposure only the sands infrequently contain cut and fill structures. The topset/foreset contact in the distal part of the delta is very flat.

The decrease in the number of channel deposits from proximal to distal is interpreted as the result of an increasing number of stream anabranches in the downstream direction. In the proximal part of the delta, stream flow was more incised and there were fewer but larger channels. The braided pattern increased in the distal direction as the streams fanned out over the spreading delta. As well as increasing in number, the channels were probably shallower and broader. This relationship has been observed by others. McDonald and Banerjee (1971, p. 1284) noted that meltwater discharge flows on to the Peyto outwash delta in one large channel and the number of anabranches increases in the distal direction. Smith (1978) studied a lake (Hector) with low sediment input but the Balfour stream also changed from a single stream to braided near the mouth of the Balfour outwash delta (Smith, 1978, p. 742). Boothroyd and Ashley (1975, p. 198) state that on the Yana and Scott outwash fans, one or two incised channels on the upper fan spread out downstream into a braided pattern.

The proximal to distal change in the topset/foreset contact reflects this change in stream pattern. The topset/foreset contact occurs at the base of the inflow water (Gustavson *et al.*, 1975, p. 268) so that the larger incised streams on the proximal part of the delta indent the topset/foreset contact. In the distal part of the delta, the streams shifted laterally and eroded the top of the foreset beds to a common depth, producing a planar, erosional unconformity. McDonald and Banerjee (1971, p. 1284) noted that on the distal part of the delta at Peyto Lake, stream discharge swept from one side of the delta to the other every few years.

The channel deposits in the proximal part of the delta are in units 1 and 2. When unit 3 was deposited, the ice front had receded farther northward from the delta and the 'proximal' part of the delta had become more 'distal' in the outwash system. Thus, the vertical sequence in the topset beds at the proximal end of the exposure displays the lateral proximal/distal variation in stream pattern.

Foreset Facies

Description

Proximal

The foreset beds on the western end of the exposure dip 20 to 25° to the southeast (B, Fig. 5.7), and are generally similar in grain size to unit 1 but finer than units 2 or 3, although there are some coarse beds (sample 77-10, Fig. 5.5). Bed thicknesses range from several centimetres to 50 cm. Approximately 29 percent of the pebbles are sedimentary in origin. The foreset facies is surprisingly thin, approximately 2 m, as bottomset beds (C, Fig. 5.7) occur well up the bank.

Mid Delta

The foresets are poorly exposed in the mid part of the delta and no bottomset beds are visible because talus usually extends to or near the topset/foreset contact. Most of the exposed foresets dip to the southeast at about 25° while some dip to the south or southwest so that the exposure is a strike section for these beds. In general, the foreset beds are coarser than elsewhere, being as coarse as unit 2 in many places. Pebbles of sedimentary rock type attain a maximum value of 49 percent in the mid foresets. At one location, a penecontemporaneous normal fault, with a dip of 60°E and a displacement of 1 m, cuts the foresets and unit 2 of the topsets, but not unit 3. Movement of the underlying, unexposed bottomset beds is the probable cause of the fault.

Distal

The foreset beds in the distal section of the delta dip 20 to 30° to the southwest (Fig. 5.8). The beds are finer grained than elsewhere in the delta which makes the topset/foreset contact easy to place. The topset/foreset contact is a planar angular unconformity, conspicuous by its lack of relief (A, Fig. 5.8). Medium to coarse grained sand is the dominant component of the beds but there is some coarser sediment in the grain size distribution (sample 77-16, Fig. 5.5). About 22 percent of the pebble gravel is sedimentary in origin. Bedding thickness ranges from about 5 to 50 cm, but most beds are 10 to 20 cm thick. The beds are exposed up to a vertical distance of 2.5 m.

A peculiar sand 'dune' with a height of 25 cm was developed less than 1 m below the topset/foreset contact in a shallow depression on a foreset bed (Fig. 5.9). The cross stratification dipped to the northwest, or up the foreset bed which dipped about 20°SE. The sand at the base of the dune was laminated essentially parallel to the foreset bed, but the laminae steepened upwards until they reached an angle of 32° to the foreset bed. The laminae had an angular lower contact but were convex at the top of the dune and continued down dip essentially parallel to the surrounding foreset beds. Succeeding laminae had tangential to concave lower contacts (Pettijohn *et al.*, 1973, p. 352) and gradually infilled the lee side of the dune.

Interpretation

Grain Size

Processes involved in the deposition of foreset beds will be discussed in the chapter on the delta at Spencers Island, so only the general

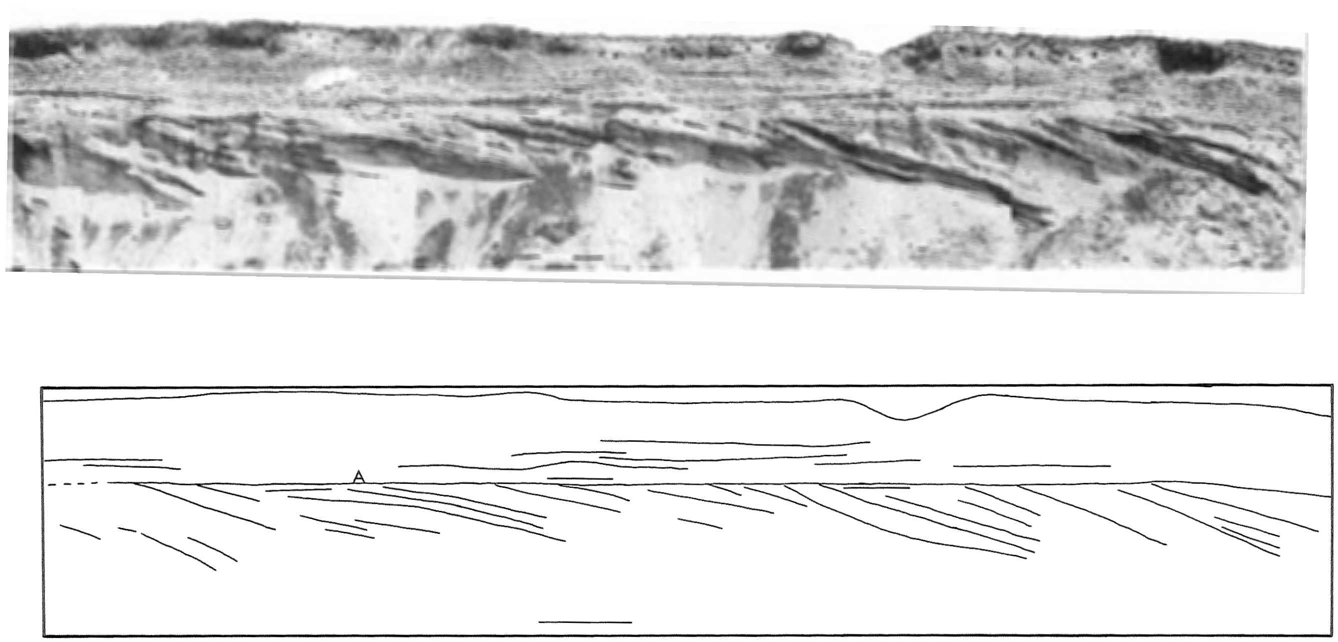


FIGURE 5.8. Photolog of distal part of exposure, Lower Five Islands. Planar topset/foreset contact (A). Several scales (1.5 m) marked on line drawing and on subsequent line drawings of photologs.

FIGURE 5.9. Sand dune developed in upper foresets. Shallow depression, with steep side up dip, in underlying bed.

characteristics of the foresets will be discussed here. The foreset beds are coarsest in the mid delta area where the proportion of sandstone is greatest, but the foresets are also composed of gravel in the proximal area and can be coarse there as well (sample 77-10, Fig. 5.5). They are composed of sand in the distal part of the exposure, so the foreset beds generally fine in a proximal to distal direction.

Pebble Lithology

The proportion of Cobequid rock types in the pebble fraction, from proximal to distal, is 71 percent, 51 percent and 78 percent. The abundance of sandstone in the mid delta not only coarsens the foresets but decreases the proportion of Cobequid rock types. Cobequid rock types are most abundant in the distal part of the delta, but are 10 percent less abundant than in the corresponding topsets. This may be partly due to the grain size difference between the distal topsets and foresets.

Pebble counting was difficult in the distal foresets because of the fine grain size. Sediment coarser than -2.5ϕ had to be searched for and the pebbles that were counted were generally finer than elsewhere. For this reason, all pebbles coarser than -2.5ϕ were counted at every deposit within the Minas terrace. The sandstone pebbles that were found in the distal foresets were often very thin and chip like. Pebbles, cobbles and boulders of sandstone in the topset beds are easily recognized because the sandstone splits readily along bedding planes. Many cobbles and boulders are really loose bundles of smaller tabular shaped pieces, so the source of the chips is readily apparent. The tabular shape of the pebbles makes their settling velocity equivalent to the settling velocity of smaller clasts with spherical shapes (or their equivalent diameter [Blatt *et al.*, 1972, p. 52] is smaller than their sieve diameter). Thus, the enrichment of sedimentary clasts in the foresets relative to the topsets could be partly due to their hydraulic behaviour as smaller particles. This theory is also applicable to the foresets and topsets in the mid delta where the foresets are enriched in sedimentary pebbles (49%) relative to the overlying (unit 2) topsets (29%).

Sand 'Dune'

The peculiar 'dune' in the foresets (Fig. 5.9) appears to have been deposited by an eddy with a horizontal axis and small radius (the dune is less than 1 m below the topset beds). The dune did not migrate up the foreset slope, as laminae farther down the foreset bed are parallel to the bedding, but it infills a localized erosional scour on the underlying foreset bed. The eddy was probably caused by strong stream outflow into the bay forming a powerful separation eddy (Pettijohn *et al.*, 1973, p. 351) at the top of the foresets. The backflow of the eddy probably eroded the scour and infilled it with the cross stratified sediment. Lower down the foreset slope, larger scale backflow must have been too weak to form bedforms that migrate up the foreset slope.

Only one dune was found so that the formation of such a small, powerful separation eddy must have been uncommon. Strong stream outflow was undoubtedly required, but should have been common as shown by the stream velocities required to move the gravel bedload ($\simeq$1 to 4 m/sec). The formation of this type of eddy must have been contingent upon other factors, one of which appears to be strong and immediate mixing between the lower part of the outflow and the sea water in the Minas Basin.

Bottomset Facies

Description

Proximal

At the extreme western end of the exposure the foreset beds flatten into subhorizontal sandy bottomset beds (C, Fig. 5.7), with the transition 3 to 4 m above the base of the bank. The bottomsets are made up of red clayey sand beds, 1 to 2 cm thick, interstratified with sand beds, 2 to 3 cm thick. Towards the east the bottomset beds disappear under talus.

Approximately 100 m east of the west end of the exposure, bottomset beds reappear. They are exposed for a length of about 70 m and where they first appear, they extend to 2.5 m above the base of the bank. The lower 1.5 m are interstratified sand and silty clay (A, Fig. 5.10) which are overlain by 1 m of sand bottomsets (B, Fig. 5.10) that extend up into the foresets (C, Fig. 5.10). The sands are up to medium grained in coarseness. Although the clay content of the lower beds is significant, it is much lower than the clay content of the bottomset beds at a key exposure at Spencers Island.

The bottomsets are subhorizontal over most of this section but there are two areas where the beds steepen to dips of 15 to 20°W. At the first 'steepening' (D, Fig. 5.10), the bottomsets rise to 3.5 m above the base of the bank. The lower bottomsets are disrupted and display low amplitude folds (10 cm) that are overturned to the west (E, Fig. 5.10). At the second 'steepened' section (F, Fig. 5.10), the upper metre of bottomsets is not disrupted but 40 cm of the upper strata

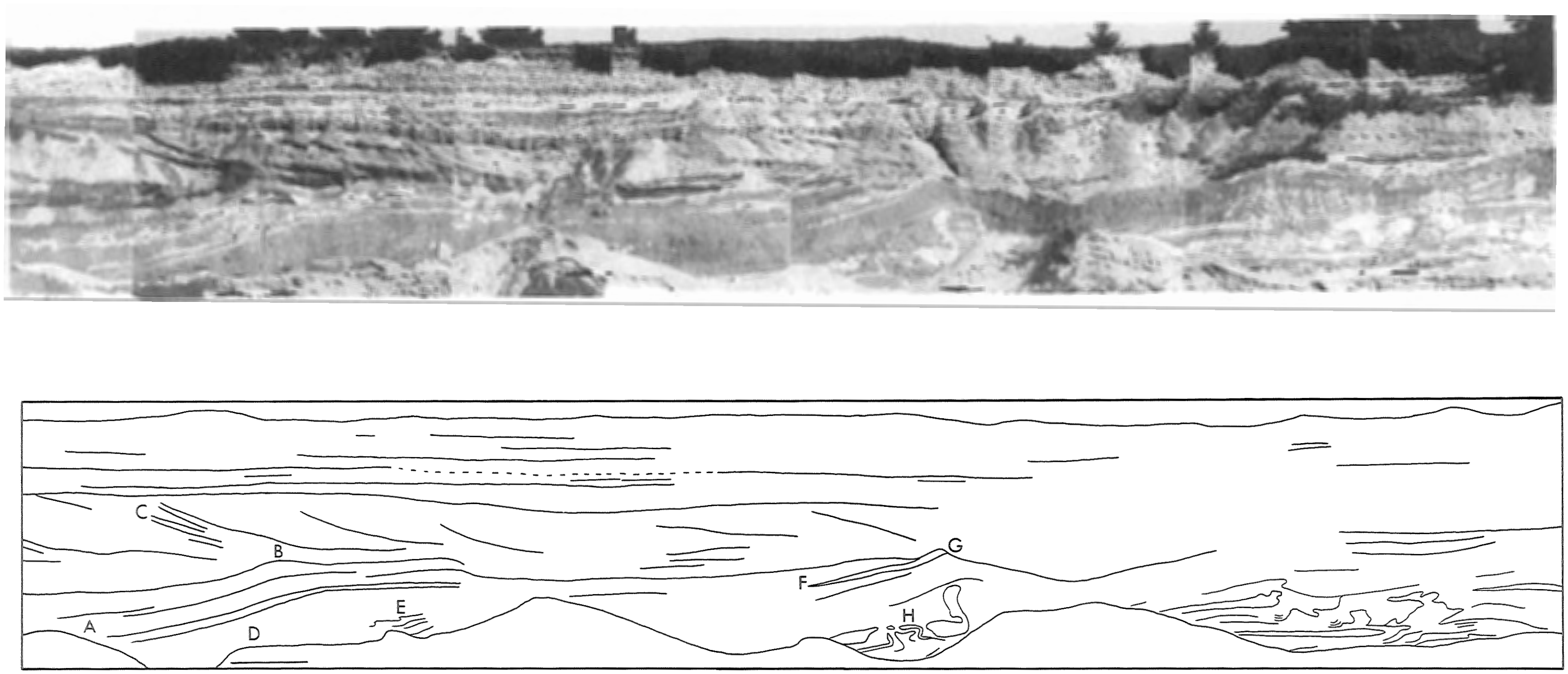


FIGURE 5.10. Photolog of proximal part of exposure, east of figure 5.7, Lower Five Islands.

are truncated at the up dip end (G, Fig. 5.10). Below this (H, Fig. 5.10) the bottomsets are folded and contorted. The sand bottomsets that overlie the clayey bottomsets to the west pinch out at this point.

East of this, the bottomsets rise gradually to a height of 5 m above the base of the bank (Fig. 5.11). The upper 1 to 1.5 m of the section are subhorizontally stratified (A, Fig. 5.11), but the lower part is disrupted with folds and contortions in the beds (B, Fig. 5.11). The disrupted interval ends abruptly against a deposit of poorly exposed gravel that extends eastwards for 35 m. Although the gravel is foreset gravel, it is discussed here because of its interrelationship with the bottomsets. The contact between the gravel and the bottomsets was poorly exposed but the bottomsets are contorted (C, Fig. 5.11) and the gravel is not interbedded with the bottomsets at the contact. Most of the gravel is massive but disoriented blocks of foreset beds (D, Fig. 5.11) are visible in the upper part. The upper metre of uncontorted bottomsets laps up onto the gravel (E, Fig. 5.11) thinning to about 0.5 m and becoming sandier while doing so, until the bottomsets underlie the fluvial topset gravel. Exposure is poor but the bottomsets appear to extend down the eastern side of the deposit (F, Fig. 5.11), below the succeeding eastward dipping foreset beds.

Distal

A long (60 m) section of bottomset beds, with up to 3 m of vertical exposure, occurs in the distal part of the delta (Fig. 5.12). The bottomsets appear immediately east of a talus covered notch and extend from the base of the bank to the topset gravel, thereby eliminating the

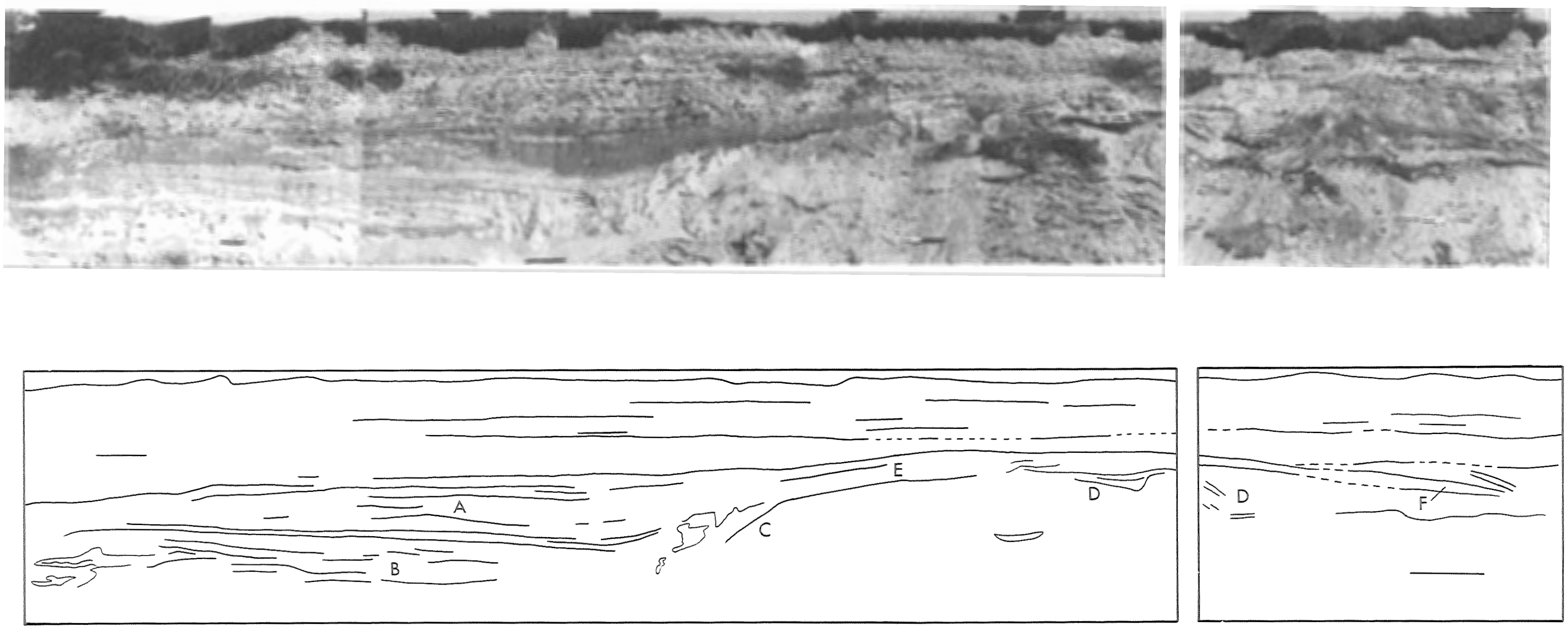


FIGURE 5.11. Photolog of proximal part of exposure, east of and adjacent to that of figure 5.10, Lower Five Islands.

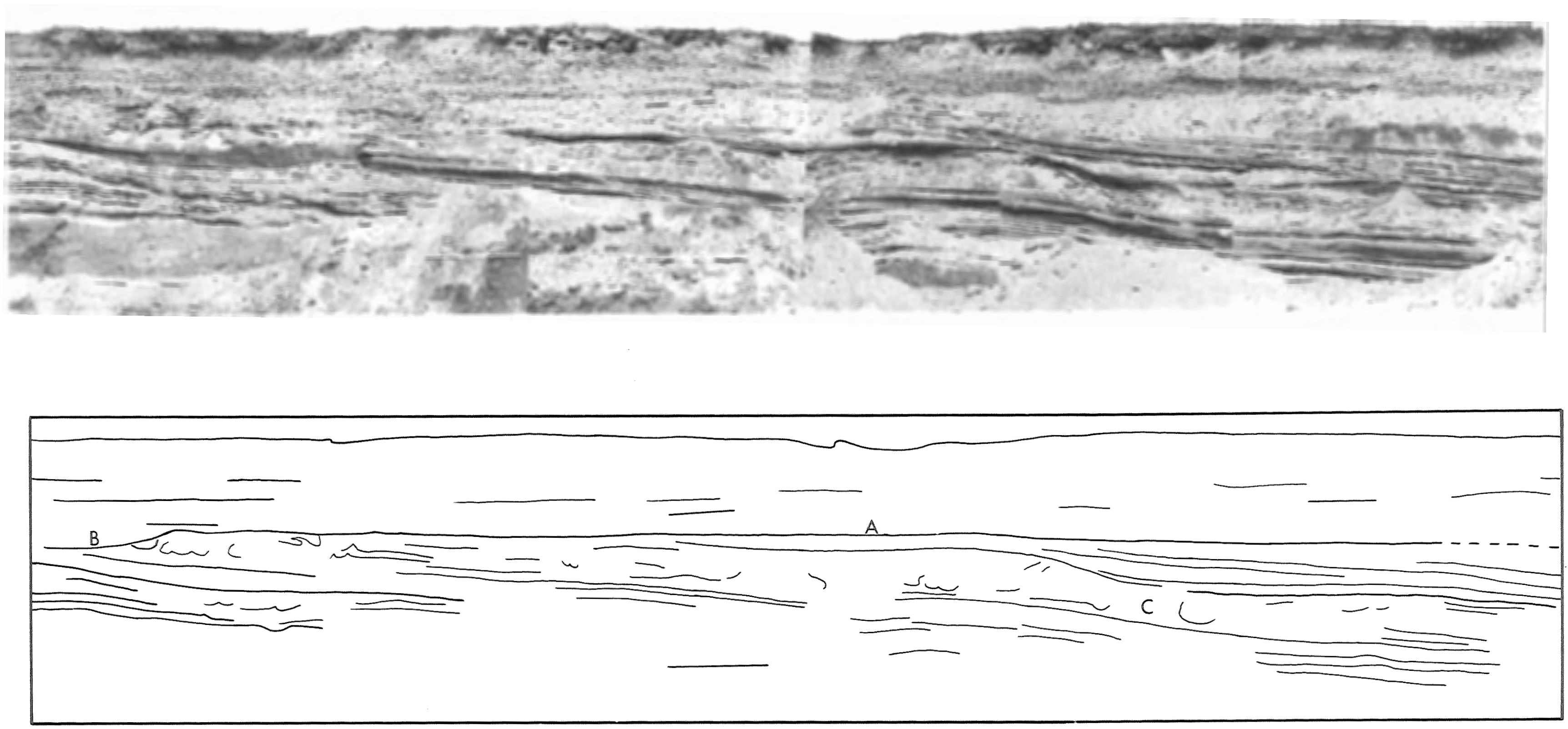


FIGURE 5.12. Photolog of distal part of exposure, east of that in figure 5.11, Lower Five Islands.

foreset beds. West of the notch, there is no indication of bottomsets and the foresets are exposed to 2 m below the topset/foreset contact. The topset/bottomset contact (A, Fig. 5.12) is an angular unconformity that is planar for all but a 5 m interval at the west end.

The bottomset section is composed of parallel laminated sand beds, up to 20 cm thick, that dip shallowly ($\approx 35°$) to the southeast. The sands are generally fine to very fine grained but can be as coarse as medium grained. Interstratified with the sand beds are red clayey sands that are usually less than 1 cm thick (laminae) but can be up to 6 cm thick. The distance between the red laminae varies but is usually several centimetres to 10 cm. Some of the sand beds have long wave length (18 to 25 cm), low amplitude (1.5 to 2 cm) ripples with rounded tops. The ripples are fairly symmetrical with a continuous lamination and they are interpreted as wave ripples (Reineck and Singh, 1975, p. 45).

There are several disturbed intervals, from 30 to 80 cm thick, in the bottomsets. At the base of the intervals the bedding is disturbed so that the stratification is irregular and uneven in thickness. Above the base, the degree of disruption increases to the point where strata are not recognizable.

One of the disturbed intervals occurs at the topset/bottomset contact. At and west of the up dip end of the interval, topset gravel extends 50 cm below the topset/bottomset contact (B, Fig. 5.12) and the contact between the gravel and the disrupted sand is sharp (Fig. 5.13), Stratification within the disrupted sand indicates vertical structures that are not overturned either down or up dip. The disturbed interval, generally 50 to 60 cm thick, extends down into the bottomsets (C, Fig. 5.12) until it is covered with talus at the bottom of the bank.

The bottomset beds end abruptly to the east where they are erosionally overlain by foreset beds, dipping 20 to 25°SE (Fig. 5.14), that continue to the end of the section. The foreset beds cut out 1.5 m of bottomsets before flattening to about 5° and disappearing under the talus.

Interpretation

The bottomsets are only visible in several short sections along the exposure and where they are exposed, the lower part is covered by talus. The bottomsets extend well up the bank, ranging from about 4 m below the topset/foreset contact to the contact itself. Thus, where the bottomsets are exposed, the foresets prograded into rather shallow (4 m or less) water. The build up of bottomsets indicates that the underlying bedrock and/or till is relatively high.

The predominant component of the exposed bottomsets is sand, at both the proximal and distal ends of the exposure. The presence of clayey sand beds within the cleaner sands means that there were periods of significant clay deposition as well. As the clay content of the bottomsets is greater lower in the section, clay deposition was greater farther from the foresets in deeper water.

The sudden appearance of the distal bottomsets to the east of a talus covered notch and their abrupt termination where foreset beds overlap them is problematical. The contact between the foreset and bottomset beds is sharp and erosional (Fig. 5.14) but it seems unreasonable that a thickness of 2 m of bottomset sand could be eroded and replaced by

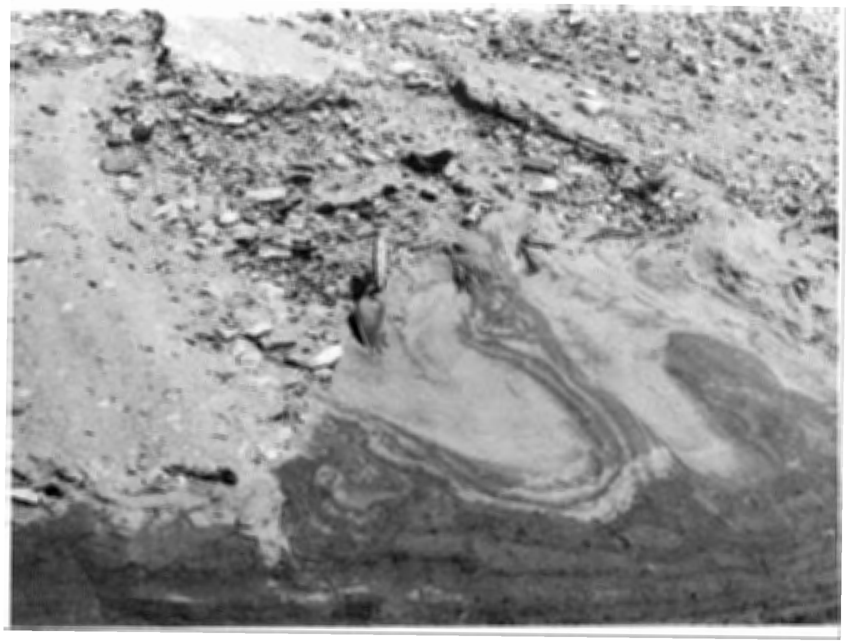


FIGURE 5.13. Sharp contact between topset gravel (left of trowel), extending down into bottomset sand, and disturbed interval (right of trowel). Darker sand (wetter) outlines vertical structures in disturbed interval. Dark laminae within sand are clay rich.

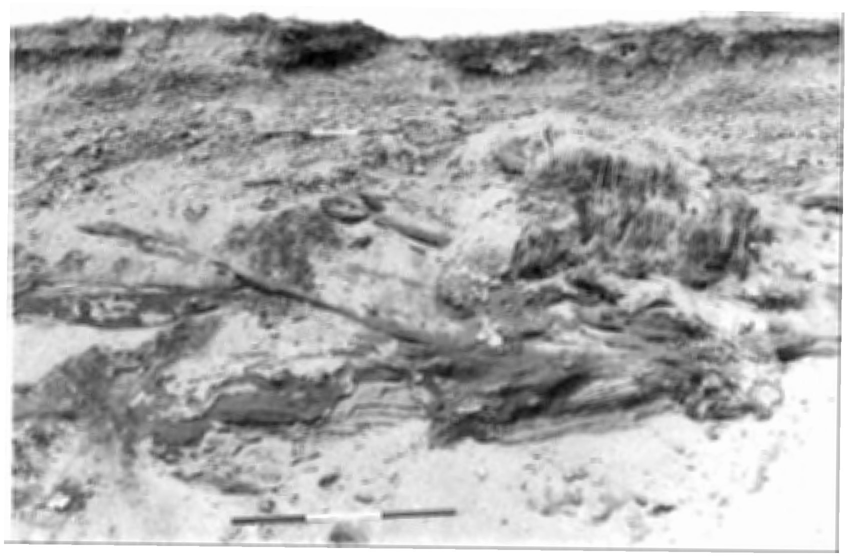


FIGURE 5.14. Foreset beds overlapping and truncating bottomset beds.

foreset beds. There is no evidence of such strong current activity anywhere along the exposure. The abrupt termination of the bottomsets may be tied in with the cause for such a build up of bottomsets.

The buildup of bottomsets to the height of the topset/foreset contact probably indicates an underlying bedrock and/or till high. As was mentioned previously, the thinness of the foreset beds wherever the bottomsets are exposed implies an underlying 'basement' high. The basement highs produced shallow water that was shallowed even further by bottomset deposition. The eastern end of the bottomsets was undoubtedly eroded but it probably coincides with a drop off in the underlying basement so that only the periphery of the bottomsets, and not a complete section, was eroded.

Structure

The disruption of the bottomsets in the proximal and distal sections of the delta shows that the bottomsets were unstable and significant amounts of movement occurred after they were deposited. The bottomsets near the proximal end of the exposure, shown in figures 5.10 and 5.11 have been particularly disrupted, especially in the lower part. It is difficult to tell the direction of movement of the bottomsets but the folding indicates a shortening in the east-west direction. The folded bottomsets at H in figure 5.10 are overlain by unfolded sediments and a core through this section would give the mistaken impression that the folded sediments are overlain by undisturbed sediments. The unfolded sediments have been uplifted and eroded on the up dip end. The unfolded/folded sequence could be due to two generations of movements or to relatively competent/incompetent sediments.

The massive gravel to the east of the bottomsets (Fig. 5.11) does contain some blocks of gravel with recognizable but disoriented foreset bedding, but it appears that the gravel has slumped to the extent that most of the bedding has been destroyed. Topset gravel is possibly included in the slump. The beds in the lower part of the bottomsets abut, but do not interfinger with, the slumped gravel. Therefore, the bottomsets were deposited after, and not contemporaneously with, the gravel.

However, the gravel may have slumped either before or after the deposition of the lower bottomsets. If the slump was first, then the foresets failed a second time after the bottomsets were deposited, for the contact is disrupted. If the slump occurred after the lower bottomsets were deposited, it means that the delta advanced with a lobate front as the foresets to the west of the bottomsets dip to the southeast. The bottomsets were deposited between two lobes. In either case, after the bottomset/foreset contact was disrupted (C, Fig. 5.11), the slumped gravel was below the level of the water in the bay and the upper bottomsets (E, Fig. 5.11) were deposited right over the top of the slump and down the other (east) side. The bottomsets on top of the slump are slightly disturbed, indicating minor movement at a later stage.

The disrupted bottomsets in the distal section also show east-west shortening, probably because of a down dip (eastward) movement. Failure of the beds was along a well defined zone (C, Fig. 5.12) and the topset gravel at the up dip end probably foundered when failure occurred

Slumps have been observed on the foreset slope of modern Gilbert type deltas and have been interpreted as the cause of mounds in the lower foreset zone at the base of the classical foresets (Fulton and Pullen, 1969, p. 789; Gustavson, 1975b, p. 455). Gilbert (1975, p. 1710) interprets a field of mounds at the base of the foresets as slumped foreset sediment. Smith (1978, p. 747) also interprets mounds on the lake floor as slumped sediment and cores from the mounds revealed both disturbed and undisturbed laminations. However, it was previously mentioned that uncontorted laminations could be mistaken for undisturbed laminations. Delta foresets and nearby scree slopes are given as likely sources of the slumps (Smith, 1978, p. 748). As well as the mounds, Fulton and Pullen (1969, p. 789) interpret ridges in the bottomset zone as pressure ridges forced up by the foreset beds.

Thus, sediment movement in the foreset and bottomset zones appears to be common in Gilbert type deltas. Fulton and Pullen (1969, p. 790) attribute sediment failure to overloading at the top of the foreset zone. Gilbert (1975, p. 1709) suggests that the shear strength of the sediment is below that required for stability on the lower slopes (2 to 8°) of the Lillooet Delta. Gustavson (1975b, p. 455) gives several possible reasons for failure, the first of which is sediment compaction and de-watering of the delta sediments. Although the cause of the sediment failure at Five Islands is not known, the author favours failure because of underconsolidation (Gustavson's first possibility).

The bottomsets probably became overloaded (undercompacted) when the foresets prograded over them and they reacted to this load by faulting or flowage. Movement would have been generally towards the prodelta area, a zone of lesser pressure. This process is similar to, but on a much smaller scale than, the overloading of prodelta clays at large deltas, such as the Niger and Mississippi, where mass movement is towards the offshore area (Elliott, 1978, p. 138). Failure in the bottomset beds would disturb, and possibly cause a slump of, the overlying foreset beds.

Postglacial Emergence

The elevation of the topset/foreset contact decreases from the proximal to the distal portion of the delta. At the proximal end, the contact is 15 m above mean sea level and at the distal end, it is 12 m above mean sea level.

EAST OF LOWER FIVE ISLANDS DELTA

Description

East of the mapped delta margin between North River and Shad Brook is a thin postglacial deposit exposed in a 2 m bank along the shoreline. The base of the deposit varies from about 30 cm to 1 m thick and consists of flat lying red clay interstratified with silt and fine sand. The clay beds are usually 1 to 3 cm thick and the silt/sand beds are thinner. A sharp erosional contact separates the red unit from an overlying gravel unit. The contact is essentially horizontal except for a short interval on the east end of the exposure where the clay unit thickens to 1 m and the gravel thins from 1.5 to 1.0 m. The gravel is closed-work and massive but occasional discontinuous sand beds impart a rough horizontal stratification. The deposit does not extend inland and as such is not a mappable unit.

Interpretation

The upper gravel unit is similar in texture and structure to unit 3 of the topset beds at Five Islands and to the glaciofluvial gravel exposed in the deposits to the east of Five Islands. The gravel is interpreted as glaciofluvial in origin. The clay/sand rhythmites below the gravel are similar to bottomset beds exposed at deltas farther west in the terrace. This unusual sequence will be discussed later in the thesis when the model for the deposition of the terrace is proposed.

SEDIMENT INFILLED WEDGES, LOWER FIVE ISLANDS

Borns (1965, p. 1224) described ice wedge casts in the Minas terrace and noted that the largest ice wedge casts were found at Lower Five Islands. The identification of ice wedge casts is important because ice wedges form only in areas of permafrost (Bloom, 1978, p. 354). As formation of the outwash terrace requires large scale melting of glacial ice, Borns (1965, p. 1225) felt that the ice wedge casts formed in a cold spell following the deposition of the terrace and correlated the cold spell with the Valders readvance of the classical sequence. Therefore, Borns (1966, p. 53) and Swift and Borns (1967, p. 703) concluded that the deposition of the terrace was complete by Valders time.

In the fall of 1975, two sediment filled wedges were exposed at the distal end of the shoreline exposure of the delta at Lower Five Islands (Fig. 5.15). By the summer of 1977, the wedges had disappeared due to erosion of the bank. The wedges (3 m long x 2 m wide at top, 2.5 m long x 0.5 m wide at top) extended from the soil disturbed top of the topset beds to 1.5 m and 1.0 m, respectively, into the foreset

FIGURE 5.15. Sediment infilled wedges, distal part of exposure, Lower Five Islands. Wedges widest at top and pinch out in foresets.

beds before pinching out. The wedges were composed of topset gravel that was massive in the middle and vertically stratified along the sides. Some of the beds adjacent to the wedges were downwarped while others were undeformed. Borns' (1965, p. 1224) description and drawings of the ice wedge casts show the same characteristics as the wedges described above, including a vertical or near vertical stratification and the downwarping of adjacent beds.

In a recent review, Black (1976) suggests that many ice wedge casts have been misidentified. There are no unequivocal criteria for the identification of ice wedge casts outside present permafrost regions (Black, 1976, p. 11), so identification of these features must be regarded as tentative rather than conclusive.

There are several features of the sediment wedges at Lower Five Islands (including those described by Borns, 1965, p. 1224) that lend doubt to their origin as ice wedges. Deformed beds adjacent to the wedges are downwarped, whereas ice wedge casts typically have adjacent strata that are upturned (Black, 1976, p. 11). Walters (1978, p. 50) points out that seasonal frost crack wedges (no permafrost) typically have downwarped strata along the contact. Black (1976, p. 12) suggests that ice wedge casts should have a horizontal stratification whereas vertical stratification or fabric is more common to tensional crack or seasonal frost crack infills (Black, 1976, p. 18). However, Walters (1978, p. 51) regards the stratification type as nondiagnostic.

Probably the largest source of doubt as to their origin as ice wedge casts is the lack of polygonal structure. Borns (1965, p. 1224) states that it was impossible to determine if the casts were parts of polygonal fracture systems. The sediment wedges observed by the author were single wedges at least 100 m apart along the exposure. These wedges were first observed in the fall of 1973 and they were single wedges then as they were in 1975 after significant shoreline erosion (measured in metres) had taken place. As shoreline erosion gives a three dimensional view of the wedges, at some time the wedges should have been in pairs if they were part of a polygonal structure. The exceptions to this would be at the initial and last edges of the polygon, but the three dimensional exposure makes these possibilities unlikely.

Walters (1978, p. 52) tentatively identifies sediment wedges in New Jersey as ice wedge casts and the main criteria for doing so is the associated polygonal pattern. The absence of such a pattern at Lower Five Islands leaves reasonable doubt as to whether the sediment wedges are ice wedge casts. Therefore, the existence of postdepositional permafrost is uncertain as is the correlation of the wedges with Valders time.

OVERVIEW

LOWER FIVE ISLANDS

The southeastward dip of the foreset beds, the slope of the upper surface and the location of the delta indicate that the main discharge of meltwater and sediment was out of the Harrington River valley. Discharge from the North River valley was also probably important. The topset beds of the delta are similar in texture and structure, and therefore in process, to the glaciofluvial gravel exposed to the east of Five Islands. Streams proximal to the ice front were large (up to 20 m wide and 2 m deep) and incised. Channel deposits with coarse gravel throughout suggest periods of overtopping of the channels and flooding of the upper surface of the delta. Towards the distal part of the delta, the streams fanned out into a braided pattern. Grain size of the gravel topset beds decreases slightly in a distal direction while the foreset beds fine from gravel to sand. The ice margin receded northward during the deposition of the delta and released different rock types as it crossed bedrock boundaries.

Swift and Borns (1967, p. 696) found molds of *Portlandia glacialis* (synonym for *Portlandia arctica*, Wagner [1975]) in the delta at Lower Five Islands. The environmental implications of this pelecypod will be given in the discussion of the delta at Spencers Island.

The exposure at Lower Five Islands is the 'type' section for the Five Islands Formation (Swift and Borns, 1967, p. 703). Unit 3 of the topset beds appears to be equivalent to Swift and Borns' (1967, p. 700) Saints Rest Member while units 1 and 2 are equivalent to their topset beds. The stratigraphy of the Minas terrace, including the definition of the Saints Rest Member, will be discussed in chapter 14.

FIVE ISLANDS

A large (≃400 m x 150 m) irregularly shaped depression occurs just north of the highway in the deposit at Five Islands. It is probably a kettle that has had the west side eroded by the Bass River of Five Islands. This indicates that large blocks of ice persisted in the outwash until after deposition on the upper surface had ceased.

The slope of the upper surface and location of the deposit imply that meltwater was discharging out of the valleys of the East River and

Bass River of Five Islands. Gravel exposed in the pit along the road to New Britain is similar to unit 3 of the topset beds at Lower Five Islands and is interpreted as glaciofluvial in origin. There are no good shoreline exposures of the deposit but gravel exposed at the low bluff at Sand Point is glaciofluvial in structure and texture. This seems unusual because the topset/foreset contact is 4 m above the level of high tide at the distal end of the delta at Lower Five Islands. However, the deposit at Five Islands is relatively long (over 4.5 km) compared to the length of the delta at Lower Five Islands (just over 2 km). As the topset/foreset contact dropped 3 m in less than 1 km along the exposure at Lower Five Islands, the topset/foreset contact may have dropped below the present level of high tide at Sand Point.

It is known from the delta at Lower Five Islands that during the initial stages of the formation of the Minas terrace in this area, the sea was 12 to 15 m above the present sea level. Therefore, if bedrock and/or till is below that elevation for part of the deposit, deltaic deposition should have occurred. South of the highway the underlying bedrock is probably below the level of high tide ($\approx$8 m) as estuaries border the deposit on both sides. North of the highway the bedrock in the river valleys rises, so there is less probability of deltaic sediments. Conclusive proof of a strictly glaciofluvial or a glaciofluvial deltaic origin can only be attained by exposure of the internal structure of the deposit. Based on the sea level data from the delta at Lower Five Islands, the deposit is tentatively mapped as a delta.

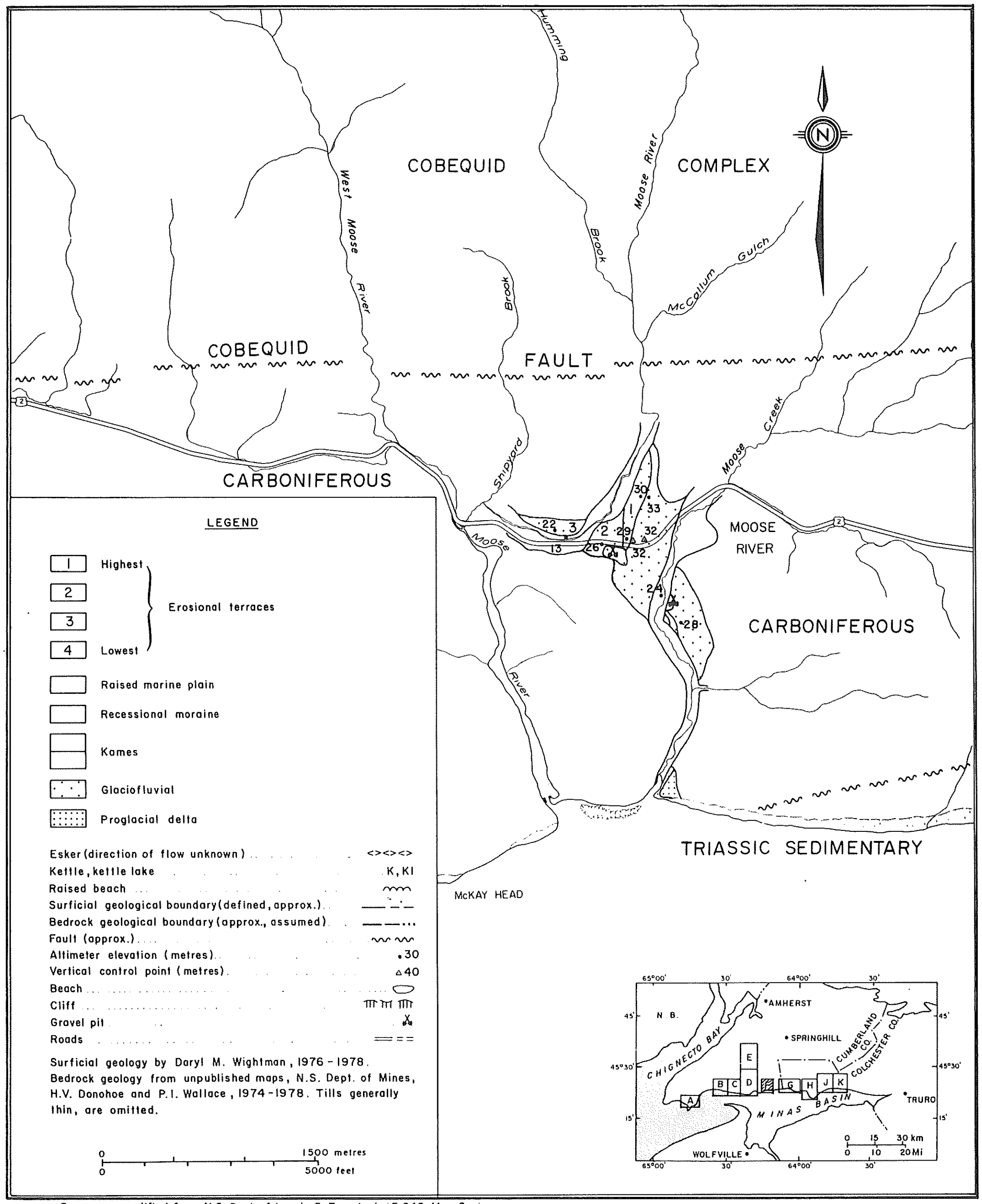
CHAPTER 6. MOOSE RIVER

SURFICIAL GEOLOGY

Moose River is a small community of several houses located approximately 7 km west of Lower Five Islands, 10 km east of Parrsboro and 2 km north of the Minas Basin. The surficial geology consists of an outwash plain at the site of the community and a very small outwash deposit at the mouth of Moose Creek (Fig. 6.1). Swift and Borns (1967) do not mention Moose River per se but they do map a deposit there (Swift and Borns, 1967, p. 695). They do not use it for terrace elevations (p. 709) so it is not known if they mapped the deposit as deltaic or glaciofluvial.

The inland outwash plain lies along, and on a bedrock low between Moose Creek and Moose River, both of which flow over bedrock. From the elevations of the rivers and the upper surface of the plain, it appears that the maximum thickness of the outwash is about 10 m. Bedrock is the Upper Carboniferous Parrsboro Formation, which also underlies the small deposit at the mouth of Moose Creek. The plain reaches an elevation of 33 m to the north and slopes down to 28 m in the south, while the surrounding bedrock hills reach minimum elevations of 75 m. Bedrock elevations rise rapidly at the Cobequid Fault to 225 m or more. The upper surface of the outwash plain slopes seaward at approximately 6.0 m/km (Fig. 6.2). There are three terraces below the upper surface adjacent to the Moose River and the first terrace has a seaward gradient of roughly 6.0 m/km, the same as the upper surface.

The small shoreline deposit occurs on the east side of the mouth of Moose Creek and rapidly thins out to the east against rising till, which



Map F. Surficial geology of the Moose River area, Cumberland County, Nova Scotia.

FIGURE 6.1.

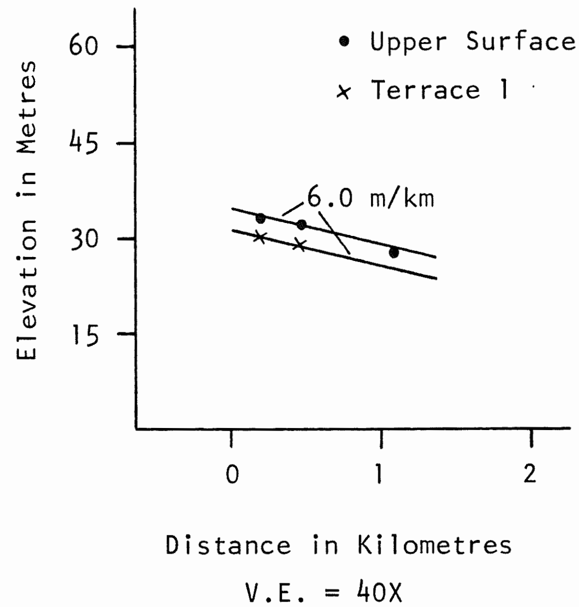


FIGURE 6.2. Gradients of upper surface and terrace level 1 of glaciofluvial gravel at Moose River. Profile essentially north-south.

also pinches out, and bedrock. No elevation was obtained for the upper surface.

SEDIMENTOLOGY

Exposure in the inland outwash plain was limited to gravel pits and these were not actively being used in the summer of 1977 so exposure was poor. The most recent pit is on the east side of Moose Creek and the upper 1.5 m was exposed in 1977. The gravel is closedwork with a crude horizontal stratification.

The shoreline deposit was poorly exposed in the summer of 1977 as talus extended well up the bank, but general observations could be made. The upper 5 m of the deposit is closedwork gravel with a crude horizontal stratification. Below this is 1 to 2 m of subhorizontally interbedded sand and gravel while the lower part of the deposit is composed of sand beds that dip from 10 to 25°S.

INTERPRETATION

The gravel exposed in the pit in the inland plain is similar in structure and texture to the glaciofluvial gravel exposed to the east of Moose River and accordingly, it is interpreted as a braided stream deposit. As bedrock underlies the gravel at a shallow depth, there is little chance of deltaic sediment occurring beneath the glaciofluvial gravel.

The shoreline deposit, although very small, is deltaic in structure. The upper coarse gravel beds are glaciofluvial in structure and texture whereas the lower beds are finer in grain size and have foreset type bedding. The 1 to 2 m 'mixed' interval that occurs below definite topset and above definite foreset beds has characteristics of both facies and is thus difficult to interpret. This disorganization is probably the result of the particular part of the delta that is exposed. The deposit is virtually the last vestige of the proximal part of a delta and the earliest deltaic sediments would be slightly chaotic, for slumping and effects due to proximity to ice would disrupt the sedimentation. From rates of shoreline retreat observed elsewhere along the Minas terrace (in the order of metres per year), one wonders how long this deposit will survive. The deltaic (foreset and bottomset) sediments may not extend to the northern edge of the deposit so shoreline erosion may in future leave only glaciofluvial sediments.

POSTGLACIAL EMERGENCE

The topset/foreset contact of the delta, placed as closely as circumstances permitted, is 16 m above mean sea level. Due to the location of the exposure (extreme proximal), there is some uncertainty as to the accuracy of this contact.

DRAINAGE IN THE MOOSE RIVER AREA

The inland outwash plain is situated along and between two river systems, Moose River and Moose Creek. The plain is the divide between the two rivers but the elongation of the deposit towards the south and the slope of the upper surface in the same direction implies discharge from the Moose River across the divide into Moose Creek. Terraces adjacent to the Moose River preclude observing the slope of the upper surface in that area.

The bedrock low over which the outwash lies and the alignment of Moose River north of the plain and Moose Creek south of the plain suggest that Moose River originally flowed directly southward. An eastern tributary of the West Moose River system probably captured the Moose River and diverted it westward to join the West Moose River. Moose Creek, formerly a tributary of the Moose River, is now the sole occupant of the valley eroded by the Moose River. River capture took place some time before the last glaciation as the elevation of the bedrock where Moose River flows westward (13 m) is much lower than the bedrock to the east over which Moose Creek flows (24 m).

OVERVIEW

Meltwater discharge deposited an inland outwash plain that covers a bedrock low between the Moose River and Moose Creek valleys. Some of the meltwater followed the original course of the Moose River and crossed the bedrock low to flow out of Moose Creek where a delta was deposited. The remaining meltwater probably flowed out of Moose River but if a delta was deposited there, erosion has completely removed it as well as almost all of the Moose Creek delta. After the upper surface of the outwash plain was deposited, stream flow from Moose River into Moose Creek was halted and the Moose River assumed its present course.

The structure and texture of the glaciofluvial sediments indicates a similar depositional process to that of other glaciofluvial sediments to the east. Relatively shallow, rapidly shifting braided streams laid down coarse gravel with lower flows depositing a sand matrix. The deltaic sediments are poorly exposed but the foreset facies is composed of sand while the topset facies is composed of gravel.

CHAPTER 7. PARRSBORO

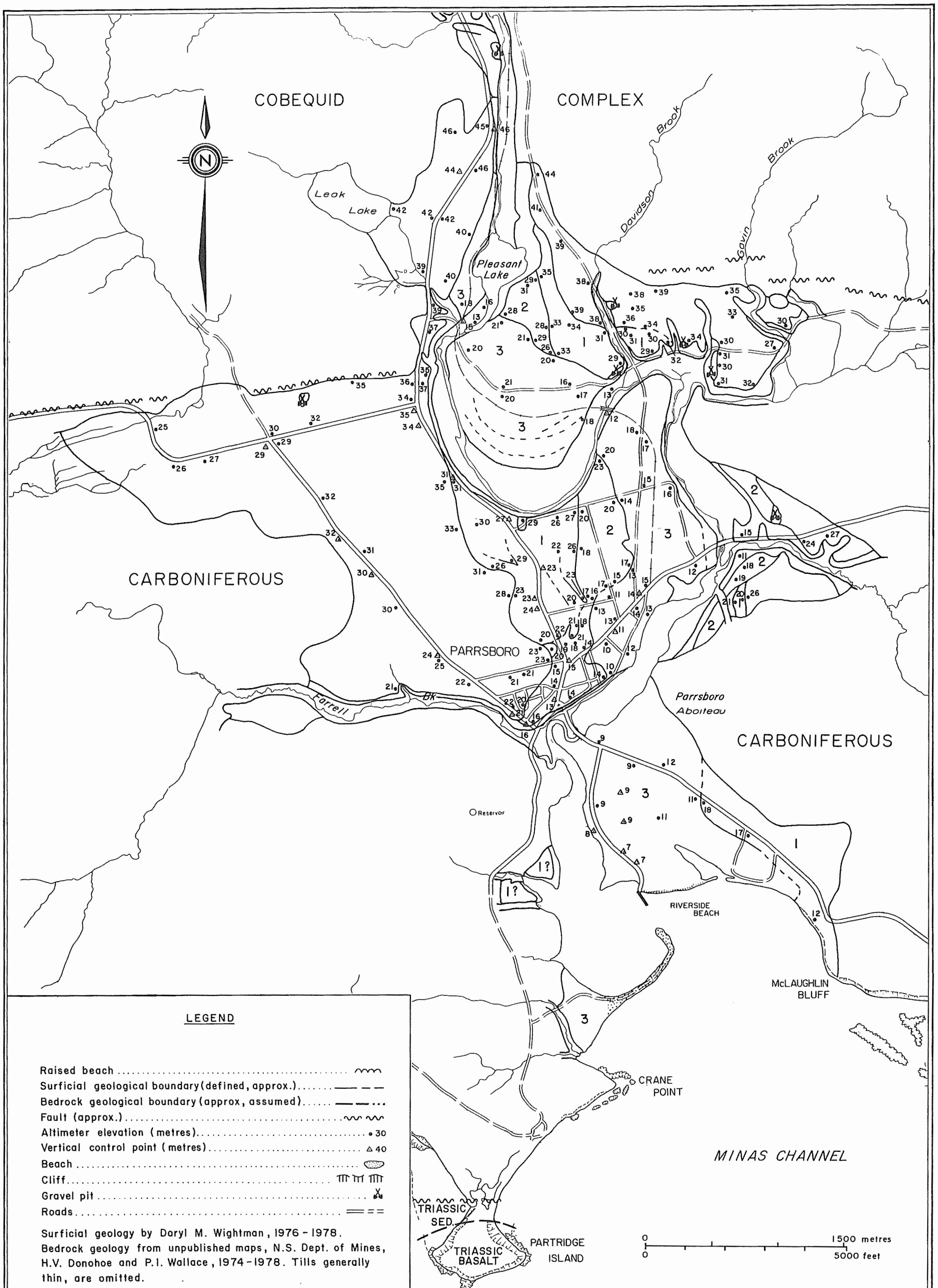
SURFICIAL GEOLOGY

Parrsboro is located in the western half of the north shore of the Minas Basin, about 80 km west of Truro and 40 km east of Advocate Harbour at the west end of the Minas Basin. However, it is centrally located within the deltas of the Minas terrace, being about 40 km from Advocate Harbour to the west and from Economy to the east.

The surficial geology at Parrsboro comprises one of the largest deltas in the outwash terrace (up to 6 km long and 6 km wide), shown on map D in the map pocket at the back of the thesis. The main delta forms a broad flat plain that lies between the Cobequid Fault and the Parrsboro Harbour. In the Parrsboro Gap, the delta extends approximately 2 km north of the fault on the west side of the valley and almost that far on the east side. It terminates where the Parrsboro Gap narrows to a width of roughly 0.3 km between the 150 foot (46 m) contours on the 1:50,000 N.T.S. map 21 H/8.

South of the Cobequid Fault, the underlying bedrock is the Upper Carboniferous Parrsboro Formation. North of the fault on the west side of the valley the deltaic sediments overlie a thin strip of Silurian to Devonian Portapique-Parrsboro Succession and north of that, the Silurian(?) Advocate Succession. On the east side of the valley north of the fault it overlies the Portapique-Parrsboro Succession.

The central part of the delta reaches an elevation of 39 m at the Cobequid Fault and inland along the Parrsboro River it rises to 46 m (Fig. 7.1). On the west side of the Parrsboro River, the fault line



Base map modified from N.S. Department of Lands & Forests 1:15,840 Map Series.

FIGURE 7.1. Map D'. Elevation data for the Parrsboro area, Cumberland County, Nova Scotia.

scarp attains elevations of 90 m or more and bedrock is hilly to the north while on the east side of the river, a high plateau (225 m) lies immediately north of the fault. Along the westernmost part of the delta, the delta thins out over a bedrock lowland below 30 m in elevation and here the margin of the delta is conjectural. South of this the Parrsboro Formation is hilly, reaching elevations of over 100 m, and the south-western limit of the delta is constricted.

Hills of the Parrsboro Formation also border and indent the eastern margin of the delta, especially to the south, resulting in an overall narrowing of the delta towards Parrsboro Harbour. The delta extends out into the valley of the East Parrsboro River and is roughly 15 m above the valley floor. South of this the delta margin follows the Parrsboro River but crosses the river at Newton Brook. Between Newton Brook and the East Parrsboro River, bedrock reaches over 60 m in elevation. South of Newton Brook the bedrock is lower, reaching 50 m, and the delta widens slightly at the Minas Basin shoreline. Two small deposits of the delta occur on the west side of the Parrsboro Harbour, one at Salter Brook and the other at an unnamed brook between Salter and Farrell Brooks.

The upper surface of the delta slopes seaward but also towards the west and the east away from the higher central portion. It attains an elevation of 46 m on its proximal end and slopes down to an elevation of 20 m north of the estuary and 12 m east of the estuary, if that surface is the upper surface. The seaward gradient of the upper surface (Figs. 7.2, 7.3) is roughly 5.0 m/km.

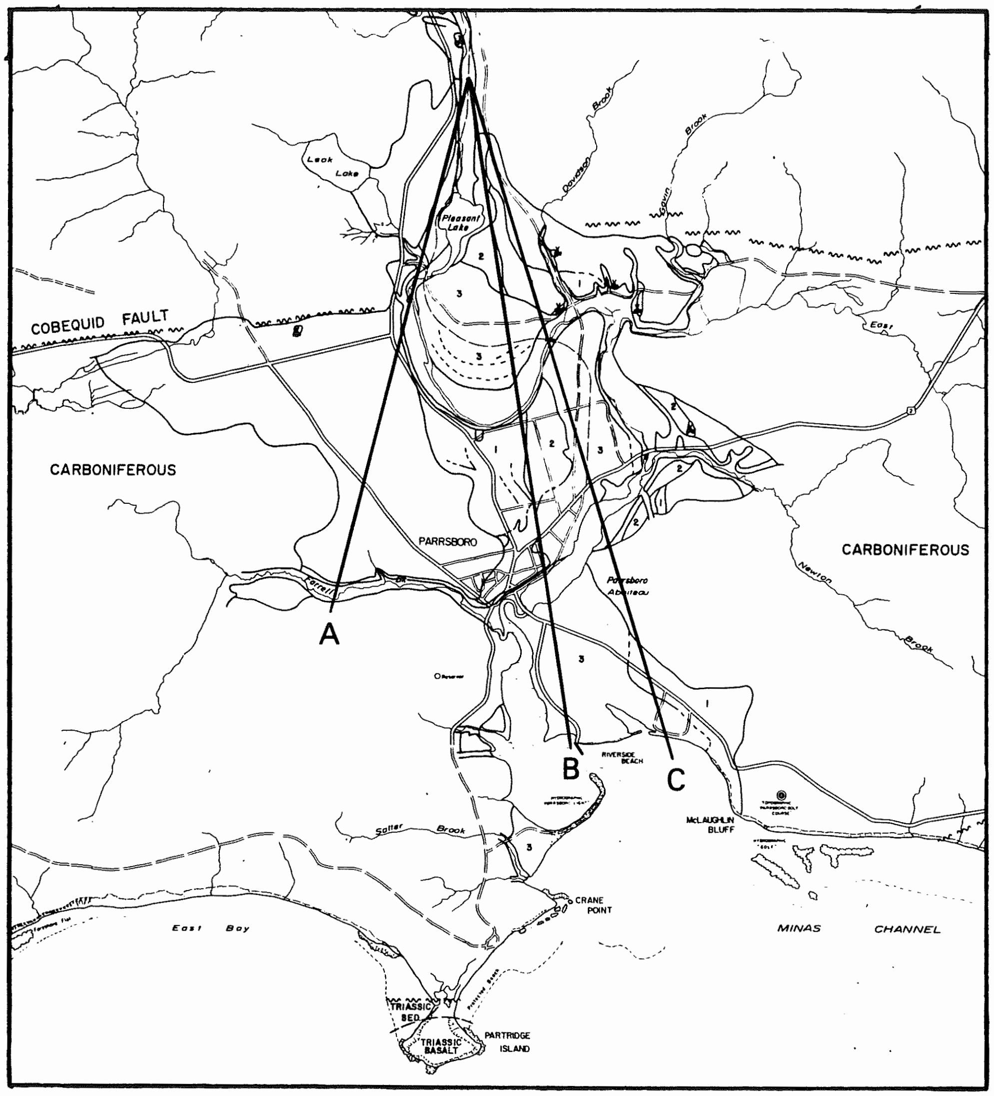


FIGURE 7.2. Locations of profiles A, B and C, figure 7.3.
Scale = 1:75,000.

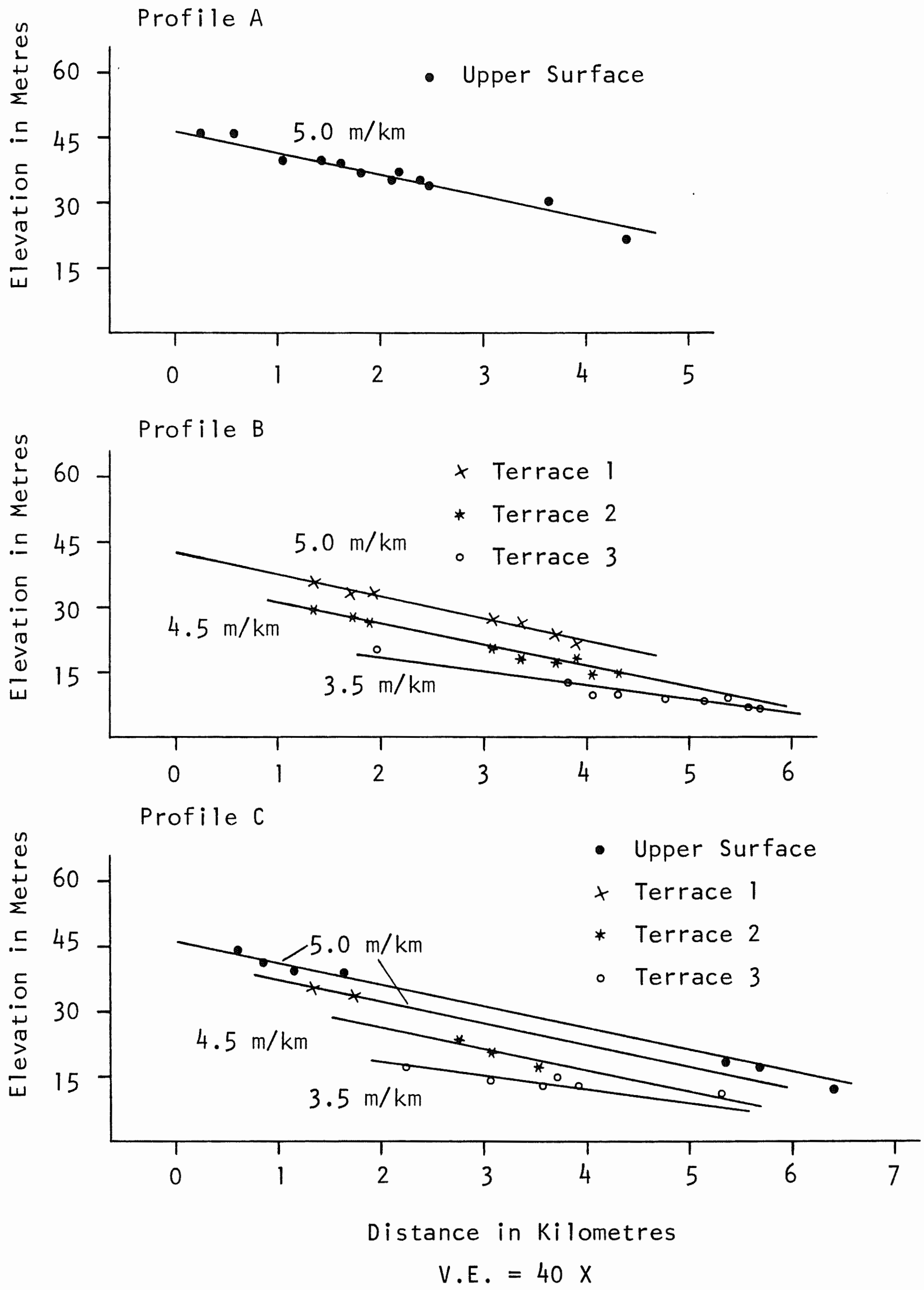


FIGURE 7.3. Correlations and gradients of the upper and terraced surfaces of the delta at Parrsboro.

TERRACES

The most noticeable feature of the Parrsboro delta is a well developed system of terraces. The terraces are generally wider than the terraces at other deposits and are fairly easy to distinguish (Figs. 7.4, 7.5) except where man has altered the landscape (e.g. in the town). Three terraces are developed on both sides of the river but the main terraces on the east side are north of those on the west side. The terrace scarps generally range from 3 to 6 m in height and the southern ends of the scarps are usually only slightly lower than the northern ends.

The first order terrace on the west side of the river has more relief than is characteristic of terraces or upper surfaces because of several isolated highs and paleochannels. The third order terrace displays a somewhat rolling surface with a higher part in the central region that is almost as high as the second order terrace. On the east side of the river, the third order terrace is the largest and in the southern part of this terrace there are ridges that follow the meander bend of the present river. Several terraces occur on the portion of the delta at the mouth of Newton Brook and on the east side of the Parrsboro estuary, there are two surface levels, with the upper one possibly representing the upper surface of the delta (Fig. 7.3). The gravel over much of this surface might be thin as bedrock appears to lie at a shallow depth. The lower terrace, which is at or just above the level of high tide along the edge of the estuary, correlates fairly well with the third order terrace to the north (Fig. 7.3).

FIGURE 7.4. Northward view along scarp between terrace 2 (lower) and terrace 1 (to left). Photograph taken at south end of terrace 1.

FIGURE 7.5. Northward view of junction between terraces 1 and 2, roughly 0.5 km east of south part of Pleasant Lake. Photograph taken from road on terrace 3.

The seaward gradients for the first and second order terraces, 5.0 m/km and 4.5 m/km, are approximately the same as for the upper surface. The gradient calculated for the third order terrace is 3.5 m/km, but this is very approximate as the gradient is lower and the terrace is not as flat as the others. This is noticeable just north of the present meander in the Parrsboro River where the elevations tend to decrease towards the southeast but the slope is interrupted by ridges that follow the meander bend.

The gradients were calculated in a north-south direction but it is evident from the elevations that the terraces have an east-west component in their gradients. The third order terrace in the south tilts towards the estuary (west) as well as sloping to the south and other terraces have a lower (first order terrace west of river) or higher (third order terrace west of river) central portion.

LAKES

There are no kettles evident on the surface of the delta at Parrsboro but there are two lakes of substantial depth on or partly on the delta. Pleasant Lake is part of the Parrsboro River system and as such would be expected to have a depth comparable to the Parrsboro River. The surface elevation of the lake is roughly 13 m but the lake attains a depth of 8 m (Fig. 7.6) so that the bottom of the lake is only 5 m above mean sea level, well below high tide. The Parrsboro River is 1 to 2 m deep near the lake.

Leak Lake occurs on the northwest edge of the delta at an elevation of just under 42 m. It has one small brook entering it from the northwest

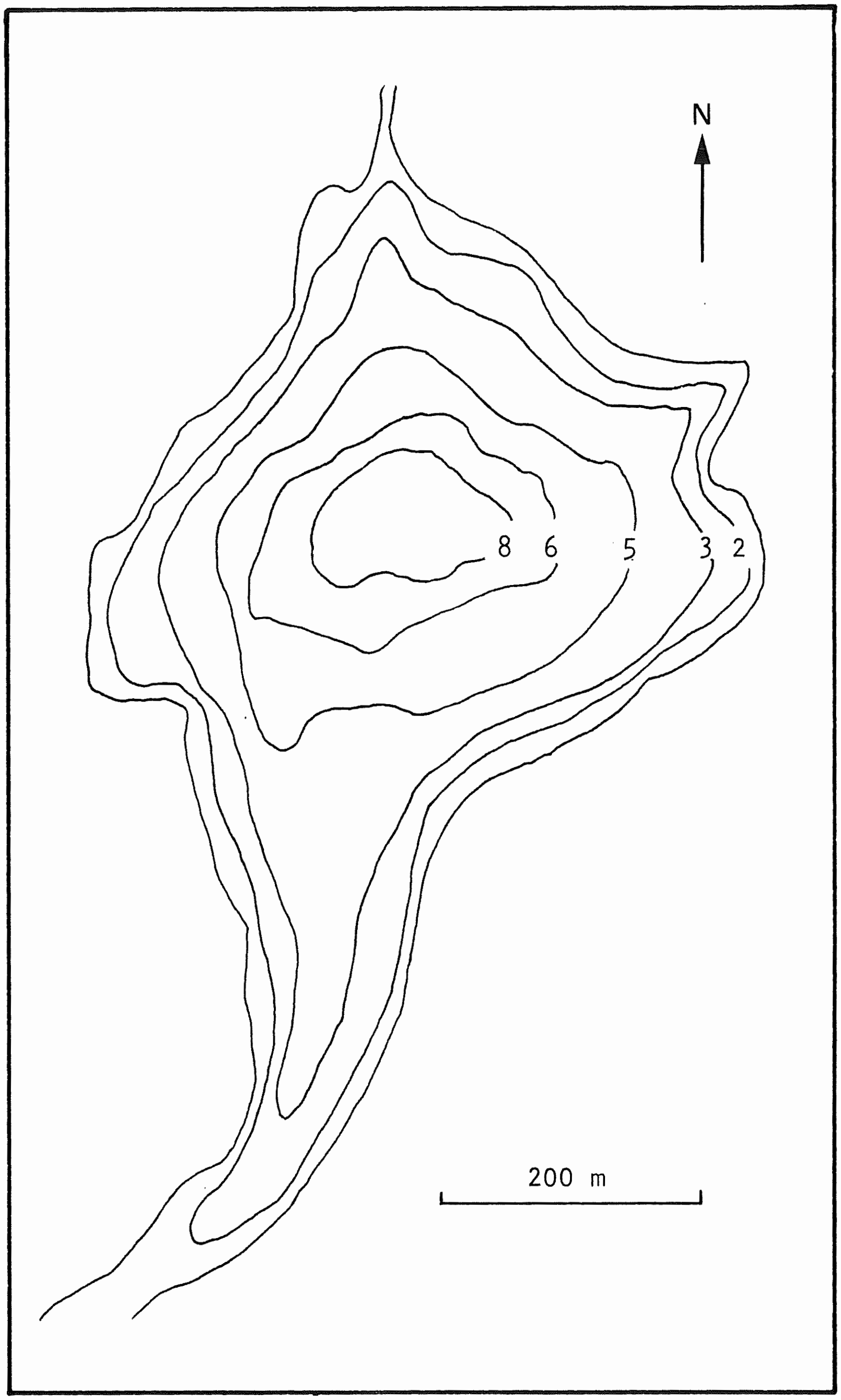


FIGURE 7.6. Bathymetry of Pleasant Lake. Five foot contour intervals have been converted to nearest metre. (After N.S. Dept. of Lands and Forests unpublished report 73019.)

and a small brook tributary to the Parrsboro River flows from it to the southeast. The watershed area for the lake is small as it is rimmed by the Cobequid Highlands on all sides, except the southeast side which faces the delta. There is a shallow bathymetric gradient on the east side, but very steep gradients elsewhere and the maximum depth is over 13 m (Fig. 7.7). The author cored the complete postglacial lake sediment column once and took three short cores across the postglacial/glacial sediment boundary, all of which are discussed in chapter 16.

SEDIMENTOLOGY

The delta at Parrsboro is poorly exposed because the bulk of it is inland, with little delta being eroded by the sea. The small deposit at the mouth of the unnamed brook between Farrell and Salter Brooks has red clay bottomset beds exposed along the beach and red clay bottomsets are also poorly exposed at several locations along Farrell Brook. Exposures in the main delta are in gravel pits and the extent and quality of exposure is dependent upon how recently it has been excavated.

The only gravel pit (Durant's pit) that exposes some of the marine part of the delta (foreset beds) is at the southern end of the first terrace east of the Parrsboro River where 5 m of foreset beds were exposed in the summer of 1976 (Fig. 7.8). The gravel foreset beds are finer grained than the topset beds and some of the gravel is openwork (Fig. 7.9). Dip is about 25° to the south. The topset beds are 5 m thick and are composed of closedwork (sand matrix) gravel that is coarser than the foreset beds. In the lowest metre, the gravel is finer with some cross stratification and cut and fill structures are abundant. The upper

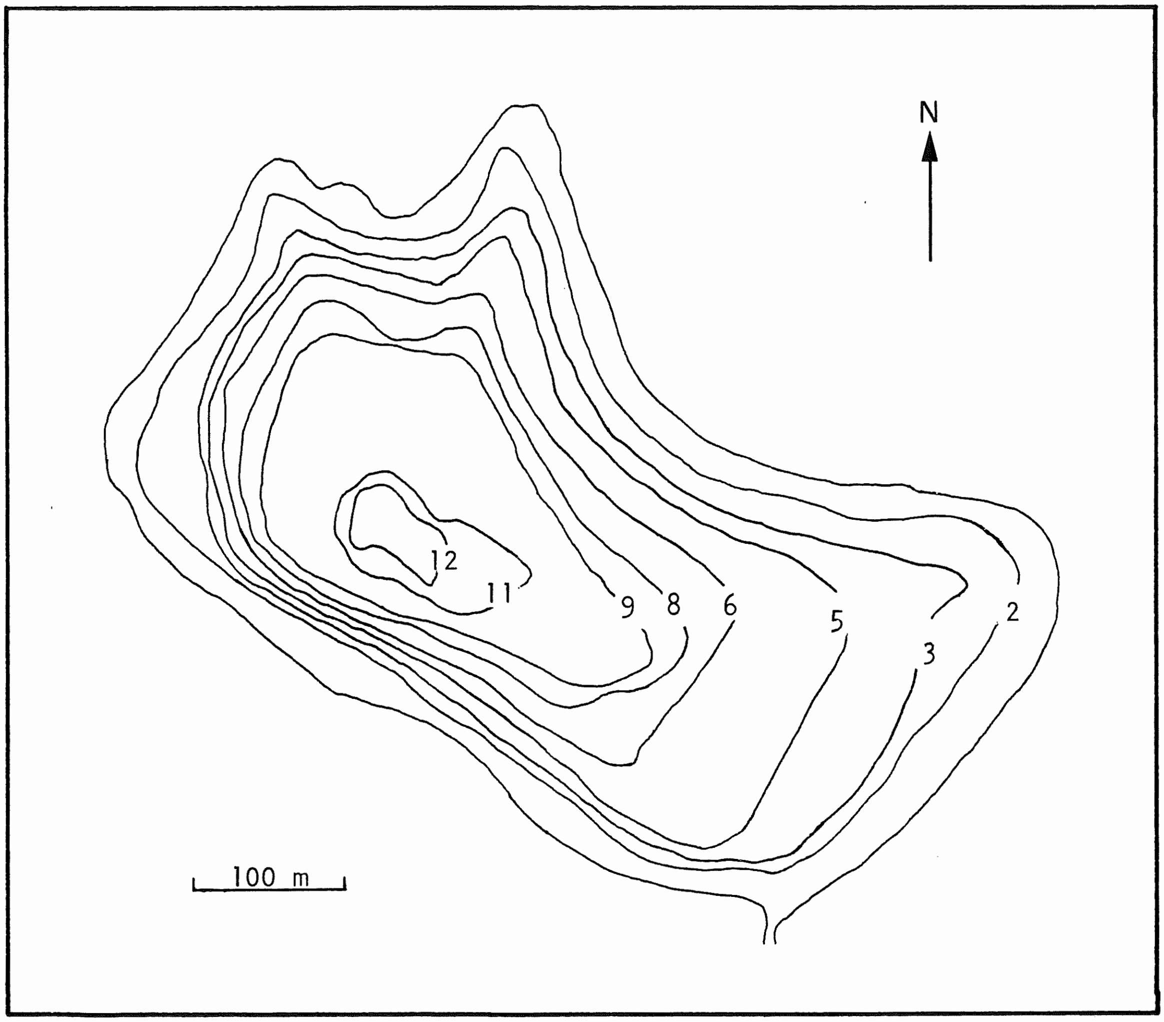


FIGURE 7.7. Bathymetry of Leak Lake. Five foot contour intervals have been converted to nearest metre. (After N.S. Dept. of Lands and Forests unpublished report 73018.)

FIGURE 7.8. Topset and foreset beds exposed in Durant's pit, Parrsboro. Topsets very coarse towards top, foresets dip south. Note vertical weathering profile of topset beds. Scale is 1.5 m.

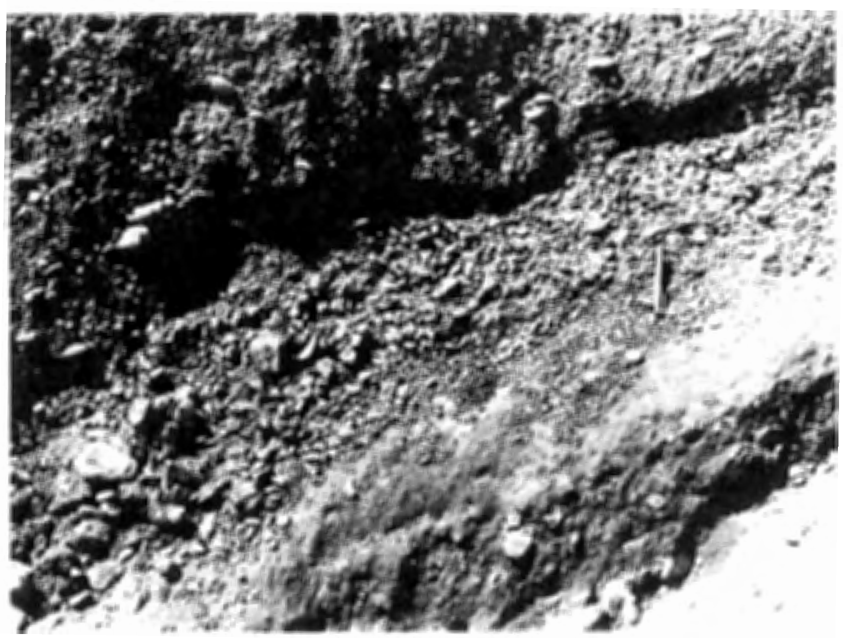


FIGURE 7.9. Openwork foreset gravel at Durant's pit, Parrsboro. Pen for scale.

2.5 m is very coarse and massive, with boulders up to 40 cm in diameter. The topset/foreset contact is horizontal and planar.

In two gravel pits to the east of Durant's pit, only glaciofluvial gravel is exposed. The top of both pits is at the upper surface of the delta. The first pit (Gilbert's pit, north of the road) was started in the summer of 1976 and has pit faces up to 8 m high. The gravel is closedwork with a medium to coarse grained sand matrix, and finer grained ($Mz = -3.22\phi$, sample 76-11) than the topset gravel at the top of Durant's pit, although it also coarsens slightly to the top. Some sand beds, generally lens shaped, are present. Large boulders (up to 40 cm x 30 cm x 15 cm) do occur but are not common. Crude horizontal stratification is predominant but there are some southward dipping tabular cross beds, up to 1 m thick, and possible channel shaped features (up to 4 m x 1 m). Bedrock, and perhaps some till, appears to underlie the pit.

The pit farthest to the east (Kernohan's pit) is finer grained and better stratified than the previous pit and is not as deep (≃6 m). The lower two thirds of the exposure is predominantly thickly bedded (0.5 to 1 m) and the bedding planes are erosional. A less discernible horizontal stratification with fewer sand beds exists in the upper part of the pit.

Another pit (Forkenham's pit) is situated north of the aforementioned three pits and occurs on the upper surface of the delta. It is less than 4 m deep and has not been excavated recently so exposure is poor. The gravel is massive and very coarse, like the upper topset beds in Durant's pit directly to the south. Channel shaped deposits up to 1 m deep and several metres wide occur within the gravel.

INTERPRETATION

The glaciofluvial gravel of the Parrsboro delta is similar in structure and texture to that of the other deposits to the east. The coarse grain size, rudimentary bedding and cut and fill structures imply deposition by fast flowing, braided streams. There are some tabular cross beds and they appear to have been deposited in an eroded low or paleochannel. This is similar to the cross beds at Little Bass River and supports the theory suggested at Little Bass River and Lower Five Islands that flow depth exceeded channel depth at least periodically. Except for the top of Durant's pit, the coarsest gravel occurs in the pit (Forkenham's) that is the most proximal, as is expected.

The gravel at the top of Durant's pit is almost too coarse, with numerous clasts over 20 cm in diameter, to be fluvial. Some of the boulders may have been transported in ice blocks, as the proximity to, and presence of, ice has been clearly indicated in several of the deposits to the east. However, very large clasts (up to 1.8 feet [55 cm], b-axis) can be transported by glacial meltwater streams (Fahnestock, 1963, p. A30). The problems of large clasts in fluvial sediments and the scatter of data about the predicted grain size of bedload are discussed by Church and Gilbert (1975, p. 37-40) and will not be recapitulated here. While occasional large clasts may have been fluvially transported, their abundance and clast to clast contact may be partly explained by the location of the pit. The top of Durant's pit is a terraced surface and the coarse clasts in the outwash may have been concentrated into an erosional lag as the terracing took place.

In Durant's pit, the weathering profiles of the topset and foreset facies differ noticeably, as they do in the delta exposures at Lower Five Islands and Moose River. In the glaciofluvial topset beds, the coarse nature of the pebble framework, the sand matrix and the sub-horizontal to horizontal bedding combine to make the gravel relatively stable so that it has an almost vertical weathering profile. Glacio-fluvial deposits east of Five Islands confirm this steep weathering profile. In contrast, foreset beds weather to a much lower slope as the 25 to 30° dip of the beds, the finer grain size and common but not abundant openwork beds make the foresets relatively less stable. Talus frequently blankets the foresets while the topsets remain exposed.

The bottomset beds, where they are exposed, are much finer grained than at Five Islands. The bottomsets are composed of massive clay to interbedded clay/sand and the interbeds are in the order of several centimetres thick. Along Farrell Brook, bottomsets are exposed within several metres of the surface of the delta, indicating the presence of bedrock at a shallow depth. Near its western limit, the delta probably prograded into very shallow water.

POSTGLACIAL EMERGENCE

The elevation of the topset/foreset contact at Durant's pit is 22 m above sea level.

ORIGIN OF TERRACES

Goldthwait (1924, p. 100) observed the terraces in the Parrsboro delta and concluded that because the concave outlines of the terrace scarps face the Parrsboro River, rather than the shoreline, they had been cut by the Parrsboro River. Swift and Borns (1967, p. 699) made a reconnaissance map of the terraces and implied that the terraces were eroded by braided channels (p. 698).

The author agrees with both of these conclusions. The curved form of the terrace scarps and their general alignment parallel to the Parrsboro River suggest that the terraces were produced by lateral migration of the river during downcutting. The paleochannels that are evident on the tops of some of the terraces indicate an abundance of stream channels, which is an attribute of a braided stream. The isolated highs on some of the terraces (such as on terrace 1) are the result of incomplete erosion across the top of the terrace, which is more compatible with a braided stream rather than a meandering stream that swept across the top of the terrace. As well, the characteristics of the outwash indicate deposition by a braided stream and the coarse gravel beneath terrace 1 at Durant's pit is also consistent with this model.

The terraces generally slope both seaward (south) and towards the Parrsboro River (east or west). This, combined with the fact that the two major sets of terraces are not opposite each other, makes it difficult to determine if they are paired terraces. The terraces correlate fairly well (Fig. 7.3) but there is scatter amongst the data, which are only accurate to ± 1.5 m and the fit of the data is dependent upon the line of profile. As an example, on the west side of the river, terrace 1 reaches 31 m in elevation while terrace 1 on the east side of the river is considerably north of that, yet only reaches 35 m. The good correlation between the two terraces in figure 7.3 is because the terrace on the west side of the river slopes eastward to where profile B was taken. To call the terraces paired or unpaired depends upon the degree of detail in the correlation. In a broad sense, the two major sets of terraces are paired as each set has three terraces of roughly the same height. But in a detailed sense they are not paired because the heights on a particular terrace level do not follow a consistent pattern in relation to those on the opposite side of the river.

Over the proximal and mid delta areas, the gradients of terraces 1 and 2 are similar (5.0 and 4.5 m/km) to that of the upper surface of the delta (5.0 m/km). Therefore, the streams that flowed over those surfaces must have had a similar equilibrium gradient which implies similar flow characteristics. Terrace 3 has a lower gradient (3.5 m/km) which might reflect a decrease in stream energy.

The ridges on terrace 3 north of the meander in the Parrsboro River represent levees that were left behind as the meander shifted southwards. It appears that at that stage, the Parrsboro River was a meandering stream much the same as it is now. However, terrace 3 south of the Parrsboro Aboiteau displays several paleochannels, one of which forms the indentation on the shoreline east of Riverside Beach. It is possible that this surface represents part of terrace 2 as well as terrace 3 because there is considerable scatter of the data on this terrace (Fig. 7.3).

The terraces on the deposits elsewhere in the study area are similar to the terraces on the Parrsboro delta. They generally have a curved form and parallel the river that flows through the deposit. The terraces are considered analagous to the terraces at Parrsboro — that is, they were eroded by braided streams after deposition on the upper surface of the deposit had ceased. At every deposit, including the Parrsboro delta, there are places where the upper surface of the deposit drops abruptly down to the level of the present river without any intervening terraces. Where the river valley is narrow, this is the result of little lateral migration of the river during downcutting. Where the river valley is considerably wider than the river, the absence of terraces is due to the elimination of any former terraces by lateral erosion of the present river. The cause of the terracing and its implications will be discussed in a later chapter that outlines the postglacial model for the north shore of the Minas Basin.

ORIGIN OF PLEASANT LAKE AND LEAK LAKE

The location of Leak Lake beside the delta, the depth of the lake and the occurrence of kettles at the deposits east of Parrsboro suggest that a large mass of ice prevented glacial outwash from infilling that area. The depth of Pleasant Lake is more surprising. The lake is located in the central part of the valley, surrounded by outwash, and appears to be simply a widening of the Parrsboro River. However, the depth of over 8 m, compared to the 1 to 2 m depth of the adjacent reaches of the Parrsboro River, indicates that the lake is not a product of the river system. A postglacial sinkhole is one possible origin for the lake as calcareous siltstone outcrops west of Parrsboro where several small upfaulted blocks of the Carboniferous Windsor Group occur in the Parrsboro Formation (Peter Wallace, personal communication, 1979). However, the bedrock underlying Pleasant Lake is the Silurian(?) Advocate

Succession and the Silurian to Devonian Portapique-Parrsboro Succession which are supposedly not calcareous here. It appears unlikely that the depth of Pleasant Lake is the result of a postglacial sinkhole.

Another possible explanation is that Pleasant Lake is also a kettle lake. This requires a large, deep mass of ice that persisted until deposition had ceased not only on the upper surface of the delta but also on at least one, or probably more, of the terraces. This is a significant amount of time but earlier melting would result in the infilling of the depression. After the block of ice did melt, the kettle was sufficiently deep for subsequent fluvial deposition, continuing to the present day, to have shallowed but not infilled the depression.

A kettle seems to be the most plausible origin for Pleasant Lake, but it is problematical. The size and depth of Leak Lake, and possibly Pleasant Lake, suggest that large blocks of glacial ice were stranded in the outwash and did not melt until well after deposition in that area had ceased.

OVERVIEW

The kettle origin for Leak Lake, and possibly Pleasant Lake, indicates that the delta at Parrsboro was deposited in close proximity to dissipating glacial ice. Pleasant Lake occurs just north of the Cobequid Fault and if it is a kettle lake, it might indicate a former ice front position aligned with the Cobequid Fault. The delta fans out south of the fault and the constriction of the valley at the fault may have created an ideal location for an ice front. If this was the situation, the outwash adjacent to and north of Pleasant Lake is probably not deltaic but strictly glaciofluvial. Even if the ice front was farther north near the edge of the delta plain, bedrock may be high enough to have prevented deltaic sedimentation in all but the central part of the valley as the glaciofluvial gravel exposed in the pits east and north of Durant's pit appear to be underlain by bedrock and/or till.

The southward dip of the foreset beds at Durant's pit, the slope of the upper surface and the location of the delta indicate that the main supply of meltwater discharged out of the Parrsboro River valley. Some meltwater discharge was probably from the smaller surrounding streams, such as Newton Brook, but the slope of the upper surface is clearly related to the Parrsboro River valley. The bulk of the delta lies south of the Cobequid Fault in a bedrock low that narrows towards the shoreline and causes the delta to do the same.

To the east and west, the delta abuts or thins out over bedrock everywhere but in the East Parrsboro River valley where the delta ends abruptly in the centre of the valley. Because of its height above the valley floor (over 15 m) and abrupt termination, it appears that the delta was impeded from building eastward. An ice plugged valley seems to be the most plausible explanation for this and the flat valley bottom covered with fertile soil suggests that a lake may have occupied this site after the ice melted.

The similarity in texture and structure between the glaciofluvial gravel at Parrsboro and the glaciofluvial gravel to the east of Parrsboro suggests that the delta was built by fast flowing braided streams. The bottomset beds, where they are exposed, show that clay deposition was important in the prodelta areas. After deposition on the upper surface had ceased, fluvial erosion removed a large central portion of the delta and formed three major orders of terraces, all of which are above the present Parrsboro River. Relict channels and the topography of the terraces imply erosion by braided streams, which formed a coarse lag on the terrace surfaces. Correlation of terraces of the same order is tenuous as elevations vary across individual terraces.

CHAPTER 8. DILIGENT RIVER

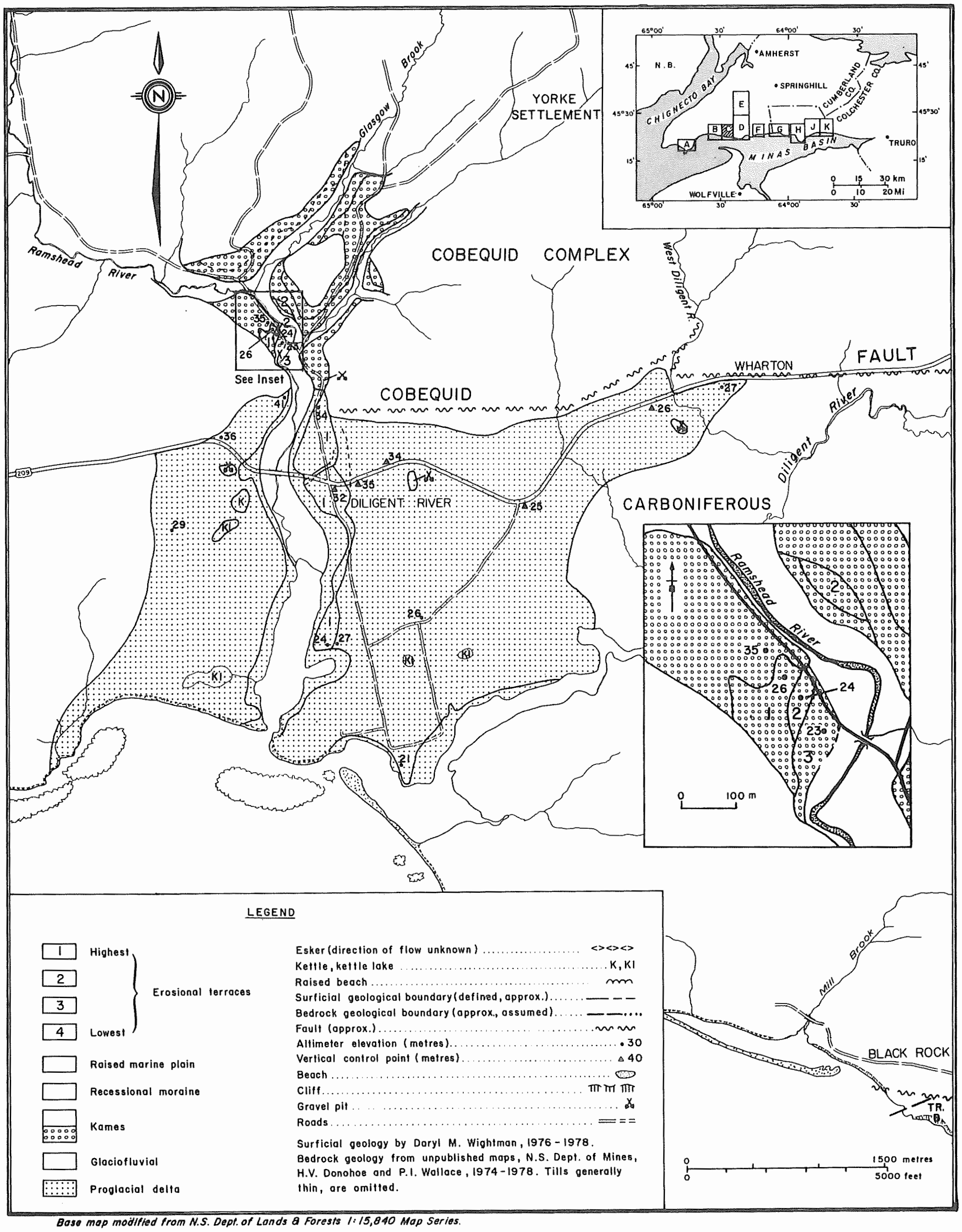
The surficial geology at Diligent River, located about 7 km west of Parrsboro, consists of two genetic units. An outwash delta forms a broad flat plain extending from the Cobequid Fault south to the Minas Basin on both sides of the Ramshead River. On the north side of the fault, adjacent to but up valley from the delta, is a series of hillside gravel deposits that extend into several tributary valleys of the Ramshead (map C, Fig. 8.1), but they were mapped only in the area immediately north of the delta.

DELTA

SURFICIAL GEOLOGY

Most of the delta lies on the east side of the Ramshead River, but it attains its highest elevation, 41 m, on the western part beside the Cobequid Fault. The Cobequid Highlands rise to over 135 m at the fault, so the northern boundary of the deposit is clearly defined. The delta slopes seaward from the Cobequid Fault but as it does so it also slopes westward and eastward away from the higher central area along the Ramshead River. In both these directions the delta thins out over a bedrock lowland that is below 30 m in elevation, making the margins of the deposit difficult to place. The delta overlies the Upper Carboniferous Parrsboro Formation.

A narrow but persistent terrace occurs on the east side of the Ramshead River and north of the highway a paleochannel is evident on its surface. The gradients of the upper surface and of the terrace are



Base map modified from N.S. Dept. of Lands & Forests 1:15,840 Map Series.

Map C. Surficial geology of the Diligent River area, Cumberland County, Nova Scotia.

FIGURE 8.1.

7.0 m/km and 6.5 m/km (Fig. 8.2). There are two groups of kettles on the delta and the largest group to the south has an east-west alignment. It consists of a kettle lake on the west and two kettle ponds on the east side of the river, while an embayment in the Ramshead estuary appears to be the remnant of a fourth kettle. Although the kettle ponds are small, the westernmost kettle pond on the east side of the river is almost 11 m deep.

SEDIMENTOLOGY

The delta at Diligent River is exposed poorly at several places along the shoreline and in three inland gravel pits. On the west side of the mouth of the Diligent River there are several exposures of red clay bottomset beds but topset and foreset beds are barely visible. Between the Diligent and Ramshead Rivers the delta is approximately 6 to 10 m thick and overlies bedrock. The bottomset beds are sandy and the topset beds (1.5 m thick) tend to be coarser than the southward dipping foresets. The exposure is poor and location of the topset/foreset contact has a fair degree of uncertainty.

There is semicontinuous but poor exposure from Ramshead River west to the first point (Christy Point) and the bank may be slumped along its entire length. The best part of the exposure occurs towards Christy Point where the delta overlies bedrock. The bottomset beds here are interbeds of clay/silt and the clay is frequently bright red on the bottom and darker on the top. The topset beds are gravel and the foreset beds are sandy, dipping towards the south.

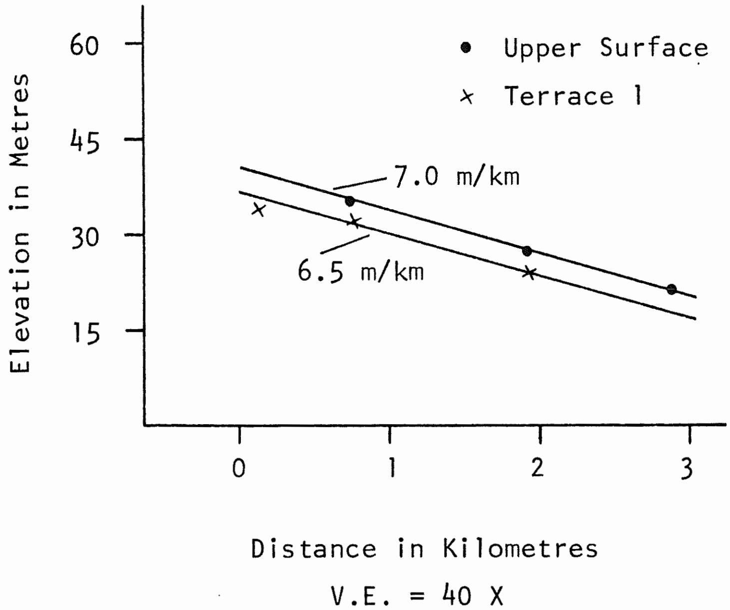


FIGURE 8.2. Gradients of the upper and terraced surface of the delta at Diligent River. Profile essentially north-south along east side of Ramshead River.

Along this exposure are some very peculiar blocks of black cemented gravel. The black gravel is in lens shaped bodies, approximately 4 m x 0.5 m, that tend to occur at the same height. Underlying and overlying gravel beds erode more rapidly, causing the lenses to stand out along the bank but some have slumped part or all of the way down the bank. All but one of the black lenses appeared to be topset gravel (the other was a black cemented foreset sand).

There are three gravel pits near and south of the highway, two on the east side and one on the west side of the Ramshead River. The pits have not been used recently but the gravel that is exposed is closework and massive to crudely horizontally bedded.

INTERPRETATION

The texture and structure of the gravel exposed in the inland pits are similar to those of the glaciofluvial gravel elsewhere in the terrace and a similar depositional environment, that of braided streams, is inferred. As at Parrsboro, clay deposition was important in the pro-delta area. Clay deposition was greatest in deeper water (more than several metres) because where the delta overlies bedrock and the contact is part way up the bank, the bottomsets are sandy. The peculiar black cemented gravel also occurs at Port Greville and will be discussed in that chapter.

POSTGLACIAL EMERGENCE

The topset/foreset contact was measured at the exposure on the east side of the headland between the Ramshead and Diligent Rivers. As mentioned previously the structure is not well exposed and location of the contact involves some degree of uncertainty. The elevation of the 'contact' is 18 m.

HILLSIDE GRAVEL DEPOSITS

SURFICIAL GEOLOGY

The gravel deposits north of the delta have both a hummocky upper surface and a relatively flat upper surface. The deposits flank rivers and lie against the hills of the Silurian(?) Advocate Succession in the Cobequid Highlands. Both types of deposits lie below 60 m while the surrounding bedrock rises to over 120 m.

Immediately north of the delta on the east side of the valley, the gravel deposits are hilly with some of the hills over 50 m in elevation. North of this, between the unnamed brook that flows through Yorke Settlement and Glasgow Brook, and on the north side of the Ramshead River-Glasgow Brook confluence, there are gravel terraces with relatively flat tops lying between 46 and 61 m in elevation (from 1:50,000 topographic map). The terrace between the unnamed brook and Glasgow Brook has a partial erosional terrace in the north central part.

An interesting set of low level terraces occurs just west of the confluence of the brook that flows through Yorke Settlement and the Ramshead River (inset, Fig. 8.1). The terraces are fairly flat and are composed of gravel. The scarps below terraces 1, 2 and 3 are curvilinear and parallel to the Ramshead River (Fig. 8.3) but the northern edge of terrace 1 (elevation 26 m) is indented with two embayments that open south. The embayments are perhaps slightly lower than the rest of the terrace (air photo interpretation).

Immediately north of terrace 1 is a series of hilly gravel deposits that rise to over 25 m. Part of the gravel deposits sweeps around and encompasses the east side of the terrace, although there is very little of this part left due to erosion by the Ramshead River and to the construction of the gravel road. Another terrace on the east side of the valley across from these terraces has an undetermined elevation and a shape similar to terraces 2 and 3. It is tentatively correlated with terrace 2 (air photo interpretation).

SEDIMENTOLOGY

There were no exposures in the gravel deposits until the summer of 1977 when one of the hilly deposits just north of the delta on the east side of the Ramshead River was excavated (Fig. 8.4). The deposit is composed of bouldery gravel with large foreset beds dipping towards the west. The pit is not exposed well enough to observe the sedimentological relationships at the top of the foreset beds but if a topset/foreset contact exists, it appears that it is not constant in elevation.

INTERPRETATION

The hillside gravel deposits are interpreted as ice disintegration features (Flint, 1971, p. 207) that fit into the subdivision of ice contact stratified drift. Although there is only one exposure to prove that the deposits are composed of stratified drift, the surficial sediment is gravel and the topography and location of the deposits are consistent with ice contact stratified features.

FIGURE 8.3. Southwestward view along scarp below terrace 1, just north of Diligent River. Tree in centre of photograph on terrace 2.

FIGURE 8.4. Westward dipping foreset beds in gravel deposits immediately north of delta on east side of Ramshead River.

The relatively flat topped deposits on the north side of the Ramshead River-Glasgow confluence and between Glasgow Brook and the brook that flows through Yorke Settlement are interpreted as kame terraces. Kame terraces are deposited by streams flowing between a glacier and the side of a valley (Flint, 1971, p. 209). The situation is slightly different for the southernmost terrace where streams deposited drift over a bedrock low between two valleys. Kame terraces are constructional forms but a small incised terrace on the corner of the southernmost kame terrace shows that the level of the streams dropped and part of the original surface was eroded. This drop in stream level was probably due to the melting of the valley ice.

The hummocky deposits occur in the same position as the kame terraces, along the sides of the valleys, and are therefore probably similar in origin. The gravel pit in the deposit just north of the delta (Fig. 8.4) exposes foreset beds dipping to the west which indicates that streams flowed westward down the side of the valley and deposited a kame delta in a ponded body of water. As the kame terraces formed along ice filled valleys, the pond probably formed between ice in the valley and the side of the valley.

Many of the other hillocks are probably not kame deltas. They may represent kame terraces that contained large blocks of ice that later melted, leaving a hummocky topography. Alternatively, the kame terraces may have been floored by ice which later melted, disrupting the upper surface. The hillocks may contain sediment deposited by flowing water, standing water or ice. Without exposures it is impossible to know the origin of every hillock so the general term kames is appropriate. An ice filled valley and relatively ice free hills was probably common to all the kames.

ORIGIN OF LOW TERRACES NORTH OF DELTA

The terraces that occur near the confluence of the unnamed brook that flows through Yorke Settlement and the Ramshead River are unusual and interesting. Terraces 2 and 3 appear to be cut by the Ramshead River as their scarps are curvilinear and parallel to the Ramshead River. However, the northern edge of terrace 1, which is a contact with kames, is oriented more east-west. It appears that the two embayments opening south were not eroded by the Ramshead River, especially as part of the kames extends around the eastern side of the terrace. Whether terrace 1 is erosional or depositional is not clear but it must have been formed by water which did not follow the present course of the Ramshead River. The terrace does not appear to postdate the kames, as it does not grade upward into them, but rather the kame material appears to postdate the terrace. This suggests that the last ice in the kames melted out after the deposition of the terrace and that the terrace was deposited by meltwater. Terraces 2 and 3 were later eroded into terrace 1 by the Ramshead River. The sediment is mapped as kame material although the more general term outwash may be more appropriate.

OVERVIEW

The alignment of the kettles in the southern part of the delta may indicate a former ice front or large strip of ice that spalled off the former ice front. This seems to be quite far south for an ice front position as it is near the southern limit of the delta. If it is a former ice front position, the sediments north of that are probably not underlain by deltaic (foreset and bottomset) sediments.

The southward dip of the foreset beds and slope of the upper surface of the delta indicate that the main supply of meltwater was discharging out of the Ramshead River valley. Secondary discharge out of the West Diligent River valley was probably important. Up valley (Ramshead River) from the delta, kames and kame terraces formed along the sides of the valley between the bedrock and glacial ice in the valley. Areas of outwash may have formed in front of some of the kames. The Ramshead River later eroded terraces in some of the up valley sediments as well as a narrow terrace and its present valley through the delta.

CHAPTER 9. PORT GREVILLE AREA

Port Greville is located approximately 6 km west of Diligent River and 15 km east of Spencers Island. The surficial geology consists of an outwash delta complex with hillside gravel deposits that extend up valley from one of the deltas (map B, Fig. 9.1), a situation similar to that at Diligent River.

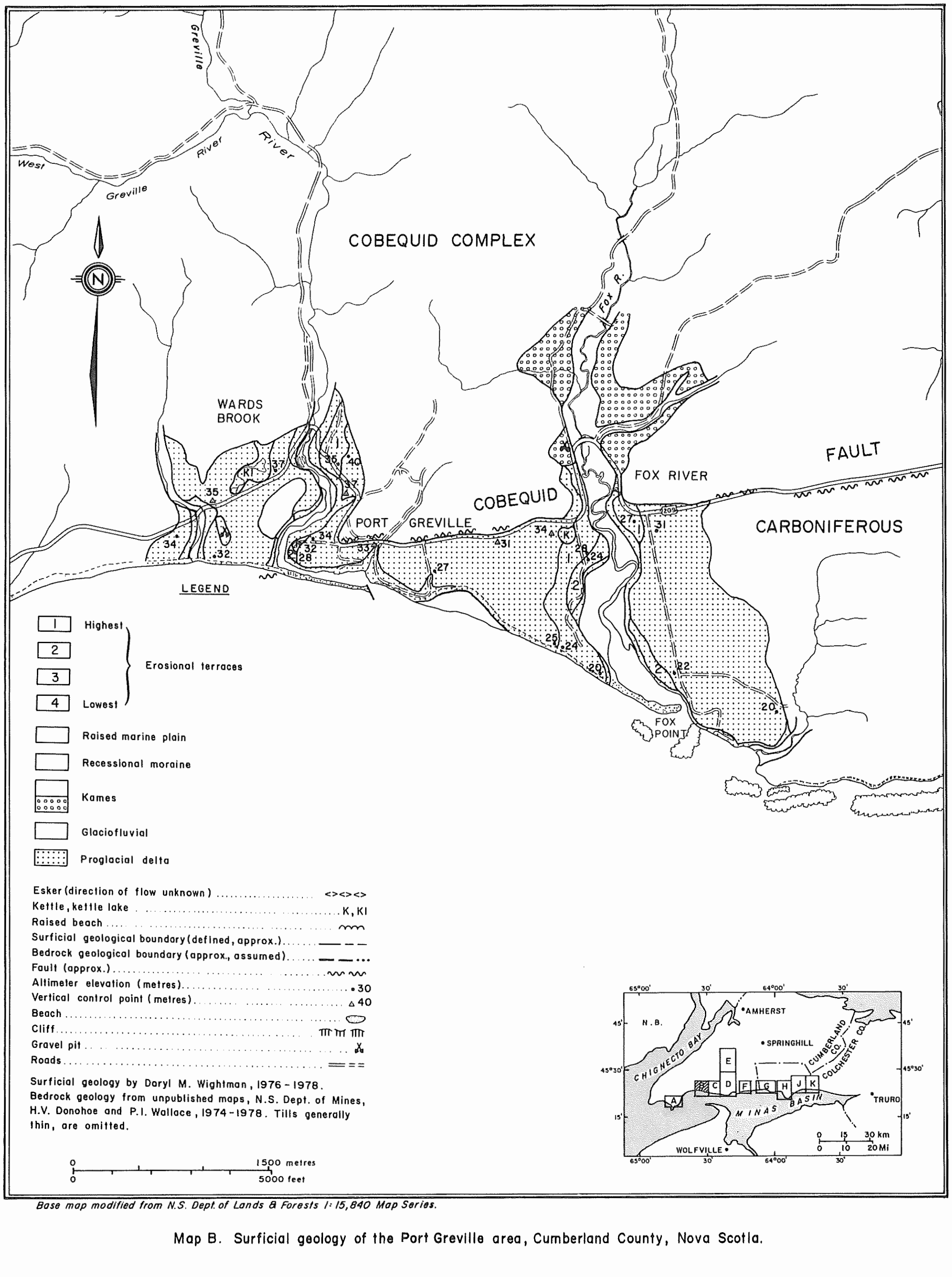
DELTA COMPLEX

SURFICIAL GEOLOGY

The delta complex extends from Wards Brook in the west to Fox River in the east and consists of two main deltas, lying near the mouths of the Greville and Fox Rivers, connected by a narrow (less than 0.5 km) belt of deltaic sediments in Port Greville. Erosion by the Greville and Fox Rivers has divided the complex into three parts, but elevations of the upper surfaces show the continuity of the deposits.

The delta complex lies immediately south of the Cobequid Fault in Port Greville and Fox River. The maximum height of the complex is 34 m, while the Cobequid Highlands rise abruptly to 100 m or more at the fault. The deltaic sediments are fairly thick between Port Greville and Fox River (>12 m along the shoreline). East of Fox River they overlie the Upper Carboniferous Parrsboro Formation and pinch out to the east against it. Bedrock is below 30 m in this area.

In Wards Brook, the deltaic sediments cross the Cobequid Fault, which lies seaward of the shoreline, and overlie the Silurian(?) Advocate Succession. To the west, the sediments pinch out against bedrock which



Base map modified from N.S. Dept. of Lands & Forests 1: 15,840 Map Series.

Map B. Surficial geology of the Port Greville area, Cumberland County, Nova Scotia.

FIGURE 9.1.

rises to over 46 m and to the north, the delta abuts bedrock which rises sharply to over 100 m. A bedrock knoll (50 m high) breaks the surface of the delta in Wards Brook.

Several small terraces have been eroded in the delta complex and the largest (terrace 1) is located on the west side of the Fox River. Unlike the situation at Parrsboro, the scarp between terrace 1 and the upper surface decreases markedly in height from north to south (6 m to 1 m). A large kettle (100 m x ≈10 m deep) that lies on this terrace was previously described by Borns (1965, p. 1225). Another large kettle is situated north of the highway in Wards Brook and it contains a small pond, only 1.5 m deep.

The deltas in the Port Greville area are rather short in the direction of progradation (south) and elevation data are sparse so that gradients calculated for the upper surface and terraces are not very accurate. A north-south profile was taken along the west side of the Fox River and the gradient of the upper surface is relatively high at 10.5 m/km (Fig. 9.2), but the gradient of terrace 1 is slightly lower at 9.0 m/km. Elevation data on terrace 1 show that the terrace has an eastward as well as a southward slope.

SEDIMENTOLOGY

The delta sediments are poorly exposed along the shoreline in Wards Brook and east of the Fox River. In both areas the sediments overlie bedrock, with the contact part way up the bank, and the exposure is relatively thin and poor. A pelecypod mold was found in the bottomset beds that overlie bedrock just west of the unnamed brook (Wards Brook?)

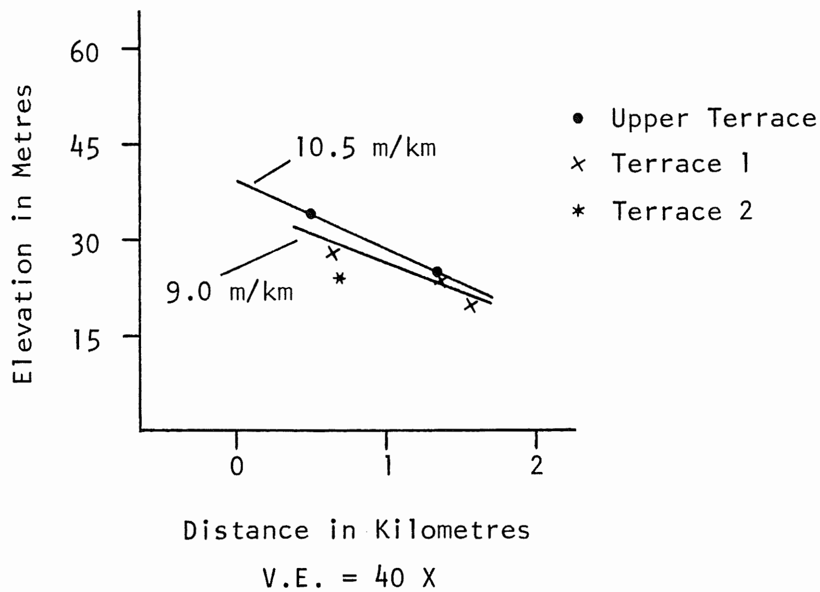


FIGURE 9.2. Gradients of the upper surface and terrace 1 of the delta at Fox River. Profile essentially north-south along west side of Fox River.

that flows through Wards Brook. It was identified by Dr. Frances J.E. Wagner as *Portlandia arctica*, the fossil that is the most abundant at Spencers Island. The paleoenvironmental implications of this pelecypod will be discussed in the chapter on Spencers Island.

A small inland exposure occurs beside Bill Merriam's house in Port Greville, located on the west end of a road that makes a loop from the main highway to the shoreline. However, one of the longest (≃700 m) and best shoreline exposures in the terraces occurs just west of the Fox River and the bulk of the chapter deals with these sediments that are considered to be part of the Port Greville delta proper.

Shoreline Exposure, Port Greville Delta

The shoreline exposure west of Fox River slopes from an elevation of 25 m on the west end to 20 m on the eastern end and is broken by small gulleys into 7 sections, the most easterly being the longest (≃300 m). The topset beds, which coarsen upwards, are generally coarser than the foreset beds. The strike of the foresets varies from 040° to 165° and the trend of the bank is 120°. As the trend of the bank intersects the range of strikes, some beds dip to the west, some to the east and others have no apparent dip as the bank is a strike section. Most of the beds dip to the east (strikes less than 120°) indicating that the west end is more proximal, which is consistent with its location as the west end is closest to the Cobequid Fault. There are no bottomset beds related to the main phase of progradation exposed at the present time along the bank, although Swift and Borns (1967, p. 696) imply that there are bottomset beds at Port Greville.

Multiple Foreset Beds

A striking feature of the internal structure is the sets of foreset beds that occur one above the other. Swift and Borns (1967, p. 696) refer to this as the "stacking" of two or three deltaic lobes. There is an area towards the proximal end where there is only one set of foreset beds, but towards the distal end, up to 3 sets of foreset beds overlie each other. Stacked foreset beds also occur at Spencers Island but they are not as extensive as at Port Greville.

Type A Surfaces

There are two types of surfaces that separate the foreset beds. The first type of surface (type A) dips relatively steeply, from 5 to 10°E, while the second type (type B) dips more shallowly, usually less than 1°E. There are two of the steeply dipping surfaces and five shallowly dipping surfaces. Both types of surfaces are contacts between vertically juxtaposed pulses of foreset sedimentation. The contact is not always between sets of foreset beds because in places the overlying foresets develop bottomsets and/or the underlying foresets have remnants of topset beds.

The first steeply dipping surface occurs near the proximal end of the section on the eastern side of the third gulley from the west (surface 1, Fig. 9.3) where it is approximately 2 m below the topset/foreset contact (A, Fig. 9.3). The surface is a sharp, planar contact between an underlying, eastward dipping (25 to 30°E) set of foresets (Fig. 9.4) at least 4 m thick and 1.5 m of overlying 'mini' foresets that dip 20 to 25°W. The thicknesses of the foresets change because the

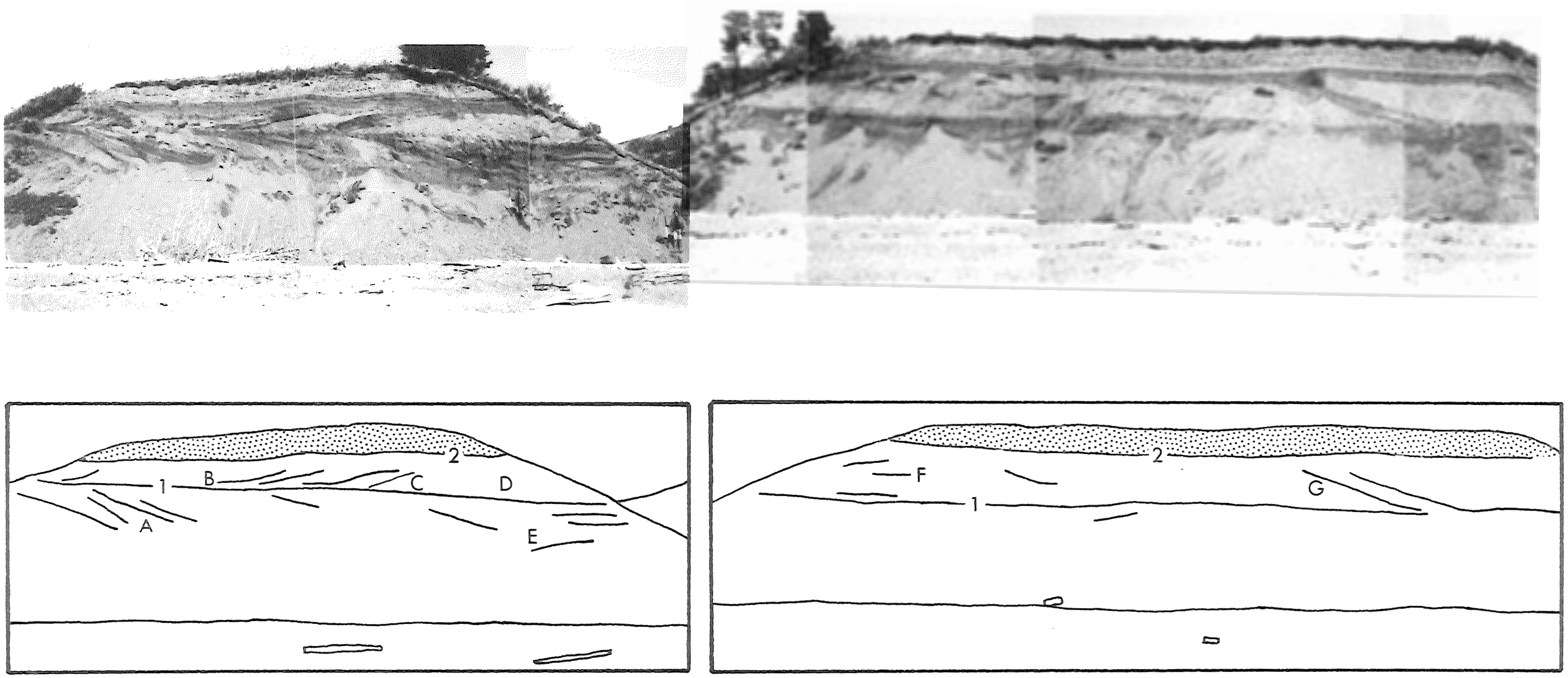


FIGURE 9.3. Photolog of part of exposure at Port Greville. Fourth gulley from west end of exposure between the two sections. Stippled pattern represents uppermost topset gravel, usually coarse.

surface slopes eastward at about 10° and as it does so, some of the overlying foresets flatten to produce sandy bottomset beds (B, Fig. 9.3). Farther east, the westward dipping mini foresets are replaced by a deposit of gravel (C, Fig. 9.3) that is massive but with a possible stratification dipping to the east. Below the massive gravel (D, Fig. 9.3) the formerly well bedded foresets are chaotic and blocks of foreset beds have various strikes and dips (Fig. 9.5).

Another gulley east of the massive gravel breaks the exposure but surface 1 is again visible east of the gulley. It flattens to a dip of about 2°E but comes to within 5 m of the top of the beach. On the west side of this section, the foresets above surface 1 dip to the west or are almost flat lying (strike section) (E, Fig. 9.3) while on the east side they dip to the east (F, Fig. 9.3). Below the surface the foresets are poorly exposed but appear chaotic. East of this, there is another gulley (5th from the west) and surface 1 is not traceable beyond that point.

A second surface of the more steeply dipping type occurs farther to the east but poor exposure makes it traceable only over a short distance.

Type B Surfaces

The second type of surface (shallow dipping) that separates foreset beds occurs near the top of the exposure in figure 9.3 (surface 2) where it lies within the topset facies at its inception. Surface 2 slopes eastward at roughly 1.5 m/100 m and east of the section shown in figure 9.3, another type B surface (3) appears 1.0 to 1.5 m above

FIGURE 9.4. Foreset beds (F) below surface 1, mini foresets (M) and bottomsets above it. Uppermost topset gravel quite coarse.

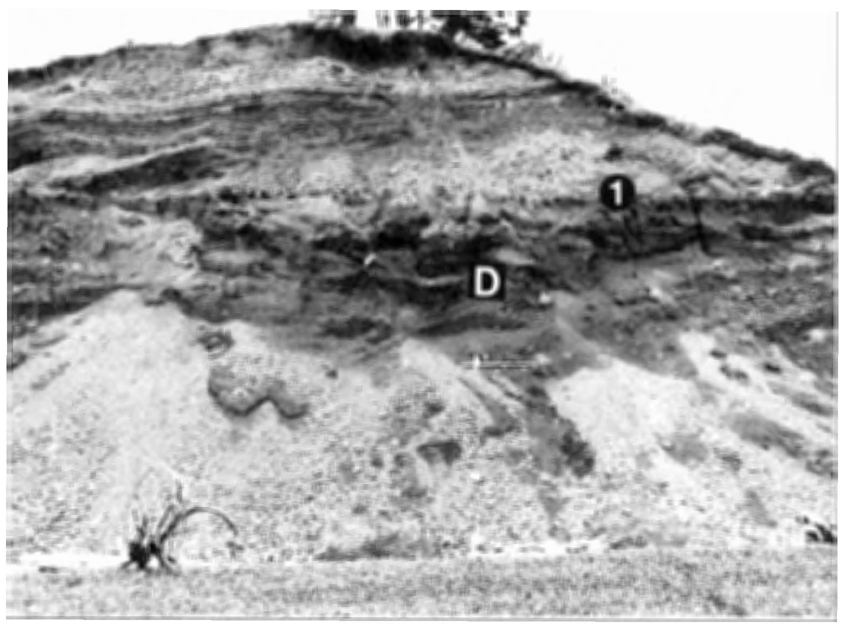


FIGURE 9.5. Disrupted foreset beds (D) below surface 1.

surface 2. Here, the foreset beds below surface 2 extend to the bottom of the exposure as surface 1 is no longer visible. Surfaces 2 and 3 are separated by eastward dipping mini foreset beds that interfinger with sand and clay bottomsets and although the distance between the surfaces increases to over 2 m in an easterly direction, the foresets shorten due to the build up of bottomsets.

Farther to the east, two more type B surfaces (surfaces 4 and 5) occur in the same en echelon manner. Where they first appear, they are from 1 to 1.5 m above the underlying surface and are overlain by 1 to 1.5 m of coarse topset gravel and underlain by mini foreset and bottomset beds. The surfaces dip eastward at a gradient of 1.5 percent and as they get lower in the section, mini foreset and bottomset beds appear above them. Towards the distal end of the exposure, the bottomsets thicken to the point where there are no mini foresets between the surfaces (A, Fig. 9.6). At the eastern end of the exposure, surface 4 becomes difficult to trace but surfaces 2 through 5 steepen to dips of 5 to 8°E. In effect, they become type A surfaces. Surface 5 is overlain by a set of foreset beds (B, Fig. 9.6) that lengthen to the east as surface 5 dips eastward.

The sedimentological relationships of the surfaces vary both along and between surfaces. The two type A surfaces are sharp, reasonably planar contacts between two sets of foreset beds (Figs. 9.4, 9.5) for the most part but bottomset beds, up to 30 cm thick, of interlaminated sand and red clay overlie the surface in some places.

Surface 2 is the most laterally extensive of the type B surfaces (shallow dipping) and was examined more carefully than the other surfaces.

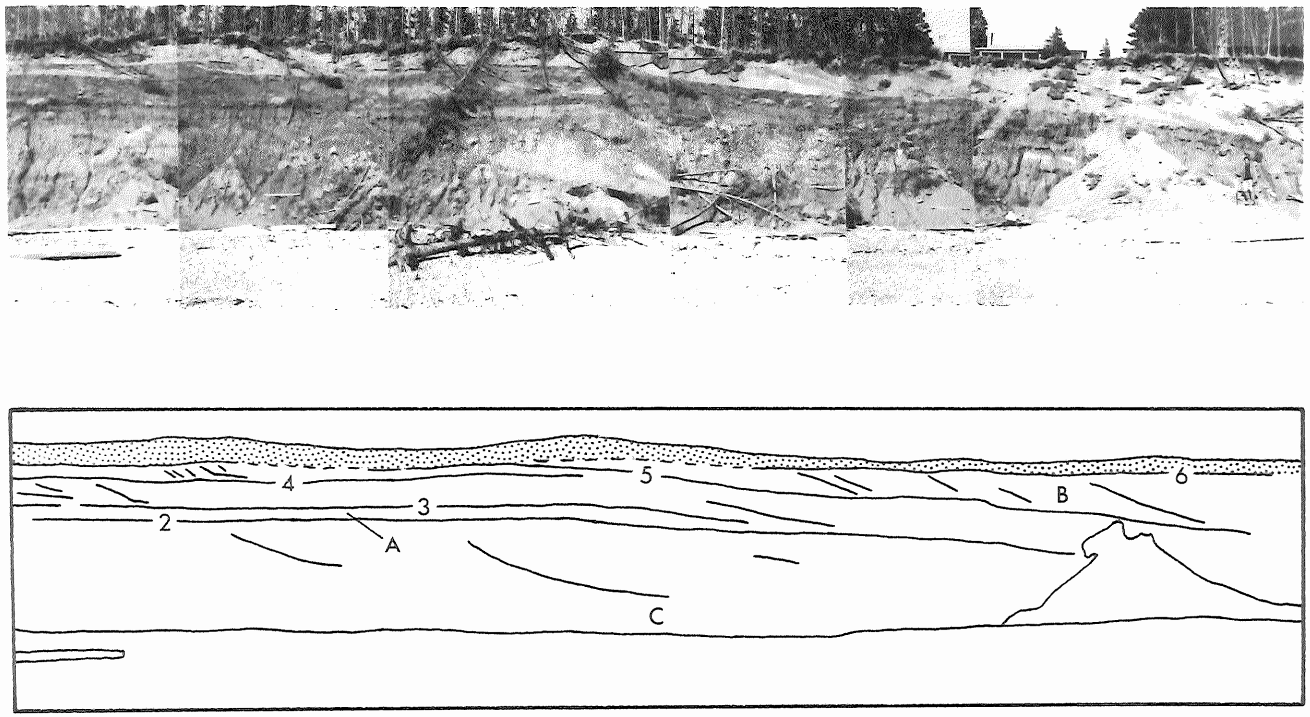


FIGURE 9.6. Photolog near distal end of exposure, Port Greville. Stippled pattern represents uppermost topset gravel, usually coarse.

At the proximal end where it lies within the topset facies (Fig. 9.3), it is a relatively planar contact between 1.5 m of underlying fine gravel and 2 m of overlying coarser gravel. East of this where surface 3 appears, surface 2 is erosional and it is underlain by 30 to 50 cm of topset beds (Figs. 9.7, 9.8). The surface has undulations in the order of 10 cm that cut out some of the topset beds. Below the topset beds are 2 to 3 m of foreset beds that dip to the west and above the topset beds are 20 to 50 cm of interstratified sand and red clay (bottomset beds). The clay laminae are 0.5 to 1.0 cm thick and are more closely spaced at the bottom where the sand has approximately the same thickness. The sand thickens upwards to a maximum of 5 cm and thus so does the spacing of the clay beds (Fig. 9.8). Overlying the interstratified sand and clay are eastward dipping gravel foreset beds, some of which are normally graded (Fig. 9.7), that underlie surface 3.

The identification of topset beds below surface 2 is based on: 1. a steeper weathering profile than the underlying foreset beds; 2. a crude horizontal stratification, the most common sedimentary structure of glaciofluvial gravel; 3. an angular erosional contact between the flat lying beds and the underlying foreset beds, similar to a topset/foreset contact; and 4. grain size distribution as outlined below.

Figure 9.9 shows the grain size distributions for samples of the topset (77-20) and foreset (77-21) beds that underlie surface 2, just east of figure 9.8. For comparison, sample 77-23 is from the uppermost topset gravel (stippled in Figs. 9.3, 9.6) and sample 76-11 is from the glaciofluvial gravel in Gilbert's pit at Parrsboro. The similarity in grain size distribution of the samples from the topset beds and the

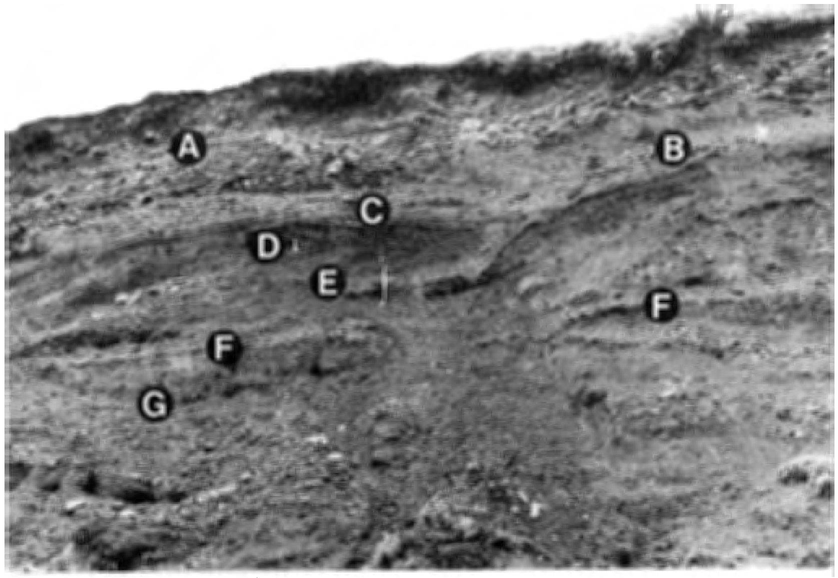


FIGURE 9.7. A. Coarse upper topset gravel above surface 3. B. Coarse gravel at base of foresets below surface 3. C. Bottomset beds of interstratified sand and clay above surface 2. D. Topset gravel below surface 2. E. Foresets dipping west. F. Type A surface not traceable laterally. G. Foresets dipping east, disrupted in places.

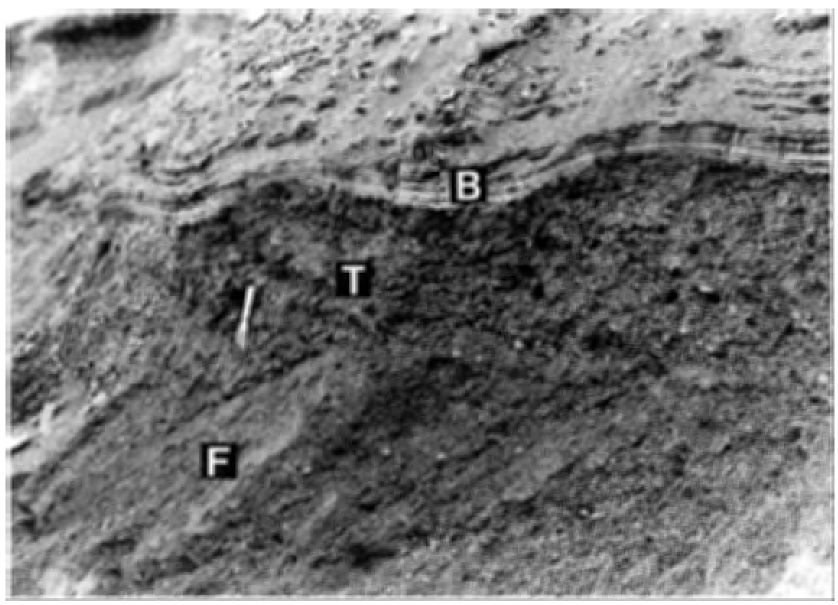


FIGURE 9.8. Topsets (T) overlain by interstratified sand and clay bottomsets (B) with erosional and undulating contact. Foreset beds (F) below topset beds, bottom of paint scraper at topset/foreset contact. Topsets have more vertical weathering profile than foresets. Paint scraper in same position as in figure 9.7.

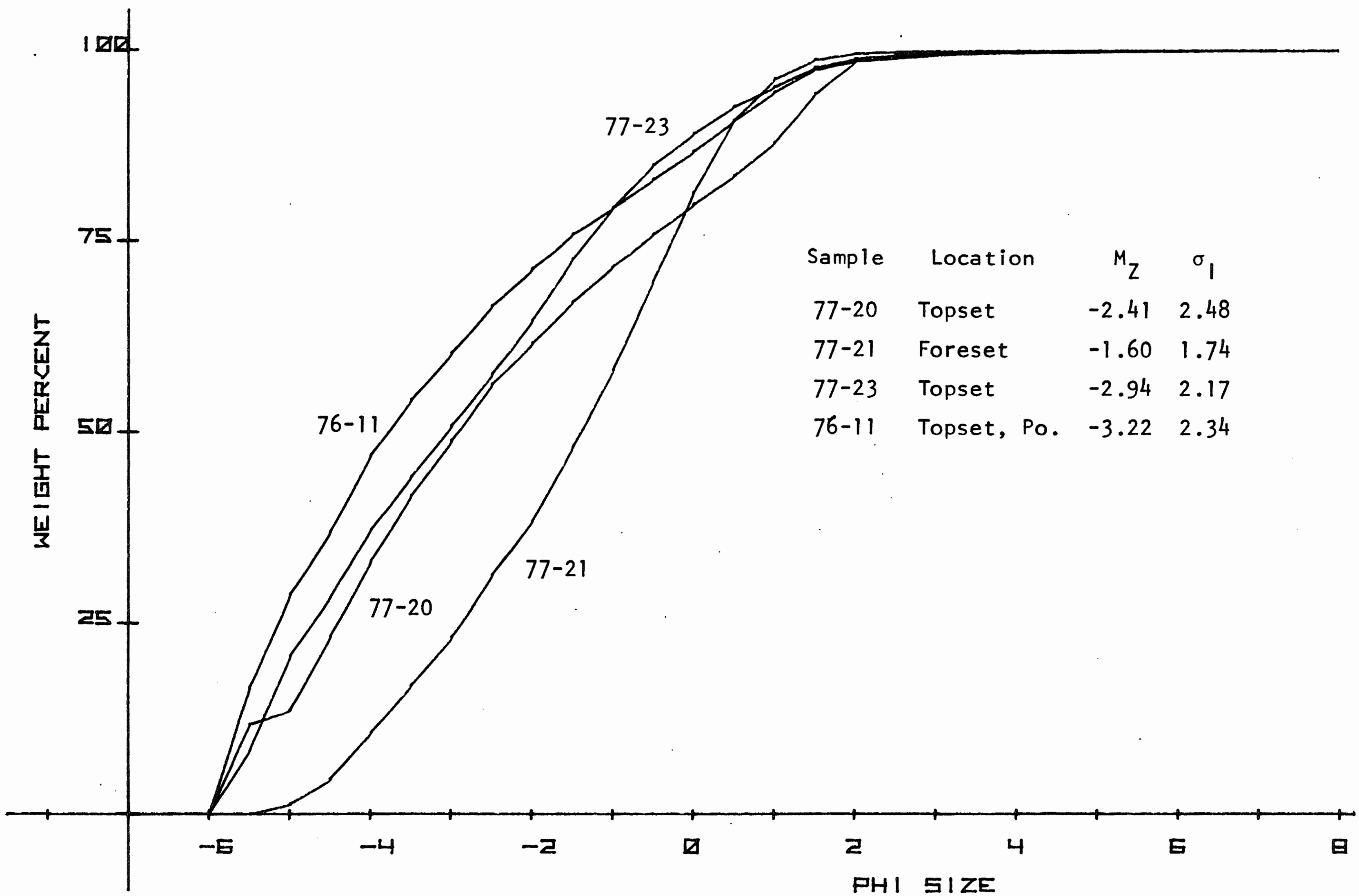


FIGURE 9.9. Grain size analyses, foreset and topset facies, Port Greville and Parrsboro.

difference of the cumulative curve for the sample from the foreset bed are readily apparent. The topset gravel tends to be overrepresented in the coarser fractions (positively skewed) as compared to the foreset gravel. Samples 77-20 and 76-11 are particularly characteristic of topset gravel in that they have a "sag" in the cumulative curve above 75 percent. This is a result of a secondary mode in the medium to coarse grained sand.

Continuing to the east, the topset beds gradually thin out but as they do so, they are overlain in places by a graded openwork gravel. East of the topset bed pinch out, surface 2 is marked by the contact between the overlying bottomset beds and the underlying foreset beds (eastward dipping) except in places where patches of the openwork gravel exist. Figure 9.10 shows the cumulative plots of samples from the topset (77-20) and foreset (77-21) gravel and from the base (77-26) and top (77-25) of the openwork gravel. The gravel is clearly graded and different in grain size distribution from the topset or foreset gravel. The distance between surfaces 2 and 3 decreases to 1 m east of the topset pinch out and bottomset beds comprise the whole interval, thereby eliminating the foreset beds between the two surfaces.

In the central part of the exposure, some of the main foreset beds, underlying surface 2, have a lower dip (14°E) than usual and do not extend to the topset beds. Instead, they pinch out against the preceding foresets, about 1.5 m below the topset beds, and thicken abruptly down dip. The foreset beds appear to have slumped along a bedding plane and the succeeding foresets prograded over the slump.

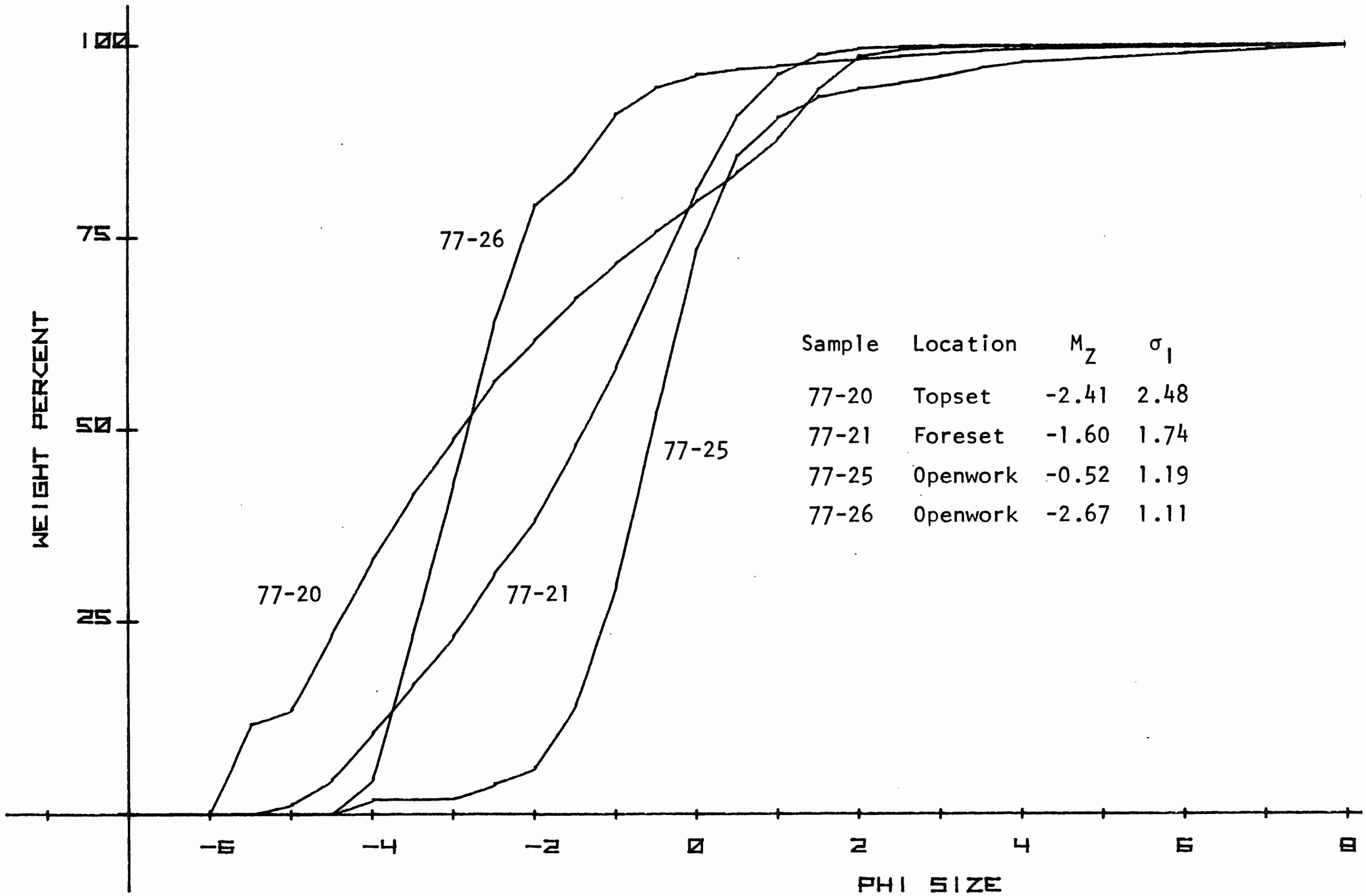


FIGURE 9.10. Grain size analyses, topset facies, foreset facies and openwork gravel.

Cemented Sediment

In the distal section of the exposure, the lower part (just above the beach) of one of the foreset beds (C, Fig. 9.6) was cemented with a black substance so that it was a friable sandstone-conglomerate rather than gravel or sand. A sample from the bed was analyzed by atomic absorption and it contained 0.78 percent MnO, determined on a dry weight basis. Presumably, manganese precipitation from groundwater flow in the Holocene has cemented the sediment.

Inland Exposure

The small exposure on the east side of Bill Merriam's house displayed an almost continuous section of seaward (S) dipping foreset beds in the spring of 1977. The exposure did not extend to the top of the bank and topset beds were not visible.

INTERPRETATION

The topset and foreset beds exposed at the Port Greville delta are similar to those at other deltas. The topset gravel is coarse, closed-work, crudely bedded and the underlying topset/foreset contacts are planar angular unconformities with relief less than 1 m. The foreset beds are slightly finer grained and dip about 25° seaward. Within a set of foreset beds the dip can change from westward to eastward which shows that the delta had a lobate front during construction. This is probably the result of an uneven distribution of streams on the top of the delta and of streams occasionally switching locations.

The multiple sets of foreset beds make the Port Greville delta distinctly different from the other deltas. It was mentioned previously that Swift and Borns (1967, p. 696) called these stacked deltaic lobes. As seen from the description there are two types of surfaces over which the foresets prograded, implying that there are at least two separate causes for the multiple foresets.

Type A Surfaces

The type A surfaces are caused by slumping or failure of the foreset beds. The foreset beds beneath the surface are chaotic (Fig. 9.5), with the amount of disruption increasing in a distal (down slump) direction. The slumps did not occur on a single plane, such as a rotational slump; much of the movement was internal, disrupting the bedding. Slumping of the foreset beds was observed and discussed in chapter 5 on the delta at Lower Five Islands. The cause of the failure is not known but failure in the underlying bottomset beds (due to undercompaction) is a distinct possibility. If the bottomsets failed, they moved very suddenly, causing the foresets to collapse.

The upper surfaces of the slumps slope fairly steeply (5 to 10°) seaward and there are no identifiable topsets on the foresets. Some of the topsets may have slid off the foresets during the slump and other topset beds may have been removed later by currents and waves. The type A surfaces were always rather sharp planar surfaces which is not characteristic of the top of a slump block with internal movement so erosion does seem necessary to plane the top of the slump. After the foresets slumped, the subaerial edge of the delta was moved inland and the succeeding foreset/bottomset beds prograded over the slump.

The slump of the foreset beds along the bedding plane is probably unrelated to movement in the bottomsets and therefore some failures did occur strictly within the foresets. The upper surface of the slump, which is a small scale type A surface, is uneven but there are no recognizable topset beds overlying the foresets. The slump probably occurred just after the deposition of the beds when there was little topset gravel overlying the foresets.

Type B Surfaces

A type B surface, at its inception, always occurs within the topset facies or represents the topset/foreset contact but it drops in elevation to the east until it is overlain by a set of foreset beds and a higher topset/foreset contact. Continued elevation drop to the east brings in other type B surfaces in an en echelon manner. To produce type B surfaces and the resultant multiple foreset beds, the relative position of the sea level must be changed. This can be accomplished either by changing sea level or by changing the level of the delta.

To explain a type B surface by sea level changes requires a slow drop in sea level followed by a rapid rise in sea level that moved the subaerial edge of the delta inland and caused a progradation of foresets over the submerged portion of the delta. This sequence would have to be repeated 4 times as there are 4 type B surfaces. As well, the height of the uppermost topset/foreset contact drops from 22 to 19 m from the proximal to the distal end of the exposure. Thus, each time the sea rose quickly it rose to a slightly lower elevation than the previous time. There seems to be no reason for this complex pattern of sea level changes, especially as it is not evident at the other deltas. At Lower Five Islands, the elevation of the topset/foreset contact indicates that there was a slow, undisturbed relative drop in sea level during the deposition of the delta.

A simpler interpretation is to retain the same model of sea level change that was observed at Lower Five Islands, which also explains the decrease in elevation (3 m) of the uppermost topset/foreset contact from proximal to distal. The pulses of foreset sedimentation are then generated by relatively rapid changes in the level of the delta top superimposed on the relatively slow drop in sea level.

Surface 2 extends to the distal end of the delta so that the delta prograded to near its present limit before any changes in relative position between the delta and sea occurred. Then the delta subsided and the upper surface was flooded to a point well back along the exposure. The topset beds were completely eroded on the distal end, where the depth of overlying water was greatest and the topset beds would have been thinnest. On the proximal end, the topsets were eroded but not completely as some topsets remain. The graded gravel that occurs on top of some topset and foreset beds is probably wave reworked deltaic gravel and the grading implies decreasing energy. The coarser gravel may have been deposited during a storm with the finer gravel during the later stages of the storm.

After the upper surface was flooded and the subaerial edge of the delta had moved well inland, the streams crossing the delta deposited a second set of foreset beds that prograded over the submerged surface. Deposition of sand and clay bottomset beds on the submerged surface shallowed the water and shortened the prograding foreset beds. At the distal end of the submerged surface, the bottomset beds had built up sufficiently to preclude deposition of foreset beds. The delta subsided a total of four times and it was only after the last one that the delta built out beyond its original limits. Each time the delta subsided, the subaerial limit of the delta moved inland but not as far as the previous time, which produced the en echelon overlapping of the type B surfaces.

The subsidence that caused the fourth type B surface (surface 5) was of greater proportions than the previous ones as it bent the other surfaces downward at the distal end of the delta. As a result of this, the foreset beds overlying surface 5 are much thicker than the foreset beds that overlie the other surfaces. This subsidence may have involved collapse of the foreset beds but this is not evident as the foreset beds are poorly exposed at the distal end.

The cause of the subsidence that formed the type B surfaces is conjectural but whatever the cause, it was able to affect large areas of the delta as the surfaces extend for over 300 m along the exposure. The foreset beds are well bedded below most of the surfaces, so slumping of the foresets was not the cause. It appears that the subsidence is related to the bottomset beds.

It is known from the delta at Lower Five Islands that movement can occur within the bottomset beds. It is also known that at the Port Greville delta, foreset beds collapsed and the bottomset beds may have been involved in, or the cause of, this movement. The possible reasons for instability in the bottomsets have been discussed in dealing with the sedimentology of the delta at Lower Five Islands. The possibility favoured by the author, which has also been mentioned by authors studying modern Gilbert type deltas, is undercompaction. As the gravel foreset and topset beds prograde over the bottomsets, the clay and sand beds become undercompacted with respect to the greater overlying weight. As the bottomset sediment compacts, which involves the expulsion of water, failure may occur and the sediments flow towards the offshore area, a zone of lesser pressure. This type of flowage is common in the large deltas such as the Mississippi (Elliott, 1978, p. 139). The net effect of such offshore flowage is to lower the surface of the delta. Even if flowage does not occur, the compaction of the sediments causes subsidence (Friedman and Sanders, 1978, p. 501) but this is usually associated with abandoned delta lobes.

The amount of subsidence gives an indication as to whether compaction or flowage could have produced the subsidence. On the distal end of the exposure, surface 2 is about 3 m below the uppermost topset/foreset contact. Assuming a compaction of 15 percent, the bottomset beds would have to be 20 m thick to produce a subsidence of 3 m. The thickness of bottomsets is unknown and these figures are just estimates but this does seem to be unrealistically thick when compared to the thickness of the exposed bottomsets at other deltas. The thickness of the bottomsets is expected to be less than the combined thickness of the foreset and topset beds, and if the foresets extend to the base of the bank, the combined topsets and foresets would be 15 to 20 m thick. It seems that offshore flowage of the bottomsets, perhaps combined with subsidence, is more feasible.

It is not known whether delta subsidence or sea level changes produced the multiple foreset beds, but delta subsidence is favoured for several reasons. At the other deltas, there is no evidence for rapid sea level changes, especially rises in sea level, and sea level should have been uniform in the Minas Basin. Delta subsidence is less complicated and the proposed mechanism for the subsidence is consistent with evidence at other deltas. The bottomsets at Port Greville are not exposed so there is no firm evidence to corroborate delta subsidence. However, the interpretation is thought to be the most reasonable.

INSTABILITY OF THE PORT GREVILLE DELTA

The collapsed foreset beds and possible flowage of the bottomsets lead to the question of why the delta at Port Greville is so unstable. The delta at Lower Five Islands was unstable in the foreset and bottomset facies but not to the extent that multiple foresets were deposited. The one obvious difference between the two deltas is the thickness of the foresets. At Five Islands, the foresets prograded into rather shallow (in the order of 2 m) water where the bottomsets are exposed, and the build up of bottomsets even precluded foreset deposition in one area. Bottomsets are not exposed at Port Greville, at least at the present time, but foresets do extend to the base of the bank in several places. Thus, the foresets were prograding into water that was at least 15 to 20 m deep in some places. The greater length of foreset slope increases the probability of oversteepening on the upper foresets, which causes slumps or slides. The greater thickness of foreset beds also increases the weight on the underlying bottomsets, which increases the degree of bottomset undercompaction and instability.

The delta at Spencers Island also has multiple sets of foresets, although they are not as extensive or as well developed as at Port Greville. The foresets at Spencers Island also prograded into relatively deep (≃15 m) water. It seems that the instability of the deltas is at least partly related to water depth.

POSTGLACIAL EMERGENCE

Only the uppermost topset/foreset contacts were used for obtaining the marine limit. At the inland exposure, the foresets attain an elevation of 26 m and the topset/foreset contact was not exposed. Near the proximal end of the exposure, the elevation of the topset/foreset contact is 22 m and at the distal end, it is 19 m.

HILLSIDE GRAVEL DEPOSITS

SURFICIAL GEOLOGY

The hillside gravel deposits occur north of the delta in the Fox River valley and in some of its tributaries. The deposits flank the rivers (Fig. 9.11) and lie against the bedrock hills of the Silurian(?) Advocate Succession which rise above 75 m in elevation. The deposits are themselves hilly but the tops of the deposits are fairly uniform in elevation at about 30 m. Elevations of over 40 m are attained where the deposits thin out against the bedrock hills. Thus, the gravel deposits are similar in elevation to the upper delta surface, but the rolling hummocky topography of the gravel deposits is markedly different from the flat deltaic surface.

SEDIMENTOLOGY

The gravel deposits are poorly exposed along the eastern side of Fox River along the large bend in Highway 209 (Fig. 9.11). A small but recent exposure occurs on the west side of the river just beyond the northern limit of the delta (Fig. 9.12). The deposit is composed of closedwork gravel with a few sand beds. Bedding is poorly developed but where it is apparent the dip directions are widely discrepant. For example, in the middle of the pit, large scale gravel cross beds (several metres long) dip north while sand beds on the south side of the pit dip south. There is no definite topset/foreset relationship and abrupt changes in grain size are common.

INTERPRETATION

The hillside gravel deposits are similar in topography (hillocks) and location (up valley from a delta) to the kames at Diligent River, and they are interpreted as kames. At the time of kame formation, the bedrock hills probably had little ice cover and the Fox River valley was full of stagnant ice. Stream flow along the sides of the valley between the ice and bedrock deposited stratified drift of various forms. Hillside wash deposited kame deltas in ponded water along the valley sides, one of which is the exposed kame. These ponds probably had unstable water levels and the kame deltas would lack stable topset/foreset contacts. The kames do not have a well developed terrace top but they are generally of the same height. They may have formed as a terrace but later ice melt out disrupted the terrace and caused the present rolling topography.

FIGURE 9.11. View northeastward across Fox River valley at gravel deposits (G). Tops of deposits nearly uniform in elevation, higher bedrock (B) in background. Highway bridge over Fox River at left of photograph.

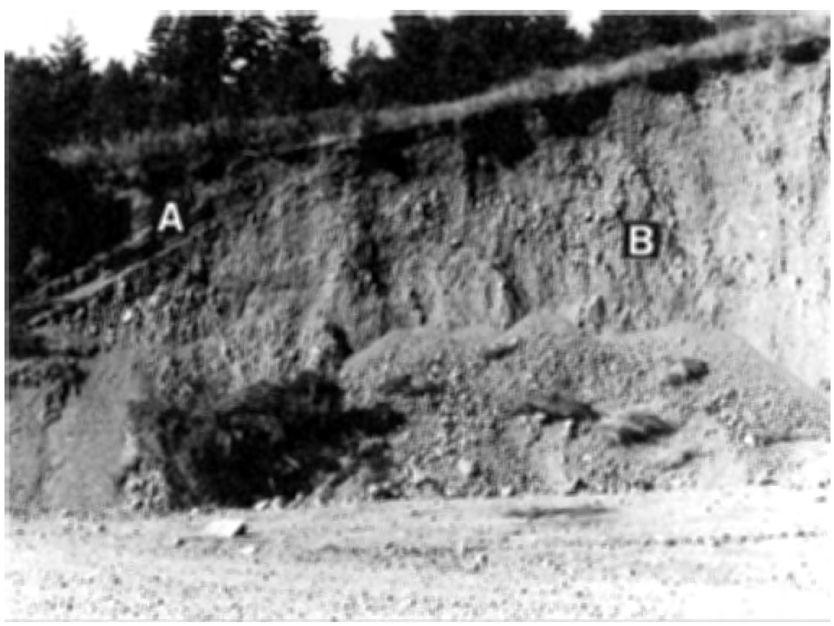


FIGURE 9.12. Borrow pit in gravel deposits just west of bridge over Fox River on Highway 209. A. Sand with a southward component of dip. B. gravel with a northward component of dip.

OVERVIEW

The kettles in Wards Brook and Fox River indicate that the deltas were deposited in close proximity to dissipating ice. The highest elevations on the deltas occur near the Greville River and Fox River valleys, which indicate that the main meltwater discharge was out of these valleys. However, the southward dip of the foreset beds at the inland exposure (less than 250 m south of the Cobequid Fault) implies that some discharge was unrelated to these two valleys, at least in the early stages of delta formation. Later discharge was out of the Fox and Greville River valleys as there is a slight low in the upper surface of the delta in Port Greville, between the two river valleys. The direction of dip of the foreset beds along the shoreline is consistent with this interpretation.

The structure and texture of the gravel show that the deltas were deposited by fast flowing braided streams. The orientation of the foreset beds west of the Fox River indicate that the delta had a lobate front during progradation. Vertically juxtaposed sets of foreset beds are interpreted as the result of delta subsidence, although the evidence for this is not conclusive. Up valley from the Port Greville delta, kames formed along the sides of the Fox River valley. After the peak of delta aggradation, the Greville and Fox Rivers eroded terraces and their present river valleys in the deltas. A kettle near Fox River indicates that the ice that formed the kettle melted after erosion of the terrace, which is a conclusion drawn previously by Borns (1965, p. 1225).

CHAPTER 10. SPENCERS ISLAND

SURFICIAL GEOLOGY

Spencers Island is located about 35 km west of Parrsboro and 6 km east of Advocate Harbour with the village facing eastwards towards the head of the Minas Basin. The surficial geology consists of a small delta that may be the most important deposit along the north shore of the Minas Basin. Along the shore the delta extends from Mahoney Brook to just south of the village of Spencers Island (Fig. 11.1, Advocate Harbour map) and it reaches less than 1 km inland at its maximum width.

The Upper Carboniferous Parrsboro Formation underlies the delta and bounds it to the west and south. The delta is highest (36 m) inland along the Mahoney Brook and has a seaward (E) gradient but it also decreases in elevation to the south. Mahoney Brook and two smaller brooks to the south have eroded valleys in the delta, but there are no terraces which is anomalous as all the other deposits have at least one terrace. The small size of the delta and lack of comprehensive elevation data preclude an accurate assessment of the gradient of the upper surface. However, estimates can be made using the elevations near the limits of the deposit. An east-west profile, taken through the central part of the delta, has a seaward gradient of 4.5 m/km (Fig. 10.1) but is based on only two data points. A north-south profile taken from Mahoney Brook has a steeper gradient of approximately 8 m/km.

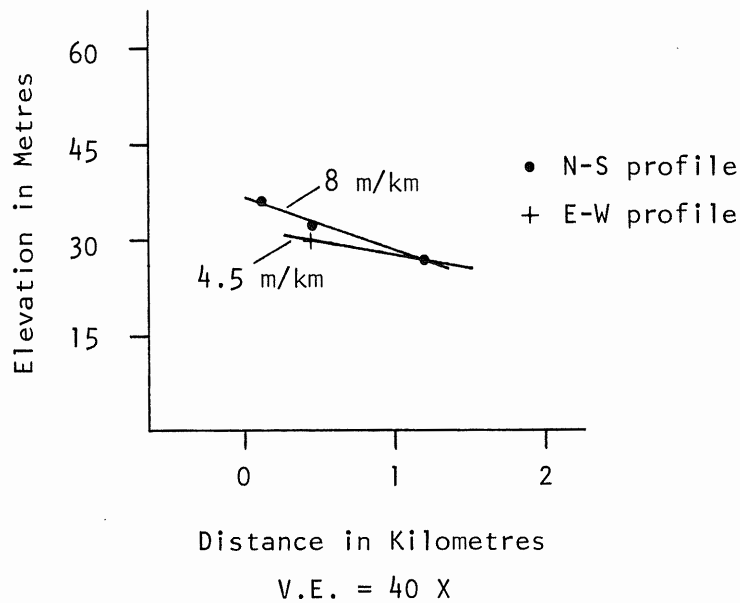


FIGURE 10.1. Gradients of the upper surface of the delta at Spencers Island.

SEDIMENTOLOGY

The topset/foreset/bottomset structure of the delta is exposed in several locations between Mahoney Brook and the unnamed brook to the south that passes through Spencers Island village. The exposures are the result of shoreline erosion oversteepening the bank and causing large slumps. A large slump in the bank just east of elevation 27 on the surficial geology map (Fig. 11.1) resulted in very good exposure in the fall of 1974 (Fig. 10.2). Fortunately, detailed sedimentological observations were made at that time and the descriptions that follow in this section are from this exposure. The bank has deteriorated since then and the bottomset beds that were so well exposed at that time have been eroded and covered by slumps of foreset beds. To the north, the deposit thins because it is underlain by a thick (up to 8 m) red brown till. Slumps more recent than 1974 have exposed parts of the delta overlying the till but the extent and development of the structure, particularly the bottomset beds, is not as good as in 1974 (Fig. 10.3).

A more proximal exposure of the delta occurs on the east side of Highway 209 in the Mahoney Brook valley, but this has also deteriorated since 1974. A small part of the delta occurs to the south of the unnamed brooks and is poorly exposed but it is also thin as it overlies bedrock or till with the contact well up the bank.

TOPSET FACIES

Description

The topset section of the Spencers Island delta is thin and fine grained compared to the topset beds at the deltas to the east. At the

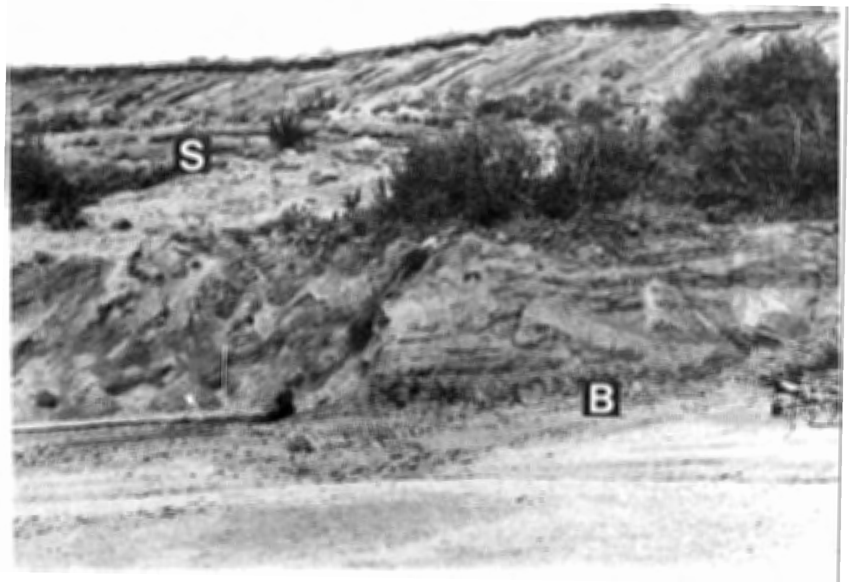


FIGURE 10.2. Exposure on southern end of Spencers Island delta just west of spit. Thin topset beds, erosional topset/foreset contact (marked by black line on right side of photograph), foresets exposed by slump block (S), bottomsets (B) in lower right centre. Scale is 1.5 m.

FIGURE 10.3. Exposure farther north (towards Mahoney Brook) than in figure 10.2. Foresets (F) are finer grained than in figure 10.2 and sigmoidal in shape. Till (T) in lower part of photograph. Bank is approximately 23 m high.

southernmost exposure (examined in 1974), the topset section was 1 to 1.5 m thick. Twenty six beds were described in the field and 16 (62%) were composed of gravel while the rest were sand beds. The gravel was relatively fine grained and the sand was coarse to very coarse grained (Fig. 10.4). Large clasts (1 to 12 cm) were occasionally scattered in the beds with the largest in the gravel beds. Lateral changes in grain size were frequent such that individual beds could not be traced laterally for more than 10 m, sand beds being especially susceptible to lateral pinch out.

The thickness of the beds ranged from 2 to 23 cm and averaged 9.6 cm. The sand beds were slightly thinner than the gravel beds, averaging 6.6 cm. Subhorizontal to horizontal bedding planes were fairly well developed and were frequently marked by an increase in silt and fine sand content. Of the subhorizontal gravel beds, only one complete bed and the base of another were openwork; the rest were closedwork with a medium to very coarse grained matrix. No measurements were taken but many of the flat pebbles were noticeably imbricated with their maximum projection (a-b) planes dipping to the northwest. The sand beds were massive to parallel laminated.

Tabular cross beds were common (6 out of 26) and usually consisted of alternating sand and openwork gravel beds, the latter being graded. The sand cross beds were massive to faintly parallel laminated. Individual cross beds were from 3 to 10 cm thick, up to 1 m long and dipped from 15 to 20°SE indicating a paleoflow in that direction. The sets of cross beds were approximately 0.5 m thick. Above one of them was a lens shaped occurrence of massive sand, interpreted as a channel deposit,

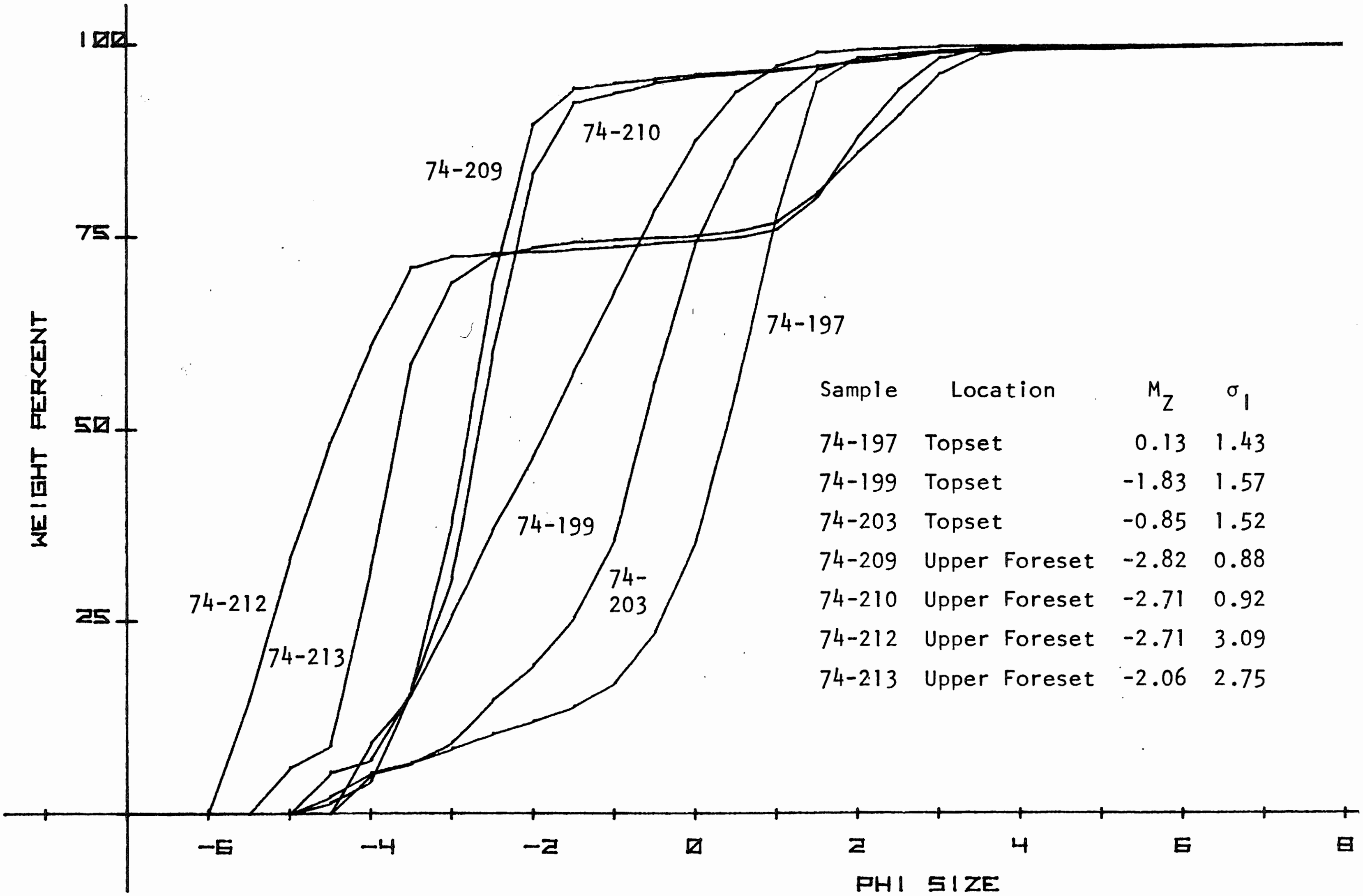


FIGURE 10.4. Grain size analyses, topset and upper foreset facies.

graded from very coarse grained to coarse grained. The sand was matrix free and had an erosional lower contact. The width and depth of the deposit were 60 cm and 10 cm.

The topset/foreset contact was a flat lying erosional unconformity between the steeply dipping (25 to 34°SE) foreset beds and the essentially horizontally stratified topset beds.

Interpretation

The horizontal stratification and closedwork gravel are typical of the glaciofluvial gravel discussed previously, and the depositional mechanism is interpreted to be braided streams. Pebble imbrication, especially of the larger clasts (Rust, 1975, p. 244) is also consistent with this type of stratification and deposition by braided streams (Church and Gilbert, 1975, p. 62). The only two horizontal beds that were openwork or partially openwork (at the base) were also graded. The process responsible for this is probably that suggested by Smith (summarized by Collinson, 1978, p. 23) in which the graded beds are the product of a waning flow and the finer gravel at the top of the bed inhibits the infiltration of fines at low discharge.

The tabular cross beds appear to have been deposited on the slip faces of inchannel bars but the morphology of the bars is unknown. The thickness of the tabular cross beds indicates that water depths greater than 0.5 m existed in some channels. The alternation of gravel and sand cross beds probably reflects changes in stream velocities, with sand deposited at lower velocities than the gravel. The openwork nature of the gravel cross beds is probably due partly to the grading and partly to their inclined attitude.

Sand beds were common in the topset facies (10 out of 26) which suggests that stream power was often reduced and sand was transported while gravel was immobile. Lower stream power was probably due to lower velocities and shallower depths. The lateral impersistence of the sand beds suggests that sand transportation was localized, probably in the deeper parts of the system. The small deposit of channel sand supports this. However, some of the sand beds were erosionally truncated so this may not always have been the case.

Although the horizontal sand beds are not cross stratified, they may have been deposited under either lower or upper flow regime conditions. The sands are coarser than 0.6 mm (sample 74-197, Fig. 10.4) which precludes the formation of ripples (Simons *et al.*, 1965, p. 52). At this grain size, parallel lamination can form in either lower or upper flow regime conditions (Blatt *et al.*, 1972, p. 121).

Boothroyd and Ashley (1975) provide a good summary of changes in texture and structure along a longitudinal profile on an outwash plain. Proximal or upper fan areas are dominated by longitudinal bars, pebble imbrication and coarse, crudely stratified gravel. The mid fan is finer grained and contains tabular cross beds from deposition on the slip faces of bars. Lower (distal) fan areas are sandy and have a greater variety of bedforms and structures.

The relative abundance of sand beds, grain size of the gravel and the presence of tabular cross beds in the topsets compares favourably with the coarse to fine gravel mid fan position of Boothroyd and Ashley (Collinson, 1978, p. 25). The exceptions are in the grain size of the tabular cross beds and in the structure of the sand. The cross beds are supposed to be composed of sand while at Spencers Island, the tabular cross beds are predominantly composed of fine gravel. The horizontal sand beds are not cross stratified.

The topset facies at Spencers Island is anomalously thin (1.0 to 1.5 m) compared to the topset facies at other deposits, implying that the streams flowing across the top of the delta did not aggrade to the same extent as they did on the other deltas. The small size of the delta, the absence of terraces and the thinness of the topset facies lead to the possibility that the entire upper surface has been terraced or eroded. If this is the case, the topset/foreset contact could be below its original height. The contact is not unreasonably low compared to the height of the contact at the nearest delta (Port Greville) so if the upper surface was eroded, the erosion was restricted mainly to the topset beds. As erosion also occurs during the deposition of the topset beds, it cannot be determined if the topsets have been eroded after the peak of aggradation, but the thinness of the topset beds makes this a distinct possibility.

Although the main exposure (1974) was located at the distal portion of the delta, the delta is very small and in fact the exposure is close (0.5 to 1.0 km) to the proximal limit of the delta. The short distance suggests that the exposed topsets should be more proximal in nature. One explanation for this is that the delta at Spencers Island was part of a shorter overall system, just as the Yana outwash fan has similar characteristics to, but is much shorter than, the Scott outwash fan (Boothroyd and Ashley, 1975, p. 200). However, there was coarser gravel in some of the underlying foreset beds (Fig. 10.4). The fine grain size of the topset beds is part of a larger problem of grain size discrepancies between the topset and foreset beds, a problem common to all of the deltas in the Minas terrace and it is discussed below.

GRAIN SIZES OF TOPSET AND FORESET FACIES

The topset/foreset contact is an erosional unconformity at all of the deltas in the Minas terrace and its genesis has been discussed in the chapter dealing with the Five Islands delta. During delta construction, the streams aggrade in an upstream direction so the unconformity is formed at the outer margins of the delta. Because the topset beds were deposited after the foreset beds at any given locality, variance in stream discharge or sediment supply can cause differences in grain size between the two facies.

The topset beds at the Spencers Island delta are finer than the foreset beds at the 1974 exposure which is unusual as the topset beds are generally coarser grained than the foreset beds. It has been suggested that the topset beds may have been eroded and thus represent a later phase of sedimentation. Finer sediment may have been supplied at that time as the dissipating ice would have been farther from the delta, but if the topsets have been eroded, a coarse lag would be expected. There were scattered coarse pebbles and cobbles in the topsets, but not of a concentrated nature. In any case, the topset sedimentation was finer than the earlier foreset sedimentation and differences in stream discharge and sediment must have occurred.

At the other deltas in the Minas terrace, the topset beds are generally coarser than the foreset beds. This is not unique to the

Gilbert deltas of the Minas Basin. Work on recent Gilbert type deltas (Fulton and Pullen, 1969, p. 789; Born, 1972, p. 62; Gilbert, 1974, p. 1699) and Pleistocene deltas (Gustavason *et al.*, 1975, p. 268) confirm the relative coarseness of the fluvial gravel which was predicted by earlier workers (Elliott, 1978, p. 97). The grain size difference between the topset and foreset beds is particularly noticeable at the distal end of the exposure of the Lower Five Islands delta. There the topset beds are composed of fairly coarse pebble gravel and the foresets are composed of sand. Most of the foreset beds are planar, at least in the upper part, and the topset/foreset contact is a sharp unconformity. If gravel was transported to the mouths of the streams, it seems that some of the gravel would roll down the foreset slope and be incorporated in the foreset sand, probably at the base of the foreset slope. The lower part of the foresets are covered, so it is possible that some coarser gravel exists near their base. But even if some gravel does occur in the lower foresets, the depositional process limited the amount of gravel deposited on the foreset slope.

One possible explanation for this is that streams transported sand at faster rates than the gravel so the distal ends of the streams were composed of sand. The sand bed load was deposited as foreset beds and the slower moving gravel later capped the foresets. While this is perhaps possible, it has not been observed on modern Gilbert type deltas. It also requires an unrealistically high rate of delta progradation, as the gravel in the stream beds never catches up with the prograding delta front.

Another possibility is that the erosion of the top of the foreset beds that produced the planar unconformity has obscured the sedimentological relationship that existed between the topsets and foresets at the time of deposition. This possibility is suggested by the foreset beds that overlie the till at the Spencers Island delta (Fig. 10.3). The foreset beds here are composed of finer gravel and have a sigmoidal shape as well as a lower dip angle (20 to 27°) than the coarser foresets (25 to 34°). The lower dip angle and concavity of the lower part of the beds is to be expected as the finer grain size increases the proportion of suspended load (versus bed load) sedimentation and shortening the foreset slope has the same effect (Pettijohn *et al.*, 1973, p. 352). However, the foresets also have a curved upper part which gives the overall sigmoidal shape. Some of the sand foreset beds at Lower Five Islands also display this feature and it is possible that an increase in suspended load deposition also causes a curvature in the upper foresets. Perhaps this is not the cause of the curvature, but whatever the cause, it does seem to be related to grain size as the curvature is less pronounced in the coarser foresets (Fig. 10.2).

If the foreset beds had a curved transition zone (convex up) with the stream beds, the coarse gravel may have been trapped on this 'platform' while the finer sediment was moved seaward onto the steeper part of the foreset slope. A curved transition zone gradually increases the flow depth, which reduces stream velocity and the tractive shear stress. Also, fresh water tends to enter marine basins as overflows which further reduces tractive force with depth. Thus, the shear stress would decrease over the length of the transition zone and it might have fallen below the critical value for moving the coarser sediment. Thus, coarser sediment was no longer moved by the stream and the low slope of the zone was insufficient for sediment avalanches (Fig. 10.5). The emergence of the deltas and continual shifting of the streams across the upper surface caused part or all of the curved transition zone to be eroded. If this was the case, the transition zone was not extensive as stream erosion would not have removed much of the upper foresets.

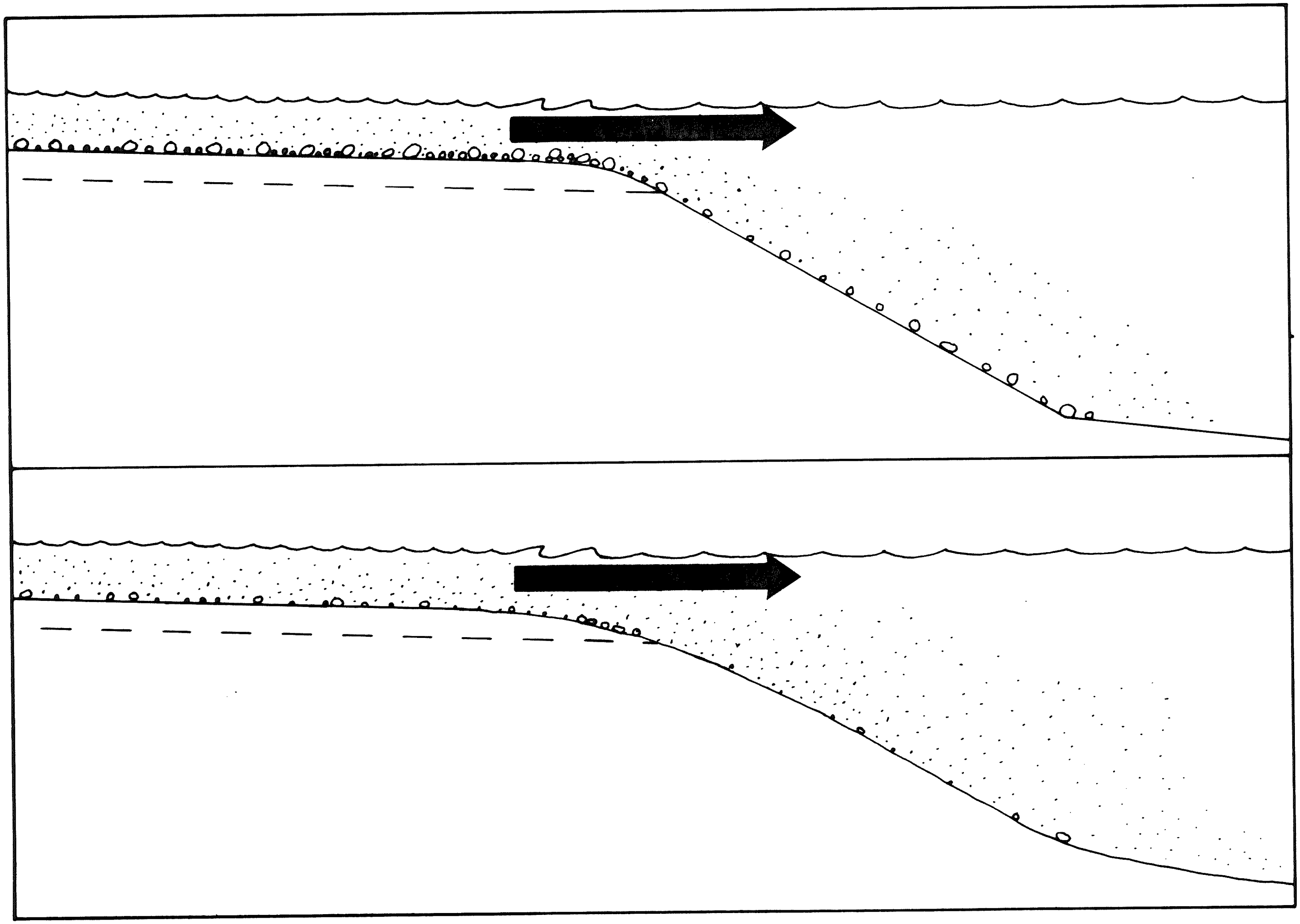
This is rather hypothetical but it is a possibility. Differences in grain size between the topsets and foresets are to be expected as this depositional environment is highly variable in sediment and melt-water discharge and there is a temporal difference between the topsets and underlying foresets where the contact is an angular unconformity. However, the consistent coarseness of the topsets as compared to the foresets suggests that the sediments are size sorted at the edge of the delta.

FORESET FACIES

Description

The strike of the foreset beds at the 1974 exposure varied substantially from 50 to 100° but the dip (25 to 34°) was always to the south, similar to the direction of dip of the tabular cross beds in the topset facies. The foreset section was approximately 15 m thick and various slump blocks exposed up to 3 m of the top and 5 m of the bottom of the foresets, leaving the centre of the beds unexposed.

FIGURE 10.5. Schematic diagram showing different topset/foreset transitions. Top — foreset and topset beds both coarse grained. Fairly angular contact between foresets and topsets, gravel rolls down foreset slope. Bottom — foreset beds finer grained, curved contact between topsets and foresets. Pebbles arrested on upper foreset slope because of reduced stream velocity along bed (increased depth) and low slope of upper foresets. Later erosion of shifting streams as delta is uplifted produces topset/foreset contact at level of dashed lines.



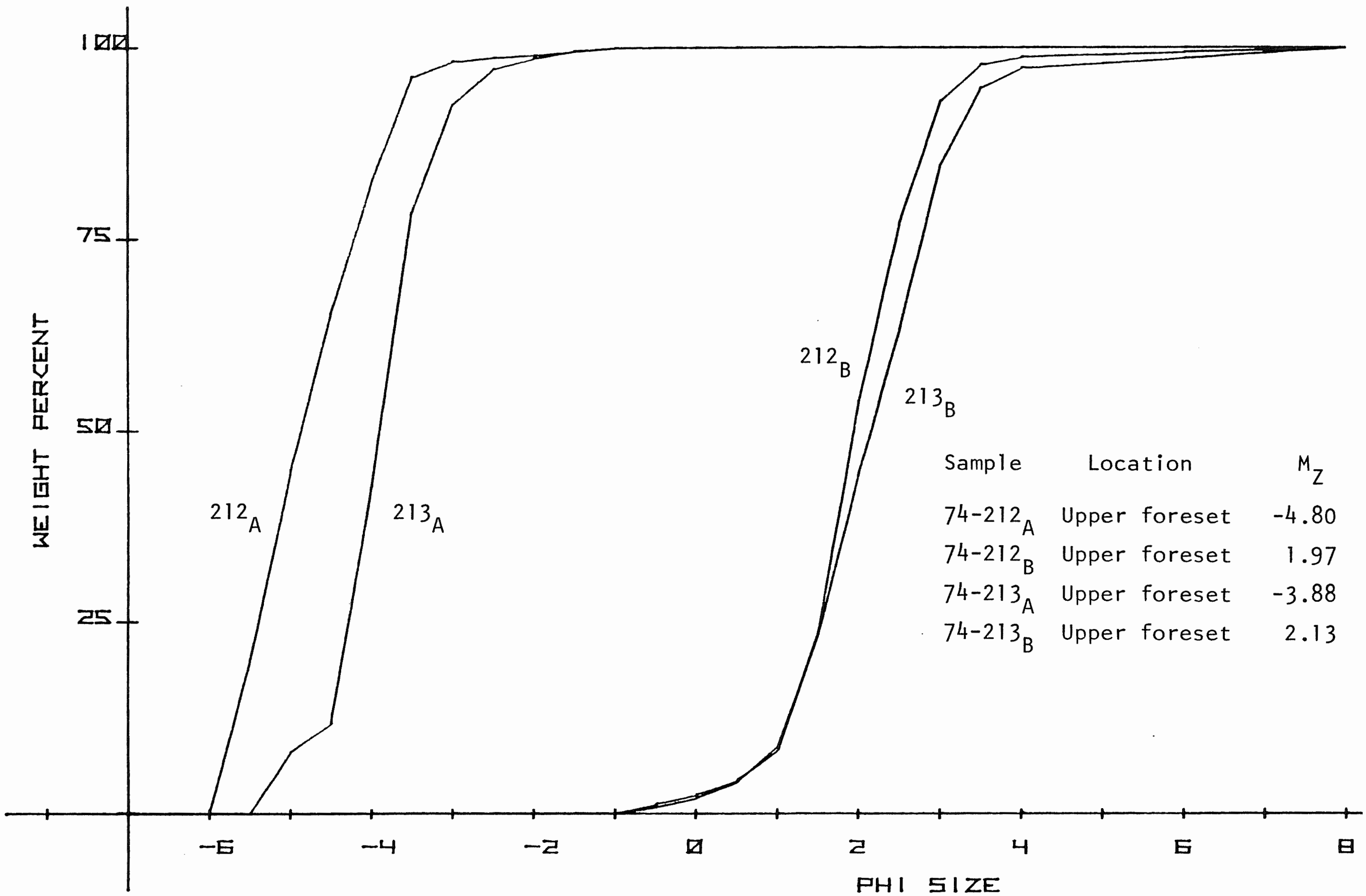
Upper Foresets

Twenty three foreset beds at the top of the section were recorded in detail in the field. The foreset beds were coarser than the topset beds as only one of the beds was composed of sand (it was discontinuous, pinching out down dip) and the gravel was relatively coarse (Fig. 10.4). Clasts of up to 15 cm were scattered throughout the beds and one 25 cm cobble occurred at the topset/foreset contact. Lateral changes in grain size within strata were frequent. Most of the changes were haphazard but 4 beds coarsened down dip as compared to 2 beds that coarsened up dip.

Most of the foresets were composed of closedwork gravel with a fine to medium grained sand matrix but there were 3 beds and the upper part of a fourth that were openwork. The openwork gravel was always graded (Fig. 10.6) but samples 74-209 and 74-210 are from the bottom and top of an openwork bed and they are only slightly graded (Fig. 10.4). The bed that was openwork at the top was inversely graded. Two of the closed-work beds were graded (Fig. 10.7) and they were coarser in the gravel fraction (samples 74-212, 74-213, Fig. 10.4) than the openwork graded beds. However, their poor sorting (very poorly sorted [Folk, 1974, p. 46]) results in smaller mean grain sizes. The shape of the cumulative curves and poor sorting imply that these samples are composed of two better sorted end members. Samples 74-212 and 74-213 were split at -1ϕ and replotted as separate samples (Fig. 10.8) to confirm this. The gravel fractions are moderately to moderately well sorted and the sand fractions are moderately sorted.

FIGURE 10.6. Graded openwork gravel in upper foreset beds. Photograph taken parallel to dip of 30°S. Scattered coarser pebbles in finer part of bed, basal contact sharper than upper contact.

FIGURE 10.7. Graded closedwork gravel in upper foresets. Basal contact sharper than upper contact.



Sample	Location	M_Z	σ_I
74-212$_A$	Upper foreset	-4.80	0.77
74-212$_B$	Upper foreset	1.97	0.77
74-213$_A$	Upper foreset	-3.88	0.66
74-213$_B$	Upper foreset	2.13	0.89

FIGURE 10.8. Grain size analyses, foreset facies. Replot of samples 74-212 and 74-213 split at -1ϕ.

The beds ranged in thickness from 55 to 37 cm and averaged 16.2 cm. Bedding contacts ranged from distinct to gradational but in general were not sharp. The three graded openwork beds had a sharp lower contact and a more gradational upper contact. Approximately one third (8) of the beds thickened or thinned along dip and three of those pinched out down dip while one pinched out up dip. Four of the beds contained flat pebbles that were imbricated up dip.

A peculiar channel shaped lens of openwork gravel occurred approximately 1.5 m below the topset unconformity. The lens, 70 cm long and 10 cm deep, was aligned with the bedding. The gravel was graded from approximate modal diameters of 3 cm to 1.5 cm, with an erosional lower contact and a distinct upper contact (Fig. 10.9).

In the upper foresets, there were two surfaces of discontinuity that separated vertically juxtaposed sets of foreset beds. The first surface, marked by a thin (5 cm) sand bed, dipped 14°SE and extended from the topset/foreset contact into the middle foresets that were covered by the slump block. The steepness of the surface makes it similar to the type A surfaces described at Port Greville.

The second surface also originated at the topset unconformity. It dipped 8°SE before flattening to 3°SE approximately 1 m below the topset facies. The surface separated the second set of foreset beds from a relatively thin (over the width of the exposure) third set which prograded over the second set. The surface was marked by a bottomset sand bed, 3 to 15 cm thick, that was thickest farthest to the south and thinned towards the topset unconformity. The sand was fine to medium grained with scattered pebbles up to 2 cm.

Foreset beds below the second surface were oversteepened to a dip of 36 to 37°S while the overlying 'mini' foreset beds only dipped 20°S (Fig. 10.10) and were slightly finer grained ($M_z \simeq -2\phi$). Towards the north end of the surface, some of the 'mini' foresets had an abrupt angular contact with the underlying sand bed while others flattened at the contact producing a small bottomset bed. Farther south, the bottomset beds were better developed and consisted of coarse to very coarse grained sand with scattered pebbles. The bottomset facies increased in thickness to the south as successive beds were deposited, even though individual beds thinned to the south.

There was a lens shaped deposit 50 cm wide and 20 cm deep, interpreted as a channel deposit, immediately below the second discontinuity surface where it dipped 3°S. The bottom 10 cm of the deposit was gravel, modal diameter approximately 0.5 cm, with a fine grained matrix and scattered pebbles to 3 cm overlain by a fine to medium grained sand. The gravel consisted of channel fill cross bedding while the sand was trough cross bedded.

Lower Foresets

Foreset beds that were exposed at the bottom of the section were similar to those exposed at the top of the section but there were several differences. The lower part of the foreset beds (24 were described in the field) had a greater range in thickness, from 1 to 45 cm, but the average thickness (11.7 cm) was less. Less common were graded beds (8% vs. 22%) and openwork gravel (3% vs. 16%). Beds containing pebbles with an up dip pebble imbrication were about the same in abundance (16% vs 18%).

FIGURE 10.9. Lens of openwork, graded gravel aligned with bedding in upper foresets.

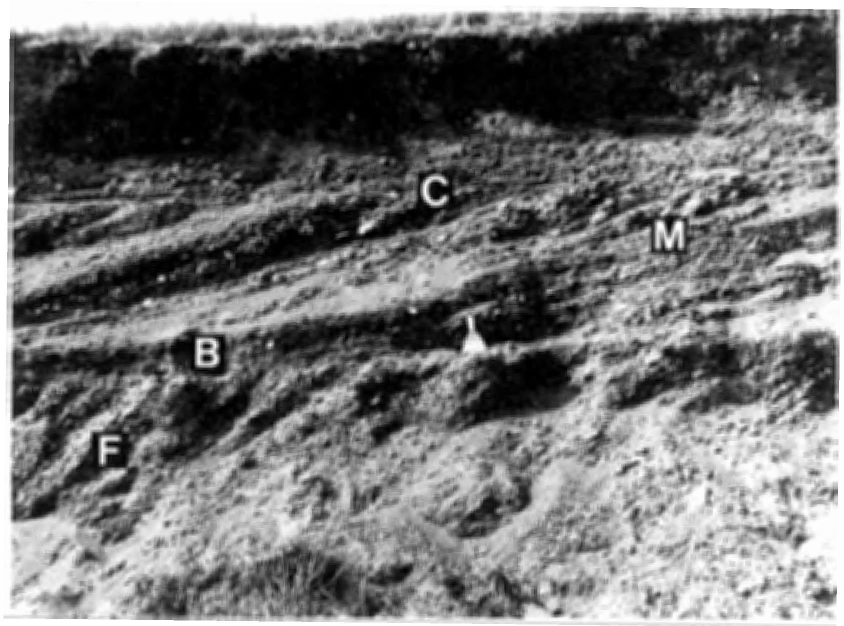


FIGURE 10.10. 'Mini' foresets (M) and bottomsets (B) deposited on over-steepened foresets (F). Bottomsets are thicker on left (south) side of photograph. Planar, erosional topset/foreset contact (C) above mini foresets.

The grain size of the gravel was similar, but sand beds were more common (21% vs. 4%). Some of the gravel beds were coarser at the base, as shown by samples 74-220, 74-221 and 74-222 in figure 10.11. The samples were taken at the base of the bed, at 1 m and at 2 m above the base, respectively. Sand as matrix and in the sand beds was generally medium to coarse grained (Fig. 10.11). Sand beds were mainly stratified parallel to the bedding plane but one occurrence of cross stratification inclined up the bedding plane was noted (Fig. 10.12). The cross stratification was 20 cm long, 2 cm thick and dipped 11° to the bedding. It was underlain by sand laminated parallel to the bedding. The lower contact of the sand bed was sharper than the upper contact and the gravel bed beneath the sand bed had pebbles imbricated up dip (Fig. 10.13).

One of the most noticeable differences between the lower and upper foresets was the occurrence of "red beds" in the lower foresets. The red colour came from a silt and clay matrix (10 to 15%) and the beds were extremely bimodal as they lacked fine gravel ($< -3\phi$) and sand (samples 74-224, 74-225, Fig. 10.11). Grading, either inverse or normal, was usually absent (74-224, bottom; 74-225, top). The mud coated the pebbles and did not completely infill the porosity except where it was usually concentrated at the base of the beds. The beds had a sharp lower contact, less distinct upper contact and both the gravel and clay pinched out up dip. The distance between beds varied from 0.3 to 1.2 m.

Some foreset beds ended abruptly at the bottomset beds, but others flattened to dips of 5 to 8°S before pinching out over the bottomset beds (tangential contact). The transition zone normally spanned several metres and included sand and sandy gravel from the bottomsets that extended into

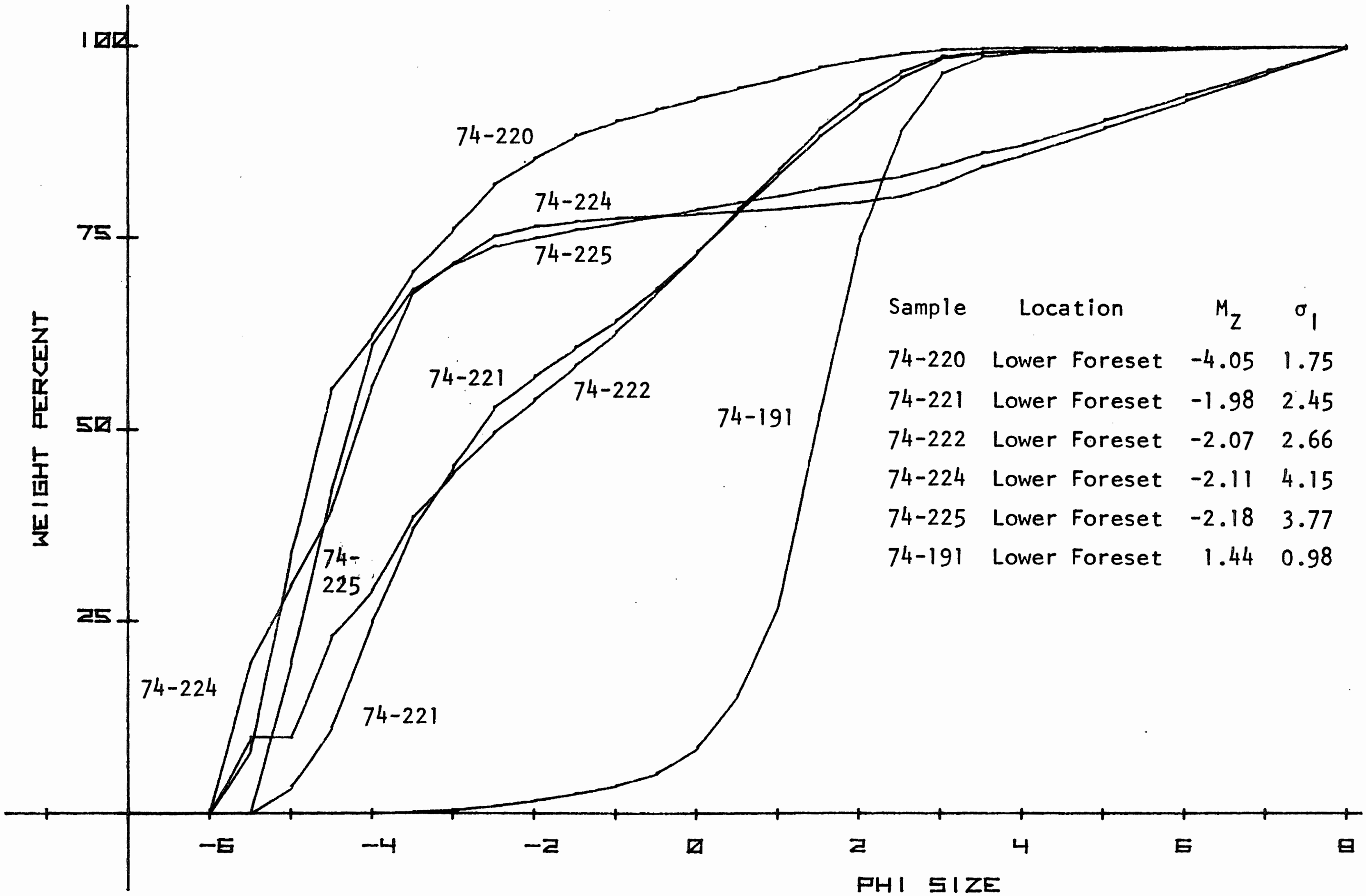


FIGURE 10.11. Grain size analyses, lower foresets.

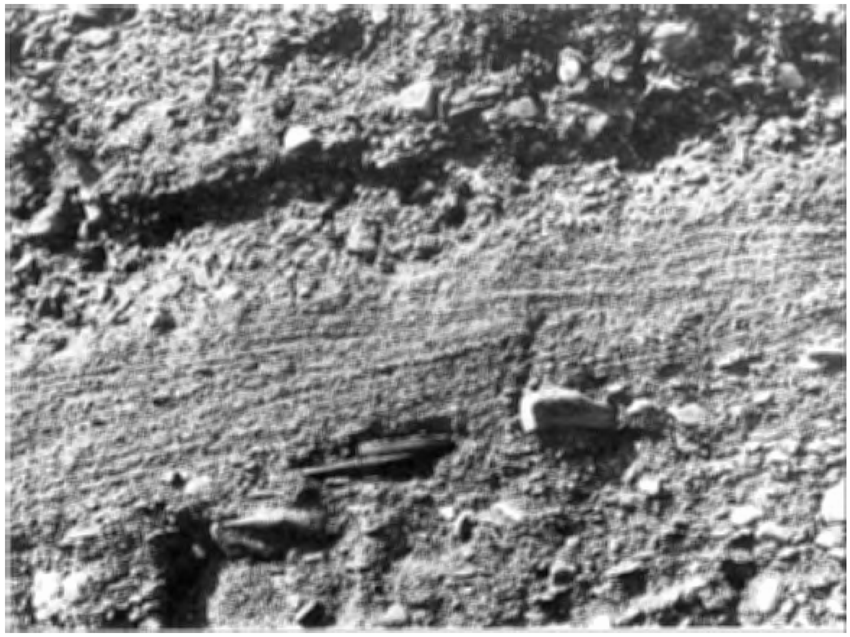


FIGURE 10.12. Sand bed in lower foresets. Bottom 2 cm laminated parallel to bedding. In top 3.5 cm, low angle cross stratification (11° to bedding) inclined up the bedding.

FIGURE 10.13. Pebbles imbricated up dip (to right). Photograph taken parallel to bedding (25° dip). Sand in figure 10.12 continuous with, but up dip from, sand at top of photograph.

the foreset facies. Gravel beds with the aforementioned red clay matrix extended into the bottomset beds where the gravel content pinched out while the clay continued as a clay bed.

Interpretation

The characteristic feature of Gilbert type deltas is the steeply inclined foreset beds. Steeply inclined foresets are the result of bed load deposition (Collinson, 1978, p. 65) so the structure of the delta predicts that sediment avalanching is the main sedimentary process on the foreset slope. The grain size of the foreset beds in the Minas terrace, mainly pebble gravel with some coarse sand, agreeds with this because it would have been transported to the delta front as bed load (Church and Gilbert, 1975, p. 46).

Allen (1970b) developed a theory for the avalanching of granular solids on the foresets of ripples, dunes and Gilbert type deltas. Avalanching occurs where the rate of deposition is higher on the upper slope than on the lower slope. As sediment is supplied to the upper slope, the angle of the slope increases to the critical angle and failure occurs. The sediment avalanches down, leaving the slope at the initial, smaller angle. Avalanche frequency depends mainly on the rate of sediment supply and the difference between the slope necessary for failure and the slope after failure. As the rate of sediment supply increases, avalanches become more frequent until avalanching is continuous and sediment on all parts of the slope is in a persistent state of downward motion. When sediment supply is low, small slides are arrested on the upper part of the slip face and large, infrequent avalanches occur.

The steepness of the foreset beds (25 to 34°) implies that sliding was almost continuous and therefore the rate of sediment supply was very high. The avalanching (sometimes referred to as sliding) of granular solids down a slip face is a natural grain flow (Middleton and Hampton, 1973, p. 19). In grain flows, the coarsest grains should accumulate at the base of the slope, the effect of gravity, and at the outer part of each bed, the zone of least shear (Brush, 1965, p. 30). Thus, the beds are normally graded with respect to the length of the bed and inversely graded with respect to the thickness of the bed. Middleton (1970, p. 267) suggests that the inverse grading is due to a kinetic sieve mechanism; small grains fall downward between the large grains, displacing the large ones upward. Sallenger (1970, p. 559) agrees with Bagnold (1954) and Brush (1965). Whatever the cause, the effect is the same.

Upper Foresets

The sedimentological characteristics of avalanche deposits are not particularly well developed in the upper foresets and there are two probable reasons for this. The avalanches originate on the upper part of the slope (Allen, 1970b, p. 350) and are deposited on the lower slope. Secondly, the high sedimentation rate and steep slope favour continuous sliding and therefore large slides were infrequent. The lack of sharp bedding contacts and changes in bedding thickness and grain size along dip are consistent with continuous, small scale avalanches that were not large enough to develop the overall characteristics of grain flows. Many avalanches were probably arrested on various parts of the slope. Some beds (4) did coarsen down dip, in accordance with the theory, while 2 beds coarsened up dip.

Four of the 23 beds that were examined in the upper foresets had pebbles that were imbricated up dip. This could be the result of a current flowing down the foreset slope but the imbricated pebbles were the largest pebbles in the bed and were not supported by pebbles their own size, a characteristic of current oriented pebbles (Church and Gilbert, 1975, p. 60). Eynon (1972, p. 43) ascribes pebble imbrication to subequal sized pebbles rolling down a foreset slope in braided stream deposits. Pebble imbrication can also be formed in avalanche deposits and this will be discussed in the interpretation of the lower foresets. Thus, the imbricated pebbles are not conclusive of sporadic current activity on the foreset slope.

Grain size changes between and within beds on the upper foreset slope are the result of different sizes of sediment being transported to the delta front and are therefore probably related to changes in stream discharge or location. Thus, graded beds are the result of a decreasing stream flow supplying progressively finer sediment to the avalanche slope.

The coarseness of the gravel and the lack of sand beds (only one) are compatible in that under the high stream discharge necessary to transport the gravel, the sand would be transported in suspension. As the gravel was deposited on the foreset slope, the sand, already in suspension, would be carried out into the sea for some distance. Sand deposited at lower discharges would not be carried out so far and would infiltrate the coarse gravel to become matrix. The following flow velocity calculations quantify this concept.

The closedwork gravel bed that was split into two fractions and replotted (Fig. 10.8) is convenient for hydrodynamically comparing the gravel and sand fractions. The calculations were made in the same manner as for the glaciofluvial gravel at Portapique and the results are summarized in table 11. Calculations were made for -3ϕ and -5.5ϕ bed load, the two limits of the gravel. The mean grain size of the gravel is -4.8ϕ, which was used for K, the height of the roughness projections. Flow depth was taken as 1 m, although it may have been greater than this.

The minimum velocity necessary to move -3ϕ gravel is 1.0 m/sec and at that velocity, the maximum grain size of the suspended load is 0.7 mm (coarse grained). The minimum velocity necessary to initiate movement in -5.5ϕ gravel is 2.5 m/sec and at that velocity, the maximum grain size of the suspended load is 1.5 mm (very coarse grained). Thus, the 2ϕ (0.25 mm in diameter) sand that is the matrix for the gravel (Fig. 10.8) is hydrodynamically incompatible with the gravel and probably represents infiltration at lower velocities.

This can be checked by doing the reverse procedure, using the same equations. Assume that 2ϕ sand is just being maintained in suspension, and therefore would settle out if the velocity should drop. As U^*, the shear velocity, has to be equal to w, the settling velocity, to keep sediment in suspension, U^* for 2ϕ sand is 2.3 cm/sec, the settling velocity of 0°C (Blatt *et al.*, 1972, p. 54). Rearranging equation (5) to solve for τ_o gives

$$\tau_o = \rho U^{*2} \qquad (9)$$

TABLE II. Hydrodynamic calculations for the foreset gravel at Spencers Island.

Bedding Type	Bed Load (ϕ)	τ_o (dynes/cm^2)	K ϕ	K mm	Flow depth (m)	f	U (m/sec)	Fr	U_{slope} (4.5) (m/sec)	Fr_{slope}	U_{slope} (8.0) (m/sec)	Fr_{slope}	U* (cm/sec)	W mm	W ϕ
Foreset	-3.0	74	-4.8	27.9	1	.056	1.0	0.32	-	-	-	-	8.6	0.70	0.51
Foreset	-5.5	430	-4.8	27.9	1	.056	2.5	0.80	2.5	0.80	3.3	1.05	20.7	1.50	-0.58
	0.0	5.29	2.0	0.25	1	.0145	0.5	0.16	-	-	-	-	2.3	0.25	2.0
	0.0	5.29	-3.0	8.0	1	.0355	0.3	0.10	-	-	-	-	2.3	0.25	2.0

and a value of 5.29 dynes/cm^2 is obtained. Velocities were calculated for flows over a 2ϕ sand bed (K = 0.25 mm) and -3ϕ gravel bed (K = 8.0 mm) and the results are listed in the lower half of table II. The calculated flow velocity over the sand bed is 0.5 m/sec and over the gravel bed is 0.3 m/sec. Therefore, the flow velocities necessary to keep 2ϕ sand in suspension are fairly low compared to the flow velocities required to deposit the gravel. This corroborates the previous suggestion that the gravel framework and sand matrix were deposited under two different flow conditions.

Flow velocities obtained using the gravel bed load were also checked by calculating flow velocity using the gradient of the upper surface of the delta. As the gradient varied from 4.5 to 8.0 m/km, calculations were made using both these values. The results (2.5 and 3.3 m/sec) are in fairly good agreement with the calculated flow velocity for the -5.5ϕ gravel, but as at Portapique, they tend to be higher.

The flow parameters listed in table II must be regarded with some discretion as the calculations are for stream flow and the sediment is from the upper foresets. As the streams reach the delta front, their velocities are checked as they enter the standing body of water. This factor combined with a possible asymptotic transition from the stream bed to the foreset bed, especially in the finer grain sizes, hinders the deposition of the coarsest stream sediment in the foreset beds. Because of this, the stream velocities listed in table II are minimums.

The peculiar channel shaped lens of openwork gravel (Fig. 10.9) in the upper foresets is not an avalance type deposit. The shape of the deposit, its erosional basal contact (it sits in a "low" in the underlying bed) and the openwork nature of the gravel together suggest deposition by a current. If it was deposited by a current, the current was not flowing straight down the foresets because a cross section of the deposit was exposed on the dip section of the foresets. The current must have had a sinuous path or been deflected from proceeding straight down the delta front. Such currents must have been rare as the deposit was the only one observed in the foresets of all the deltas in the Minas terrace. Norman Smith (1976, personal communication) has observed similar deposits in foreset beds of glaciolacustrine deltas.

The general process of sediment deposition at the front of the Spencers Island delta may be summarized as follows. At flow velocities of 1 to 3 m/sec, pebble gravel was moved as bed load and sediment up to a coarseness of medium grained sand was transported in suspension. When the streams reached the delta front, the sand in suspension was carried into the sea where currents and turbulence determined its site of deposition. The coarser sediment moved along as bed load avalanched down the foreset slope. At stream velocities as low as 0.3 to 0.5 m/sec, 2ϕ sand was just maintained in suspension and coarser sand was moved as bed load. The bed load sand avalanched down the foreset slope, infiltrating the coarser sediment. The deceleration of the stream at the delta front probably caused the medium to fine grained sand to settle out of suspension and some of it also infiltrated the gravel on the upper foreset slope.

While variable discharge is characteristic of this environment, the sand could also be deposited in areas away from direct stream discharge as three dimensional mixing took place. The lack of sand beds suggests that periods of low flow lasted only long enough to infill the gravel or that sand beds were eroded by a succeeding high flow deposit.

As most of the gravel at the top of the foresets would be openwork when deposited, preservation of the openwork texture would depend upon: 1. sealing off the gravel from later sand infiltration; or 2. deposition of enough gravel that later sand infiltration could not fill all the voids. These two conditions must have occurred infrequently because openwork gravel is not the norm. Graded beds, with finer gravel at the top, would inhibit infiltration of sand and 3 of the 4 openwork beds were normally graded. The two closedwork beds that were graded were very coarse in the gravel fraction (Figs. 10.7, 10.8) and the large pore spaces would enhance sand infiltration.

The sedimentological characteristics of the 'mini' set of foresets above the second surface of discontinuity confirm some of the processes interpreted for the large foresets. Some of the foreset beds had coarse gravel at their lower terminations, indicative of avalanche deposits. On the flatter portion of the surface, the sand that separated the two sets of foresets graded up dip into foreset beds. Therefore the sand was being deposited at the same time as the foresets. This is consistent with the sand being carried in suspension out into the sea by an overflow as the gravel slid down the foreset slope. The sand then settled out of suspension to form the massive to parallel laminated bottomsets, which accounts for the thinning of individual sand beds towards the south, but the overall thickening of the sand unit is in this same direction. The sand unit built up as the foresets prograded towards and then over the unit.

Lower Foresets

In the lower foresets, the effects of stream discharge on grain size are less obvious, probably due to reworking as the sediment slid down the delta front. Graded and openwork beds were less abundant. Some of the beds were coarser at their base (sample 74-220, Fig. 10.11) than they were up dip (samples 74-221, 74-222, Fig. 10.11) but inverse grading was only described in one bed.

The thickness (≃15 m) of the foreset facies and slope (≃30°) of the foreset beds means that the lower foresets extended approximately 26 m beyond the top of the foresets. This distance was sufficient to have had significant amounts of sand settle out of suspension and deposit on the lower foresets, probably even at high discharge, which accounts for the abundance of sand beds (21%). The sand beds in the lower foresets were massive to parallel laminated, which is consistent with deposition from suspension. The greater supply of sand enhanced infiltration and decreased the probability of preserving openwork beds.

The contact between the foreset and bottomset beds varied from angular to tangential. As predicted from theory (Pettijohn *et al.*, 1973, p. 352), the beds with an angular contact were coarse (avalanche deposits) and the tangential contacts were probably a result of the increase in sediment deposition from suspension.

As in the upper foresets, there were several beds (3 out of 19) that conspicuously had pebbles imbricated up dip. Two of the beds were much thicker (25, 45 cm) than the average (≃12 cm) while one was thinner (7 cm). One of the beds was photographed (Fig. 10.13) and although it was not noticed in the field, the pebbles appear to have their long (a) axis parallel to the dip of the imbrication. Harms *et al.*, (1975, p. 146) suggest that this type of imbrication is indicative of clast dispersive forces, such as in a grain flow, while pebbles imbricated by a current have their a-axis oriented perpendicular to the dip of the imbrication (Harms *et al.*, 1975, p. 136). The clast dispersive imbrication may also be associated with inverse grading and the gravel in figure 10.13 may also have a slightly higher concentration of pebbles at the top of the bed. Thus, the imbricated pebbles may be the result of a larger avalanche (the bed in Fig. 10.13 is 45 cm thick) that became a true grain flow. The other beds with imbricated pebbles may also have been grain flows, but the type of imbrication was not noticed in the field. If the beds were grain flows, they represent about 15 to 20 percent of the beds in the lower and upper foresets.

The sand bed with the low angle cross stratification dipping up dip (Fig. 10.12) is problematical. The low angle of the cross laminae relative to the bedding plane (11°), their slightly convex up shape and the long wave length (20 cm) are unlike cross stratification produced by ripple movement; and the cross stratification is dipping up dip. The cross stratification is similar to that of antidunes (Blatt *et al.*, 1972, p. 133) which would result from deposition of a current in the upper flow regime. The cross stratification also overlies parallel laminated sand which, because of the grain size (less than 0.6 mm, sample 74-191, Fig. 10.11) indicates upper flow regime conditions (Blatt *et al.*, 1972, p. 121) or deposition from suspension. The sharp and possibly erosional lower contact and the sharpness of the laminae are more compatible with the former concept.

The cause of such a rare, high speed current is not obvious as there is no other evidence of current activity in the lower foresets. However, there is one interesting possibility brought about because the sand bed overlies the thick pebble imbricated gravel (Fig. 10.13). If the gravel is a grain flow deposit which originated on the upper foreset slope, it is possible that sand in the avalanche was put into suspension above the gravel and followed the avalanche downslope as a fine grained turbidity current. This would in effect be one form of sediment gravity flow (grain flow) producing another (turbidity current). The position (lower foresets) of the deposits is correct and even the abundance of sand in the slide agrees with Allen's (1970b) theory on avalanche events and the interpreted process of sedimentation on the foreset slope. Larger, infrequent slides occur when rates of sediment supply are low, and when rates of sediment supply are low, sand would be deposited on the foreset slope. Overdeposition on the upper foreset slope is not the only cause of large slides. Slumps in the foresets, which have been documented at the Lower Five Islands and Port Greville deltas, would also generate large avalanches and later discussion will suggest that this happened. This series of events (slide [slump] → grain flow → turbidity current) is tenuous because the central idea of a grain flow is based only on a poorly documented pebble imbrication but it is an interesting possibility.

The 'antidune' need not be related to the imbricated gravel to be a turbidity current deposit for any large slide on the upper foreset slope could generate a turbidity current. Antidunes are rare in the geologic record (Blatt *et al.*, 1972, p. 134) and one of the few examples from the rock record is a turbidite sequence (Skipper, 1971).

The distinctive "red beds" of the lower foresets are genetically related to the bottomset beds and are discussed in the bottomset section.

Multiple Foreset Beds

In the upper foresets, the two surfaces of discontinuity were similar to those of the Port Greville delta and similarly, the two most probable causes of overlying sets of foresets are changes in sea level or changes in the level of the delta. Either one means that the surfaces were former topset/foreset contacts.

The first surface dipped steeply (14°) in the direction of delta progradation (SE) and was exposed for about 3 m before being covered by the slump blocks in the middle of the exposure. The topset/foreset contact at the second set of foresets was continuous with, and at the same height as, the topset/foreset contact over the rest of the exposure. To account for these relationships by changing sea level would require an extremely rapid drop of at least 3 m followed by a rapid increase to the same height as before the drop. The rapid drop in sea level is inconsistent with the horizontality of the topset/foreset contact, and thus stability of the sea level, elsewhere at the exposure. The subsequent rise in sea level to exactly the same height as before the drop seems improbable.

An alternative, and the favoured, interpretation is that the foresets slumped. This would make the surface equivalent to the type A surfaces at Port Greville. The foresets beneath the surface were not disrupted but the foresets under the up dip end of type A surfaces at Port Greville are also well bedded. It is at the down dip end of the surfaces that the foresets are disrupted and at Spencers Island, this was not visible. After the slump, a second set of foresets prograded over the first set.

The second surface was not as steeply dipping, about 8°SE at its point of origin, and shallowed to 3°SE over the rest of the exposure. The channel deposit at the top of the lower foresets is evidence that the lower foresets were at one time at the topset/foreset contact. The surface is only 1 to 1.5 m below the topset/foreset contact, which is once again continuous with and at the same height as, the topset/foreset contact over the rest of the exposure. To explain these relationships by changing sea level requires a complex and improbable series of events and also, changing sea level does not explain the oversteepened (36 to 37°) foresets beneath the surface.

Subsidence of the delta seems to be a more reasonable explanation for the surface and its similarity to the type B surfaces at Port Greville corroborate this interpretation. The surface was more extensive laterally and not as steep as the first surface, which implies that the subsidence was due to movement in the underlying bottomset beds. As well, oversteepening of the foresets requires downward movement of part of the bottomsets. The bottomsets probably became overloaded (under-compacted) when the foresets prograded over them and they reacted to this load by faulting or flowage. This is the process suggested for the formation of the equivalent type B surfaces at Port Greville.

There is no evidence for movement in the bottomsets at Port Greville because the bottomsets are not exposed, but this was not the case at Spencers Island. Faults, some of them penecontemporaneous, and slumped intervals are described in the next section dealing with the bottomsets. The multiple foresets and disrupted bottomsets at Spencers Island strengthen the previous interpretation that at Port Greville the surfaces of discontinuity were created by subsidence of the delta rather than by fluctuating sea levels. The pattern of sea level changes that is necessary to produce the multiple foresets at Spencers Island is different than the pattern required to produce the multiple foresets at Port Greville. The deltas are adjacent to each other in the Minas terrace and sea level changes should have been the same everywhere in the Minas Basin.

As suggested in the discussion of the multiple foreset beds at Port Greville, subsidence of the deltas, at least on the scale required to deposit several sets of foreset beds, appears to be related to water depth. The reason given for this was that greater water depth causes a higher degree of undercompaction of the bottomset beds by thickening, and thus increasing the weight of, the overlying foreset facies. Another corroborating factor is that the bottomsets deposited in deeper water tend to have a higher proportion of fines. The bottomsets at Spencers Island, the deepest bottomsets exposed in the Minas terrace, have a high (≃70 to 100%) silt/clay content and a high proportion of fines increases the susceptibility to undercompaction.

The lack of topset beds along the second surface of discontinuity is problematical. The topset beds may have slumped off of the steeply dipping surface, but the second surface is too shallow for this to happen. If the subsidence occurred when the foreset beds were near the outer margin of the delta, the overlying topset beds would have been relatively thin. But the complete absence of them except for one incised channel deposit implies that marine erosion was very thorough. Even at Port Greville, the topset beds are thin to nonexistent on the type B surfaces and why this is so is not understood.

BOTTOMSET FACIES

Description

Structures and Textures

Approximately 5 m of bottomset beds were well exposed from 2 m below high tide to 3 m above high tide. The exposure was heavily faulted with high angle normal and reverse faults showing displacements of less than a metre. Larger magnitude rotational slump faults had tilted most of the bottomset beds so that they dipped N or NW instead of S or SE and bottomset beds with a 'backward' tilt are uncovered from time to time along the beach in front of the former (1974) exposure. The large faults may have occurred at any time after deposition of the bottomsets.

The bottomset beds were composed of couplets that coarsened upwards from clay/silt through clay/sand to clay/gravel. The clay beds were red brown and had a fairly constant thickness throughout the section. They are referred to as 'clay' beds because the distinguishing feature of the beds in the field was the clay content but grain size analyses, presented later, show that silt was a significant component. The coarsening up of the silt/sand beds was accompanied by a marked thickening. The couplets or rhythmites were essentially flat lying or shallow (1 to 5°) seaward dipping (S to SE), except where they had been faulted and dipped 5 to 10° N or NW.

Clay/silt Rhythmites

Eighteen clay beds and nineteen silt beds from the clay/silt interval at the base of the section were recorded in the field. The description started where the silt beds began to be consistently present and weathered out on the exposed surface. Below this, the bottomsets were essentially clay with few silt beds thick enough to recognize. The average thickness of the red brown clay was 2.4 cm, with a standard deviation of 1.1 cm. The silt beds were thinner, averaging 1.6 cm, with a standard deviation of 1.0 cm.

The colour of the clay varied from a fairly bright red to a dark red brown. The upper and lower contacts of the clay beds varied from sharp to gradational without a tendency for one or the other. Fifty six percent of the beds had scattered clasts, usually ranging in size from granule to 2 cm and in the beds where the clasts were concentrated, the clasts were more abundant in the lower part of the bed. Over half (58%) of the clay beds contained scattered tube shaped structures, ranging from 1 to 6 mm in diameter, infilled with very fine sand. The structures were horizontal to subhorizontal and are interpreted as burrows. Marine pelecypod molds, discussed in a separate section, were found in the lower clay beds. The molds were commonly of one shell but several molds contained both valves in a living position.

The silt layers were composed of sandy, coarse silt with a red clay matrix and had scattered clasts from coarse grained to 2 cm. Half of the lower 10 silt beds had scattered clasts while 8 out of the 9 upper silts had scattered clasts and of these, 6 had the clasts concentrated at the top of the bed. Four of the silt beds reached a coarseness of sand size.

Liquefaction of Bottomsets

Twenty nine centimetres from the bottom of the clay/silt interval was a 30 cm horizon of convoluted beds. The clay/silt rhythmites were contorted into irregular shapes but weathering of the silts showed that the unit had not been homogenized and stratification still existed. 'Roll up' structures (Born, 1972, p. 21) were common.

In 1976, another section of the bottomsets was exposed along the beach and a 20 cm disturbed interval, similar to the one described above, occurred in the section. It was not the same interval as it was found at a higher elevation, higher stratigraphic interval (clay/sand) and down dip from the one exposed in 1974.

The disrupted beds represent zones of flowage within the bottomsets. Similar zones of disrupted bedding in the foreset and bottomset beds of a Pleistocene Gilbert type delta are described by Born (1972, p. 20). He attributes the zones to liquefaction of the sediment, which he discusses in some detail. Liquefaction is caused by undercompaction of the sediment and overpressuring of the intrastratal pore fluid. Rapid depositional loading can cause liquefaction.

The contorted sediments in the Spencers Island bottomsets appear to have been deformed while in a liquefied state. The liquefaction may have occurred at any time after the sediments were deposited, but it is probable that it occurred just after the foresets had prograded over those particular bottomsets. Sediments adjust (by compaction) to the weight of overlying sediments with time so the most unstable period is just after the loosely compacted sediments (a feature of rapid deposition) are subjected to an overlying load. Thus, liquefaction probably occurred in sediments near the delta front, an interpretation suggested by Born (1972, p. 21). Some of the disrupted sediments in the Lower Five Islands delta probably also represent zones of liquefaction.

Although the disrupted intervals described here are only 30 cm and 20 cm thick, the deformation of liquefied bottomset sediments is consistent with the interpretation of deltaic subsidence at Spencers Island and Port Greville. Obviously larger zones of flowage would be needed to lower the delta surface substantially and faulting would probably be involved. But the fact that liquefaction did occur strengthens the concept of bottomset failure because of undercompaction.

Clay/sand Rhythmites

Clay Thicker than Sand

Above the clay/silt interval the silt beds coarsened to sand. Forty one clay/sand rhythmites were measured in the field (Fig. 10.14). The lower 13 rhythmites were similar to the underlying clay/silt rhythmites except for the increase in grain size to sand. The clay beds were thicker than the sand beds, averaging 2.4 cm with a standard deviation of 1.2 cm. The sand beds averaged 1.3 cm thick with a standard deviation of 1.1 cm. The beds were drier (closer to high tide) than the beds in the clay/silt interval which made them easier to describe and as a result a coarsening up cycle began to be apparent.

Ten out of the 13 clay beds had a sharp, fairly planar lower contact while 10 beds had a gradational or irregular upper contact. The disturbed upper contact was caused by clay wisps protruding into the sand and lenses and stringers of sand extending down into the clay. Burrows were common (10 beds) and scattered within the clay beds. One burrow was noted that

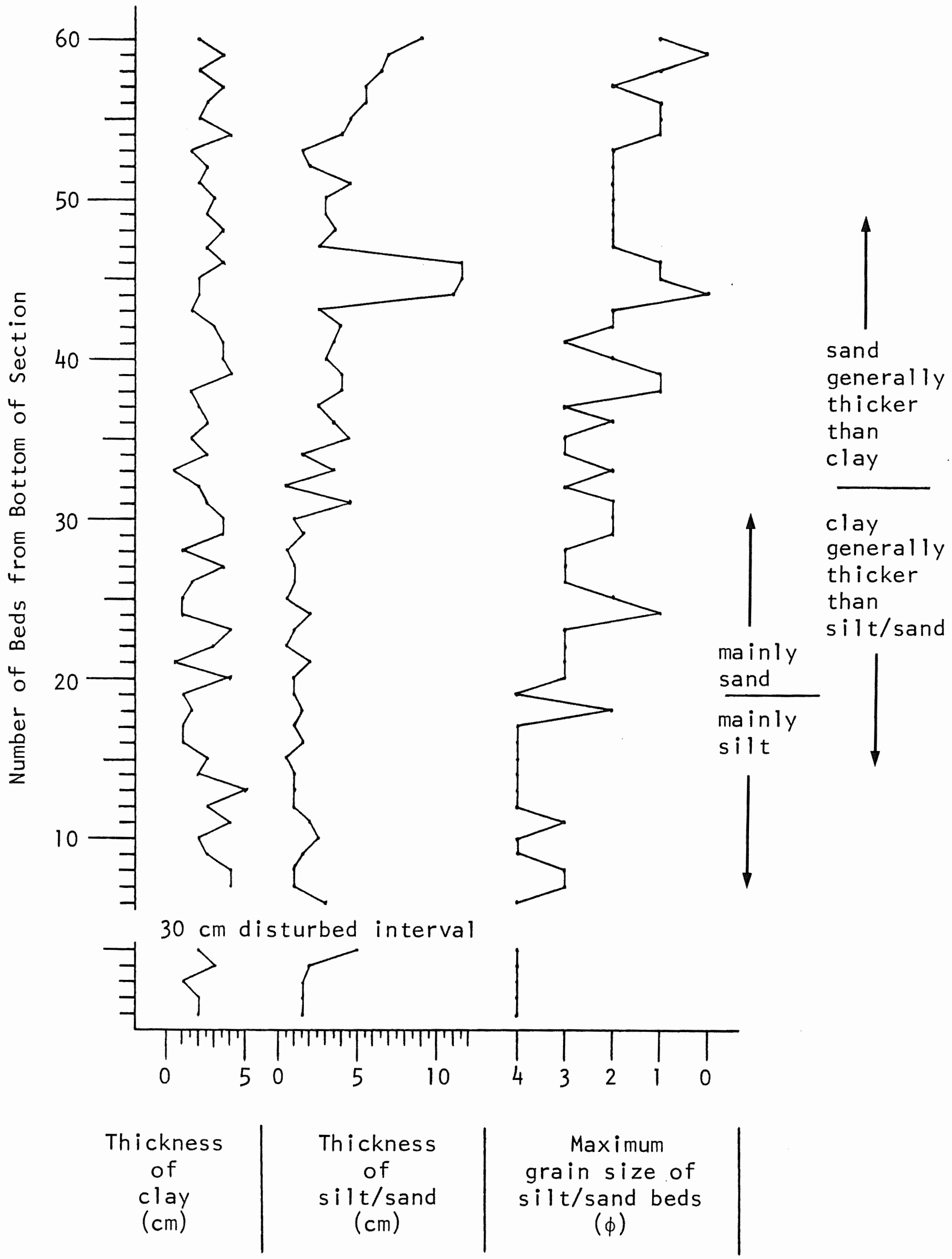


FIGURE 10.14. Thicknesses of the clay and silt/sand beds and maximum mean grain size attained in each silt/sand bed in the bottomset facies, Spencers Island. Clay thickness almost constant while silt/sand thickness and grain size increase upwards. 4ϕ = silt; 3ϕ = very fine, 2ϕ = fine, 1ϕ = medium, 0ϕ = coarse grained.

bottomed in a clay bed and extended into the overlying sand. Scattered clasts, up to 12.5 cm, were present in 4 beds and silt laminae occurred in the middle of 4 of the beds.

Eight of the sand beds were very fine grained, 4 were fine grained and one reached a coarseness of medium grained. Three of the beds were normally graded, two of those graded from very fine grained to silt, while the medium grained sand was graded to fine grained at the top. Four of the beds were inversely graded from very fine grained to fine grained including 3 of the top 4 beds in the interval. One of them had a 'cap' of very fine grained sand. All of the sands had a red silty clay matrix. Scattered clasts were very common (11 out of 13 beds), ranging in size from coarse grained to 2 cm. The clasts were more numerous at the top of the bed.

The sharp lower and disturbed upper contact of the clays, combined with the coarse nature of the top of the sand beds, produced a coarsening up rhythmite that was accentuated by weathering. The bottom of the clay layer stood out and the clay bed weathered back slightly towards the top, with the recession continuing in the sand layer until a recessive pocket formed at the top of the sand bed (Fig. 10.15).

Sand Thicker than Clay

The upper 28 clay/sand rhythmites showed a pronounced thickening upwards of the sand bed while the clay bed remained at much the same thickness as in the clay/silt interval (Fig. 10.14). The average thickness of the clay beds was 2.5 cm with a standard deviation of 0.9 cm. The sand beds averaged 4.7 cm thick with a standard deviation of 2.8 cm, so there were greater fluctuations in thickness as well.

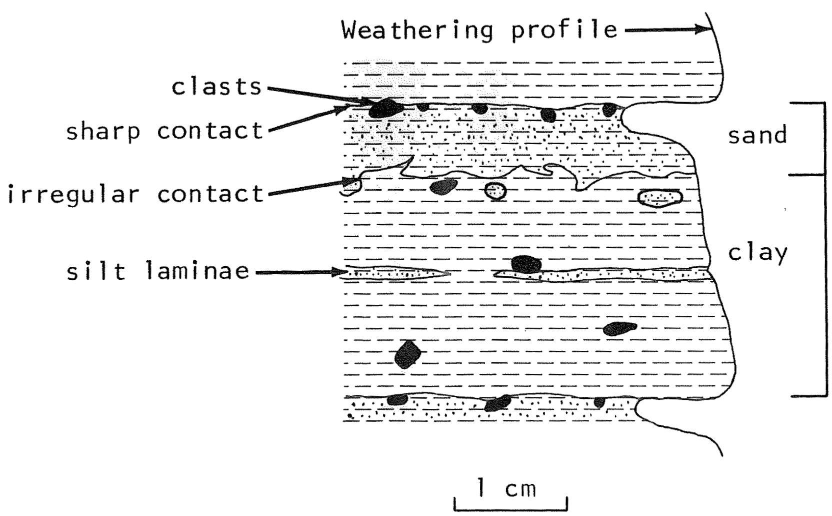


FIGURE 10.15. Top — Clay/sand bottomset rhythmites. Weathering profile controlled by grain size. Bottom — Schematic diagram of clay/sand rhythmite, clay thicker than sand.

Of the 28 clay beds, 82 percent (23) had sharp lower boundaries and of those 57 percent (13) were fairly planar (versus irregular). In contrast, only 36 percent (10) of the upper contacts were sharp and 96 percent (27) were irregular or gradational. Burrows were again common (in 50% of the beds) and of these, 86 percent (12) were found in the upper half of the bed. Most of the burrows were subhorizontal to horizontal, and some had a significant vertical component but none was at 90° to the bedding plane. Scattered clasts, from medium grained to pebble size, were found in half (14) of the beds and 79 percent (11) of those had the clasts in the lower part of the bed.

As mentioned previously, the clay beds do not consist solely of clay. Table III gives the grain size analyses for 5 clay beds and they show that significant amounts of silt and some sand occur in the beds. The beds were split into upper and lower halves (the interval numbers in Table III refer to their location in the section as measured from the bottom) for grain size analyses. The mean grain size of the beds (except for bed no. 60) was calculated using the silt and clay fractions and just the clay fractions. The sand fraction was omitted because it probably represents a different depositional process and the amount of sand in the beds, especially in the upper halves, was strongly affected by burrows. This effect was eliminated where possible.

The finest bed was lowest in the section (bed no. 13, in the clay/silt interval). There was a slight tendency for the beds to be siltier towards the top, something also noted in the field. Silt, or less commonly sand, laminae (1 mm to 1 cm thick) with a high amount of clay matrix were also present in 41 percent (13) of the beds. The laminae were

TABLE III. Grain size analyses, 'clay' beds, bottomset facies.

Bed no.	Interval	Percentage Sand	Silt	Clay	Silt/clay ratio	Mean grain size silt and clay	(ϕ) clay
60	439.5-440.5	18.85	50.13	31.02	1.62	-	-
	438.5-439.5	5.42	45.23	49.35	0.92	-	-
44	223-224	22.03	37.34	40.63	0.92	8.64	10.56
	222-223	26.92	32.57	40.51	0.80	8.72	10.55
39	190.5-191.5	14.71	40.04	45.25	0.89	8.67	10.63
	189.5-190.5	7.09	47.70	45.21	1.06	8.26	10.44
35	168.5-169.5	16.26	44.35	39.39	1.13	8.16	10.72
	167.5-168.5	12.95	44.24	42.81	1.03	8.21	10.54
13	92-94	22.90	26.46	50.55	0.52	9.33	10.68
	90-92	1.60	30.66	67.74	0.45	9.49	10.65

present in the upper two thirds of the clay bed and thin laminae were frequently discontinuous while thicker laminae were continuous. Some beds had several laminae. The laminae became thicker and coarser towards the top of the interval. Lenses and stringers of sand, many due to burrowing, were common in the upper part of the clay and contributed to the disruption of the upper contact.

The thickening upwards of the sand beds in this interval was accompanied by a coarsening upwards. The sand beds (Fig. 10.16) were mainly very fine grained to fine grained at the bottom of the interval and coarsened to medium and coarse grained towards the top of the interval. The thickness of the sand beds correlates fairly well with grain size (Fig. 10.14). Of the 4 beds that reached a maximum grain size of very fine grained sand, their average thickness was 3.0 cm. Beds that reached a maximum grain size of fine grained sand (13) averaged 3.2 cm thick; medium grained sand beds (7), 6.1 cm thick and coarse grained sand beds (4), 9.0 cm thick. No sand beds reached a coarseness of very coarse grained.

As well as an overall trend towards upward coarsening, there was a consistent coarsening upward trend in grain size within individual sand beds. This trend was evident in the uppermost sand beds of the underlying (clay thicker than sand) interval. Twenty one of the sand beds (75%), including the upper 12, coarsened upwards and all but the uppermost sand did so from a very fine grained sand at the base. The very fine grained sand (usually 1 to 2 cm thick) coarsened upwards to a fine or medium grained sand (usually 2 to 3 cm thick). Over half (12) of the coarsening up sands had a cap (0.5 to 1.5 cm thick) of finer grained sand, usually very fine grained at the bottom and fine grained at the top of the interval.

In the middle of the interval, there were three successive sands (bed nos. 44-46) that were very thick (11 to 11.5 cm). They coarsened from a very fine grained sand (2 to 3.5 cm thick) at the base to a medium grained (2) or coarse grained (1) sand (6.5 to 7.5 cm thick) capped by a finer grained sand (1 to 1.5 cm thick) at the top. Figures 10.17 and 10.18 show this fine/coarse/fine sequence.

Of the 7 beds that did not coarsen upwards, 3 had a uniform grain size of very fine grained. Thus, 24 (86%) of the sand beds were very fine grained at the base. Of the remaining 4, 2 were fine grained throughout, 1 was graded from medium to fine grained and the other coarsened upwards from a fine grained sand at the base. Four of the sands that coarsened upwards were not inversely graded as there were fairly sharp contacts between the different grain sizes. Three of the distinct contacts were between the finer lower sand and the coarser middle sand.

The other bed in which a sharp change in grain size occurred was the third bed (bed no. 58) from the top of the interval. The common very fine, to medium, to fine grained graded sequence was interrupted at the top by a normally graded bed. This bed occurred above 1.5 cm of fine grained cap (sample 74-253, Fig. 10.19) and had a sharp lower contact that marked a change in colour from red brown to yellow brown as well as a change in grain size. The bed was 3 cm thick and graded from medium (sample 74-254, Fig. 10.19) to fine grained with a clay matrix at the top. There was an unusually high concentration of clasts at the top of the bed which had the effect of coarsening the fine grained sand to medium grained

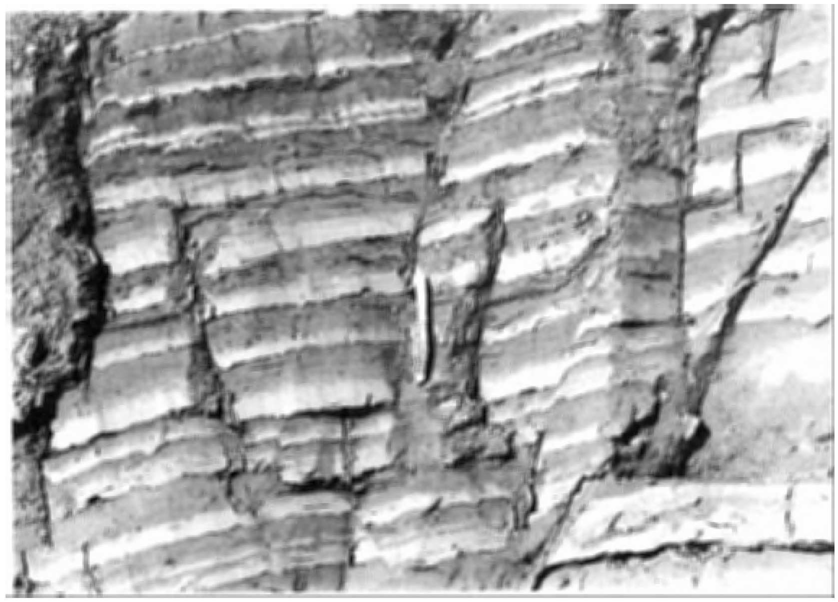


FIGURE 10.16. Faulted clay/sand rhythmites in bottomset facies. Sand thicker than clay, lower clay contact sharper than upper contact. Scattered clasts most common in lower clay/ upper sand (knife blade rests against pebble).

FIGURE 10.17. Sand bed no. 45, 11.5 cm thick. Very fine grained with 15.4% red clay matrix at bottom grading to medium grained (1% of matrix) in middle and then finer with 3.2% red clay matrix at top. Underlying clay bed has sharp lower contact, disrupted upper contact. Prominent sand infilled burrow (B) is connected to overlying sand in 3rd dimension (U shape).

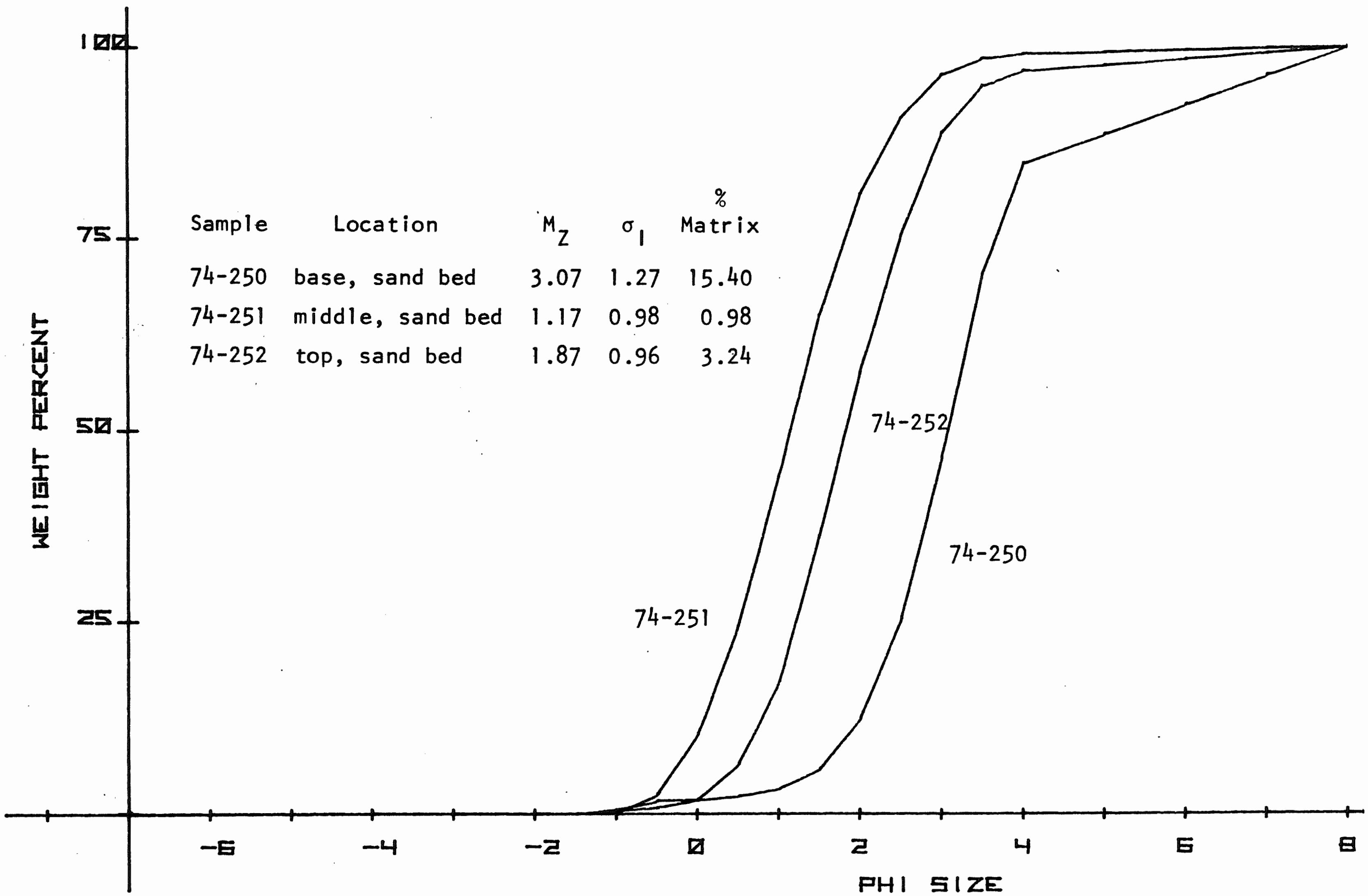


FIGURE 10.18. Grain size analyses, bottomset facies.

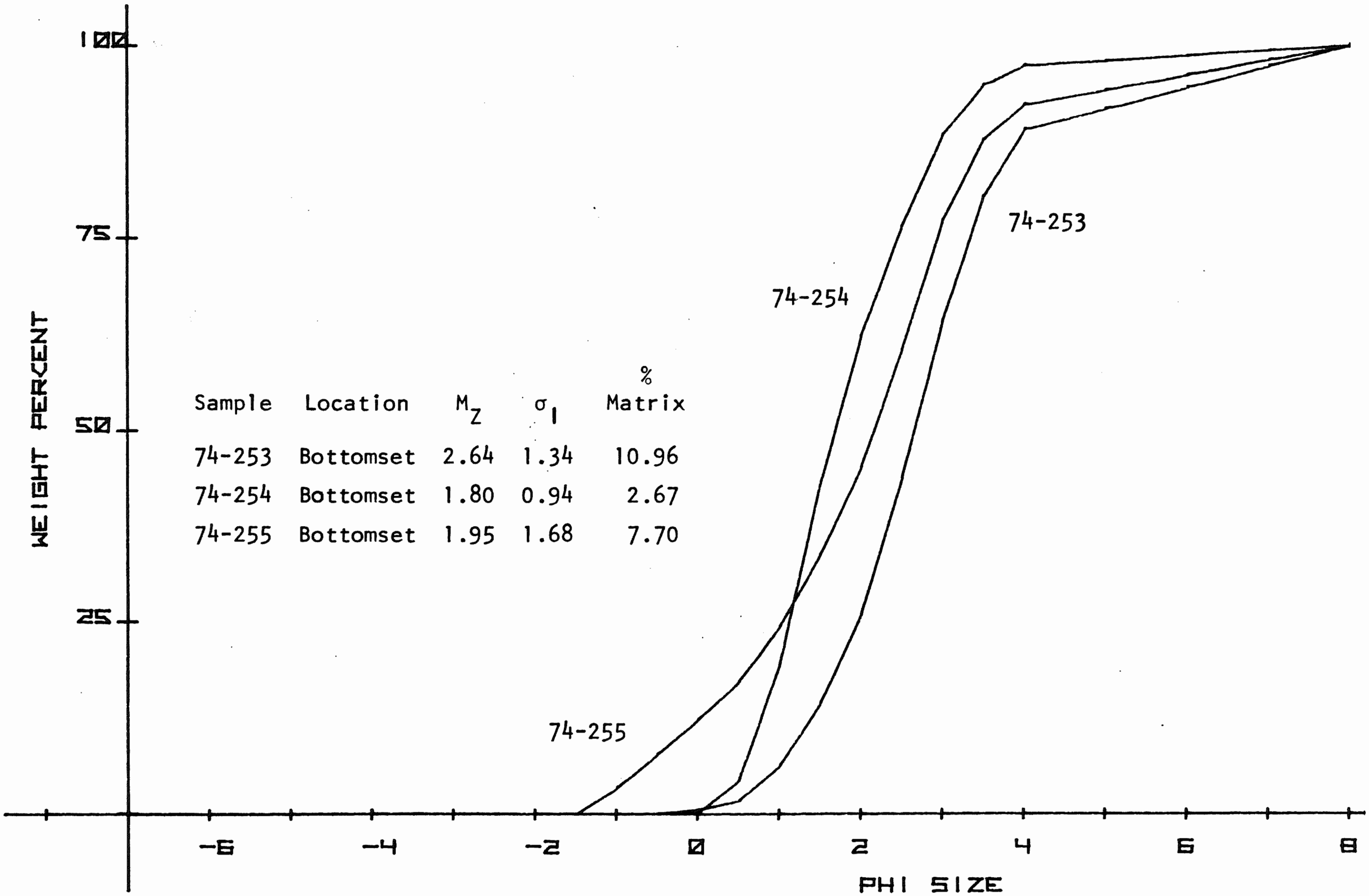


FIGURE 10.19. Grain size analyses, sand bed no. 45, bottomset facies.

in places (sample 74-255, Fig. 10.19). The lower contact was sharp (Figs. 10.20, 10.21) but it was also slightly undulating and some of the clay matrix from the underlying sand appeared as 'wisps' in the bottom of the layer. The contact may have been erosional but a considerable thickness (1.5 cm) of the underlying cap remained.

A red clay matrix was present in the thin sand beds and in the lower and upper parts of the thick sand beds. These were also the finer grained sands, so in fact the clay matrix was size specific. In the thick beds, the lower part had the most matrix, the middle was almost matrix free and the upper part had some matrix (Fig. 10.17), with the matrix giving the yellow brown sand a red tinge. Scattered clasts were once again very common in the sand beds (in 23 [82%] of the beds) and in almost all of those (19, 83%) there was a higher clast concentration at the tops of the beds.

Although definite changes in grain size occurred within the sand beds, the layers of a particular grain size were not always constant in grain size or thickness. In particular, the coarse sands in the middle to upper parts of the thick beds contained parallel to wavy laminae of slightly different grain sizes but the laminae were not laterally persistent. Even when laminae were not discernible, the large flat grains were oriented parallel to the bedding and effected a faint stratification. The thickness and lateral persistence of the coarse sands themselves varied considerably, as shown by the coarser sand below the sharp contact in figure 10.20. Beds with a uniform grain size were massive and cross stratification was conspicuously absent in all of the beds.

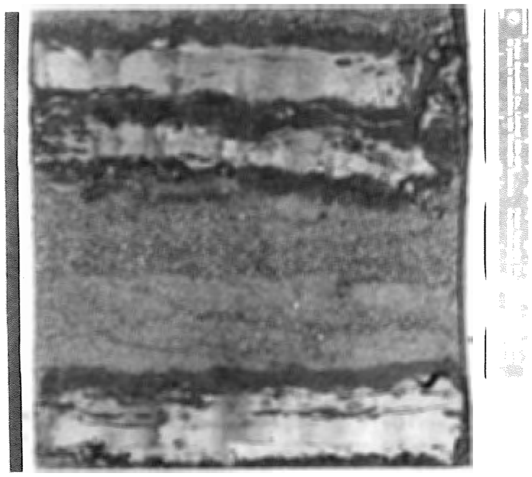


FIGURE 10.20. Inversely/normally graded sand with red clay in lower and upper parts (10.96% matrix) overlain, with a sharp contact, by a sand graded from medium (2.7% matrix) to fine grained (7.7% matrix). Clasts concentrated at top of sand bed. Underlying clay bed has abundant sand stringers at top. Sand bed, 1 cm thick, in overlying clay bed.

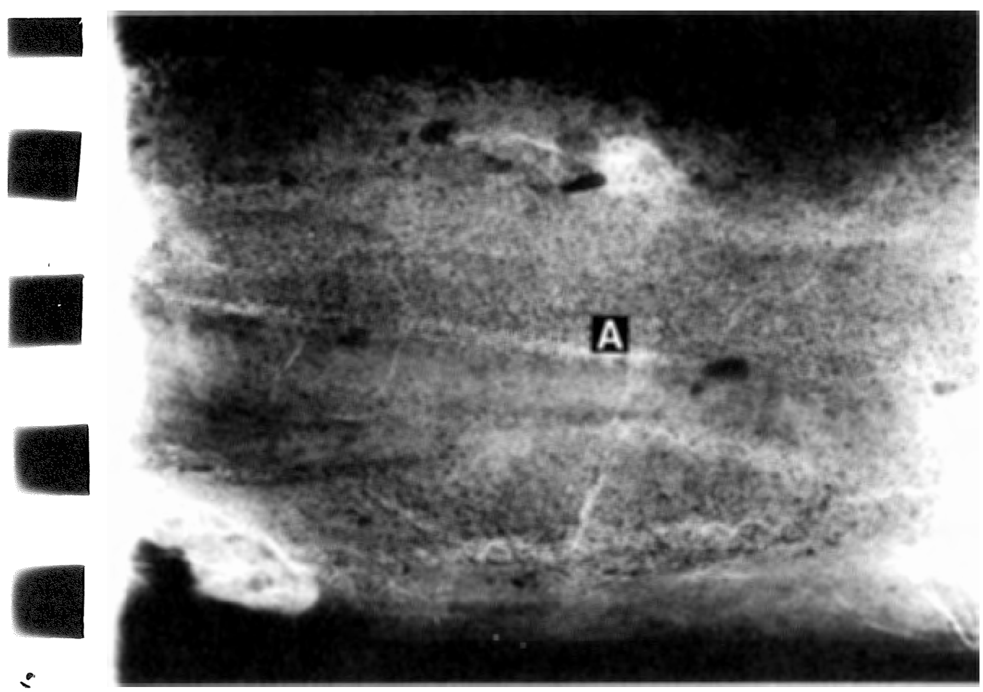


FIGURE 10.21. X-radiograph of sand in figure 10.20. Sharp contact (A) at base of graded bed. Abundant clasts at top of sand bed. Faint cross lamination(?) in lower sand at left. Black squares are 1 cm.

The bottomset section exposed in 1976 was also measured and it contained 22 clay and 21 sand beds. In the upper 15 beds the sand was generally thicker than the clay (Fig. 10.22). The rhythmites generally had the same characteristics as those described previously, but there were several differences, perhaps due to the location of the exposure (or to a more observant eye!) that facilitate a more complete description of the bottomset beds.

The clay beds were abundantly burrowed as before and some of the burrows were studied in detail. Those that were examined carefully went from the sand layer down into, but not through, the underlying clay bed. The burrows usually flattened in the clay bed to a subhorizontal to horizontal position. They were infilled with well sorted sand and some of the burrows had an enlarged cavity in the horizontal position (Fig. 10.23).

Pebbles were found scattered throughout the clay and sand beds and the concentration of pebbles at the top of the sand beds was once again evident. However, many of these protruded into the clay and were in an "unstable" position with respect to the sand. In other words, it was a sharp or pointed part of the pebble that was in contact with the sand and if the clay were to be removed, the pebble would roll over to a "stable" position (Fig. 10.24). One elongated pebble was found extending from the sand into the clay layer with its a-axis vertical ("standing" in the clay).

The sand beds were similar to those previously described, with inversely and inversely-normally graded beds approximately equal in number to those that had a uniform grain (8 vs. 9) but increasing in proportion to the top of the section. The most noticeable difference was an increase in normally graded (2) and coarse clean sands and gravels (3) that

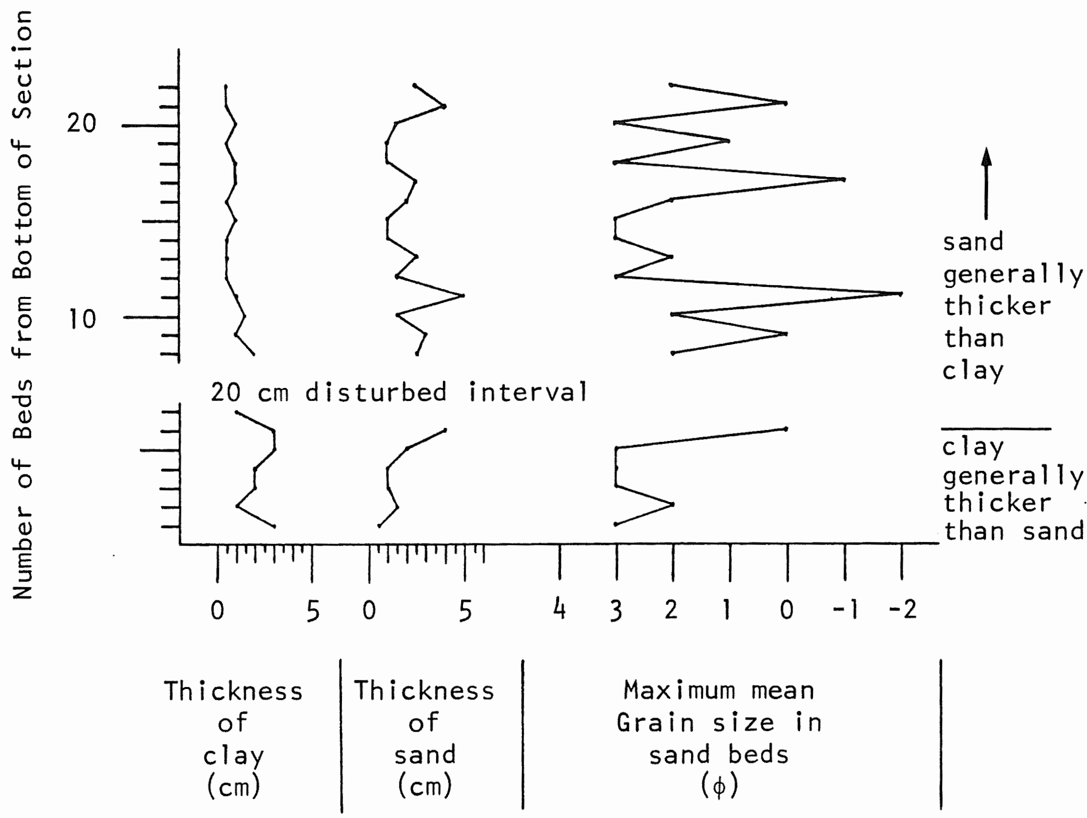


FIGURE 10.22. Thicknesses of the clay and sand beds and maximum mean grain size attained in each sand bed. Thicknesses and grain sizes of the sand beds are very irregular. Sands (5) coarser than 1ϕ are discussed in text. 4ϕ = silt; 3ϕ = very fine, 2ϕ = fine, 1ϕ = medium, 0ϕ = coarse, -1ϕ = very coarse grained; -2ϕ = granular gravel.

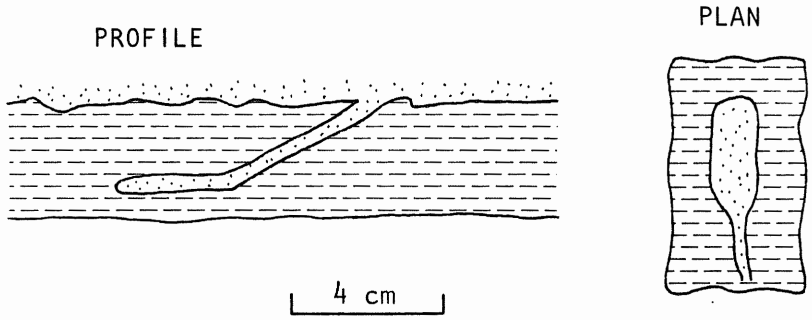


FIGURE 10.23. Sand infilled burrow in clay bed. Enlarged cavity in horizontal position. Burrow does not extend into sand bed below.

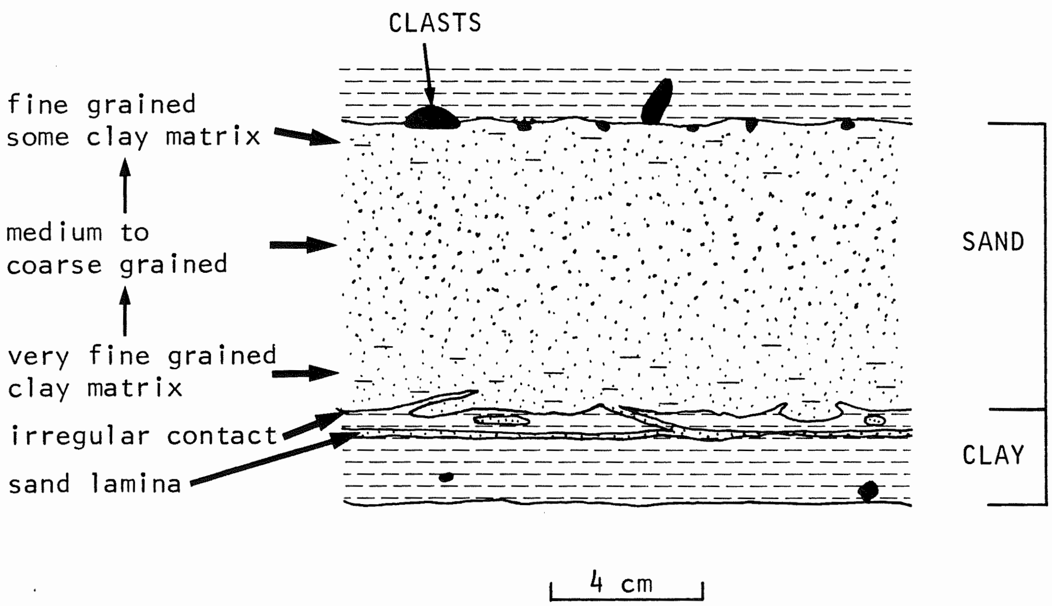


FIGURE 10.24. Schematic diagram of clay/sand rhythmite, sand thicker than clay. Sand is inversely-normally graded. Sand infilled burrows in clay. Some clasts in "unstable" position.

occurred within the sand beds, similar to the graded bed in figure 10.20. All had a sharp lower contact and only 1 of the 5 occurred at the bottom of the sand bed such that the lower contact was with the underlying clay. That bed was graded from coarse to fine grained sand (4 cm thick in total) and had a very red colour from an unusually high percentage of red clay matrix. It was overlain by 1 cm of very fine grained sand with a gradational contact. The other graded bed was 2 cm thick, occurred above 1 cm of very fine grained sand and was graded from coarse to medium grained sand (Fig. 10.25). The upper part immediately underlying the clay bed had some clay matrix. These two graded beds were conspicuous as they were located low in the section (bed nos. 6 and 9, respectively) and yet were coarse grained (surrounding beds were very fine to fine grained).

Two of the 3 coarse ungraded layers occurred at the top of sand beds. One was only 3 mm thick and essentially consisted of a blanket of granules while the other was 1 cm thick and very coarse grained. Both were matrix free. The third ungraded layer was a coarse grained sand, with little matrix, that occurred in the middle of a clayey, very fine grained sand.

Clay Mineralogy

Four samples from the clay beds were taken for x-ray diffraction analysis. The preparation technique and method of mineral determination and abundance are described in Piper and Slatt (1977). All of the samples were from the clay/silt interval in the 1974 exposure. Three clay beds were sampled and one of the beds was sampled twice; at the bottom where it was reddish brown and at the top where it was brown.

The semiquantitative analyses show a consistent clay mineralogy in the beds, even where the colour of the clay varied. Illite is dominant and varies from 73 to 75 percent, chlorite is next in abundance (14 to 19%) followed by kaolinite (8 to 11%). Montmorillonite is absent.

The relative abundances of the clay minerals are consistent with their general abundances on the eastern margin of Canada (Piper and Slatt, 1977). Piper and Slatt attribute high grade metamorphic terrain as the source for illite and low grade metasedimentary and metaigneous terrain as the source for chlorite, the two most common clay minerals. Possible sources given for kaolinite are the Lower Carboniferous and Triassic red beds of the Maritime Provinces, while the Triassic red beds are suggested as a source for montmorillonite. Montmorillonite is absent or rare in eastern Nova Scotia, as well as in other parts of the Atlantic provinces.

Bottomset/foreset Transition Zone

The transition zone between the bottomsets and the foresets was exposed in a slump block (1974) and although the exposure was not laterally continuous, it did show the characteristics of the transition. Interfingering of some of the bottomset and foreset beds resulted in gravelly bottomset beds and sandy foreset beds, although the interfingering only covered several metres (Fig. 10.26). Several faults unrelated to the slump cut the bottomset beds and most of these were high angle (>45°) normal faults with displacements less than 1 m. Some of the faults did not cut the overlying foreset beds, so they occurred before the foresets were deposited. Faults that did affect the foreset beds only cut the foresets immediately above the bottomsets, so the faulting occurred just

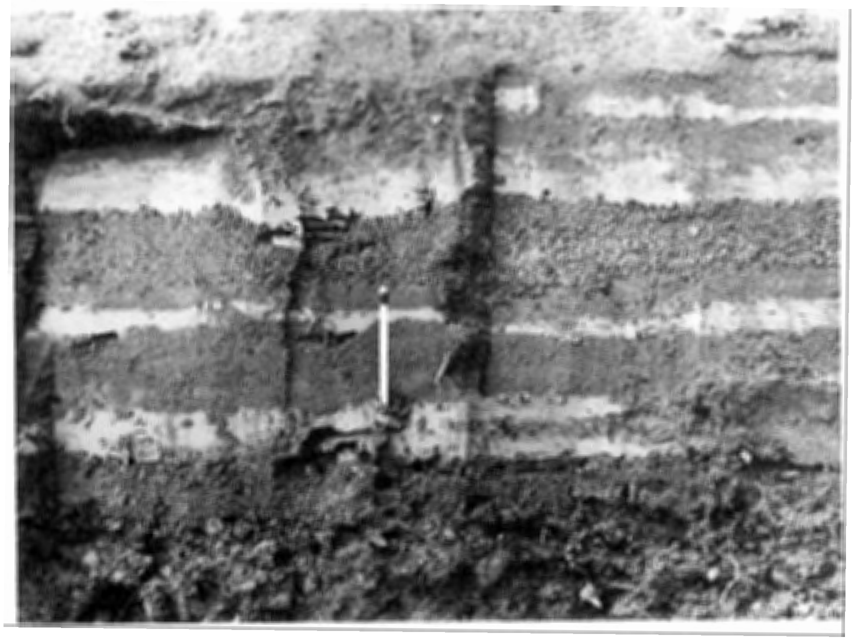


FIGURE 10.25. Graded sand bed, above match, in bottomset facies. Graded sand much coarser than surrounding sand.

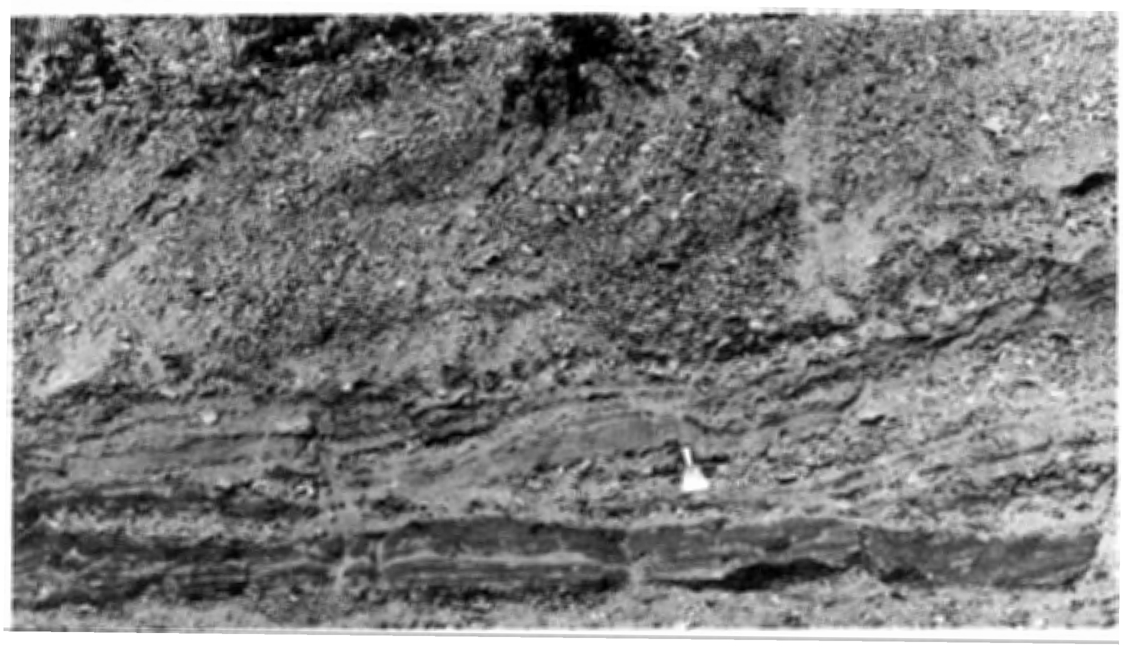


FIGURE 10.26. Slump exposing bottomset/foreset contact. Foreset beds have tangential contact with bottomset facies. Bottomset sand extends several metres into foreset facies.

as the foresets prograded over the bottomsets. This is consistent with the previous interpretation of instability in the delta front area.

Five clay/sand couplets were present in the transition interval. The clay beds were thicker than usual, averaging 3.3 cm, with a standard deviation of 1.5 cm. One bed 6 cm thick had 2 cm of pure clay overlain by 4 cm of gravelly (Mo ≃ 2 cm) clay. However, the gravel was discontinuous as it pinched out several metres down dip (SE) and was not traceable up dip because of a fault. Another bed (1.5 cm thick) also consisted of red gravel, modal diameter (Mo) 1.5 cm but with pebbles up to 4 cm, with a sandy clay matrix. The bed thickened to 8 cm up dip and became one of the 'red beds' described in the lower foresets. Down dip the gravel pinched out while the clay continued as a clay bed. Two of the clay beds had clean clay at the bottom that became silty towards the top and the last clay bed was an interlaminated silty clay and fine grained sand.

The 5 sand beds were also thicker than usual (6.7 cm average) but the average was significantly increased by one very thick (13 cm) bed that also created a large standard deviation (3.3 cm). This latter bed was inversely graded from very fine grained (1 cm thick) to medium grained (10 cm thick) and was capped by 2 cm of gravel graded from 0.5 cm (Mo) to very coarse grained sand. The graded gravel was thickest (7 cm) up dip, pinched out several metres down dip of the section and was overlain by an almost continuous layer of pebbles, Mo ≃ 2 cm. The medium grained sand had scattered pebbles to 2.5 cm and the top 8 cm was very clean (no matrix) and laminated (1.2 mm thick) parallel to the bedding. Four out of the 5 sand beds were inversely graded but only 1 of the beds had a fine grained cap. All of the beds thinned down dip (SE) and thickened up dip.

Interpretation

The most striking and characteristic feature of the bottomset beds is their rhythmic nature, to which Swift and Borns (1967, p. 696) alluded. Any process oriented interpretation of the rhythmites must explain two features: 1. the coarsening upward sequence of a) massive clay with few silt interbeds at the top overlain by b) clay/silt couplets with the silt thickening and coarsening upwards into c) clay/sand couplets with the sand coarsening and thickening upwards to the bottomset/foreset contact; and 2. the relatively constant thickness of the clay beds despite variation in the thickness of the coarser beds. Thus, the process was both persistent and regular.

Edwards (1978, p. 422) states that "Glaciomarine deposits lack rhythms." Three reasons are given for this: 1. glacialmeltwater is less dense that sea water, which results in overflow and widespread dispersal of suspended sediment; 2. clays flocculate in salt water; and 3. seasonal influences may be reduced by currents and sediment sources in addition to streams.

Edwards (1978, p. 422) uses Gadd's (1971) data on Late Wisconsinan sediments in the St. Lawrence Lowland to suggest that glacial varves die out as they pass upward into marine laminated muds. However, this is not clear from Gadd's (1971) work. Gadd (1971, p. 48) states that at one locality 1.5 ft (0.5 m) of unfossiliferous varved silts grade upward into 2.5 ft (0.75 m) of similarly banded sandy silts, containing marine foraminifera, that are overlain by massive marine silty clay. However, at other localities the thickness of the marine rhythmites is signifcant and judging from the photographs (Gadd, 1971, p. 65), they could be as thick as 5 m or more.

Gadd (1971, p. 64) refers to the marine rhythmites as the stratified facies of the Champlain Sea Clay. The Champlain Sea existed from approximately 12,500 to 9,500 yr BP (Prest, 1977, p. 11) in the St. Lawrence Valley. Gadd (1971, p. 64) describes the marine rhythmites as follows. The fine and coarse material are regularly alternated, giving the appearance of varves, but there is no graded bedding. Contacts between the layers are sharp. The fine layers are clay or clayey silt, similar to the massive marine clays, while the coarser layers are very well sorted, fine to medium grained sand. In some places, the banding is caused by interbedded clay and silty clay and to these he proposes the name 'pseudo varves.' The similarity between Gadd's (1971) descriptions and the rhythmites of the bottomset facies in the Minas terrace is striking.

One possible explanation for the rhythmites is that they are turbidites. However, the sedimentological characteristics of the rhythmites do not fit Bouma's sequence (Friedman and Sanders, 1978, p. 393). The inverse to normal graded pattern, lack of well defined lamination and absence of cross stratification in the sand and sharp contact between the sand and overlying clay do not fit Bouma's A-D divisions.

The inverse grading in the lower part of the sand beds is more akin to grain flow deposits (Middleton and Hampton, 1973, p. 29) but the base of the sand beds is not erosional. Also the normally graded upper part is not characteristic of grain flows and would therefore have to be deposited by another process. However, there is no consistent sharp contact within the sand beds.

Grain flows or turbidity currents do not explain satisfactorily the sedimentological characteristics of the bottomsets and they also fail to explain the rhythmic nature of the couplets. Grain flows or turbidity currents are spasmodic events and the relative uniformity of thickness of the clay beds suggests a regularity in timing.

One of the simplest ways to explain the rhythmic nature of the bottomsets is that they are seasonal in nature. The coarse layers are deposited during summer months of high meltwater and sediment discharge and the fine layers represent winter months of low or no meltwater discharge. Thus, they would be varves, couplets that were deposited in one year (Ashley, 1974, p. 304).

Glaciolacustrine varves are the result of seasonal changes (Embleton and King, 1975, p. 551). During the summer months, glacial meltwater carries sediment into the lake, and silt or sand is deposited on the lake bottom, probably by underflows (Gustavson, 1975, p. 261). During the winter months, the meltwater no longer reaches the lake or is much reduced and the finer lake sediment settles out of suspension. The fine layer (generally clay sized) caps the coarse summer layer to form a fining upwards couplet and the clay layer itself is graded because the finest sediment takes longer to settle out of the water (Ashley, 1975, p. 310).

Gadd (1971, p. 64) does not refer to the marine rhythmites as varves, even though the banding is similar to that of the underlying freshwater varves (p. 48). He calls the glaciomarine rhythmites 'pseudo varves' where the grain size is similar to that of freshwater varves. His discussion implies that the reasons he does not call the marine rhythmites varves are partly sedimentological and partly due to the fact that they are marine. However, the term varve actually has no genetic implications but simply means an annual layer and glaciomarine varves would be expected to have different characteristics from glaciolacustrine varves. The bottomset rhythmites at Spencers Island and at the other deltas in the terrace are interpreted as varves because the depositional process proposed for their formation is seasonal in nature and one couplet is deposited per year. Glaciomarine varves formed under the proposed mechanism differ in many aspects from the classical glaciolacustrine varves, but they are still varves.

Beneath the varves, the massive clay with a few scattered silt layers in the upper part is typical of what most authors (Embleton and King, 1975, p. 551; Edwards, 1978, p. 422) predict for glaciomarine situations, because of the flocculation of clay in salt water. In the rhythmites above this, the silt/sand layers coarsen and thicken upwards with the encroachment of the foresets until the foresets prograde over the bottomsets. Thus, the quantity and grain size of the sediment supplied to the bottomsets increases as the delta front approaches. The deposition of the summer silt/sand beds is adequately explained by the process derived for the deposition of the foreset beds.

Deposition of the Coarse Layer

The model for the deposition of the coarser summer beds is as follows. In the summer season, high meltwater and sediment discharge were supplied to the delta and as the meltwater streams reached the delta front, coarse sediment (gravel) avalanched down the foreset slope. Almost all of the deposits on the foreset slope, including the upper part, showed no evidence of current activity. Therefore stream discharge continued out into the water as an overflow, carrying the suspended sediment with it in a surface plume. Some of the sand settled out on the foreset slope, especially at low flow, but at high flow, some of the sand (finer than coarse grained) was carried beyond the foresets and settled out onto the pro-delta bottomset beds. For any given discharge, the distance away from the delta front to which sediment coarser than clay was carried was inversely porportional to grain size. Silts were carried farther out into the sea than sand, and fine sand was carried farther out than coarse sand. As the delta prograded, progressively coarser sediment was supplied to any given area until the foreset beds prograded over that area.

As the rhythmites first became recognizable with coarse, sandy silt, the extent of the 'suspension fall out' zone was the distance to which coarse silt to very fine grained sand was transported. The quantity, and thus thickness, of sediment supplied to any given area within this zone would also increase closer to the delta front. An area being supplied with sand from suspension would accumulate a greater thickness of sediment in one meltwater season than a more distant area being supplied with silt for several seasons. The larger diameter of sand grains means that for a given number of grains, a sand bed is thicker than a silt bed. Because of three dimensional mixing away from the delta, the area being supplied with silt is much larger than the area being supplied with sand. As discharge varies, an area in the 'sand' zone would receive coarse silt at low discharge, while an area formerly in the 'silt' zone would be removed from the 'suspension fall out' zone and receive less coarse sediment. Thus as the delta front prograded, any given area in the prodelta region would progressively receive a greater amount of, as well as coarser, sediment from suspension. This explains the correlation between sand thickness and grain size. Beyond the 'suspension fall out' zone, the deposition of silt and clay sized sediment would be uninterrupted by coarser sediment so that massive silty clays would accumulate.

The massive to faintly parallel laminated structure of the sands is consistent with deposition from suspension. The absence of cross stratification or well developed parallel lamination in all grain sizes from coarse silt to coarse sand indicates that bottom currents were weak to nonexistent. Variations in discharge and/or differences in turbulence in the water mass probably caused the thin, discontinuous laminae of slightly different grain sizes. More marked changes in grain size were caused by fluctuations in discharge at the delta front and to the switching of streams from one part of the delta to another, something that was suggested in previous discussions of the topset facies and of the topset/foreset contact. Other authors (Gustavson *et al.*, 1975, p. 279; Ashley, 1975, p. 318) have also proposed the switching of meltwater discharge from one part of a delta to another as a mechanism to explain prodelta grain size changes for a given area.

The common inverse/normal grading pattern must be related to the systematic way that discharge changed during the melting season. The very fine grained sediment at the base of the sand beds implies that early in the melting season the discharge was relatively low. The sediment coarsens upwards and reaches a maximum grain size 1/2 to 3/4 of the way up the bed, when meltwater discharge reached a maximum later in the season. The thin cap of finer sediment at the top of the bed indicates that meltwater discharge decreased late in the season, but this took place fairly rapidly. The absence of a fine grained cap on many beds means that meltwater discharge was cut off abruptly, probably due to a rapid seasonal change. Variations in this grain size pattern could result from changes in a particular melting season or from the shifting of streams. For instance, a sand bed that was normally graded could be the result of streams that gradually shifted away from that area during the summer season.

The presence of medium grained sand in many of the beds in the upper part of the section means that at high discharge the overflow was strong enough to transport medium grained sand beyond the toe of the foresets. The settling velocity of a quartz sphere 0.5 mm in diameter (upper limit of medium grained) in water at 0°C is about 5.7 cm/sec (Blatt *et al.*, 1972, p. 54). It would take the quartz sphere 263 seconds to settle 15 m, roughly the height from the bottomset beds to the topset/foreset contact, and the toe of the foreset beds extends about 26 m beyond the top of the foresets (30° slope). Accordingly, a minimum seaward current of 9.9 cm/sec is required to transport the 0.5 mm sphere beyond the foresets during its theoretical settling time (ignoring turbulence).

Coarse grained sand only occurs in the uppermost beds and thus is not transported far beyond the toe of the foresets. Similar calculations for 1.0 mm sand show that seaward velocities of about 20.8 cm/sec are required to transport the sand to the base of the foresets. Therefore, the average seaward rate of transport of the sand in suspension generally did not exceed 21 cm/sec but did attain average velocities in the range of 10 to 21 cm/sec over the length of the foreset slope.

These calculations are very rough and do not take into account several factors such as turbulence, the non-sphericity of the sand grains and the greater density of sea water. However, they are reasonable approximations and it is interesting to note that Gustavson (1975a, p. 256) measured underflow velocities of 18 cm/sec on the bottom of a proglacial lake.

The three very thick, coarse sand beds (bed nos. 44-46, Fig. 10.14) were probably caused by an extremely warm season, and therefore very high discharge, and/or a direct location in front of the outpouring streams. Locations equidistant from the delta front but away from the area of stream discharge would receive less sediment (and finer) as if they were more seaward.

The proposed mechanism of stream overflow (hypopycnal inflow; Bates, 1953, p. 2125) for the deposition of the foreset and bottomset beds during the melting season is based on the sedimentary structures and textures of the beds. But it is also the favoured process when the densities of the inflowing stream water and sea water are considered. The density of sea water at 15°C is 1.025 (CRC Handbook, p. F-3), so at temperatures between 4°C and 15°C, it would be slightly more dense. Gustavson (1975a, p. 261) observed overflows, interflows and underflows in Malaspina Lake and concluded that the three types of flow were dependent upon the relative suspended sediment content of the lake water and of the inflowing meltwater. The suspended sediment content of the major meltwater streams near Malaspina Lake ranged from 0.135 grams per litre to 4.7 grams per litre (Gustavson, 1975a, p. 256). The suspended sediment content of two underflows were 1.55 and 1.86 grams per litre while the suspended sediment content of the Yana outwash stream had a high of about 3.0 grams per litre (Boothroyd and Ashley, 1975, p. 198). If water temperature is taken as 4°C in order to give the maximum density (1.0), a suspended sediment concentration of 4.7 grams per litre raises the density of fresh water to only 1.0047. A suspended sediment concentration of more than 25 grams per litre is required to make inflowing stream water more dense than sea water. As Kuenen (1951, p. 75) suggested, it appears improbable that meltwater would be heavier than sea water.

Coarse Graded and Ungraded Sands

The coarse graded sands (Figs. 10.20, 10.25) in the bottomset beds and graded gravel in the bottomset/foreset transition zone present a special problem because such abrupt, laterally persistent changes in grain size are infrequent. Especially significant is the fact that 2 of the beds in the 1976 section (bed nos. 6 and 9, Fig. 10.22) occurred low in the bottomset facies but were coarse grained. The sharp, possibly erosional lower contact, normal grading, low matrix content (all but one), absence of 'wispy' laminae, anomalously coarse grain size and poor correlation of grain size and bed thickness (characteristic of the suspension fall out deposits) suggest a different depositional process.

The three coarse ungraded layers described in the 1976 section had the same characteristics as the graded beds but were coarser (coarse grained to granule) and lacked grading. They were probably deposited by the same process as the graded beds, but may represent more proximal equivalents, and thus are included in the interpretation of the graded beds.

Two possible explanations for the graded beds are deposition from a strong tractive bottom current or deposition from a sediment gravity flow (Middleton and Hampton, 1973, p. 1) of which a turbidity current seems the most plausible. Both processes could produce a sharp, erosional basal contact and deposit coarse sand with little matrix. The erosional contact could at least partly explain the anomalous thinness of some of the sand beds, considering their grain size, by erosion of some or all of the preexisting sand that was deposited from suspension.

The lack of well defined parallel lamination and absence of cross stratification is incompatible with deposition from traction. The grain size of sample 74-254 (less than 0.6 mm) is such that ripples would form in the lower flow regime and plane beds would form in the upper flow regime (Blatt *et al.*, 1972, p. 121). Coarse sands of the 'suspension fall out' zone contain thin laminae of slightly different grain sizes but these beds have uninterrupted grading. The grading of the sands is similar to Bouma's 'A' division and the lack of matrix indicates rapid deposition, which is consistent with a turbidite origin.

If the beds are 'turbidites,' only the A division is present. The overlying very fine to fine grained sand may be the equivalent of the D to E divisions because deposition occurred during the summer season when much of the sediment in suspension was silt and sand. The small size of the turbidity currents, inferred from the thickness and extent of the beds, may have inhibited or precluded the deposition of the B and C divisions (parallel and cross stratified sediment) which are deposited by the body and tail of the turbidity current.

The similarity of the grain size distributions of the graded bed in figure 10.20 and the coarse sand in the 'suspension fall out' sands (Fig. 10.17) is striking (Fig. 10.27). The grain size curves imply similar depositional processes, but the structure of the sands suggests otherwise. An examination of the source for a 'turbidite' sheds some light on this problem.

If the graded beds are turbidites, initiation of the event would probably be a slump on the foreset slope. The sand on the slope would be put into suspension to form the density current and be resedimented on the prodelta slope. Thus, the source of the turbidites would be sand that was initially deposited from suspension and therefore their textures should be similar. Furthermore, they are both deposited from suspension, although one is much more rapid than the other. As the source beds lack silt and clay, these would also be absent in the 'turbidite' deposit. The graded bed that occurred directly above the clay layer had an unusually high concentration of clay matrix. This is consistent with a turbidity current that either occurred at the start of the melting season or eroded the existing sand down to the clay layer. The turbidity current would stir up some of the underlying clay and incorporate it as matrix.

The graded beds were rare as compared to the other sand beds, which is consistent with the spasmodic initiation of turbidity currents. The turbidity currents were probably caused by large infrequent slides on the foreset slope. The large slides could be the result of overdeposition

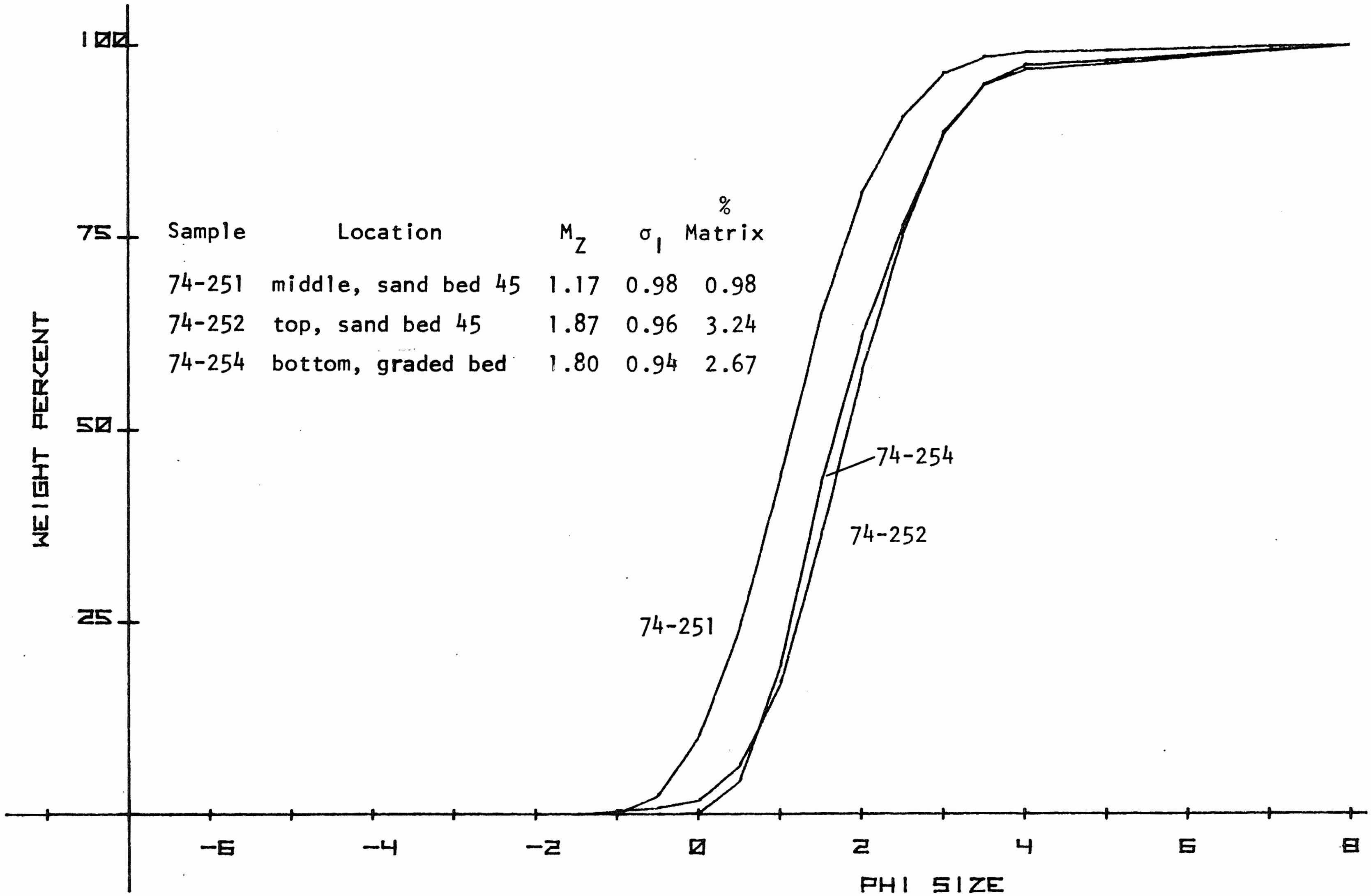


FIGURE 10.27. Grain size analyses, bottomset facies.

on the upper foreset slope or slumping of the foresets due to failure within the foresets or failure of the bottomset beds.

As turbidity currents have been proposed as the depositional mechanism for the antidune(?) cross stratification in the lower foresets, the possibility of turbidity currents on the prodelta slope is consistent with what has been previously described and interpreted. If the 'antidune' was deposited by a turbidity current, the current must have continued out across the bottomset beds.

Large slides on the upper foreset slope have been suggested as the cause for the turbidity currents. In the case of the 'antidune,' the slide may have initiated a grain flow and perhaps the grain flow helped generate the turbidity current, as was proposed in the discussion of the lower foresets. Thus, a common process, sediment gravity flows caused by large avalanches, can explain the graded beds in the bottomsets and the antidune(?) and imbricated gravel in the lower foresets. The relationships between the observed phenomena in the different parts of the delta is tenuous and can not be taken as fact. But it is worthy of consideration as it links probable processes on the foreset slope to possible processes on the prodelta slope and in so doing explains several problematical features.

Although the graded sands are largely an unresolved problem, a turbidity current is the favoured process. The similarity in grain size curves of sands apparently deposited from different mechanisms is thought to be an example of how misleading grain size curves can be. In commenting on the use of grain size curves in fluvial sedimentology, Miall

(1977, p. 55) states "curve analysis still does not appear to provide unambiguous environmental interpretations."

Deposition of the Fine Layer

While a few of the sand beds are controversial in origin, the bulk of the sand beds are consistent with a process of gravel avalanching down the foreset slope and an overflow carrying some silt and sand in suspension out into the sea. The clay beds in the rhythmites pose a more interesting problem. The persistence of the clay/sand couplets has been attributed to seasonal changes with the coarse silt/sand being deposited in the summer season. The fine silt and clay would also be introduced into the water mass during the summer, so the separation into sand and clay beds within the 'suspension fall out' zone must be accounted for. There appear to be two probable mechanisms. Deposition of fines is relatively constant but the deposition of coarse silt/sand is so high in the summer that the fines are diluted. Alternatively, the deposition of fines is somehow inhibited in the summer and unimpeded or enhanced in the winter.

If deposition of the fines is constant, then the amount of matrix in the sand beds would indicate the relative lengths of time over which the two have been deposited. Samples 74-250, 74-251 and 74-252 (Fig. 10.18) were taken from the bottom, middle and top of a sand bed 11.5 cm thick and they contained approximately 15 percent, 1 percent and 3 percent, respectively, of sediment finer than sand. The weighted average of matrix over the entire bed is about 5 percent, which means that the sand contains the equivalent of a clay bed roughly 0.6 cm thick. The thickness of the overlying clay bed was roughly 2 cm thick; therefore, the deposition of the sand bed represents between 1/3 and 1/4 of the total amount of time required for the rhythmite to form. Similar calculations for 2 other rhythmites (nos. 34 and 35, Fig. 10.14) show that the deposition of the sand bed represents only between 1/5 and 1/8 of the total amount of time represented by the rhythmite.

The calculations have been expressed in terms of proportions instead of months because the clay may come out of suspension very quickly in the winter season, because of flocculation, and thereby create a hiatus in the late winter. Even if deposition were spread over the entire year, the summer season would be only between 1 and 3 months long. This seems to be unreasonably short especially as deglaciation is occurring. With summer/winter season in the preceeding proportions, glacial growth or at least preservation would be expected. Thus it appears that although fines were deposited in the 'suspension fall out' zone in the summer, deposition was inhibited and the rate was increased in the winter.

Whitehouse *et al.*, (1960) have shown that the flocculation of illite, chlorite and kaolinite, which produces settling rates of 10 to 15 m/day (.011 to .018 cm/sec), occurs at chlorinites as low as 1 $^{o}/oo$. As molds and casts of marine pelecypods (discussed later in the chapter) occur in the bottomset beds, it is unlikely that salinities were lowered at any time to levels where physicochemical flocculation would not occur. Drake (1976) discusses flocculation and stresses the importance of biological production of aggregates and fecal pellets on clay deposition, especially in estuaries. Little is known about the biological productivity in the Minas Basin in the Late Pleistocene, but if biological mechanisms were important in forming composite particles, they would have been at least as active, if not more active, in the summer as in the winter.

Drake (1976, p. 139) also notes that physicochemical flocculation is not effective below concentrations of 200 mg/l. The amount of clay deposited in each clay bed (≈2.5 cm) suggests that the concentrations in the prodelta area were at least this high. Assuming that only 1 cm of the 2.5 m clay beds was present in a suspended state in a 15 m water column at the start of the winter season, the clay concentration (assuming a density of 2.0 for the clay and 50% clay content) would be .0007 grams per litre. This is over 3 times the minimum concentration (.0002 gm/l) required for effective physicochemical flocculation.

Ashley (1975, p. 310) found that the clay layers in freshwater rhythmites showed a constant decrease in mean grain size (of about 2ϕ, p. 308) from bottom to top and concluded that flocculation was not significant in clay sedimentation. The marine clay beds at Spencers Island are very similar in grain size from bottom to top (Table III). There is a very slight tendency for the clay fractions of the beds to fine up, but it is too small to be significant (≈0.10ϕ). Therefore the lack of grain size change provides evidence that there was flocculation (the smallest particles were not settling out last). As Drake (1976, p. 146) suggests, "flocculation by purely physicochemical means may be significant where rivers are directly tributary to the coast." Thus, flocculation probably did occur and this must be incorporated into any theory on fluctuating rates of clay deposition. Some insight on this problem is gained by examining the process already proposed for the deposition of coarse silt/sand in the summer.

As the streams flowed out from the delta and transported silt and sand in suspension, clay would also have been carried along in the freshwater current (overflow). Farther out, as the current slowed and mixing took place in three dimensions, the clay would come into contact with the salt water and flocculate but the remanent currents would still continue to carry the composite particles seaward. Where the currents and turbulence were low, the flocculated particles would settle out on the prodelta slope. Thus the strong freshwater discharge would prevent flocculation from occurring right at the mouths of the streams and the overflow would also transport the flocculated clay seaward. Even after the clay had flocculated, the settling velocities of the composite particles (.01 to .02 cm/sec, Whitehouse *et al.*, 1960, p. 48) is an order of magnitude lower than the settling velocity of very fine grained sand (0.18 cm/sec, Blatt *et al.*, 1972, p. 54) and is approximately equal to that of fine to medium grained silt. Thus, the particles would be transported beyond the sand because of their low settling velocity. The silt/clay rhythmites were only recognizable where the silts were coarse, so where the finer silts were being deposited, there was no differentiation into rhythmites. Fine silt and clay were deposited together as a massive unit, which agrees with the settling velocities which show that the flocculated clay probably behaved like fine silt. The 'suspension fall out' zone was previously defined as the distance to which coarse silt to very fine grained sand was transported in suspension.

Within the 'suspension fall out' zone, the clay along the bottom of the freshwater overflow would come into contact with the salt water and flocculate. Currents and turbulence permitting, the clay would settle out in that zone. Also, some of the flocculated particles farther seaward might be transported back into this zone by a return flow of water beneath the outflow. Some of the flocculated particles might be broken by the turbulence and/or sand grains and remain in suspension for longer periods of time. The greater percentage of matrix in the finer lower and upper portions of the thick sand beds is consistent with the overall process. The coarser sand in the middle is indicative of higher discharge and therefore stronger overflow of fresh water and higher rates of sedimentation. The stronger overflow carries the clay farther seaward and the higher rate of sedimentation dilutes whatever clay is deposited. Therefore, the coarse sands have less matrix than the finer sands.

The relatively thin cap of fine grained sediment on most sand beds, or the lack of such a cap, means that the transition from summer to winter was fairly rapid. In the winter, the supply of meltwater and sediment probably stopped. Less turbulence and no outflow of water in the 'suspension fall out' zone would allow an even distribution (in the pro-delta area) of the fines in suspension and the fines then settled out of suspension to form the clay bed above the summer sand. Some silt and clay may have been derived during the winter by marine processes, such as shoreline erosion of tills or clayey sediments, but this form of input was probably small. The bulk of the clay was probably introduced into the basin by streams during the summer. The presence of silt in the clay also indicates that current activity may have been sufficient to suspend silt for much of the year, an observation and conclusion also made by Ashley (1975, p. 316). However, some of this silt was probably derived from the wave reworking of coarser sediments, such as the topset and foreset beds, on the shoreline. The low clay content of these beds precludes the supply of clay in this manner. There is no evidence of major reworking of the topset and foreset beds by waves.

Many clay beds contained a lamina of coarse silt to very fine grained sand in the middle to upper part of the bed. This could be the result of a warm spell during the winter temporarily reestablishing meltwater and sediment discharge. It could also be the result of a slump putting sediment into suspension. Both possibilities are also sources for the fine silt scattered throughout the clay beds. The position of the laminae (there were never any laminae in the lower centimetre of the clay beds) means that silts were never deposited in the early part of the winter or that clay deposition was very rapid at the end of the summer season. The latter concept is consistent with the expected time of highest clay concentration in suspension and therefore seems more probable.

The lenses and stringers of sand in the upper part of the clay beds were probably due to burrowing. There may have been two types of burrows, one predominantly horizontal and the other with a greater vertical component, but not as steep as 90°. It is not known what organism or organisms did the burrowing, but they always burrowed in and not through the bottom of the clay bed. The burrows were an important part of the beds as they disrupted the clay/sand contact. Because the sand/clay contact was sharp, the distinct break in the rhythmites was in the 'fall' and not the 'spring' position. The sharp contact between the sand and the overlying clay is not limited to marine rhythmites as Ashley (1975, p. 307) noted sharp silt/clay contacts in bottomset beds from Glacial Lake Hitchcock. While the clay beds themselves (at Spencers Island) did not coarsen upward, the introduction of the very fine sand into the upper part of the clay bed produced a coarsening upward sequence from the base of the clay beds (fall) to the top of the overlying sands (fall).

The isolated clasts of very coarse sand to pebbles that occurred in both the sand and clay beds are interpreted as ice rafted dropstones. The dropstones were particularly abundant in the clay and at the top of the sand. The 'unstable' position of many of the pebbles means that the pebbles were dropped after the clay was deposited and the pebbles wedged into the clay. Many of these pebbles abut the underlying sand and the observed concentration of clasts at the top of the sand bed may have been partly caused by pebbles dropping through the clay. However, some of the pebbles at the top of the sand were lying within the sand or in a stable position on the sand. It is puzzling as to why ice rafting was more common in the late fall to early winter than at other times. In any case, the dropstones indicate that the basin was not frozen over in the winter as blocks of ice were free to float, which presents another possible source for the scattered silt and sand grains within the clay beds.

The gravel beds in the lower foresets with the red silty clay matrix (red beds) graded into clay beds in the bottomset facies and are therefore the equivalent of the winter clay layer. The up dip pinch out of the beds means that clay deposition did not persist to the top of the foresets. During deposition of the fines on the lower foresets, some pebbles rolled or slid down the slope to become part of the winter deposit, but they did not travel much beyond the toe of the slope. Some of the gravel may have been ice rafted. The concentration of silty clay matrix at the bottom of the beds agrees with the previous suggestion that clay deposition was more rapid in the early winter. The thickness of the beds (5 to 10 cm) and lack of sand means that there was very little supply of sediment during the winter and the distance between the red beds, from 0.3 to 1.2 m, represents the amount of delta progradation in that area for one year.

Extent of the 'Suspension Fall Out' Zone

As there were 60 rhythmites between the massive clay and the foreset beds (exluding the disrupted interval), the delta prograded over the 'suspension fall out' zone in about 60 years. The width of the 'suspension fall out' zone is not known but the bedding thickness between the red gravels (winter) in the lower foresets can be used to derive a rough estimate. The bedding thickness varied from 0.3 to 1.2 m which equals yearly horizontal progradations (at a 30° slope) of 0.6 to 2.4 m. This gives a range of 36 to 144 m for the width of the 'suspension fall out' zone over the bottomset beds. This is a very simplistic calculation which assumes a constant angular contact at the base of the foresets and does not take into account greater variations in yearly progradation. It is difficult to judge how realistic this estimate is.

POSTGLACIAL EMERGENCE

The elevation of the topset/foreset contact at the small exposure near Highway 209 is 32 m above mean sea level. The topset/foreset contact at the 1974 slump (a more distal position in the delta) is 26 m above mean sea level but the elevation is thought to be 1 to 2 m low as the edge of the bank seems to have dropped a little.

FOSSILS

Marine bivalve molds were found in the clay/silt interval of the bottomset beds in the fall of 1974. The carbonate shell material had been leached but the chitinous periostracum remained. Some of the molds were given to Dr. Frances J.E. Wagner at the Atlantic Geoscience Centre, Bedford Institute of Oceanography, Dartmouth, Nova Scotia for identification. The most common mold was *Portlandia arctica* (Fig. 10.28) and height:length ratios indicated the presence of 2 subspecies *P. arctica arctica* and *P. arctica siliqua*. Identification of one specimen of *P. arctica portlandia* was doubtful because of poor preservation. Two other specimens were tentatively identified as *Nuculana permula* and *Macoma* sp. One larger specimen was tentatively identified as *Mya* sp. (Fig. 10.29).

The excellent conditions of the molds indicated that they were living in situ where they were collected. The assemblage suggests colder water than at present and an approximate water depth of 35 m (Wagner, 1975). If *P. arctica portlandia* is present, Dr. Wagner suggests that it would have been transported from shallower, less saline (ca. 28 $^{o}/oo$) water. The optimum habitat of the assemblage is "off the mouths of rivers carrying meltwater into the sea, or off glacier fronts, areas where large quantities of silt and clay are being deposited" (Wagner, 1977).

Three samples of the clay were processed for microfossils but only a few poorly preserved foraminifera (*Buccella, Cribrononion*) and some derived Paleozoic palynomorphs were found.

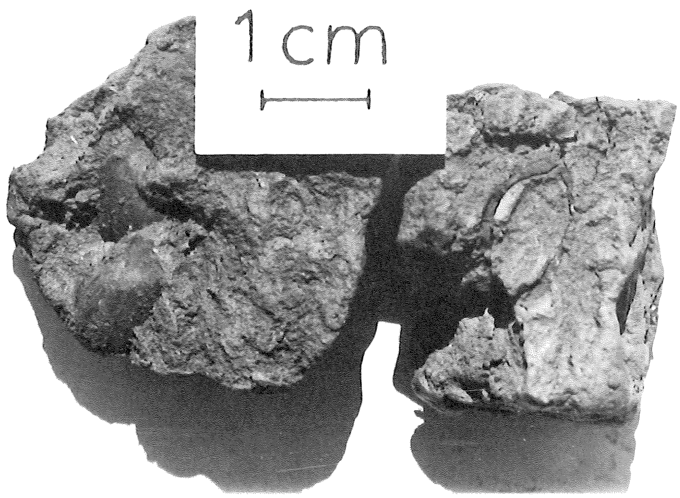


FIGURE 10.28. Molds and casts of *Portlandia arctica*. Internal molds of 2 disarticulated valves on left. Cross section, transverse to the plane of symmetry, of cast on right.

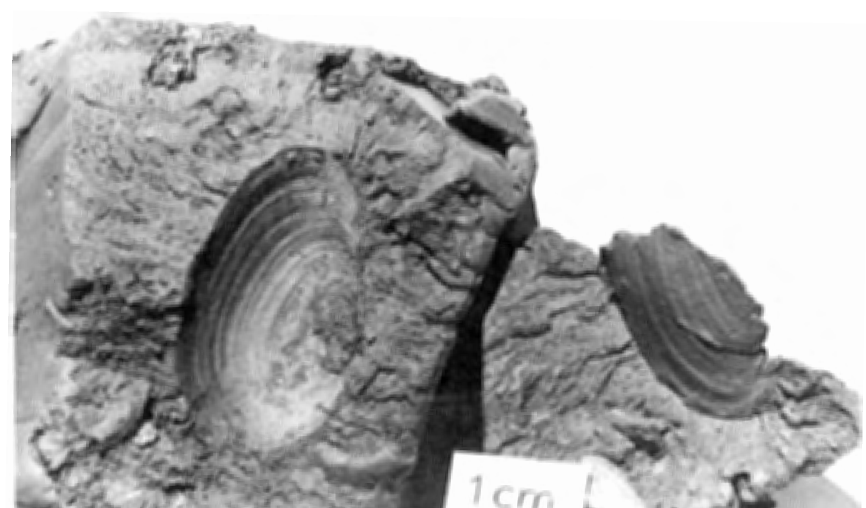


FIGURE 10.29. External (left) and internal mold of *Mya* sp. valve.

OVERVIEW

The direction of dip (from east to south) of the exposed foreset beds and tabular cross beds in the topset facies and the slope of the upper surface indicate that the main supply of meltwater that deposited the Spencers Island delta discharged out of Mahoney Brook. Minor meltwater and sediment discharge probably flowed out of the two smaller brooks to the south.

In the summer season, rapidly flowing (velocities in the 1 to 3 m/sec range) braided streams transported sediment across the delta. At the delta front, gravel and coarse sand avalanched down the foreset slope while silt and finer sand were carried out into the bay by a freshwater overflow. Coarse silt and sand settled out on the foreset slope and on the prodelta slope beyond the toe of the foresets. The distance of seaward transport was inversely proportional to grain size and the distance to which coarse silt was transported defined a suspension fall out zone. Within that zone, the grain size of the suspended sediment increased to about medium grained sand at the toe of the foresets. The grain size of the sediment supplied to any given area within the suspension fall out zone was dependent upon stream discharge and the location of the streams. Stream discharge was relatively low in the spring and reached a maximum in the mid to late summer before falling off sharply in the fall. This produced an inversely graded bed, usually very fine grained at the base, that coarsened up to fine to medium grained. A thin, finer grained sand frequently capped the coarser sand. Areas away from direct stream discharge received finer sediment than areas equidistant from the delta front but in line with direct stream discharge. Ice rafted clasts were deposited throughout the summer but reached a maximum concentration in the late fall. Seaward of the suspension fall out zone, fine silt and flocculated clay were deposited simultaneously.

Sand deposition from suspension was occasionally interrupted by coarse graded to massive sands that were emplaced with a sharp erosional contact. The coarse sands are interpreted as small scale turbidities. The turbidites were probably initiated by large slides on the foreset slope and the slides could have been the result of overdeposition on the upper foreset slope or slumps of the foresets.

During the winter season, the supply of meltwater was cut off and the fine silt and clay still in suspension spread over the prodelta area, settled out and blanketed the previous deposits. Beyond the suspension fall out zone, the deposition of fine silt and clay was the same in the winter as in the summer so a massive silty clay unit formed. Within the suspension fall out zone, the fine silt/clay deposition blanketed the coarse silt/sand. The abrupt termination of the supply of meltwater and rapid deposition of the fines produced a sharp sand/clay contact. During the mid to latter stages of fine silt/clay deposition, coarse silt or sand was frequently reintroduced into the system and formed a silt/sand layer in the clay bed. The source of the silt could have been a slump of the foreset beds or a warm spell during the winter that reestablished the flow of meltwater. Ice rafting of sediments continued as the bay did not freeze over and clasts were implanted in the clay (Fig. 10.30).

The process was repeated each year and the division into 'sand' and 'clay' beds in the suspension fall out zone produced varves. Organisms

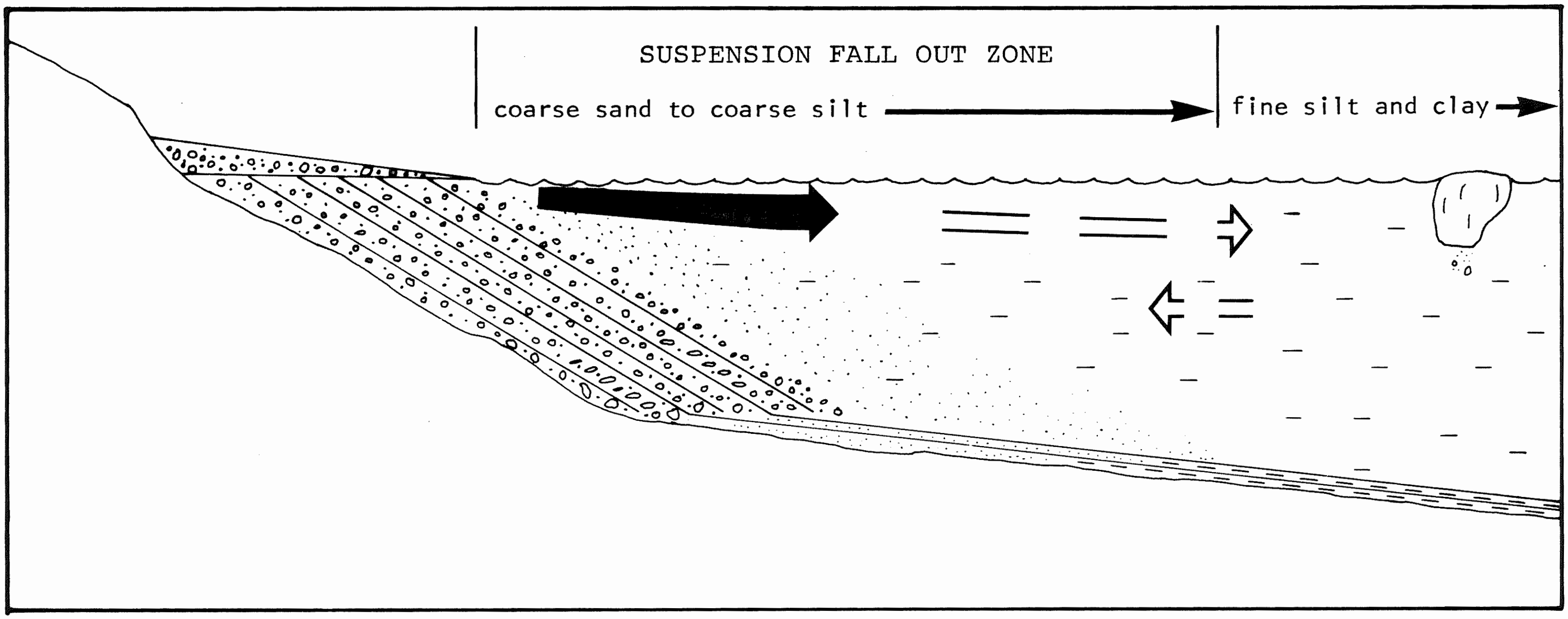


FIGURE 10.30. Schematic diagram of deposition at delta front during melting season. Bedload sediment avalanches down foreset slope, suspended load sediment settles out on foreset and bottomset slopes. Clay in freshwater overflow flocculates along bottom of overflow and at distal end of suspension fall out zone as overflow diffuses. Weak interflow may transport some flocculated clay landward. In winter, suspended silt and clay settle out and blanket summer deposits, forming varves in the bottomsets within the suspension fall out zone and lower foresets. Massive silty clay deposited seaward of varves. Ice rafted sediment deposited year round.

burrowed from the sand into the underlying clay, disrupting the clay/sand contact and introducing sand into the upper part of the clay. The sharp sand/clay contact and disrupted clay/sand contact, coupled with the coarsening up of the sand bed, produce coarsening up cycles from the base of the clay beds, or from fall to fall. This is contrary to classical freshwater varves that fine upwards from the base of the silt/sands, or from spring to spring.

Delta progradation resulted in shifting of the suspension fall out zone seaward. This produced a coarsening upward sequence in the bottomsets of massive silt/clay overlain by silt/clay varves that coarsened upward to sand/clay varves. The summer silt/sands thickened upward as they coarsened upward. Small numbers of marine pelecypods, mainly *Portlandia arctica*, lived in the clay beds of the clay/silt interval. *Portlandia arctica* is typically found in cold marine water where large quantities of silt and clay are being deposited, such as off glacier fronts.

Progradation of the foresets over the bottomsets caused overloading (undercompaction) and failure of some parts of the bottomsets. Types of failure included faulting and zones of liquefaction that underwent flowage. Failure of the bottomsets was probably the cause of delta subsidence and the deposition of foreset beds over the flooded part of the upper surface.

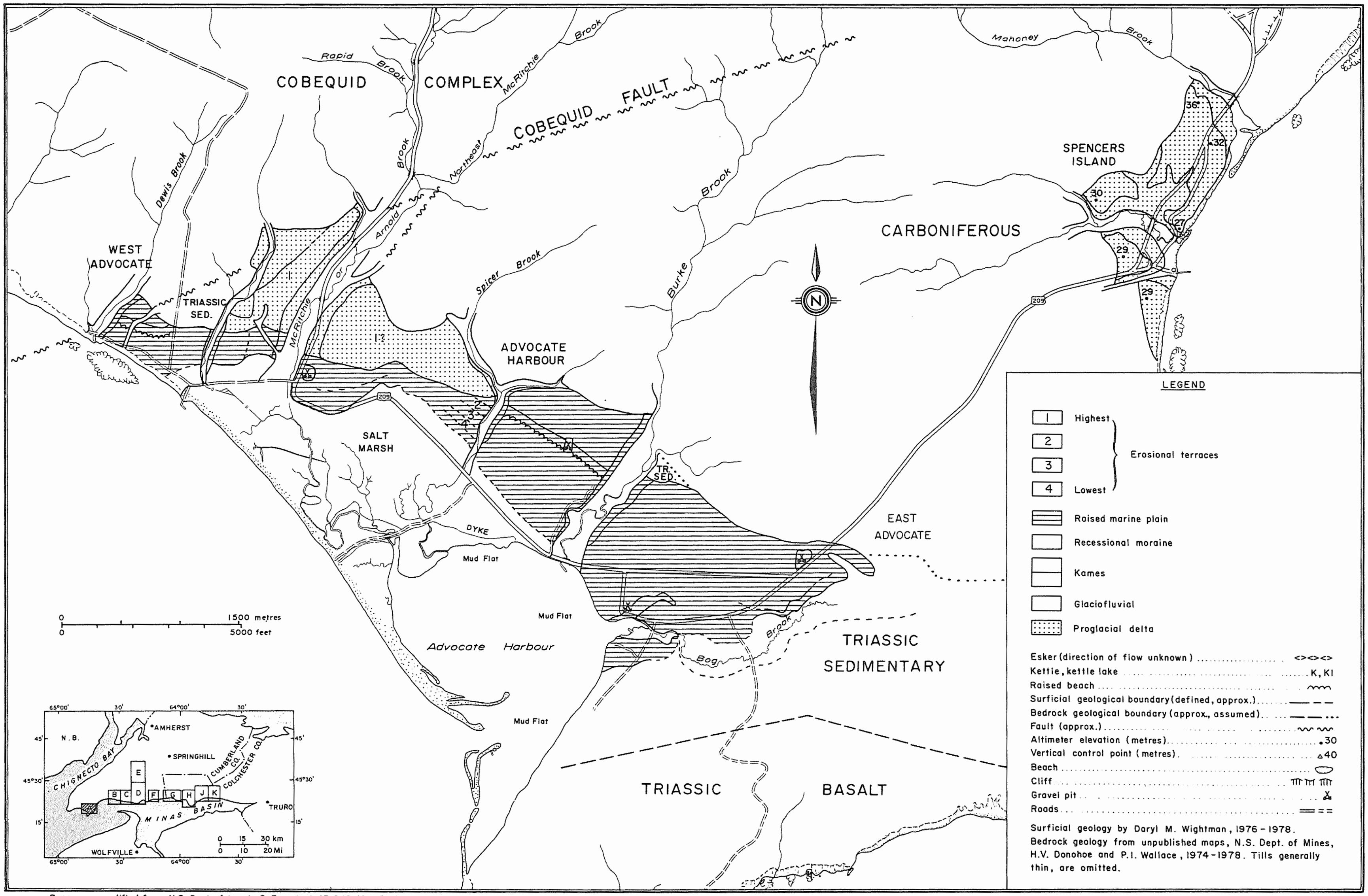
After the delta was deposited, Mahoney Brook and two unnamed brooks to the south eroded valleys through the delta. No terraces were formed, but the thinness of the topset facies implies that the entire upper surface of the delta may have been eroded to a certain extent, although the erosion did not affect the foreset beds appreciably.

CHAPTER 11. ADVOCATE HARBOUR AREA

PHYSIOGRAPHY

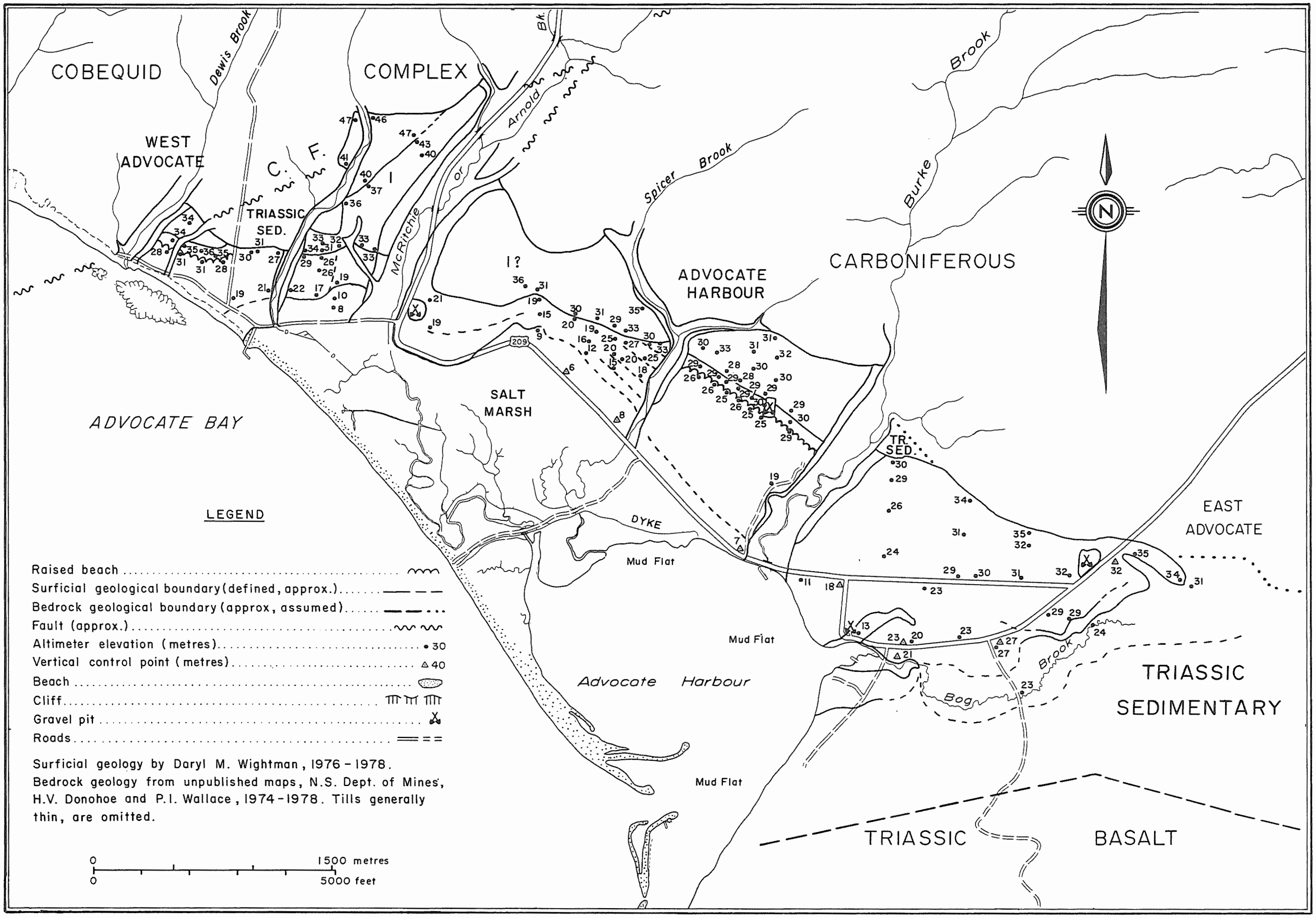
Advocate Harbour lies at the western end of the Chignecto Peninsula and faces towards the Bay of Fundy. The postglacial raised marine deposits occur on the sedimentary lowlands between the Triassic basalt to the south and the Cobequid Highlands to the north. The lowlands are composed of soft Triassic red beds and more resistant Carboniferous sedimentary rocks. The red beds form a low undulating plain and attain a maximum elevation of only 50 m while the Carboniferous rocks have more relief and attain elevations of 150 m. The raised marine deposits occur almost exclusively on the low Triassic sedimentary rocks. South of the lowlands, the Triassic basalt rises sharply to an elevation of 170 m while to the north, elevations increase at the Cobequid Fault from 110 m on the lowlands to 185 m on the highlands.

The postglacial surficial features at Advocate Harbour are composed of two genetic units (Fig. 11.1), with the highest unit being a terrace deposit in West Advocate that reaches an elevation of 47 m (Fig. 11.2). Below the terrace deposit is a raised marine plain (Swift and Borns, 1967, p. 697) that stretches from Dewis Brook in West Advocate through Advocate Harbour to East Advocate. It slopes seaward from elevations of approximately 34 m to the level of the present high tide (6.5 m). A raised beach (Swift and Borns, 1967, p. 698) with an elevation of about 29 m is developed on the plain in Advocate Harbour. Another shorter, higher (≃35 m in elevation) ridge occurs on the plain in West Advocate but it was not mentioned by Swift and Borns (1967).



Base map modified from N.S. Dept. of Lands & Forests 1:15,840 Map Series.

FIGURE 11.1. Map A. Surficial geology of the Advocate Harbour and Spencers Island areas, Cumberland County, Nova Scotia.



Base map modified from N.S. Department of Lands & Forests 1:15,840 Map Series.

FIGURE 11.2. Map A'. Elevation data for the Advocate Harbour area, Cumberland County, Nova Scotia.

TERRACE DEPOSIT

SURFICIAL GEOLOGY

The terrace deposit is located at the mouth of the McRitchie Brook valley in West Advocate. It is most evident on the west side of the brook and is of questionable existence on the east side, where gravel veneered bedrock corresponds in height to an incised terrace on the west side. Accordingly, further discussion will be limited to the deposit on the west side of the brook.

The deposit abuts the Cobequid Fault to the north and is flanked by the Triassic Blomidon-Wolfville Formations (undifferentiated red beds) on the west. The greater height and relief of the bedrock make it fairly easy to delineate the deposit/bedrock contact, although thick woods on the north end makes this recognition more difficult. On the west side of the southern limit of the deposit, there is an approximate drop off of 5 m to the plain below while near McRitchie Brook the drop off increases to 10 m or more. Along the east side of the deposit, there is a drop off of 20 to 25 m to McRitchie Brook.

The upper surface of the deposit is flat, slopes seaward and has an incised terrace that is 3 to 4 m below the upper surface. Although the deposit is small and lacks comprehensive elevation data (Fig. 11.2), gradients for the upper surface and the terraced surface, calculated along the boundary between the two, are very steep. Figure 11.3 shows the profile of the upper surface which has a gradient of approximately 17 m/km and of the lower surface, which slopes at about 11 m/km.

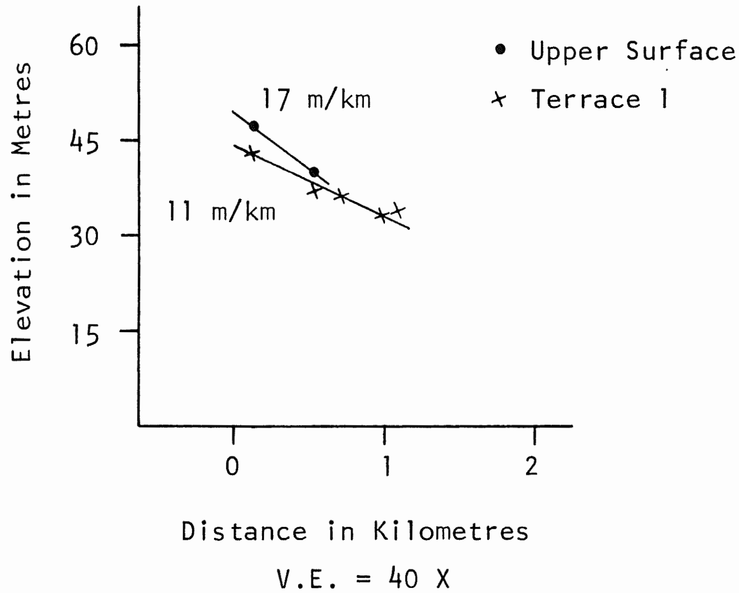


FIGURE 11.3. Gradients of the upper surface and terrace 1 on the terrace deposit, West Advocate. Profile essentially northeast-southwest along scarp between terrace 1 and upper surface.

There is no good exposure in the deposit but there are surficial indications of its composition. A fire ditch dug to a depth of 1 to 1.5 m across the northern end exposes gravel. The road that runs from the main street in West Advocate up onto the deposit, as well as the roads that cross the top of it, all have a gravel base.

INTERPRETATION

The terrace deposit is interpreted as a delta, although the interpretation is 'probable' rather than definite. It would be definite only if the topset/foreset/bottomset structure characteristic of the other deltas in the Minas terrace was exposed.

The interpretation is based on the topography, surficial composition, inferred thickness and location of the deposit, all of which are typical of deltas on the north shore of the Minas Basin. The flat seaward sloping surface of the deposit and also the incised terrace are features that characterize the known deltas. Sediment at the surface and in the fire ditch at the north end indicate that the deposit consists of gravel, as do the deltas in the outwash terrace. Moreover, the sharp drop of roughly 25 m from the incised terrace to McRitchie Brook along the east side and the approximate 10 m drop from the southeast corner of the deposit to the gravel veneered bedrock plain below suggests that the thickness is substantial. The gravel veneered bedrock rises to the west until it is within 5 m of the incised terrace, suggesting that the deposit probably thins to the west. The deposit lies at the mouth of the McRitchie Brook valley, like the known deltas, all of which are located at the mouths of rivers or large brooks. Thus, the terrace deposit is interpreted as a probable delta on geomorphological evidence which is insufficient by itself for a conclusive interpretation.

The 'delta' at West Advocate is a lesser or unknown entity in the outwash terrace along the north shore of the Minas Basin. Goldthwait (1924, p. 152) states that Chalmers (1894) listed the height of a delta at West Advocate as 130 ft (40 m). The author found that Chalmers (1894, p. 24M) had listed an elevation of 130+ ft for the highest Pleistocene shoreline at Advocate Harbour without identifying the type of shoreline deposit. Swift and Borns (1967) make no mention of a delta or terrace deposit at West Advocate. The lack of exposure in the deposit is probably the reason that it has been overlooked.

RAISED MARINE PLAIN

SURFICIAL GEOLOGY

At a lower elevation than the 'delta' is a marine plain that slopes towards Advocate Bay. The plain extends from Dewis Brook in West Advocate to Bog Brook in East Advocate.

In West Advocate, the plain is situated on the south side of the Cobequid Fault, as is the 'delta.' It is narrow (less than 0.5 km wide) and has more relief than elsewhere, with a gently rolling topography. Triassic bedrock (red beds) is near the surface and is covered only by a veneer of gravel. The highest elevation that the plain reaches (34 m) is in the northwest corner near Dewis Brook and a slightly higher (35 m) ridge that trends almost east-west is also located here. It is short (less than 0.5 km long) with a relatively steep front and a poor exposure reveals that the ridge is composed predominantly of tabular shaped pebbles. The ridge is cut off by Dewis Brook on the west end but to the east it is attached to Triassic bedrock at the Eatonville Road. From the ridge the plain slopes southward where it terminates at a 15 to 20 m sea cliff composed of Triassic red beds.

Towards McRitchie Brook, the plain decreases in elevation and slopes down to a lowland that is protected from high tide by dikes. The plain occurs south of, and adjacent to, the 'delta.' It is 5 m lower than the incised terrace on the southwestern edge and 10 to 15 m lower than the southeastern edge of the 'delta.'

East of McRitchie Brook in western Advocate Harbour, the plain remains narrow, is fairly low (maximum height ≈21 m) and lies south of the gravel veneered bedrock. There is a steep, bluff like slope from the veneered bedrock to the plain below. Here, as everywhere to the east, the plain slopes down to the lowland at the level of high tide (6.5 m).

Farther east near Spicer Brook, the plain broadens and rises in height on the northern edge until it is similar in elevation (27 m) to the veneered bedrock (29 to 35 m). Here the plain has three terraces developed on it (Figs. 11.1, 11.4). The upper terrace has an approximate elevation of 25 m, the middle terrace 18 to 20 m and the lower terrace 15 m. The raised beach, originally mapped by Swift and Borns (1967, p. 698), joins the gravel veneered bedrock just above the upper terrace. It is difficult to see this relationship clearly because of a graveyard located on the beach ridge and because of a thick growth of trees on the gravel veneered bedrock. The height of the beach ridge (30 to 33 m) is similar to the elevation of the veneered brook.

Between Spicer Brook and East Advocate, the plain broadens to its maximum width of approximately 1.25 km and has a more gentle slope. The aforementioned beach ridge is recognizable between Spicer and Burke Brooks, a distance of 1 km, where it trends northwest-southeast, parallel to the modern beach and is almost constant at 29 m in elevation. North of the ridge the plain reaches an elevation of 33 m and has a very low gradient; depth to bedrock appears to decrease northwards. The northern limit of the plain closely follows the assumed bedrock contact between the underlying Triassic Wolfville-Blomidon Formations (undifferentiated) and the Upper Carboniferous Parrsboro Formation.

East of Burke Brook, the plain continues to be broad and gentle, but has several ridges and troughs that undulate the surface. To the east and south, it thins out over Triassic bedrock while to the north the plain abuts the steep hills of Upper Carboniferous bedrock along the assumed Carboniferous/Triassic contact. The plain reaches its highest elevation of 35 m along the contact with the Carboniferous and slopes gently towards Advocate Bay.

SEDIMENTOLOGY

Several pits reveal that the plain is composed of shallow, seaward dipping (2° to 14°S) strata composed of gravel and some sand. Generally, the sand beds are close to the upper surface of the plain, overlying the gravel.

Wightman (1976) made a detailed study of the internal structure of the raised beach (Fig. 11.5), which is part of the raised marine plain in Advocate Harbour. In summary, the beach was divided into two units.

FIGURE 11.4. View northwestward from 4th terrace on raised marine plain. Arrows on sides of photograph mark 2nd and 3rd terraces. Steep slope above 2nd terrace is southern scarp of gravel veneered bedrock.

FIGURE 11.5. View northwestward along trend of raised beach, Advocate Harbour. Gravel pit (west wall) exposes internal structure, seaward dip of beds. Middle of pit face about 6 m high.

Unit 1, the lower unit, contains facies from the subtidal through the foreshore to the backshore (supratidal) and the profile of beds that are continuous through these facies is sigmoidal. The low tide terrace facies (subtidal) consists of sand or fine gravel ($M_z \simeq -2\phi$) and the beds have low dips (2° to 5°S). Coarser gravel ($M_z \simeq -3\phi$) is dominant in the foreshore facies and beds steepen to dips of 5 to 14°S. In the backshore facies, the beds flatten to dips of 3 to 7°S and the gravel is finer grained ($M_z \simeq -1.5\phi$) and better sorted than in the other facies.

Unit 2 is composed mainly of coarse gravel ($M_z \simeq -4\phi$) washed over the berm and deposited in the backshore zone of the beach as a storm ridge, which gives the raised beach its characteristic convex up shape. The coarse gravel is capped by a finer grained (sand) deposit.

INTERPRETATION

Swift and Borns (1967) interpreted the plain as a raised marine plain (glaciolittoral lithosome). The topography and internal structure of the plain and its relationship to the raised beach in Advocate Harbour corroborate this interpretation. The internal structure of the upper part is typical of beaches (McKee, 1957) in general but more specifically of the backshore zone of the beach (Wightman, 1976). Interpretation of the ridge in Advocate Harbour as a raised beach requires that the surficial sediment below it is also from the littoral zone (sensu lato). The seaward slope of the plain is also consistent with this interpretation. Where the plain rises above the raised beach, it is presumably older than the beach. The landward limit of the plain represents a very early shoreline established after deglaciation in the Late Wisconsinan.

Wightman's (1976) sedimentological study of the raised beach in Advocate Harbour confirmed Swift and Borns (1967, p. 698) interpretation of the ridge as a beach and led to several new conclusions. Because the facies in Unit 1 extend from the subtidal zone to the supratidal, the thickness of the foreshore facies represents the paleotidal range present during the formation of the beach. The thickness of the foreshore facies, and thus the maximum paleotidal range, is 3.4 m. Unit 2 was deposited during a small (1.7 m) transgression that resulted in the landward movement of the beach. The storm ridge in Unit 2 gives the raised beach its characteristic convex up shape and the cap of finer sediment (sand) represents the final stage of the transgression when the beach was a true spit.

The ridge in West Advocate is interpreted as a raised beach although it has not been studied in detail, which poor exposure will not allow at present. The interpretation is based on similarities in geomorphology and texture to the beach in Advocate Harbour and on its location on the raised marine plain.

The shape of the pebbles in the ridge is tabular, similar to the shape of the pebbles in the backshore facies of the raised beach in Advocate Harbour but exposure is not sufficiently good to determine if the gravel is openwork. The geomorphology of the deposit, a rounded ridge with a steeper seaward (S) than landward (N) slope, is similar to that of the raised beach in Advocate Harbour. The ridge is attached to bedrock on the east end and may have been attached to bedrock on the west end as well but erosion by Dewis Brook has removed the western end of the beach. The texture and topography of the deposit suggest that it was deposited as a storm ridge in the backshore facies of a beach. Post-depositional uplift and erosion have steepened the seaward (S) side of the deposit.

DISCUSSION OF SWIFT AND BORNS' (1967) MAP OF THE ADVOCATE HARBOUR AREA

Swift and Borns (1967, p. 698) recorded several raised beaches (spits) in the Advocate Harbour area (Fig. 1.2). They mapped the beach between Spicer and Burke Brooks, but continued the beach to the east of Burke Brooke. Another spit was shown in East Advocate, attached to the Triassic basalt of Cap D'Or, while a third spit was indicated in western Advocate Harbour. The beach mapped by the author in West Advocate is not shown by Swift and Borns (1967). They term the raised marine plain "outwash terrace" (Fig. 1.2) but refer to it in the text of their paper as the glaciolittoral lithosome (marine). Their "emerged wave cut bluff" is roughly similar to the edge of the 'delta' in West Advocate and western Advocate Harbour. Although figure 11.1 differs somewhat from their map of the Advocate area, the interpretations are essentially similar.

POSTGLACIAL EMERGENCE

The topset/foreset contact of deltas is potentially the most accurate indicator of former sea levels (Wightman and Cooke, 1978, p. 62), but there is no exposure in the 'delta' at West Advocate. The marine limit can not be inferred from the elevation of the *upper* surface of the 'delta' (40 to 47 m) because there is an unknown thickness of topset beds below the surface. However, the 'delta' does indicate a substantial amount of postglacial rebound and the elevation of the upper surface may be taken as a maximum for the marine limit provided that it is not an erosional surface.

Wightman (1976) concluded that the mean sea level during the deposition of Unit 1 in the raised beach at Advocate Harbour was 23.9 m above the present mean sea level. This elevation represents the midpoint of the foreshore facies, equidistant (vertically) between the subtidal and supratidal facies. The elevation of the top of the raised beach, 29 m, which represents the height of the sediments deposited in the backshore zone of the beach (typically a storm or beach ridge), does not represent the former mean sea level.

The raised beach in West Advocate has a higher elevation (35 m vs. 29 m) than the raised beach in Advocate Harbour. The marine limit for the beach is not known, as the poor exposure did not make possible a detailed facies interpretation that is necessary to fix the position of the former sea level. Assuming an equal thickness of supratidal sediments for both beaches, the beach in West Advocate would represent a former mean sea level of approximately 30 m (23.9 m + 6 m = 29.9 m). This is probably a maximum as the supratidal sediments appear to be quite thick. In any event, the raised beach in West Advocate was probably deposited during a higher sea level than the beach in Advocate Harbour because it is significantly higher (6 m) and the distance is not sufficient for crustal tilt to be invoked to explain the difference. The elevation of the former sea level represented by the raised beach in West Advocate is probably between 24 m and 30 m.

CHAPTER 12. NORTH OF PARRSBORO

One of only two passes through the Cobequid Highlands occurs north of Parrsboro while the other (Wentworth) is east of the thesis study area. The postglacial sediments in the Parrsboro Gap and in two major valleys that lie north of the gap, but are connected to it, were mapped to determine what influence the glacial ice north of the Cobequids may have had on the outwash delta at Parrsboro.

The chapter has been divided into sections to facilitate description and interpretation of 5 rather different areas. The first area is the Parrsboro Gap which extends from the Parrsboro delta to the headwaters of the Parrsboro River, a distance of less than 3 km (map D, map pocket, back of thesis). A low ridge at an elevation of about 30 m blocks the valley and forms the drainage divide between the Parrsboro River and River Hebert (map D); this constitutes the second area to be discussed. Gilber Lake abuts the north side of the dividing ridge and Newville Lake is about 4 km north of Gilbert Lake (maps D and E, map pocket, back of thesis). These two lakes form part of the headwaters of the River Hebert system and will be discussed together. The West Brook valley trends northeast from Gilbert Lake (map E) and part of this valley was studied, forming the fourth area considered in this chapter. An imperceptible drainage divide occurs on the flat valley floor near the settlement of West Brook. The last feature to be discussed is the Boars Back ridge which trends along the west side of the River Hebert valley between Newville Lake and the town of River Hebert (map E).

PARRSBORO GAP

SURFICIAL GEOLOGY

A series of hillside gravel deposits similar to those up valley from the deltas at Diligent River and Port Greville occur north of the delta at Parrsboro (map D) and extend to the drainage divide at Gilbert Lake. The deposits flank both sides of the Parrsboro River and lie against the bedrock hills of the Cobequid Complex. Along the bedrock contact, the deposits rise to almost 60 m (N.T.S. map 21 H/8) on the west side and to 55 m on the east side of the valley. On the west side of the river the deposits are more extensive and they cover much of the valley floor.

Small scattered deposits occur immediately north of the delta on the west side of the valley. Most of the deposits are hillocks but there is one esker like ridge, broken by a stream flowing out of Murray Lake, that "disappears" into the hillocks. A pit obscures the relationship between the hillocks and the ridge and where they meet they are at approximately the same height but the kames rise to the south. It appears that the kames cover the ridge rather than that the ridge goes over top of the kames.

North of this (west side of river), there is one large continuous deposit that forms a terrace (Fig. 12.1) that rises from approximately 25 m at the valley floor to over 46 m. The top of the terrace has a rolling topography punctuated by knobs and kettles. Some of the latter contain water but the permeability of the deposits is attested to by an ephemeral stream that ends on the terrace.

On the east side of the river, across from the undulating terrace, there is a continuous gravel deposit with 3 flat terrace levels on it, the lowest at 28 to 31 m, the intermediate level at roughly 45 m and the upper surface at 55 m. The terraces are fairly flat, but not quite as flat as the terraces on the deltas, and they are separated by terrace scarps.

SEDIMENTOLOGY

Gravel is exposed along the road cuts of the hillside deposits, but these exposures are poor and do not reveal any sedimentary structures. A pit immediately north of the delta on the west side of the Parrsboro River is the only good exposure. The sediment is mainly pebble gravel but changes in grain size and large boulders (up to 60 cm x 50 cm x 30 cm) are common. Most of the gravel is fluvially stratified with a crude horizontal stratification (few cross beds) so the direction of paleoflow is not readily apparent (Fig. 12.2). Previous exposures (before 1977) in the pit revealed blocks of red flow till incorporated in the upper parts of the gravel (Nielsen, 1976, p. 47). Zones of collapse and tilted sediments are also exposed in the pit.

INTERPRETATION

The hillside gravel deposits are similar in location, topography and composition to the hillside gravel deposits at Diligent River and Port Greville and the interpretation is the same; ice contact stratified drift. The deposits formed when the Cobequid Highlands were probably ice free and dissipating ice sat in the Parrsboro River valley with its upper surface probably not much higher than the height of the present deposits.

FIGURE 12.1. Northward view of hillside gravel deposits in Parrsboro Gap. Photograph taken from deposits immediately north of Parrsboro delta. Cross bar on second telephone pole marks top of rolling terrace. Parrsboro River in centre of photograph to right of closest telephone pole. Cobequid Highlands in background.

FIGURE 12.2. West wall of pit in gravel deposit immediately north of delta at Parrsboro. Sediments are fluvially stratified, especially in middle of pit. Some disrupted bedding near top of pit. Collapsed sediments on extreme right. Pit about 9 m deep, white field book in lower right.

The drift formed along the valley sides between the bedrock and the ice but became quite extensive as it almost covers the entire valley floor where the valley broadens. The exposed drift is well washed and most of the deposits were probably water deposited. As at Diligent River and Port Greville, the sediments could have been deposited by streams flowing along the trend of the valley or by streams flowing down into the valley from the highlands. The washed deposits probably vary from stream deposits to crevasse infills to small deltas. The ridge just north of the delta may be the remnant of an esker, indicating subglacial meltwater drainage and formation before the kames. It may also be stream sediment that was deposited on ice or in a crevasse and after the ice melted, the sediments were left as a ridge. The contact between the ridge and the hillock of drift has been partly removed and it is this evidence that is needed to decide if the ridge is an esker.

The inclusion of till and zones of collapse and faults in the deposits are consistent with an interpretation as ice contact stratified drift (Flint, 1971, p. 208). Some of the drift may have been ice floored or ice supported and when the ice melted, the sediments collapsed. The dissipating ice is also the source for the till. Nielsen (1976, p. 47) interprets the red till that was formerly exposed in the pit (Fig. 12.2) as having flowed out over the drift from the upper surface of the ice. Where the valley broadens and the large continuous terrace of drift was deposited on the west side of the river, the proportion of till in the sediment might be greater. Stagnant ice played an important role in the deposition of the terrace as the top of the terrace is dotted by kettles. The Parrsboro River floodplain is very narrow in this area so the drift/ice complex almost covered the valley, or perhaps did and part of it was later eroded by the Parrsboro River.

Kame is probably the best term for encompassing the variety of ways in which the drift may have been deposited. Individual hillocks are kames and the more extensive deposits are kame terraces. The ice contact drift on the east side of the Parrsboro River is a well defined kame terrace, similar to the flat topped kame terraces at Diligent River. The upper surface is probably the original depositional surface of the streams and as the ice melted and lowered, two terraces were cut into the kame material. The form and location of the fluvial terraces are such that they must have been cut by streams that flowed parallel to, but higher than, the present Parrsboro River.

The interpretation of the hillside gravel deposits as kames is not new. Goldthwait (1924, p. 26) identifies them as kame terraces and rejected the previous interpretation of marine benches by Chalmers (1894, p. 30M). Nilesen (1976, p. 47) used the more general term of moraine for the deposits but the interpretation is the same, although less detailed.

DRAINAGE DIVIDE

SURFICIAL GEOLOGY

At the southern end of Gilbert Lake, glacial material extends across the valley, creating the drainage divide between the Parrsboro River and River Hebert and also providing a natural crossing that man has used to construct a road. On the south side of the road the topography of the deposited material is hummocky and appears generally similar to the kames to the south. An unnamed Lake (Goldthwait, [1924, p. 26] calls it Summit

Lake, which will be used here) sits in a depression in the middle of the blockage. On the north side of the road, however, there is a very smooth sharp drop off to Gilbert Lake.

The drainage in this area is deranged as evidenced by the Parrsboro River and Henry Brook. The headwater of the Parrsboro River actually originates on the west side of the valley in the kames to the south and flows northward to a small pond, then east to Summit Lake and finally turns back on itself to flow south to Parrsboro. Henry Brook, on the east side of the valley, flows southwestward and is then deflected by the blockage to the northwest where it enters Gilbert Lake.

The glacial material at the divide is over 30 m above sea level for all but the most easterly part where it drops slightly below 30 m. The elevation of Summit Lake is approximately 26 m. There is little exposure in the material apart from road cuts and they show a semistratified sediment with rounded pebbles and cobbles. Thus, there is some evidence of deposition by flowing water but the silt/clay content is fairly high so the texture is somewhere between that of till and stratified outwash. Better exposures are needed to be more specific.

INTERPRETATION

The blockage across the Parrsboro River valley is interpreted as a recessional moraine. Glacial ice was located to the north of the moraine as the northern side of the moraine is steep and this is an ice contact feature. The ice in the valley was part of a larger ice mass north of the Cobequid Highlands (Nielsen, 1976, p. 193). The moraine is probably the result of a rejuvenation of the ice mass for a short period of time and the southward flow of the valley glacier built up the moraine. The south part of the moraine is indistinguishable from the large kame terrace to the south on the west side of the valley, which suggests that a large mass of wasting ice sat for some time in front of the more active ice. Summit Lake is a kettle lake that formed in this zone. The meltwater discharging to the south as the moraine was being built washed some of the moraine sediments and probably helped erode a path for the modern Parrsboro River through the glacially clogged valley. Southward discharge terminated before the moraine had been completely constructed as there is no sign of any breaching. Since deglaciation of the valley, the moraine has been the watershed for the Parrsboro Gap, separating the southward flowing Parrsboro River from the northward flowing River Hebert.

The boundary between the kame terrace to the south (west side of valley) and the recessional moraine is ill-defined because the kame terrace and the bulk of the moraine were formed in a zone of stagnant ice. The north side of the moraine abutted active ice so is genetically different but both the kames and the moraine are ice contact features.

Goldthwait (1924, p. 26) recognized the similarity between the kame terrace and the moraine and initially refers to the moraine as a kame moraine. Later, Goldthwait (1924, p. 99, 100) refers to the divide as a terminal moraine that was deposited by a tongue of ice in the valley. Thus, the interpretation presented here is similar to that of Goldthwait (1924).

The recessional moraine has deranged the drainage in the Parrsboro Gap and has left a relatively short Parrsboro River (≃7.5 km) and a long River Hebert (≃25 km). The trend of Henry Brook suggests that it flowed to the south prior to the emplacement of the moraine, which has deflected it to the north into Gilbert Lake where it is now part of the River Hebert system.

GILBERT-NEWVILLE LAKE BASIN

HILLSIDE GRAVEL DEPOSITS

Surficial Geology

In contrast to the area south of Gilbert Lake, most of the gravel deposits are on the east side of the valley. East and north of Gilbert Lake is a gravel terrace that is perched along the steep flanks of the Cobequid hills. The terrace is narrow and is broken by stream erosion and gullies where the gravel has probably slumped, but the top of the terrace is fairly flat and uniform in elevation although it is a little higher (62 m) on the north than on the south end (54 m). The road on the east side of the valley is built on the terrace.

Two gravel deposits of different elevations occur at the junction of the road on the east side of the valley (Prospect Road) and the New Canaan Road (maps D, E). The higher deposits reach elevations of over 45 m and have a hummocky surface while the lower deposits have a relatively flat surface that slopes west towards the valley. South of the New Canaan Road the contact between the two deposits is fairly sharp and occurs at an elevation of 30 m. North of the New Canaan Road the lower deposits have an elevation of 26 m some distance below the contact.

There are several scattered gravel deposits of the higher, hillock type east and north of Newville Lake. About 2.5 km northeast of the lake, the deposits have a fairly flat top at an elevation of about 38 m.

Sedimentology

A large pit (Fullerton's pit) is excavated in the higher, hummocky deposits at the junction of Prospect Road and the New Canaan Road. The pit face is very high (13 m) and exposes a very long set (11 m, vertical height) of gravel foreset beds that dip up to 34° to the west (Fig. 12.3), overlain by gravel topset beds that are horizontally stratified. The grain size of the gravel in the topsets and foresets is similar ($M_z \simeq -3\phi$ to -4ϕ). Some of the foreset beds are openwork and there are some sand beds in the foresets as well. Towards the western edge of the deposit, the foreset beds have several disturbed zones where they have slumped into various orientations.

The topset/foreset contact is erosional, very irregular and difficult to place in certain areas. It is undulating with relief up to 0.5 m. In addition, the topsets follow the contour of the hillock so that the topset/foreset contact is highest (44 m) at the top of the deposit and decreases in elevation along the sides of the hillock. Along the western side of the pit, the topsets are only 4 m above the floor of the pit and are disrupted in several places.

A small pit (pit face ≃ 5 m x 2 m) occurs in the lower deposits just southeast of Newville Lake on the east side of Prospect Road. The pit exposes steeply dipping (20 to 25°E) foresets that interfinger with topset beds (0.3 cm thick), all composed of sand. The interesting feature is that the foresets dip away from the central part of the valley.

A large pit (Newville pit) is excavated in the flat topped gravel deposits northeast of Newville Lake. As at Fullerton's pit, the exposed structure is deltaic, but there are two sets of foreset and topset beds. The lower set of foreset beds, exposed for about 5 m, are composed of very coarse grained sand to fine gravel and dip from 15 to 30° to the west. Bedding thickness varies from 2 cm to 50 cm and most of the beds are laminated parallel to the bedding but some of the beds with a lower dip (15°) are themselves cross stratified. The cross stratified sand dips up to 18° to the bedding and has a mini topset/foreset configuration with some of the topsets extending into the foreset laminae while others have an erosional contact with the foresets. The foreset laminae have an angular contact with the underlying bedding (Fig. 12.4). The cross stratified foreset beds are indicative of underflows depositing sand on the foreset slope which contrasts with the marine situation where underflows were not present.

The topset beds overlying the main set of foreset beds are 1 m thick while the uppermost topsets are 0.5 m thick. Both are crudely bedded, composed of gravel and are separated from the foresets by a planar, angular unconformity. The upper set of foresets is 1.5 m thick, composed of gravel and beds dip from 20 to 30°W, similar to the lower foresets.

Interpretation

The hillside gravel deposits are similar in location, topography, structure and texture to the deposits at Diligent River, Port Greville and south of Gilbert Lake and likewise, they are interpreted as kames. The kames were formed in a similar manner; that is, along the sides of an ice filled valley with largely ice free bedrock hills above the valley.

FIGURE 12.3. Westward dipping foreset beds exposed in Fullerton's pit. Foreset beds unconformably overlain by 1.5 m of topset beds. Man holding 1.5 m scale.

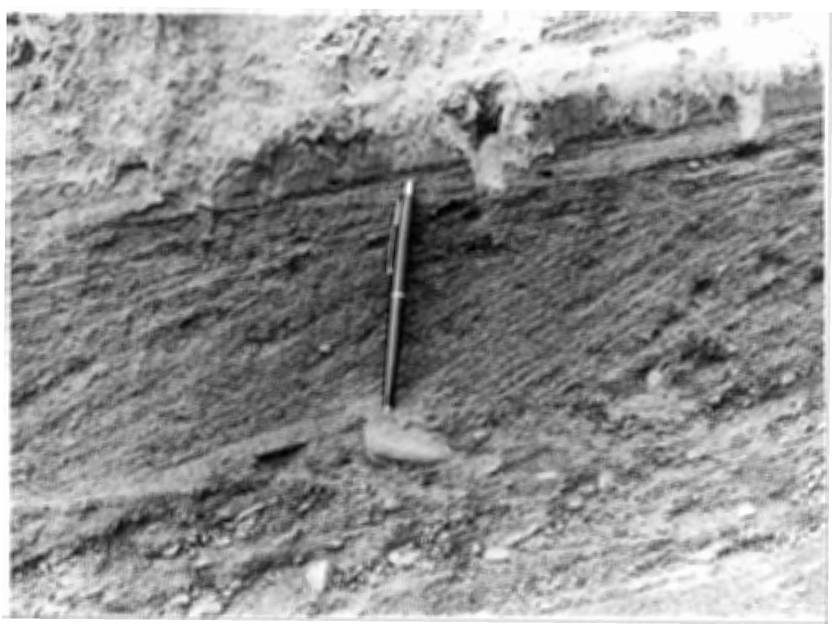


FIGURE 12.4. Cross stratified foreset beds near base of pit face, Newville pit. Some of overlying sand (above pen) extends into foreset laminae. Foreset beds dip 20°W.

The kame terrace east and north of Gilbert Lake is similar to the kame terrace southeast of Gilbert Lake, except it does not have incised terraces. The elevations of the northern terrace and the upper surface of the southern terrace are also similar and they form a southward sloping surface. Because the northern terrace is precariously perched on the flanks of the steep bedrock hills, it may have even been higher at the time of formation but has since dropped somewhat. The corresponding elevations and slope of the terraces suggest that while ice lay in the valley, before the recessional moraine was formed, water flow along the east side of the valley between the ice and the bedrock was to the south from almost as far north as Newville Lake. Goldthwait (1924, p. 100) previously made this interpretation, although he was not precise in his location of the kame terraces.

The kames exposed by Fullerton's pit and Newville pit are kame deltas. Sediment was deposited in water ponded between the ice in the valley and the bedrock hills. Delta progradation was toward the centre of the River Hebert valley and each of the kame deltas is located at the mouth of an eastward sloping stream valley that joins the River Hebert valley. Stream flow was probably from patches of melting ice in the hills. The ice would also release sediment and the streams would pick up sediment along the recently deglaciated valley sides for deposition in the kames. The kame deltas along the New Canaan Road indicate that the aforementioned southward flow along the east side of the main valley did not extend as far north as that, so probably did not extend beyond the northern limit of the kame terrace northeast of Gilbert Lake.

Wickenden (1941, p. 147) observed a sand and gravel terrace along the south side of the valley between West Brook and Newville Lake that would include the kame deltas exposed by Newville pit. He interpreted the terrace as outwash that was deposited in a lake. This interpretation is very similar to the one proposed in this thesis but Wickenden implies that the lake is fairly large and that wave action modified the form of the terrace. The present writer believes that the water was ponded between ice in the valley and the bedrock hills. There may have been many ponds at the same general elevation along the sides of the valleys, but no large lake.

The instability of the topset/foreset contact in the kame deltas is expected because the water levels of the ponds would not be as stable as sea level. Variable stream inflow would affect water level and if the water melted through part of the ice forming the pond, the water level would drop suddenly. Even if no breakthrough occurred, the water level would drop gradually as the ice melted. As well as changes in the water level, contemporaneous or postdepositional slumping or faulting of the delta sediments disrupt the contact, as it did in the marine deltas. Slumping and faulting of the sediments could be due to the melting of ice which supported sediments or to the relatively steep bedrock slopes upon which the kames were deposited. The superimposed foreset beds in the kame delta at the Newville pit may be the result of a rise in the water level of the pond or slumping of the delta, although the foreset beds were not disrupted.

The lower kame terrace at the junction of Prospect and New Canaan Road was formed after the higher kames and represents the last phase of deposition before the ice melted out of the valley. The only pit in the kames exposes sand with a deltaic structure so it was probably deposited in a small pond of water. The ice in the valley was probably not much higher than the kames and in this instance the stream flow was from the ice towards the side of the valley. The remainder of the low terrace is also probably finer grained than the higher kames, but deposition may have been more fluvial than deltaic. Elevations of the tops of kames probably approximate the elevation of the dissipating ice at the time of kame formation.

THE LAKES

Bathymetry and Bottom Sediments

Gilbert Lake lies on the north side of the recessional moraine and is part of the River Hebert headwater system. Echo soundings were taken in this lake in the summer of 1976 and the resultant bathymetric chart (Fig. 12.5) is comparable to that obtained by the Nova Scotia Department of Lands and Forests (Fig. 12.6). Gilbert Lake comprises two semi-restricted basins separated by a shallow sill (less than 2 m deep), with the northern basin deeper (over 10 m) than the southern basin (less than 6 m).

Echograms show that where the bathymetry has steep gradients (above 4 m depth in the south basin and above 8 m depth in the north basin), the lake bottom is a good reflector and is interpreted as a well compacted or sandy substrate (King, 1967, p. 693). This is especially true on the southwest side of the north basin where the gradient is very steep, the records are very sharp, and the reflector is probably bedrock or till

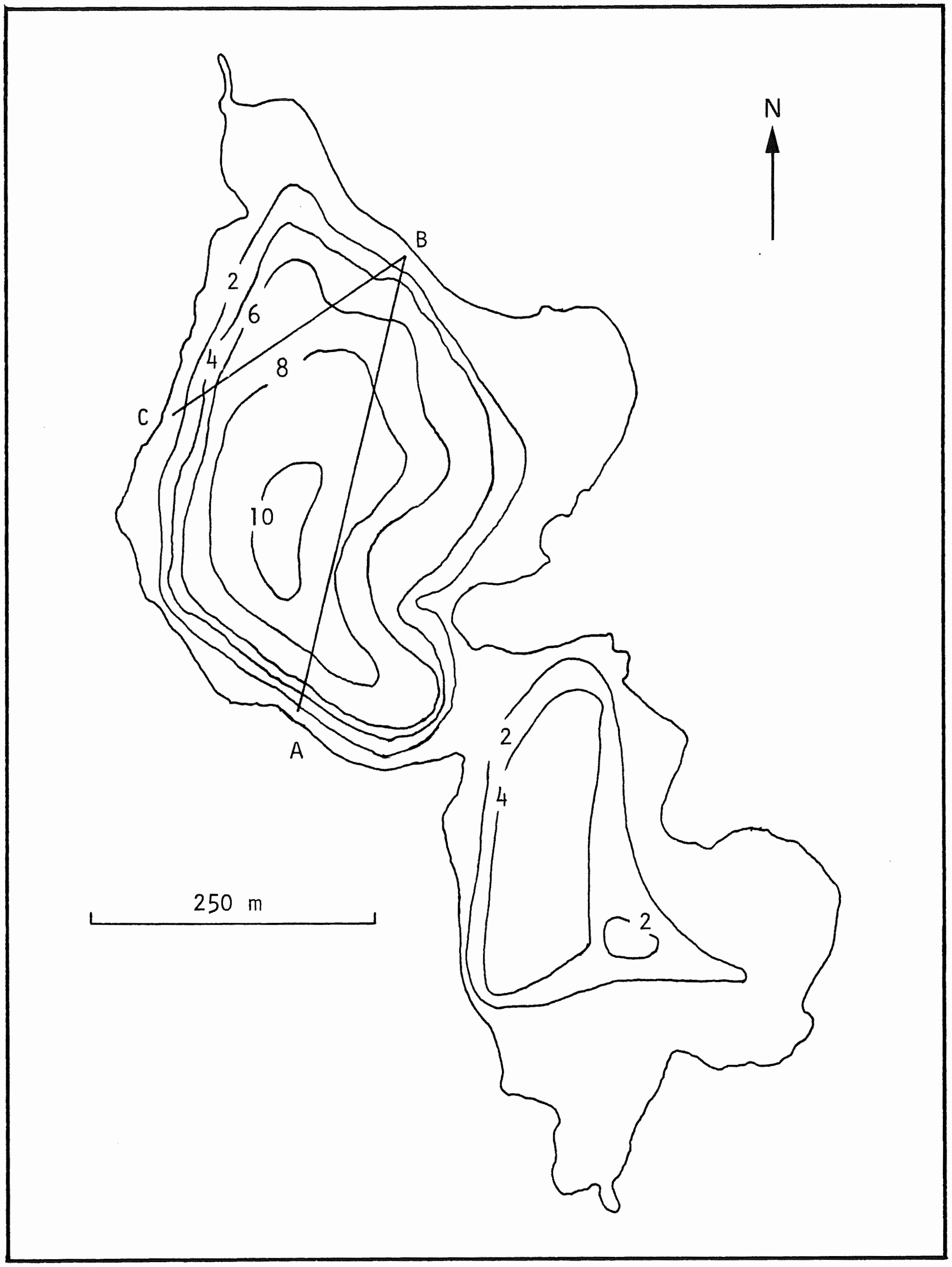


FIGURE 12.5. Bathymetry of Gilbert Lake, contour interval = 2m. Note location of profiles in figure 12.7.

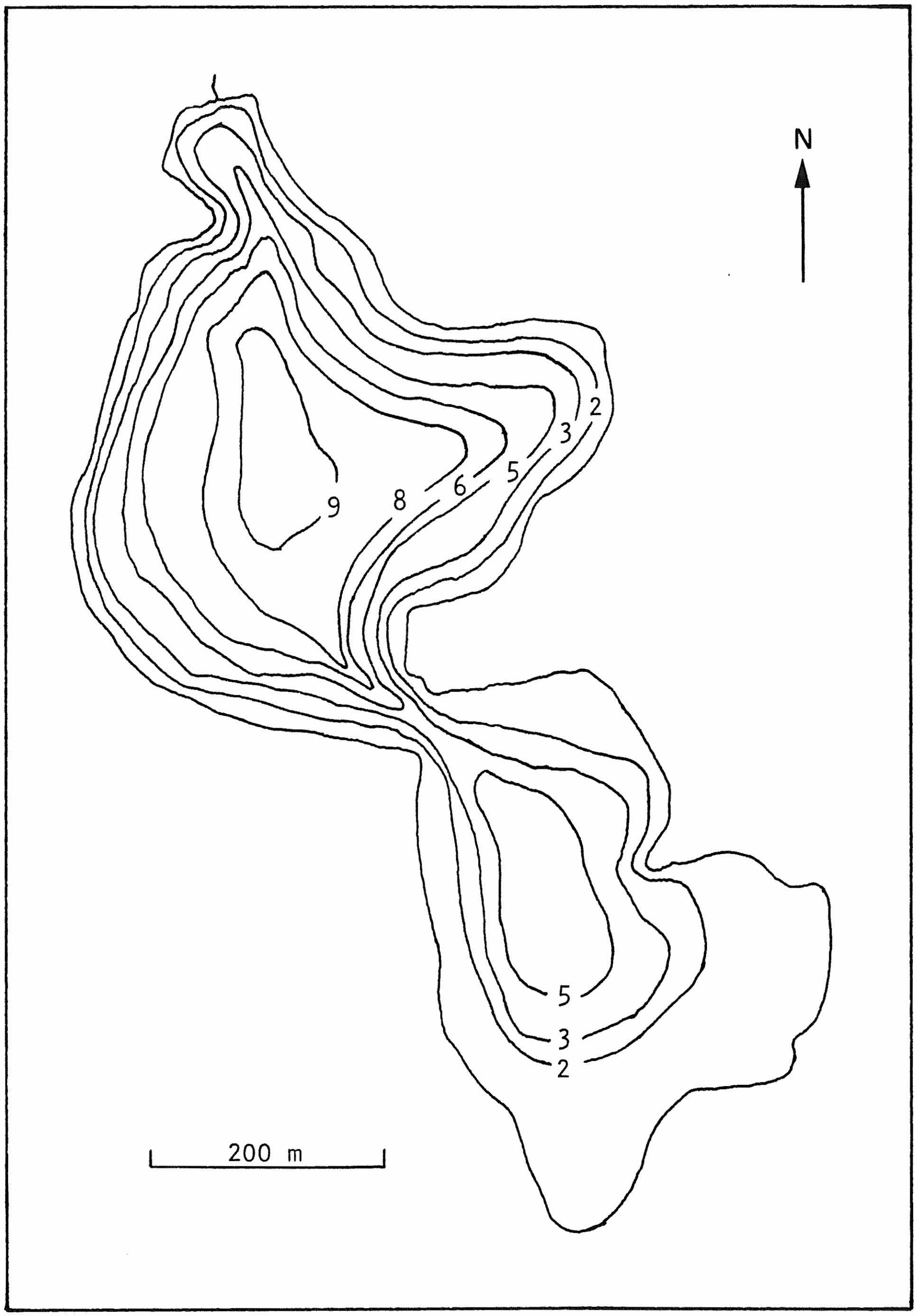


FIGURE 12.6. Bathymetry of Gilbert Lake. Five foot contour interval has been changed to nearest metre. (Modified after N.S. Dept. of Lands and Forests unpublished report 73023, 1973.)

(Fig. 12.7). The sill between the two basins is also a fairly good reflector, and the irregular topography suggests that it is a sediment veneered till. Where the gradients are lower, such as in the middle of the basins, the lake bottom is more acoustically transparent and is interpreted as loosely compacted mud. Livingstone cores, which are discussed in chapter 16, were taken in both basins and showed this to be true.

Echo soundings were also taken in Newville Lake. Despite its larger size (maximum dimensions 3.5 km x 2.1 km vs. 2.5 km x 1.3 km for Gilbert Lake) it is not as deep (just over 8 m, Fig. 12.8) as Gilbert Lake. The bathymetry and bottom sediments, as inferred from the echograms, are different in Newville Lake as well. There are shallower gradients and the lake bottom is a much better reflector, especially in the deeper parts (over 6 m) where there is a low gradient. Three distinct multiple reflections of the lake bottom are common on the records and four multiples are visible on one north-south profile. Many parts of the lake that are shallower are also good reflectors, interpreted as a sandy lake bottom.

In several places, where the bottom reflector is not quite so strong, generally between water depths of 2 to 6 m, a subbottom reflector appears at depths from 1 to 2 m. If the reflector is glacial till or bedrock, the Holocene sediments are not very thick in these areas. Below 6 m of water depth, there appears to be a depression similar to a thalweg that trends roughly parallel to the elongation of the contours. The lines of profile are poorly oriented for following this feature above 4 m but the line that is oriented the best for this purpose does show a similar feature. The thalweg trends towards the inlet and outlet positions of River Hebert and may be the result of an underflow of cold river water.

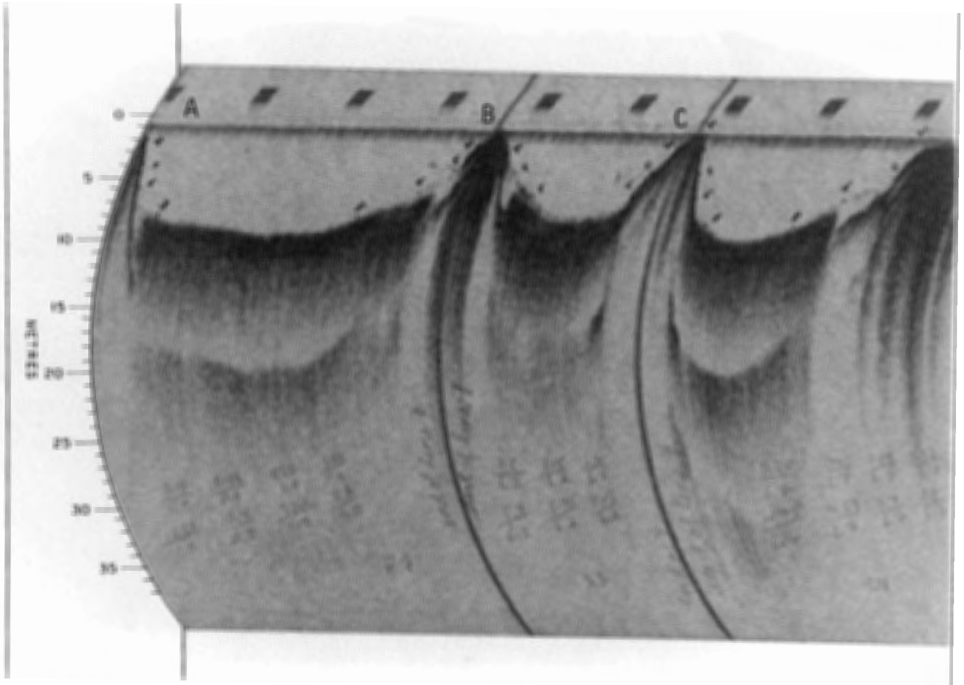


FIGURE 12.7. Echogram profiles (A-B, B-C; Fig. 12.5) across northern basin, Gilbert Lake. Sharp reflection on steep sides of lake, especially west side. Indistinct reflections on flat central parts.

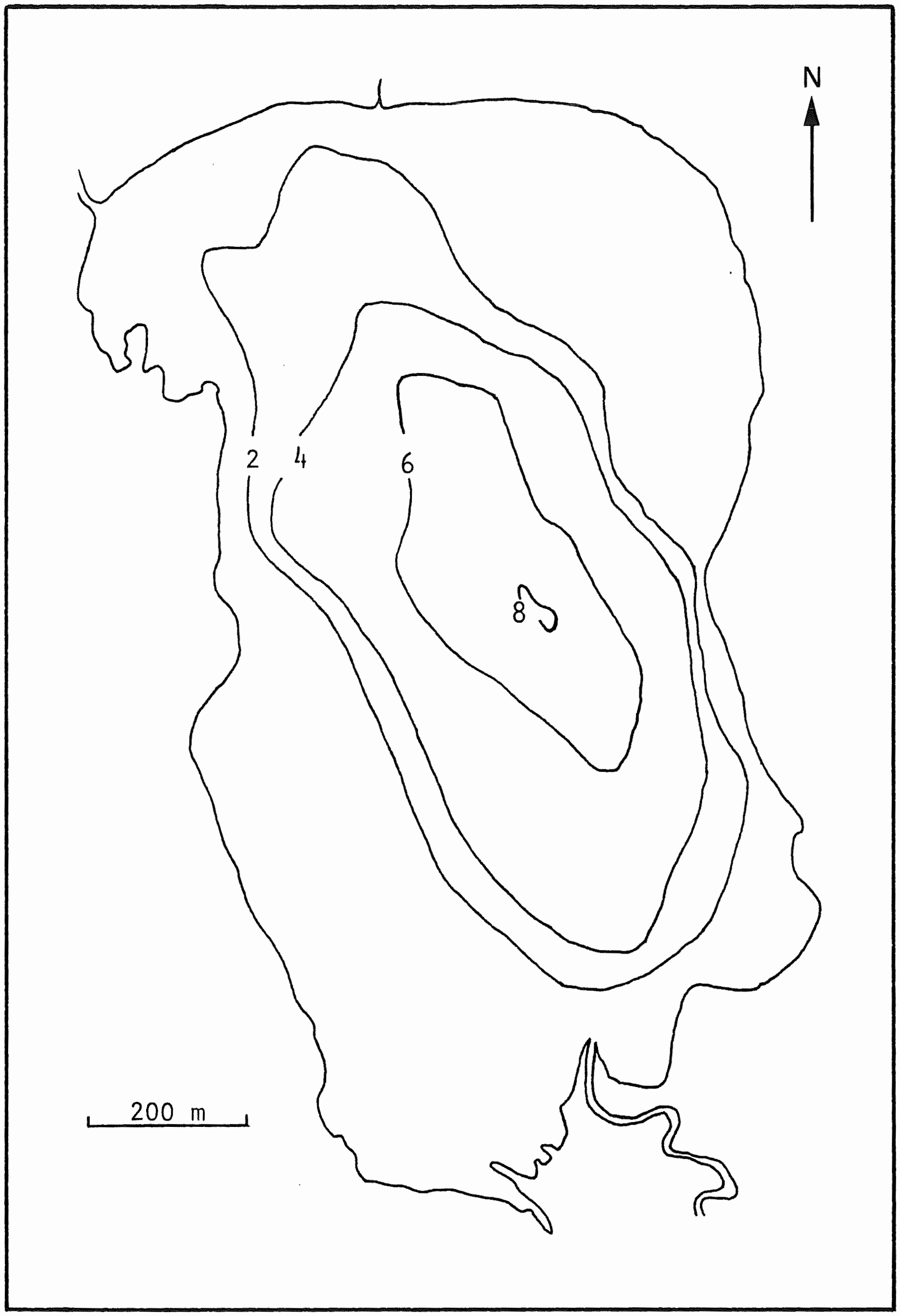


FIGURE 12.8. Bathymetry of Newville Lake. Contour interval, 2m.

There is little elevation difference between Gilbert Lake (20 m) and Newville Lake (19 m). Between the two lakes is a broad flat swampy area that encompasses most of the valley floor.

Interpretation

Gilbert Lake is formed in a glacially scooped depression immediately behind the recessional moraine. The echograms show a classic glacial U-shape, with the steep sides of the lake probably composed of bedrock or till. The till that separates the lake into 2 basins may have been sheared up into the ice as the moraine was formed. However the till was emplaced, it separated 2 blocks of late ice that occupied the present Gilbert Lake before the lake formed. The Gilbert Lake basin probably existed prior to the last glaciation, but the last glaciation deepened it and created the drainage divide at the southern end.

Newville Lake does not have a classic U-shape like Gilbert Lake, but it undoubtedly was deepened during the last glaciation. The valley is not as constricted as it is at Gilbert Lake and this may have affected glacial deepening. After the ice melted out of the valley, it appears that a much larger lake occupied this area of the valley. Gilbert and Newville Lakes are within 1 to 2 m of each other in elevation and a large swampy area separates them. After deglaciation, the lakes were probably part of a large lake with a very shallow central area, since turned into a swamp by sedimentation and growth of vegetation. The inferred limits of this postglacial lake, named Potter Lake after some of the early settlers in the area, are shown on maps D and E. Potter Lake would not have been much higher than the present Gilbert or Newville Lakes.

Chalmers (1894, p. 115M) also inferred that a large postglacial lake sat in this area after deglaciation, but he stated that the level of the lake was 30 ft (9 m) higher than the present Newville (Halfway) Lake. There is little evidence of a shoreline at this or any other elevated position. There are no terraces eroded in the north side of the recessional moraine and at Chalmers' suggested elevation (27 m), the lake would be breaching or very close to breaching the eastern side of the moraine, for which there is no evidence.

WEST BROOK AREA

HILLSIDE AND VALLEY FLOOR GRAVEL DEPOSITS

Surficial Geology

Hillside gravel deposits occur northeast of the community of West Brook along the West Brook valley. The deposits form a rolling terrace on the north side of the valley at an elevation of 30 to 45 m (N.T.S. map 28H/9W). There are also scattered deposits on the valley floor and on the south side of the valley from the community of West Brook eastward to the confluence of the Hannah and West Brooks. The deposits are best described as hillocks and the tops of the hillocks on the valley floor are between 25 and 36 m in elevation. One of the valley floor deposits that lies on the south side of the West Brook has a ridge on its northern side apparently split at the northeast end but erosion has removed the deposit beyond this point. On the east side of the confluence of the Hannah and West Brooks is a rather flat topped deposit that is situated in the centre of the valley.

Sedimentology

There are few good exposures in the gravel deposits in the West Brook area. Two pits are excavated in the flat topped deposit just east of the confluence of the Hannah and West Brooks. The northern pit is approximately 4 m deep and is poorly exposed but the southern pit is better exposed, especially in the southern part, and the pit faces vary in height from 1.5 to 4 m. Both pits reveal a medium to very coarse grained sand that is predominantly tabular cross stratified. At the southern end of the south pit the tabular cross stratification is best exposed. The beds are planar and dip 20 to 25° to the southwest with coarser sediment concentrated at the base of the strata. Bedding contacts are erosional and bed thickness is generally from 10 to 30 cm. The coarseness of the sand and abundance of tabular cross beds indicate that the sediment was deposited by shallow braided streams probably on the slip faces of linguoid or transverse bars.

Interpretation

The hillside and valley floor gravel deposits are interpreted as kames. The hillside deposits are similar to the deposits in the River Hebert-Parrsboro River valley and probably formed between ice in the valley and the bedrock hills. However, the scattered kames in the central part of the valley suggest that the dissipating ice/drift complex covered the valley floor and that washed glacial (glaciofluvial) sediment was deposited in various places, some of which have probably been eroded. The ridge on the north side of one of the kames may be part of an esker but the ridge stands above the rest of the kame and one might expect kame deposits to have covered an esker. The ridge may be a crevasse infill or stream sediment that was deposited on ice and later left as a ridge when the ice melted.

The cross strata in the exposed kame indicate that in this part of the valley, braided streams flowed southwestward, opposite to the direction of flow of West Brook which flows northeastward to join the Maccan River. There are no other exposures in kames farther to the southwest to corroborate or contradict this paleoflow direction. Interestingly, Prest (1972, p. 39) interprets that the outwash plain at West Brook was deposited by southwestward flowing streams. The streams may have initially turned and flowed southward through the Parrsboro Gap but he feels that the bulk of the meltwater entered the River Hebert valley and flowed north to the sea. Prest (1972, p. 30) suggests that this flow of meltwater was caused by ice blocking the Maccan River valley. Once free of ice, meltwater then flowed northward through the Maccan River valley.

The recessional moraine at Gilbert Lake precludes the suggested southward flow through the Parrsboro Gap but southwestward and then northward flow is possible. The paleoflow direction of the kame sediments near Hannah Brook is consistent with this interpretation, but it may also be the product of localized flow within the drift/ice complex. Exposures elsewhere in the valley, preferably to the southwest, are needed before a southwesterly flow of meltwater can be confirmed or rejected with some certainty.

VALLEY FLOOR

Surficial Geology

The valley floor from Newville Lake northeast to the community of West Brook is very flat. The drainage divide for the River Hebert and Maccan River systems occurs on this plain near the community of West Brook at an elevation of just over 30 m.

Sedimentology

Prest (1972, p. 39) states that "A significant outwash plain occurs north of the Cobequids in the headwaters of the River Hebert and the West Brook of Maccan River." It is not certain but it appears that Prest is referring to the flat valley floor in the West Brook area. However, along the banks of the West Brook, approximately 1.3 m of fine grained fluvial sediment is exposed. The permeability of the sediment is sufficiently low that the farmers in the area have buried pipes and dug ditches to drain the land (Karl Dickinson, personal communication, 1976).

A section was measured along the north bank of the West Brook (Fig. 12.9) and it revealed several interesting features. The upper 66 cm of sediment is fine grained but below that lies 30 cm of sand. The sand has a fluvial texture and structure and contains carbonaceous material near the top. Scattered tree trunks or branches, up to 7 cm in diameter, occur throughout the sand. Below that is 20 cm of black organic material with abundant wood, scattered pebbles and a fine grained sand matrix. Four centimetres of green carbonaceous clay separates that organic layer from another organic layer, with a very low sand content, that is 10 cm thick. Below that is gravel that is impenetrable with a shovel or hand sampler.

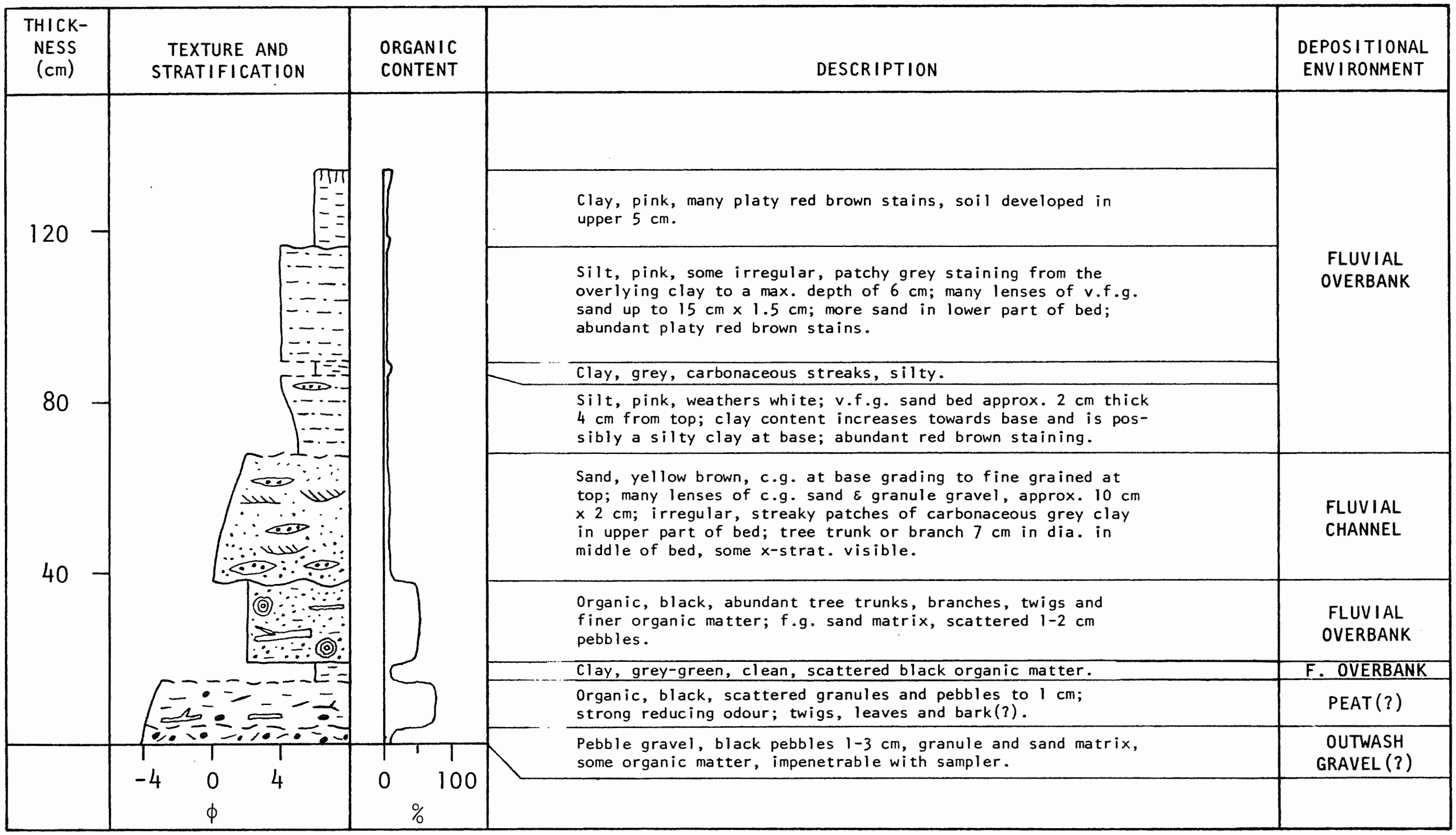


FIGURE 12.9. Section measured along north bank of West Brook. Organic content is estimated.

Interpretation

A sample (76-2) that extended from the green clay through the peat and into the top of the impenetrable gravel was sent to the Geological Survey of Canada in Ottawa. The pollen assemblage from the base of the peat layer (Dr. R.J. Mott, GSC Palynological Report No. 78-11) suggested an open treeless or sparsely treed tundra vegetation. Unfortunately, this sample was misplaced and was not radiocarbon dated. Part of the same sample was submitted to Dr. J.G. Ogden of Dalhousie University for radiocarbon dating and it yielded a date of 13,365 ± 420 yr BP (Dal-300). Because this date was unexpectedly old, a second (and different) sample (77-19) of the peat layer was submitted to the Geological Survey of Canada. The second sample (77-19) had been collected during an informal field trip and did not extend to the base of the peat layer, as the first sample had. Four pollen samples from peat sample 77-19 showed that the peat was deposited during the spruce pollen maximum (Dr. R.J. Mott, GSC Palynological Report No. 79-2). A sample from the basal part of this peat sample yielded a radioacarbon date of 9,830 ± 100 yr BP (GSC-2772) and Dr. Mott feels that this date is consistent with the pollen assemblage.

The approximate 3,500 year discrepancy between the dates (DAL-300, GSC-2772) is partly explained by the fact that the dates come from different parts of the peat layer. Mott (GSC Palynological Report No. 79-2) states that according to the pollen assemblage, peat sample 76-2 (13,365 ± 420 yr BP, DAL-300) should be slightly older than peat sample 77-19 (9,830 ± 100 yr BP, GSC-2772). But it does not appear that it would be as much as 3,500 years older. These age dates will be discussed more fully in chapter 16, but it seems that another sample will have to be collected to resolve the discrepancy in dates.

Because of the radiocarbon dates and pollen analyses from the peat layer, the impenetrable gravel below the peat is interpreted as a glacial sediment. The peat layer formed in boggy conditions (Mott, GSC Palynological Report No. 78-11) that ended with the influx of terrigneous sediment that formed the clay bed overlying the peat.

The organic layer just above the clay appears to have formed on the floodplain but near the stream channel. The sand matrix was probably washed in during floods but the bulk of the organic debris was probably derived from in situ vegetation on the floodplain. Floods that deposited the sand reworked much of the in situ organic debris and washed in other organic debris, including some or all of the larger pieces of wood. Coarse sand with fluvial structures predominates over the organics in the overlying bed and represents a channel deposit. The tree limbs and other organic debris in the sand are allochthonous. The finer sediments above this are interpreted as overbank deposits and the organic content of these sediments is relatively low.

A sample from a tree trunk or limb in the organic layer above the green clay was identified by Dr. R.J. Mott as *Pinus strobus* (white pine, GSC Wood Report No. 78-35). Another sample from the same piece of wood was radiocarbon dated by Dr. J.G. Ogden of Dalhousie University and it yielded a date of 3,745 ± 120 yr BP (DAL-265). Therefore, there is a large time gap (≈6,000 years) between the sediments below the green clay layer and those above the green clay layer but it is unlikely that the clay represents anything close to this amount of time. Some of the sediments above the clay may have been eroded and some of the finer in situ organic debris above the clay could be more than 4,000 years old. The log that was dated was at the top of the organic bed and was probably washed in from elsewhere. However, the date (DAL-265) shows that the bulk of the fluvial sediments exposed along the bank of the West Brook represent a relatively recent aggradation of the West Brook. It is a fairly broad floodplain for such a small (≈6 m wide x 0.5 m deep) brook.

Grant (1970, p. 680) points out that the level of high tide has risen approximately 12 m in the last 4,000 years in the Bay of Fundy. Most of the evidence for this comes from the Cumberland Basin into which the West Brook empties through the Maccan River. Grant (1970, p. 687) attributes the rise in the level of high tide in the Bay of Fundy to the combined effects of crustal subsidence, tidal amplification and a slow rise in sea level. Crustal subsidence and sea level rise both raise the base level for the streams entering the Cumberland Basin, which is a possible cause for the recent aggradation of the West Brook.

BOARS BACK RIDGE

SURFICIAL GEOLOGY AND SEDIMENTOLOGY

A discontinuous ridge occurs on the west side of the River Hebert valley between Newville Lake and River Hebert and much of the gravel road that follows this valley is built on the ridge. The ridge has previously been described by Dawson (1891, p. 82) and Chalmers (1894, p. 84M) and interpreted as an esker by several authors among whom are Goldthwait (1924, p. 98) and Wickenden (1941, p. 147) and with whom the present author agrees. The esker is usually 3 to 6 m above the surrounding deposits and about the width of the road. Ice contact stratified drift (kame material), which has not been mapped, flanks the esker in many places and care must be taken to distinguish between the esker and the kames. The kames are sandy and lack boulders that are common in the esker.

At the north end of the esker between Atkinson Brook and Kelley River the esker leads into a larger deposit of sand and gravel. Gravel pits obscure the relationship between the deposit and the esker and it is not clear whether the esker is part of, or is covered by, the deposit. The deposit comprises several peaked knolls with relief between the knolls reaching about 15 m. The deposit is composed predominantly of sand but there are gravel beds up to boulder size and scattered boulders, up to 60 cm in diameter, in the sand.

Small displacement (centimetres to tens of centimetres) faults are common in the sand. They are mostly high angle normal and reverse faults with the planes marked by a silty clay that stands out after weathering. Some of the faults might be due to compaction of the sediments, while the silty clay along the fault planes might be the result of intrastratal pore fluids escaping along the faults. Faults are common in ice contact sediments mainly because of the melting of the ice that supported the sediment.

It is possible that this deposit is part of, or related to, the esker but the uncertainty of this and the size and texture of the deposit favour interpreting it as kame material.

CROSS STRATIFICATION

There are several road cuts in the esker but the exposures are generally small and poor and the abundance of boulders limited the development of cross stratification. The cross stratification that was exposed dipped mainly to the south but some of the cross stratification dipped to the north, even at the same road cut.

The best exposure of cross stratification is in the gravel pit in the kame at the north end of the esker. Sand is well exposed on the north end of the east wall, near the top. Most of the sand is parallel laminated but some of it is cross stratified with dips mainly to the south, but also to the east. Climbing ripple cross stratification is present, indicative of high sediment supply (Reineck and Singh, 1975, p. 95).

PEBBLE LITHOLOGY

Pebble counts were taken at three road cuts along the esker, care being taken to ensure that the pebbles that were counted came out of the bank and were not lying loose on the ground. Vehicles can spray road gravel of various rock types into the ditch and contaminate pebble counts.

Sedimentary rock type pebbles (almost exclusively green sandstone) overwhelm all other rock types in the counts. In three pebble counts (each 110 pebbles) at the first location at the southern end of the esker, granite (2%) and quartzite (3%) are the only nonsedimentary rock types present in the counts. Three pebble counts (each 110 pebbles) at the second location farther north near Elm Brook show granite (1%), rhyolite (1%) and quartzite (2%) to be present as nonsedimentary rock types. At the northernmost location, due west of the Coldspring Brook-River Hebert confluence, two pebble counts of 160 and 161 pebbles were taken and quartzite (2%) and granite (1%) form the nonsedimentary component.

In the gravel pit in the kame deposit at the north end of the esker, the boulders were counted (118) and quartzite (2%) is the only nonsedimentary component. Finer gravel (from -3.5ϕ to -5.5ϕ) was counted (111) and it yielded the highest percentage of nonsedimentary rock, pegmatite (2%), basalt (2%) and quartzite (8%).

The sedimentary rock (conglomerate, sandstone and shale) in the esker and kame is probably derived from the underlying Upper Carboniferous Cumberland Group. Carboniferous rocks underlie and extend beyond the esker (to the north and south) and the kames which, combined with their erodibility, probably accounts for their abundance. Although quartzite, granite and rhyolite are only minor constituents, they are present and their low proportions are at least partly related to dilution by the sedimentary rocks. The granite and rhyolite pebbles are representative of the granite and rhyolite bedrock of the Cobequid Highlands north of Parrsboro (Peter Wallace, personal communication, 1978). Quartzite is less unique and there is latitude in the provenance of the basalt and pegmatite pebbles in the kame.

SUMMARY

Goldthwait (1924, p. 99) interpreted the paleoflow of the esker as southward and based the interpretation partly on the fact that there were no pebbles of Cobequid rock types in the esker. Prest (1972, p. 36) also states that the esker is composed of sedimentary rock types and implies that the paleoflow was to the south. This argument seems to be erroneous for two reasons. Firstly, the esker overlies sedimentary bedrock and its erodibility is such that basal ice would be full of sedimentary rock. All other rock types would be greatly diluted, no matter which way the water flowed. The south end of the esker ends well short (≈5.5 km) of the Cobequid Complex so even if paleoflow was to the north, there is no readily available source of Cobequid rock type pebbles in this area. The esker may have extended farther to the south, but there is no evidence for this.

Secondly, there are minor constituents of possible Cobequid origin that might have been emplaced by a northward paleoflow. Their low abundance would be the result of great dilution by the sedimentary rock types. The Cobequid type pebbles might also be from a source other than the Cobequid Highlands. Thus, the direction of paleoflow cannot be ascertained with any degree of certainty from the rock types present in the esker. The cross stratification in the sands is also ambiguous, although southward dipping cross strata does appear to predominate. More evidence is needed to interpret realistically the paleoflow direction of the stream that formed the esker.

It is possible that the deposit at the northern end of the esker is part of the esker, as the predominant dip of the cross stratified sands is in the same direction as that of the sands in the esker. Also, it occurs right at the northern end of the esker and there are no other large deposits similar to it in that area of the River Hebert valley. However, this evidence is insufficient to link the deposit positively to the esker and it is interpreted as a kame.

CHAPTER 13. PEBBLE LITHOLOGY OF THE DEPOSITS IN THE MINAS TERRACE

Pebble counts were carried out at each of the deposits in the Minas terrace so that source areas for the gravel might be located. The pebble counts themselves and a description of how and where they were taken are given in Appendix 2. The pebbles have been grouped into 10 general rock types and histograms of the counts are presented for each of the deposits. Some of the deposits have similar suites of rock types so these deposits have been grouped together to facilitate discussion. The igneous and metamorphic rock types that are present in each of the deposits are representative of the bedrock in the Cobequid Highlands directly north of the deposits. Sedimentary rocks both south and north of the Cobequids are possible source areas for the pebbles of sedimentary rock. In discussions of pebble counts from the foreset and topset facies of deltas, the frequencies of the foreset facies are listed first unless stated otherwise.

Fortunately, during the period when this work was being done, the Cobequid Highlands were being mapped by Dr. Howard Donohoe and Peter Wallace of the Nova Scotia Department of Mines. Their map was completed in 1978 and is included (in two parts) in the map pocket at the back of this thesis. It is referred to throughout this chapter, implicitly if not explicitly.

The potential sources for the rock types used in this study are set out in Table IV. The particular source may vary from area to area.

TABLE IV. Bedrock sources for the rock types used in the histograms

Rock type	Possible sources
Sedimentary	Upper Carboniferous Parrsboro Formation Upper Carboniferous Cumberland Group
Cleaved Sedimentary	Upper Carboniferous Parrsboro Formation Carboniferous Londonderry Succession Silurian to Devonian Portapique-Parrsboro Succession Silurian Advocate Succession Harrington-East River Succession North River Succession
Breccia	All rock types along faults
Metasedimentary and Quartzite	Carboniferous Londonderry Succession Silurian to Devonian Portapique-Parrsboro Succession Silurian Advocate Succession Harrington-East River Succession North River Succession Hadrynian Jeffers Succession
Granite	Carboniferous Granite Intrusions Devonian to Carboniferous Granite Intrusions
Rhyolite	Silurian to Devonian Portapique-Parrsboro Succession Silurian Advocate Succession
Andesite and Diabase	Triassic North Mountain Basalt Devonian to Carboniferous Diorite Intrusions Silurian to Devonian Portapique-Parrsboro Succession Silurian Advocate Succession North River Succession Bass River Metamorphic Suite Hadrynian Jeffers Brook Pluton
Diorite	Devonian to Carboniferous Diorite Intrusions Hadrynian Jeffers Brook Pluton

PORTAPIQUE, BASS RIVER AND ECONOMY

The pebble counts at Portapique and Bass River are similar in that mafic rock types are more abundant than felsic rock types while at Economy, the reverse is true (Fig. 13.1). The most diverse population of rock types in all of the deposits in the Minas terrace is at Portapique, but the Bass River and Economy deposits also have most of the rock types in the histogram represented. Discussion of their provenance is presented in Appendix 2.

LOWER FIVE ISLANDS

The pebble counts from the delta at Lower Five Islands are much lower in the mafic rock types (Figs. 13.2, 13.3) than the counts from the deposits at Economy, Bass River and Portapique. However, the counts are similar in that, although the delta overlies Triassic red beds, there are no Triassic sedimentary rock pebbles in the gravel. This strengthens the interpretation that the red beds were not competent enough to survive glacial erosion and fluvial deposition.

The pebble counts generally corroborate the division of the topset facies into 3 units. Units 1 and 2 are similar in lithology, but differ in the matrix, which is red in unit 1, while large clasts of sedimentary rock occur in unit 2. Unit 3 is distinctly different in lithology as granite is dominant and sedimentary rock drops to a low abundance.

Pebble counts in the foreset beds are also consistent with the lithologic divisions of the topset facies. Sedimentary rock is quite abundant and reaches a peak value of 49 percent in the mid delta area where it is overlain by unit 2 of the topset facies. Unit 2 is only marginally

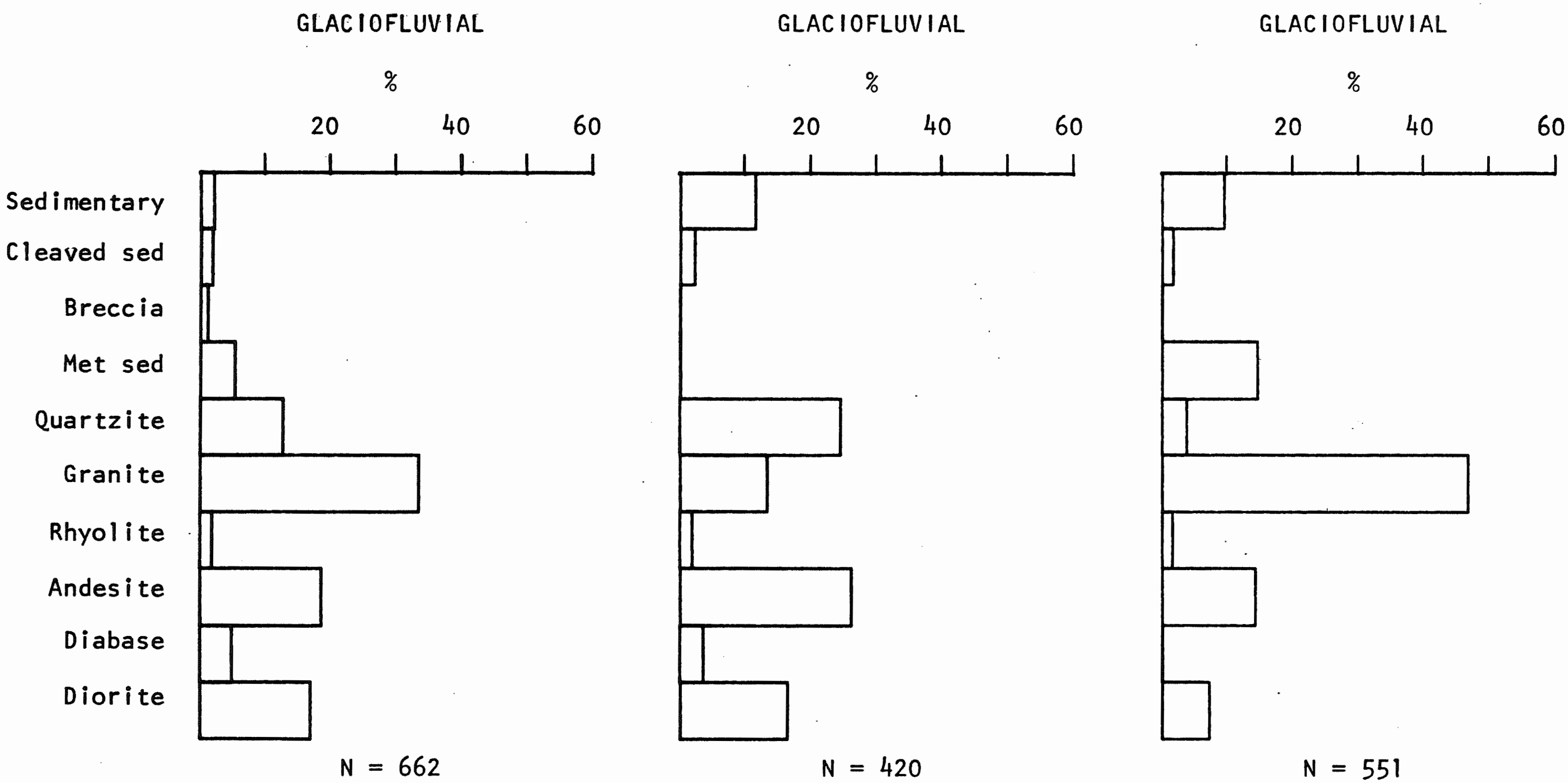


FIGURE 13.1. Pebble counts at Portapique (left), Bass River (middle) and Economy (right). N in this figure, and in all other similar figures, equals the number of pebbles counted.

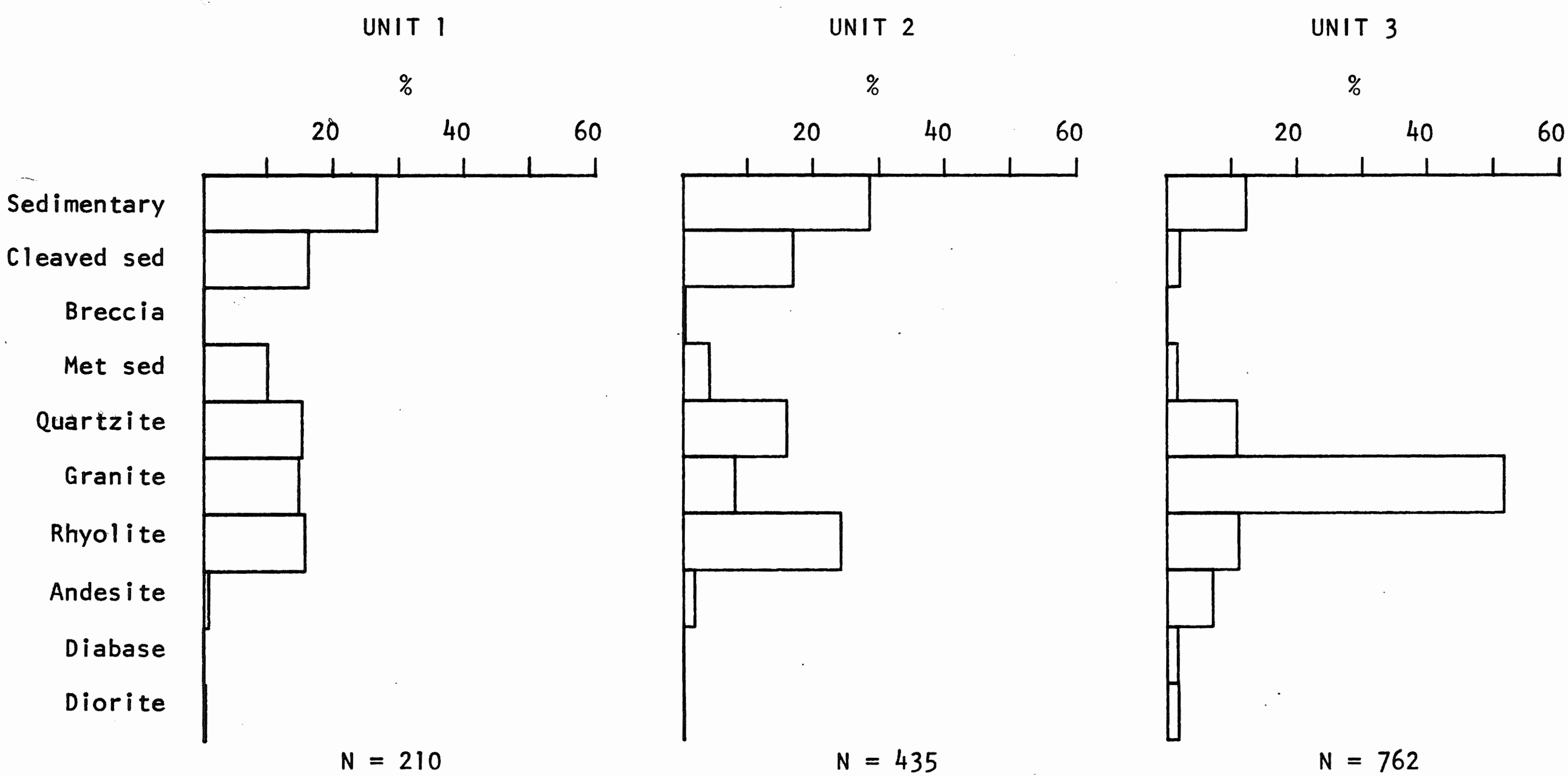


FIGURE 13.2. Pebble count from topset facies, delta, Lower Five Islands.

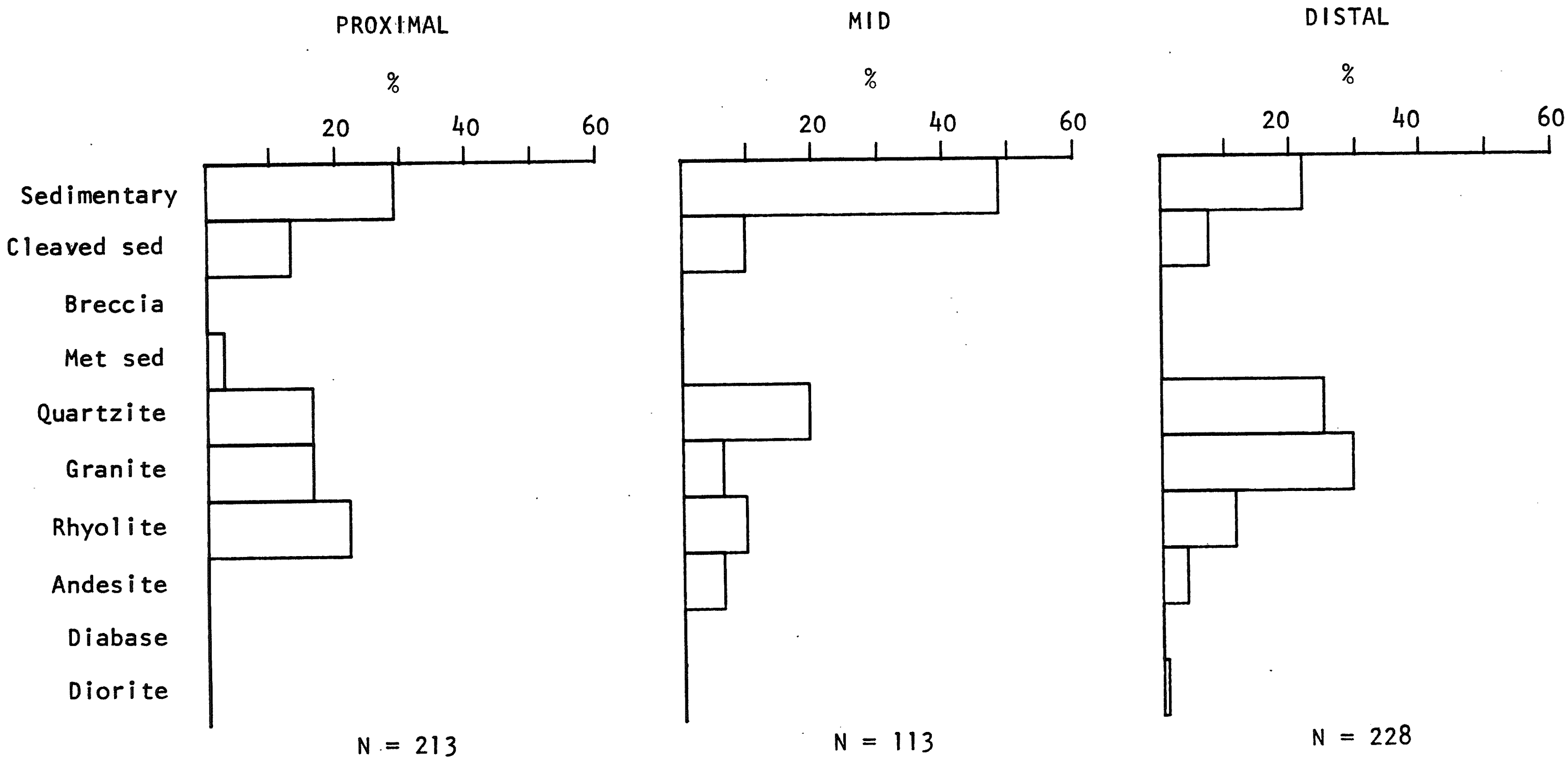


FIGURE 13.3. Pebble counts from foreset facies, delta, Lower Five Islands.

enriched in sedimentary rock as compared to unit 1 (29% vs. 27%), but as stated previously, large cobbles and boulders of sandstone are most common in this unit. It was suggested during the discussion of the delta in chapter 5 that the foresets were enriched in sedimentary rock because of the shape and size of the sedimentary clasts. The smallness (1 to 2 cm) and thin tabular shape of the clasts made them behave hydraulically as granule to sand sized particles so that many of the sedimentary clasts were deposited in the fine grained foresets. Coarser, more equidimensional clasts remained in the topsets. Thus, the enrichment of sedimentary rock in the foresets is to be expected as is the depletion of granite, which is well illustrated by the distal foresets (overlain by unit 3).

The source for the sedimentary rock in the delta is the Upper Carboniferous Parrsboro Formation that lies between the delta and the Cobequid Fault. As was discussed in the topset facies of the delta in chapter 5, the low amount of sedimentary rock in unit 3 indicates that the Carboniferous bedrock north of the Cobequids was at best a minor source area. The sedimentary rock in unit 3 might have been picked up as the streams flowed over the Parrsboro Formation or as the streams reworked the underlying outwash. By analogy, it appears that the Carboniferous bedrock north of the Cobequids is not the main source for the sedimentary rock in the deposits at Bass River and Economy either. While Carboniferous sedimentary rock is present in the deposits, it does not appear capable of withstanding glacial erosion and transportation over distances in the order of 10 km.

Granite, which is so abundant (51%) in unit 3 of the topsets, is probably derived from three Devonian to Carboniferous granite intrusions: one abutting the Cobequid Fault northwest of Five Islands, one north of the deposit in the southern part of the Cobequids and one on the north side of the Cobequids. Cleaved sedimentary rock probably comes from the Parrsboro Formation adjacent to the Cobequid Fault or the Harrington-East River and North River Successions on the north side of the fault. The Harrington-East River and North River Successions are also sources for quartzite and metamorphosed sandstone, as well as the Silurian to Devonian Portapique-Parrsboro Succession in the central part of the Cobequids. Rhyolite probably comes from the Portapique-Parrsboro Succession, which is also a source for andesite, along with the North River Succession and possibly the Devonian to Carboniferous diorite in the northern Cobequids.

MOOSE RIVER

The pebble counts at Moose River are somewhat similar to those at Lower Five Islands. The foreset beds are preferentially enriched in sedimentary clasts and the topset beds have a greater proportion of granite (Fig. 13.4). Glaciofluvial gravel in the outwash plain at Moose River is similar in composition to the topset gravel. Discussion of the provenance of the pebbles is presented in Appendix 2.

PARRSBORO

The pebble counts at Parrsboro are fairly diverse but the pebble population tends to be dominated by 2 or 3 rock types (Fig. 13.5). This trend is also noticeable at Moose River, whereas the deposits east of

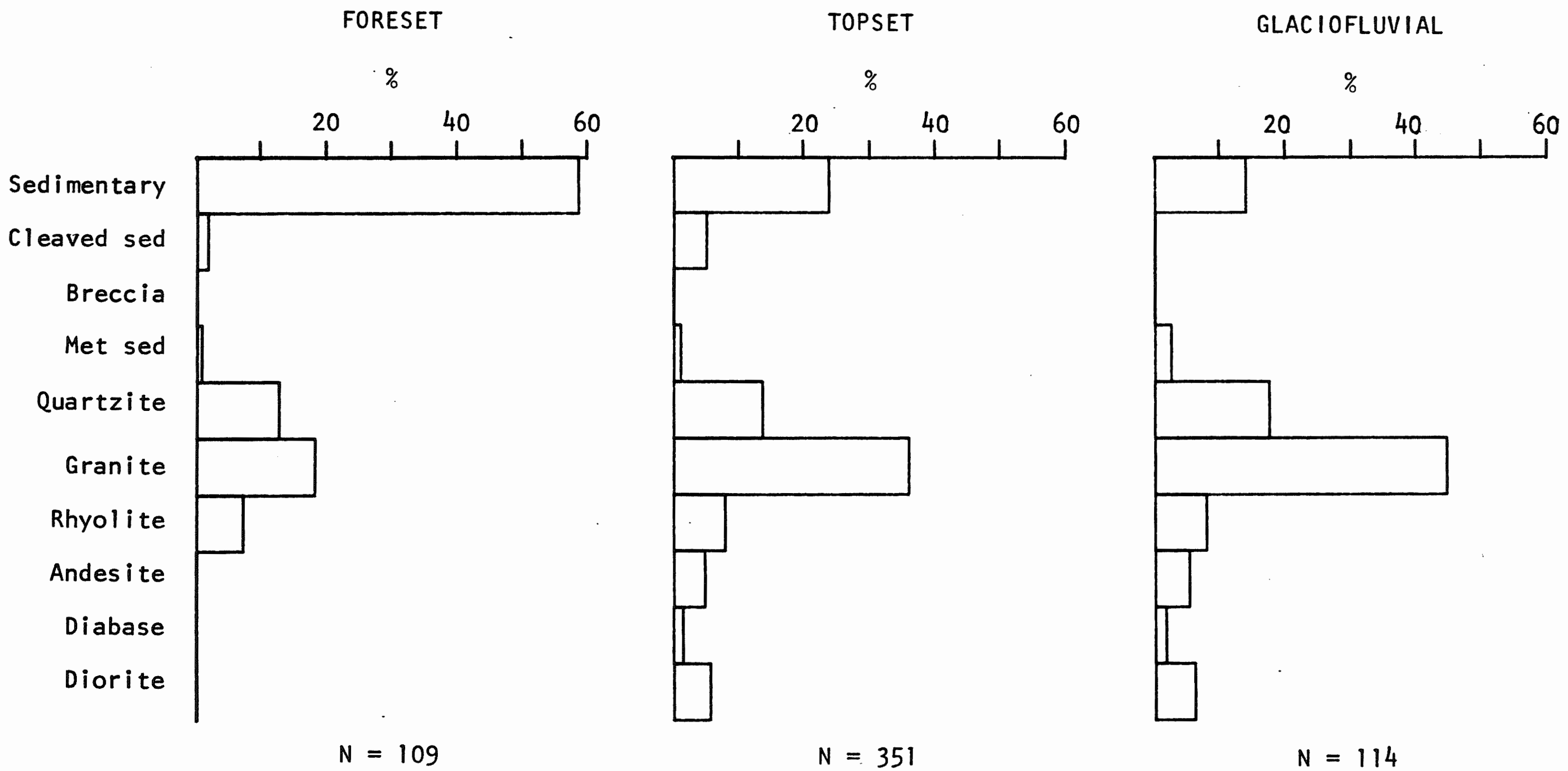


FIGURE 13.4. Pebble counts from delta (left, middle) and outwash plain (right), Moose River.

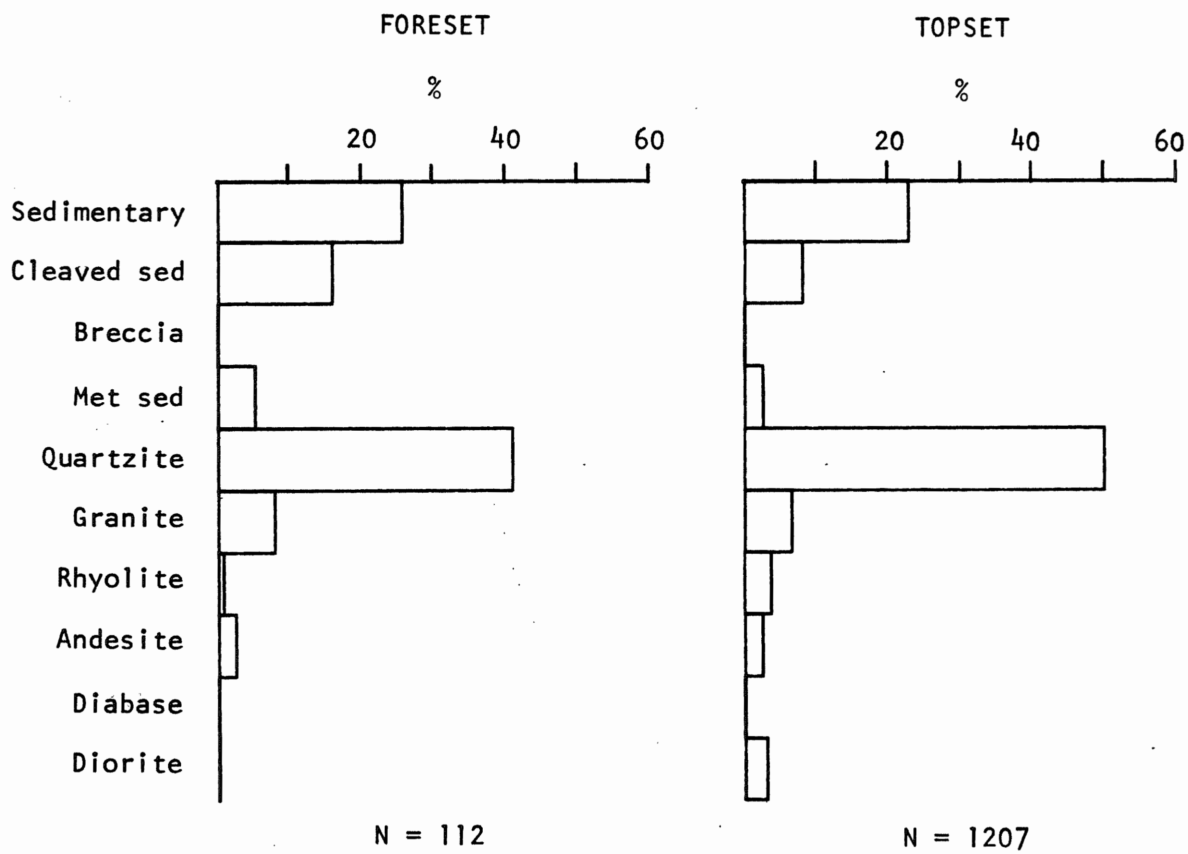


FIGURE 13.5. Pebble counts from delta, Parrsboro.

Moose River tend to have more balanced rock type frequencies. Obvious causes for this are the decreasing width and diversity of the Cobequid Highlands in a westward direction. North of Parrsboro, the Cobequids are about 7.5 km wide; north of Moose River they are about 12 km wide, and east of that the Cobequids attain a width of up to 18 km. A greater diversity in rock types accompanies this eastward broadening. Possible bedrock sources for the gravel at Parrsboro are given in Appendix 2.

DILIGENT RIVER AND PORT GREVILLE AREA

The pebble counts at Diligent River and Port Greville are less diverse than those at Parrsboro and in fact are the least diverse of all the deposits in the Minas terrace. The deposits are dominated by sedimentary rocks, quartzite and lesser amounts of cleaved sedimentary rock (Figs. 13.6, 13.7). Examination of the bedrock to the north of the deltas shows why this is so. The Cobequid Highlands are very narrow (≃3 km wide) and are composed entirely of the Silurian(?) Advocate Succession. The Advocate Succession is a source for the abundant quartzite, while sedimentary rock, which is predominant, is probably derived from the underlying Upper Carboniferous Parrsboro Formation and the wide belt of Upper Carboniferous (Cumberland Group) rocks that extends from the Cobequids to the north side of the peninsula. As the meltwater that deposited the delta discharged from the north, it seems probable that some of the sedimentary rock in the delta is derived from the Cumberland Group north of the Cobequids. The delta at Wards Brook overlies the Advocate Succession so it is highly probable that sedimentary rock in that delta is derived from north of the Cobequids.

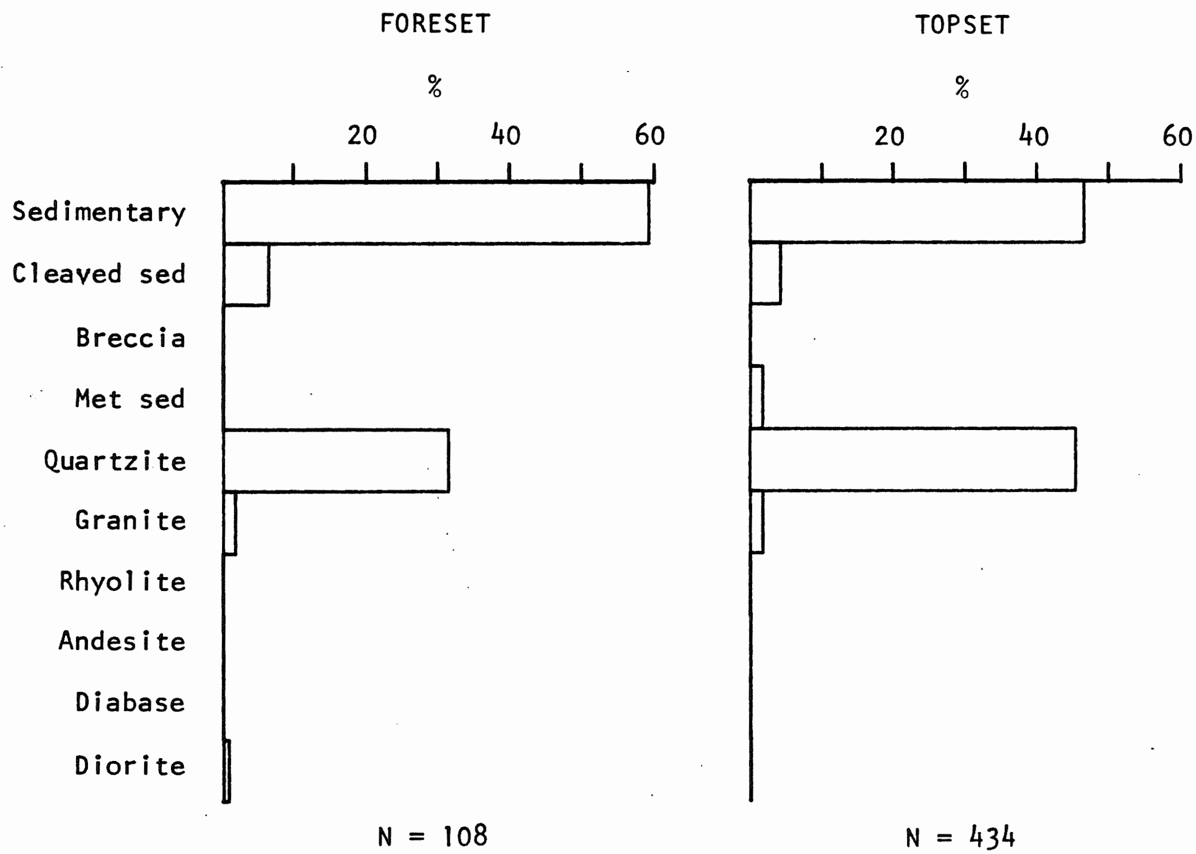


FIGURE 13.6. Pebble counts from delta, Diligent River.

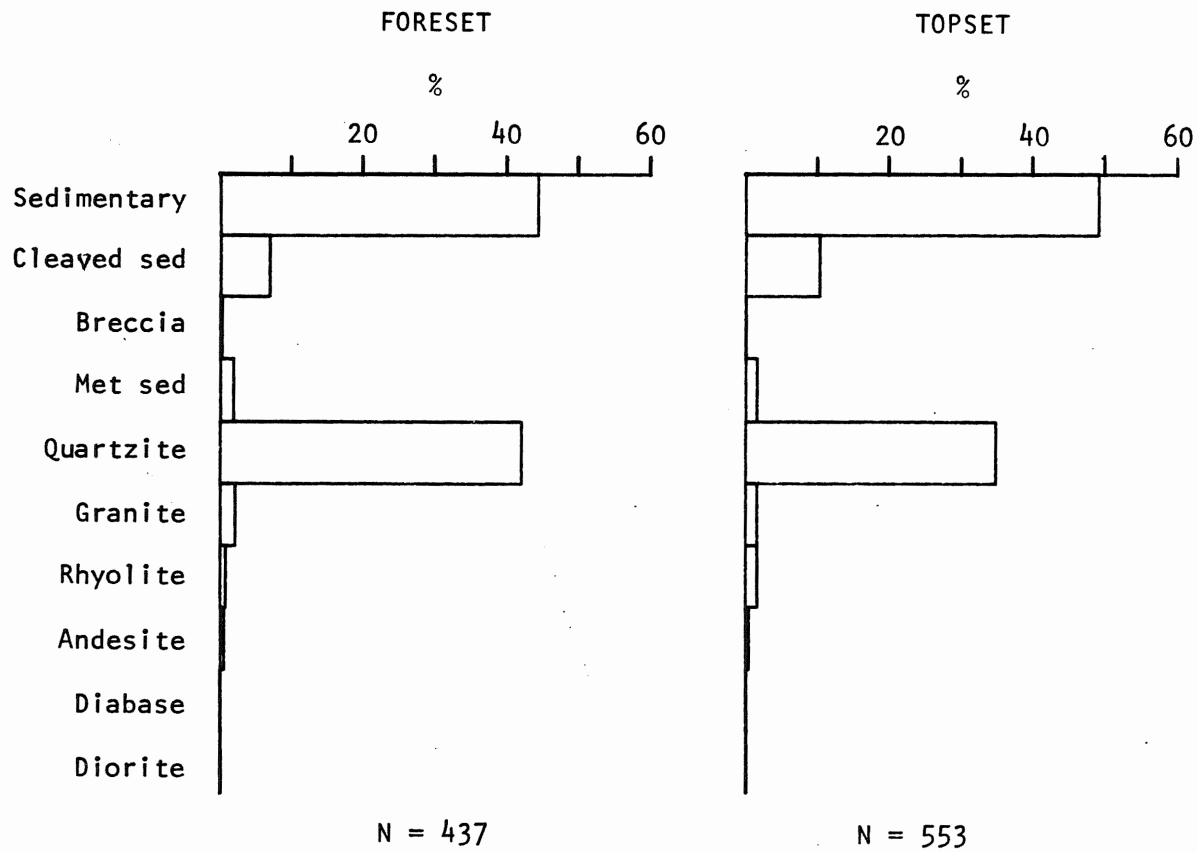


FIGURE 13.7. Pebble counts from delta, Port Greville.

In the discussion of the low abundance of sedimentary rock in the deposit at Portapique (in Appendix 2) and in unit 3 of the topset facies at Lower Five Islands, it was suggested that the Carboniferous sedimentary rocks might be largely incapable of withstanding glacial erosion, transportation and fluvial deposition over distances in the order of 10 km. This precluded the Upper Carboniferous rocks north of the Cobequids from being a reasonable source for the sedimentary rock in the deposits east of Parrsboro. It also militated against sedimentary rock from the Parrsboro Formation existing in abundance at the south end of the Portapique deposit. However, the situation at Port Greville and Diligent River is slightly different. The Cobequids are much narrower at Port Greville and Diligent River (≃3 km) than at the deposits east of Parrsboro (10 to 20 km) and thus the rocks of the Cumberland Group are much closer to the deposits. Also, the Parrsboro Formation is only about 2.5 km wide at Portapique whereas the Cumberland Lowlands due north of Port Greville are approximately 21 km wide and thus would be a better source area.

At Diligent River, the foresets contain a greater percentage of sedimentary rock (59%) than the topsets (47%), which is consistent with the other deltas in the terrace. However, at Port Greville, sedimentary rock is slightly more abundant in the topset facies (49%) than in the foreset facies (44%).

Cleaved sedimentary rock is probably derived from the Parrsboro Formation along the Cobequid Fault or possibly from the Advocate Succession. The minor amounts (less than 3%) of the other rock types such as granite and diorite are probably derived from distant bedrock sources not on the Chignecto Peninsula. The abundances of these rock types are probably indicative of the abundances of foreign pebbles in the other deposits. This suggests that most of the pebbles in the deposits of the Minas terrace are locally derived.

SPENCERS ISLAND

The pebble counts at Spencers Island are slightly more diverse than the counts at Port Greville and Diligent River, as granite is present in significant amounts (17%, 17%, Fig. 13.8). As is the situation at the other deposits where granite is present, there is a granite intrusion, in this case Devonian to Carboniferous in age, directly to the north in the Cobequids. A larger intrusion of Carboniferous granite occurs on Cape Chignecto to the west of Spencers Island, but is probably not the major source of the granite at Spencers Island. A Devonian to Carboniferous diorite intrusion is associated with this granite and diorite is not present in significant amounts at Spencers Island.

Sedimentary rock, the most abundant rock type, is probably derived from the Upper Carboniferous rocks: the Parrsboro Formation on the south side of the Cobequid Fault and the Cumberland Group north of the Cobequids. As at Port Greville, the topset beds are slightly enriched with sedimentary rock (40%) as compared to the foreset beds (35%). The Silurian(?) Advocate Succession is a source for the abundant quartzite and lesser amounts of metamorphosed sedimentary rock. Cleaved sedimentary rock is probably derived from the Parrsboro Formation adjacent to the Cobequid Fault and/or the Advocate Succession.

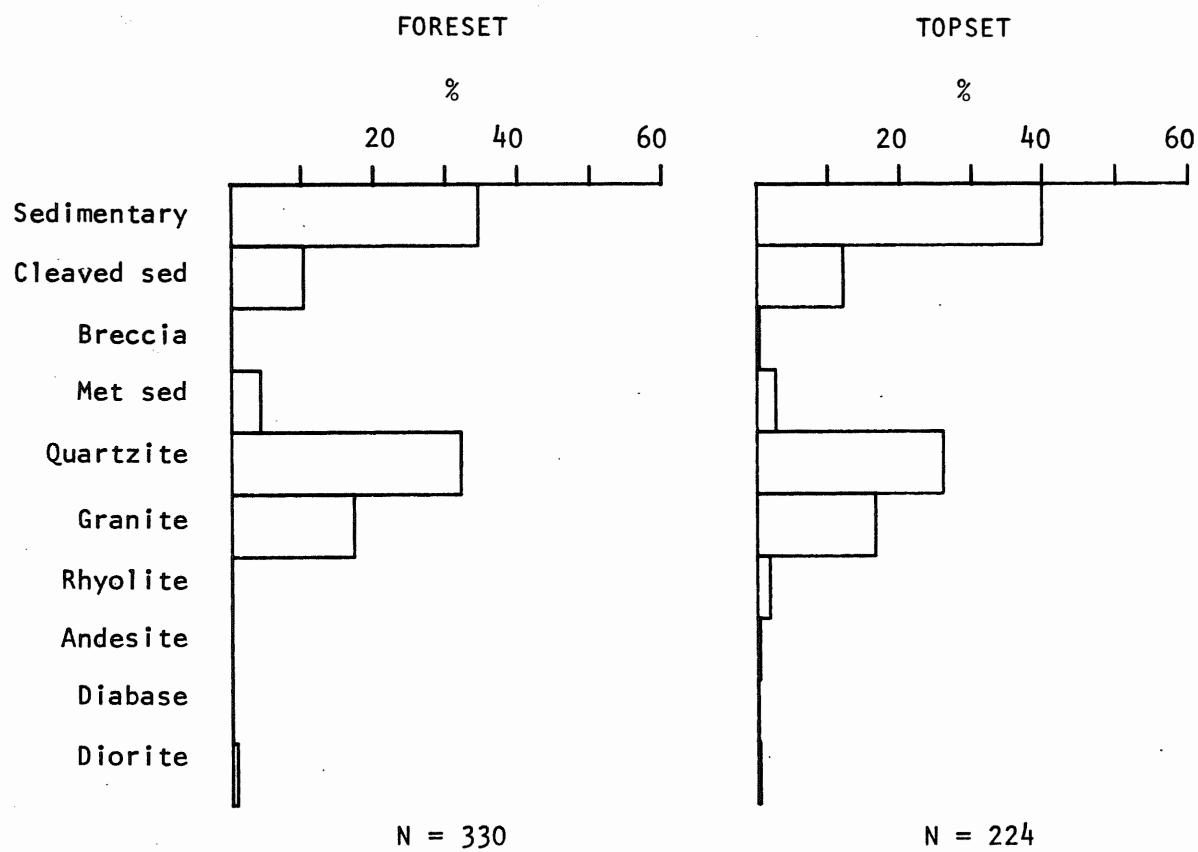


FIGURE 13.8. Pebble counts from delta, Spencers Island.

ADVOCATE HARBOUR AREA

The pebble counts from the delta in West Advocate (Fig. 13.9) are similar to the pebble counts at Port Greville and Diligent River in that the counts are dominated by quartzite and sedimentary rock with a lesser but significant amount of cleaved sedimentary rock. However, the pebble counts from the delta at West Advocate may not be an accurate indicator of the rock types present in the delta because the pebble counts were taken along surface exposures. The raised beaches in Advocate Harbour and especially West Advocate have more diverse populations of rock types than the raised delta (Fig. 13.9). The raised beach in West Advocate is the only deposit in the Minas terrace in which cleaved sedimentary rock is the most abundant rock type.

The sedimentary rock in the deposits is derived from the Upper Carboniferous rocks on the south side of the Cobequid Fault (Parrsboro Formation) and north of the Cobequids (Cumberland Group). All of the deposits except for the northeast corner of the delta overlie the Triassic Blomidon and Wolfville Formations (undifferentiated) but, like the deposits in the east end of the Minas terrace, there are no Triassic pebbles in the count. A clast of poorly lithified Triassic sand, similar to the one observed at Bass River, was exposed in the raised beach at Advocate Harbour in the summer of 1974. It was also interpreted as having been emplaced in a frozen state. The Silurian(?) Advocate Succession is the probable source for the quartzite and metamorphosed sandstone. Cleaved sedimentary rock is probably derived from the Parrsboro Formation adjacent

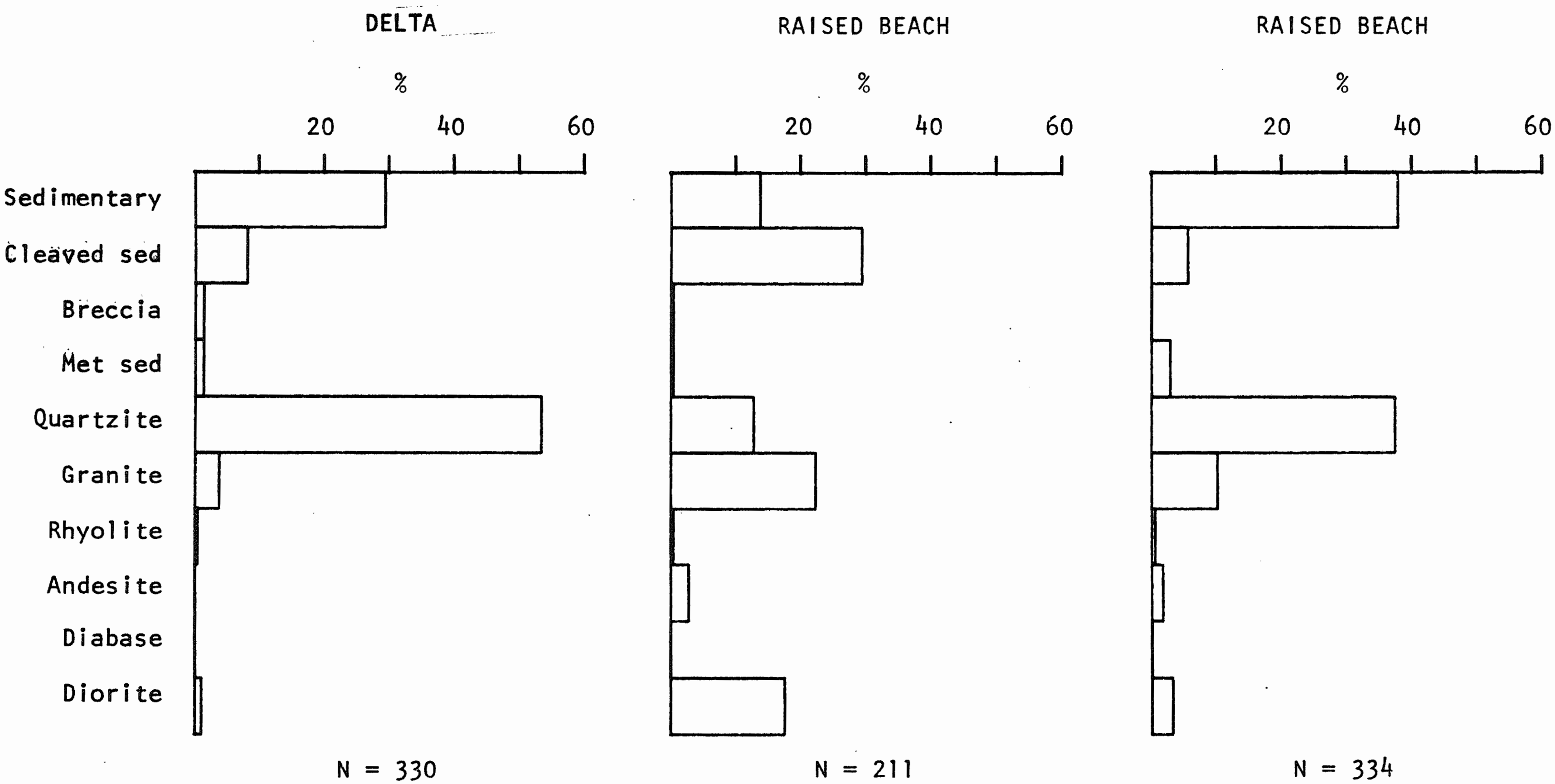


FIGURE 13.9. Pebble counts from delta, West Advocate (left), raised beach, West Advocate (middle) and raised beach, Advocate Harbour (right).

to the fault or the Advocate Succession. The small amount of granite in the delta probably has as its source the small ($\approx$1.25 km in diameter) intrusion of Devonian to Carboniferous granite northwest of the delta.

Relative to the delta, the beach in West Advocate is enriched in cleaved sedimentary rock, granite, diorite and andesite. One of the reasons for the abundance of cleaved sedimentary rock is the presence of phyllite, which makes up over half (55%) of the count. Carboniferous granite outcrops along the shore west of the beach (Cape Chignecto), as do Devonian to Carboniferous diorite and highly cleaved rocks of the Advocate Succession. If longshore drift during the formation of the beach was the same as at present (west to east), the rocks outcropping along Cape Chignecto were probably the sources of much of the sediment in the beach. Andesite might be associated with the diorite intrusion. The beach in Advocate Harbour is enriched, but to a lesser extent in these same rock types. Longshore drift is also the probable cause but the beach is farther from Cape Chignecto so the enrichment is less.

SUMMARY

The pebble counts reflect the types and proportions of bedrock that lie to the north of the deposits. To the east, the Cobequid Highlands are wide and relatively diverse in rock types and the outwash deposits are correspondingly diverse. The Cobequids narrow and become more homogeneous to the west and there the outwash deposits are dominated by two or three rock types. At Port Greville and Diligent River, the deposits are backed almost wholly by sedimentary and metamorphosed sedimentary rock and the pebble counts reflect this. The low frequencies (less than

3%) of other rock types such as granite give an indication of the probable amounts of foreign rock types in the other deposits.

Sedimentary rock from Triassic red beds were unable to withstand glacial erosion and deposition so that even where deposits overlie the Triassic red beds, there are no true pebbles of that origin in the deposit. Sedimentary rock from the Carboniferous formations does occur in the deposits, but source rocks for the pebbles are always less than 10 km away. The sedimentary clasts in the deposits are frequently split along bedding planes into several or many pieces so that they are not single clasts. The break up of the clasts undoubtedly occurred after deposition but it does indicate the relative incompetence of these materials. It is interpreted that the Carboniferous sedimentary rock was largely comminuted after several kilometres of glacial erosion, transportation and subsequent fluvial deposition.

Granite is a very good indicator rock type in the deposits. It occurs in significant amounts (>5%) in every deposit that has a granite intrusion north of it in the Cobequids. In some of the deposits, particularly at Portapique, the proportion of granite in the outwash is greater than would be expected from the proportion of granite mapped in the Cobequids. This may be the result of preferential erosion of the granite, better survival of the granite pebbles or a larger areal extent of granite than is indicated on the maps.

IMPLICATIONS ON ICE FLOW

The delineation of bedrock source areas for the deposits is very important in that it indicates a direction of ice movement on the peninsula. The bedrock source areas lie to the north of the deposits which means that the glacial ice that melted to form the outwash advanced over the Cobequids from the north. The bulk of the sediment in the outwash deposits was probably derived from the basal ice and from the underlying till. This does not mean that the only direction of ice flow on the peninsula was from north to south. But it does mean that during the last glacial episode to affect this area, the basal ice on the south side of the Cobequids had flowed from the north.

CHAPTER 14. ORIGIN OF THE DEPOSITS IN THE MINAS TERRACE

DEPOSITION OF THE TERRACE

Although the general features of the outwash terrace on the north shore of the Minas Basin were outlined by Goldthwait (1924), the accounts of Borns (1965, 1966) and Swift and Borns (1967) provided the first real evidence for the structure and origin of the deposits. This thesis study has extended the study of sedimentological aspects of the individual deposits and led to research on specific problems not previously considered. Although these investigations have in some cases refined or changed some of the earlier concepts, the general interpretation by Swift and Borns is not significantly altered and may be summarized as follows, incorporating the results of the new observations.

Dissipating ice lodged in the Cobequid hills released meltwater that carried glacial debris from the basal ice and till southward towards the sea, depositing glaciofluvial gravel in the valleys and building deltas where the streams reached the Minas Basin. The deposits collectively form a discontinuous outwash terrace that is emergent in the central and western parts and submergent on the eastern end (Swift and Borns, 1967, p. 707). Deltas are exposed where the terrace is emergent but the deltas have very little inland extension. The easternmost delta is at Lower Five Islands and farther east, only glaciofluvial gravels are exposed. At Economy, the glaciofluvial gravel is situated along the shore and may be part of a delta that was formerly exposed (Swift and Borns, 1967, p. 708). The change from emergence to submergence occurs near Economy. To the east of Economy, the glaciofluvial gravel extends well inland in the form of outwash plains along the Bass River and Portapique River valleys. The outwash plains are probably the up valley components of original delta systems but the deltas have been removed by postdepositional erosion. The submergence of the eastern part of the terrace militates against the preservation of deltas.

During the deposition of the outwash, the dissipating ice sat in the valleys and the hills were largely ice free (Swift and Borns, 1967, p. 703). The location of the deposits, the slopes of their upper surfaces and the paleocurrent directions of the sediment in the deposits clearly show that meltwater flow was confined to the valleys. The exception to this is the proximal part of the delta at Port Greville where the strike of the foreset beds at the inland exposure indicates that meltwater was flowing from the direction of the Cobequid scarp, less than 250 m to the north. In the early stages of delta formation, dissipating ice must have sat in front of, or at the top of, the Cobequid Fault (elevation $\simeq$ 90 m). After this ice melted, meltwater discharge was confined to the Greville and Fox River valleys.

Glacial meltwater and marine erosion removed most of the till in the valleys and along the shoreline, but in some areas erosion was incomplete and the outwash overlies till instead of bedrock. The patches of till attest to the former existence of a more extensive till sheet in the region. Small pockets of ice deposited outwash in bedrock lows and at the mouths of small brooks and remnants of these deposits are scattered all along the shore.

POSITION OF ICE FRONT DURING DEPOSITION

The present study suggests that there are two separate lines of evidence to indicate not only that the melting ice was very close to the deposits in the early stages of deposition, but that the ice was in close proximity to the deposits throughout their formation. The evidence is set out below.

Geomorphological Evidence

The unbreached recessional moraine that cuts across the valley at Gilbert Lake and divides the Parrsboro River and River Hebert drainage systems indicates that after the ice receded from the moraine, no melt-water was supplied to the delta. As the distance from the proximal end of the delta to the recessional moraine is only about 3 km, the distance from the ice front to the delta was never greater than this during delta construction.

Compositional Evidence

In the delta at Lower Five Islands, the three topset units and the foreset beds that directly underlie each of the units are chronostratigraphic 'packages.' The foreset beds at a given point are not exactly the same age as the overlying topset beds because they are separated by an erosional unconformity. The foresets are at least as old as, or older than, the topsets but compared to the length of time for delta formation, they can be considered as equivalent. The upper limit (time) of each of the 'packages' is the upper boundary of the topset units. Each of the topset-foreset packages represents a different and distinct period of time within the formation of the delta.

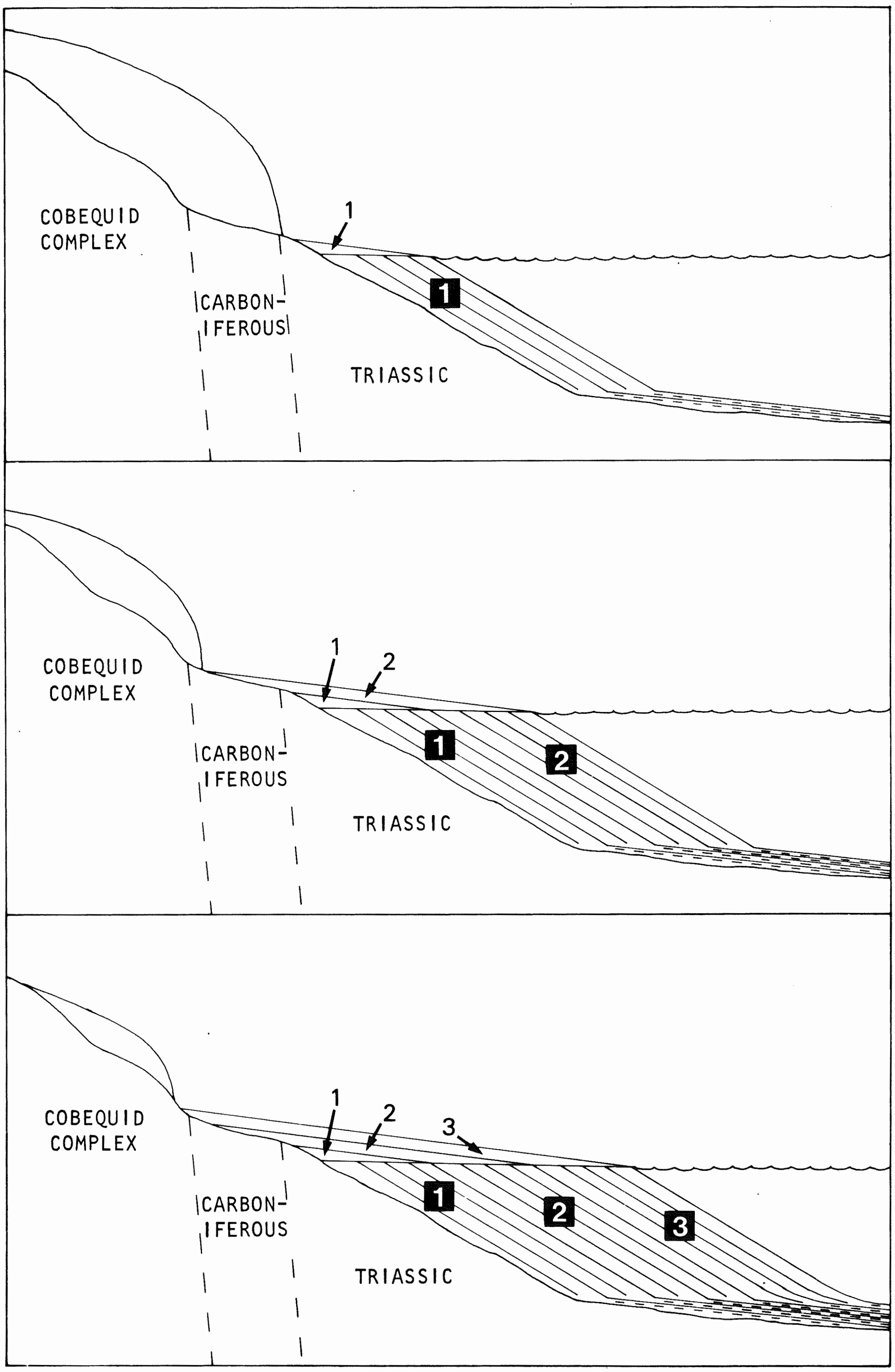
As was shown in chapter 5, when unit 1 was deposited, the delta was in its earliest stages of formation and the foreset beds were prograding seaward as the topset beds thickened upwards. The ice front then stood on the Triassic red beds (Fig. 14.1, top), which means that it lay within the present limits of the delta. As the ice front receded onto the Carboniferous rocks, the unit 2 foreset beds were deposited seaward of the unit 1 foresets and the unit 2 topset beds were deposited on top of the unit 1 topsets (Fig. 14.1, middle). The width of the Parrsboro Formation limits the former ice front to a position within 2 km of the delta during the formation of unit 2 which means that the bulk of the present delta was formed before the ice had receded to the Cobequid Fault. Unit 3 was deposited after the ice front had receded on to the Cobequid Highlands. During this time, the delta continued to build upwards and prograde seaward (Fig. 14.1, bottom).

STRATIGRAPHY OF THE FIVE ISLANDS FORMATION

Swift and Borns (1967, p. 703) proposed formal names for the units of the terrace on the north shore of the Minas Basin. The Minas terrace is designated the Five Islands Formation with the 'type' exposure at Lower Five Islands. The formation was divided by them into two members: the upper Saints Rest Member, defined as glaciofluvial and a lower marine unit, the Advocate Harbour Member, embracing the deltas of the terrace, including the topset beds, and the glaciolittoral lithosome at Advocate Harbour.

The Saints Rest Member is envisaged as an alluvial fan deposit which prograded across the Advocate Harbour Member after it had been deposited,

FIGURE 14.1. Schematic diagram showing deposition of Lower Five Islands delta. Top - margin of dissipating ice sits on Triassic bedrock as unit 1 is deposited. Middle - unit 2 is deposited after ice front recedes on to Carboniferous bedrock. Bottom - last ice sits on Cobequid Complex and unit 3 is deposited as recession and downwasting continues.


COBEQUID COMPLEX
1
CARBON-IFEROUS
TRIASSIC
1
COBEQUID COMPLEX
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CARBON-IFEROUS
TRIASSIC
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2
COBEQUID COMPLEX
1
2
3
CARBON-IFEROUS
TRIASSIC
1
2
3

uplifted and dissected to a minimum of 6 m of relief (Swift and Borns, 1967, p. 703). At the type exposure at Lower Five Islands, the distinction between the topset beds of the delta and the Saints Rest Member is based by them on composition:

> "... the upper unit (Saints Rest Member) is rich in granite and, to a lesser extent, in metamorphic clasts but is poor in sedimentary clasts... The percentages are reversed in the lower unit..."

It is clear from this description that Swift and Borns (1967, p. 700) regard my units 1 and 2 of the topset facies as the topset beds of the delta and that my unit 3 is the Saints Rest Member. It has been shown in chapter 5 that unit 3 has the same relationship within the delta as units 1 and 2. The difference in composition in the topset facies is due to a receding ice front that released different rock types as it crossed bedrock boundaries. Swift and Borns (1967, p. 700) recognize this latter fact but envisage the unit 2/unit 3 boundary as representing a distinct time gap followed by a change in depositional environment. I interpret both units 2 and 3 as fluvial (subaerial) topset beds laid down within a continuous sequence of delta building. The relief on the contact is similar to the relief commonly found between beds within the units and does not seem to be a significant unconformity. Because the units are part of the delta, they should be part of the Advocate Harbour Member as it was defined and thus, at the shoreline exposure at Lower Five Islands, the Saints Rest Member *per se* does not exist.

However, at Saints Rest, the type location for the Saints Rest Member, the glaciofluvial gravel overlies bedrock and therefore complies with the definition by Swift and Borns (1967). Also, up valley from the delta at Lower Five Islands or up valley from any of the other deltas where the glaciofluvial gravel overlies bedrock, the glaciofluvial gravel could be called the Saints Rest Member. Only where glaciofluvial gravel is part of a deltaic structure, and therefore overlies marine sediments, does it become definable as topset gravel and is therefore a part of the Advocate Harbour Member.

The problem can best be resolved by redefining the Saints Rest Member. One possible solution would be to define the Saints Rest Member as comprising nondeltaic glaciofluvial gravel and therefore the member would end at the point where the fluvial gravel becomes part of a deltaic structure (i.e. overlies foreset and bottomset beds). This would retain the original definition of the Advocate Harbour Member, which includes the topset beds. However, this solution is unsatisfactory because continuous exposure within a delta is required in order to delimit the boundary between the two members and this does not exist. A better alternative is simply to define the Saints Rest Member as glaciofluvial gravel so that the topset beds of the delta become part of the Saints Rest Member. This solves the problem of locating the proximal boundary of the topset beds.

Accordingly, it is proposed that the Advocate Harbour Member be redefined as comprising the truly marine portions (foreset and bottomset beds) of deltas and raised marine littoral gravel (including raised beaches). The Saints Rest Member is sediment of glaciofluvial origin and thus includes the topset beds of deltas, as well as the subaerial outwash plains and the glaciofluvial gravel continuous with, but up valley from, the deltas. The topset beds may be regarded as the topset facies of the Saints Rest Member.

INLAND EXTENT OF THE DEPOSITS

The distance that the deposits extend inland is dependent upon the topography of the valleys in which they occur. If a deposit is situated in a wide (≃1 km) valley with low (15 to 30 m) walls, such as the deposits at Portapique and Bass River, it extends well inland because the braided streams were able to be spread out and form a well developed outwash plain. If a deposit is situated in a narrower valley with higher walls, it is restricted in its inland extent. Even if postglacial sediment were deposited in a narrow, deep valley, the preservation potential of the deposit is low. This effect is well illustrated at Bass River, Portapique and Parrsboro.

The deposits at Bass River and Portapique extend well inland but for most of their lengths they overlie the soft Triassic red beds. At Bass River, the valley narrows and the walls become higher (by about 100 m) where the bedrock changes from Triassic to Carboniferous and the outwash ends just south of this transition. At Portapique, the deposit also ends where the hills steepen and the valley narrows, but this occurs closer to the Cobequid Fault within the Parrsboro Formation. The delta at Parrsboro terminates inland where the Parrsboro Gap narrows to its minimum width.

As the valleys generally steepen and narrow as they cross from the sedimentary lowland on to the Cobequid Highlands, the bulk of the deposits lie south of the Cobequid Fault. The most notable exception is the delta at Wards Brook where the Cobequid Fault runs right along the shoreline and the small deposit overlies an eroded part of the Silurian(?) Advocate Succession.

KAME FORMATION

As the ice dissipated, kames and kame terraces formed up valley from the deltas along the sides of the valleys. Topography also controlled the formation of kames which formed in deep valleys that are relatively wide. Kames that might have formed in narrow valleys have since been removed and valleys without sufficient wall height precluded the formation of kames. Kames were deposited up valley from the deltas at Parrsboro, Diligent River and Port Greville. The streams that flowed along the sides of the valleys forming the kames and kame terraces were in many cases major feeders of water and sediment to the deltas at the mouths of the valleys. At Parrsboro, the southward drainage of the streams that deposited the delta extended, on the east side of the valley, almost as far north as Newville Lake. This was before the ice front had receded to Gilbert Lake and formed the recessional moraine, at which point the supply to the Parrsboro River was cut off. Goldthwait (1924, p. 100) alludes to the kame terraces as being part of the deltaic drainage system.

The final deposition of the kame material occurred as the ice retreated up the valleys and left the kames perched along the sides of the valleys. During this phase of deglaciation, meltwater construction of the outwash deposits had ceased and meltwater destruction had commenced.

EROSION OF THE DEPOSITS

As the ice receded up the valleys from the deltas and the outwash plains, postglacial rebound was greater than the rise in sea level and the deposits were emerging. As the deposits emerged, meltwater eroded terraces in them and removed much of the original deposits. The cutting of the terraces may well have commenced before the ice had receded very far from the proximal limits of the deposits. For example, the kettle on terrace 1 at Port Greville means that stream downcutting was in progress before large blocks of ice had melted out of the outwash. At Portapique, there is a kettle on terrace level 2 so the kettle at Port Greville is not an isolated occurrence.

The best developed and most extensive terraces occur at Parrsboro where a large portion of the original delta has been removed. The reason for this is the existence of the Parrsboro Gap. At the other deposits, the valleys generally become narrower and shallower north of the deposits as the drainage divide (in the Cobequid Highlands except at Diligent River and Port Greville) is approached. Because the dissipating ice sat in these valleys, there was a decreasing supply of meltwater to erode the deposits as the ice wasted and receded. At Parrsboro, the valley is much larger and because of the gap in the Cobequids, ice north of the Cobequid Highlands (and thus north of the present drainage divide) supplied meltwater to the delta until the ice had withdrawn to a point north of the recessional moraine. At Spencers Island, there are no terraces and the drainage system north of the delta (Mahoney Brook) is very short. Thus, the amount of terracing was related to the supply of meltwater during uplift. However, even where the terraces are poorly developed or absent, there is a central valley eroded through the deposit in which the former parent stream flows.

Many of the small deposits that occur along the shoreline were largely eroded during the destructional phase and the remnants frequently contain only parts of the delta structure. An example of this is the delta that formerly existed on the east side of the North River at Lower Five Islands. As the delta was uplifted, streams cut down through the delta and removed the foreset beds, leaving glaciofluvial gravel directly overlying clay/sand bottomset beds. Many of the small deposits were probably eroded completely.

While meltwater was responsible for the bulk of the erosion of the deposits, it is unlikely that all of the erosion was accomplished by meltwater. The final phase of the erosion of the valleys in the deposits, and probably of the lowest terraces, occurred during the later stages of postglacial uplift when the supply of meltwater had ceased. The delta at Parrsboro may be the exception as the Parrsboro Gap enabled a larger amount of meltwater to be supplied over a longer period of time. The terrace levels narrow down to the width of the present valley floors and this attests to a decreasing stream power during uplift.

As the streams eroded terraces in the deposits, the eroded material was carried farther seaward and redeposited in the Minas Basin. Marine erosion has removed all but one of these deposits (that at Advocate Harbour), as well as significant portions of the original deposits. Present day erosion rates of up to 1 m/yr or more are attained where the deposits are exposed along shoreline bluffs. Terms such as proximal and distal are used for the exposure at Lower Five Islands but these terms are relative as only the proximal remnants of the deposits remain.

The one area where the outwash sediments that were eroded during postglacial uplift have been preserved is at Advocate Harbour. For this reason, Advocate Harbour is dealt with separately below.

ADVOCATE HARBOUR AREA

During uplift in the Advocate Harbour area, the delta in West Advocate was eroded by McRitchie Brook as it flowed across the delta and by the sea along the front of the delta. The eroded delta sediments, as well as sediment eroded from other sources such as bedrock (and probably till) were deposited on a low lying shoreline platform eroded into the soft Triassic red beds. The shoreline sediments were themselves uplifted and they now form a raised marine plain. The existence of the low lying Triassic bedrock and its location in a down drift position from the delta are the causative factors for the deposition and preservation of the shoreline sediments. Longshore drift also accounts for the topography of the plain. The plain is narrow and steep in the northwest near the delta, where erosion occurred and is broad and gently sloping to the southeast where deposition took place. The low Triassic red beds extend from Advocate Bay to the Minas Basin so at this time the Triassic basalt south of the red beds was an island.

The beach in West Advocate lies west of the delta and in the discussion of the provenance of the pebbles in the beach, it was noted that much of the sediment is from the bedrock outcropping along Cape Chignecto. The beach is a little higher than the beach in Advocate Harbour and thus is likely to be a little older. Formation of the beach probably occurred shortly after the main phase of delta deposition. The beach is slightly higher than the southern edge of the incised terrace on the delta, so the beach may have formed during the initial stages of the terracing.

The raised beach in Advocate Harbour, east of the delta, is slightly lower than the gravel veneered bedrock between Spicer and McRitchie Brooks. If the gravel veneered bedrock is part of the delta that was fluvially eroded to the level of terrace 1, the beach in Advocate Harbour formed during the final stages of, or just after, the terracing event.

Except for the two raised beaches near the top of the raised marine plain, the plain is relatively smooth. Two well developed terraces (2, 3) occur in the dominantly erosional part of the plain but the terraces are not extensive. The lack of well developed constructional beach ridges implies that uplift was fairly rapid and continuous without resting stages that would allow beach ridges to form.

CHAPTER 15. POSTGLACIAL EMERGENCE OF THE MINAS TERRACE

The position of the present sea level relative to the sea level during the Late Wisconsinan deglaciation is dependent upon the amounts of sea level rise and glacial rebound since the time of deglaciation, as well as on possible recent tectonic movements (Wightman and Cooke, 1978, p. 61). Included in recent tectonic movements is subsidence in the regions peripheral to the areas still undergoing postglacial rebound (Walcott, 1972, p. 878). Grant (1970, p. 686) suggests that submergence of the Bay of Fundy area for the last 4000 years is due in part to crustal subsidence and a slow rise in sea level. Thus, east of Lower Five Islands where the terrace is submergent, postglacial rebound has been exceeded by the combined effects of postglacial sea level rise and recent subsidence.

The large tidal range (approximately 16 m) in the Bay of Fundy is a factor in determining the position of postglacial sea levels because below high tide, postglacial sediments are covered by modern intertidal and subtidal sediments. Therefore, 'submergence' and 'emergence' are determined with respect to high tide. Thus, the deposits at Bass River and Portapique may not be submerged relative to mean sea level, meaning that the sea level to which the deposits are graded may lie between mean sea level and high tide. But as this can not be determined, the level of high tide must continue to be used to demarcate submergence and emergence. The tides in the Bay of Fundy have not always been this large as tidal amplification only began about 5000 to 6000 years ago (Grant, 1970; Amos, 1978). The rise in the level of high tide in the Late Holocene is due to tidal amplification and relative sea level rise in roughly equal proportion (Amos, 1978, p. 977).

The deposits east of Lower Five Islands may have been emergent in early postglacial times as suggested by the incised terraces and central valleys (≈10 to 15 m deep) eroded in them. If this was so, then the rate of postglacial rebound exceeded the rate of sea level rise along most of the north shore of the Minas Basin during deglaciation. However, the fluvial downcutting could also be explained by a change in the equilibrium gradients of the streams.

West of, and including, Lower Five Islands the terrace is emergent with the degree of emergence increasing to the west. The proximal to distal decrease in elevation of the deltaic topset/foreset contacts at Lower Five Islands, Port Greville and Spencers Island indicates that the rate of postglacial rebound exeeded the rate of sea level rise both during and after the formation of the deposits in the Late Pleistocene-Early Holocene. This is contrary to the interpretation of Swift and Borns (1967, p. 703) as they state that there was a relative sea level rise during the formation of the terrace.

The marine limits for the emergent deposits are plotted in figure 15.1, using the highest measured elevation of the topset/foreset contact. However, because the contact decreases in elevation from the proximal to the distal part of each delta, the location of the exposure or exposures within each delta affects the value of the apparent marine limit. A proximal exposure will yield a higher value for the marine limit than a distal exposure. This is particularly well illustrated at the Port Greville delta where the elevation of the topset/foreset contact at the

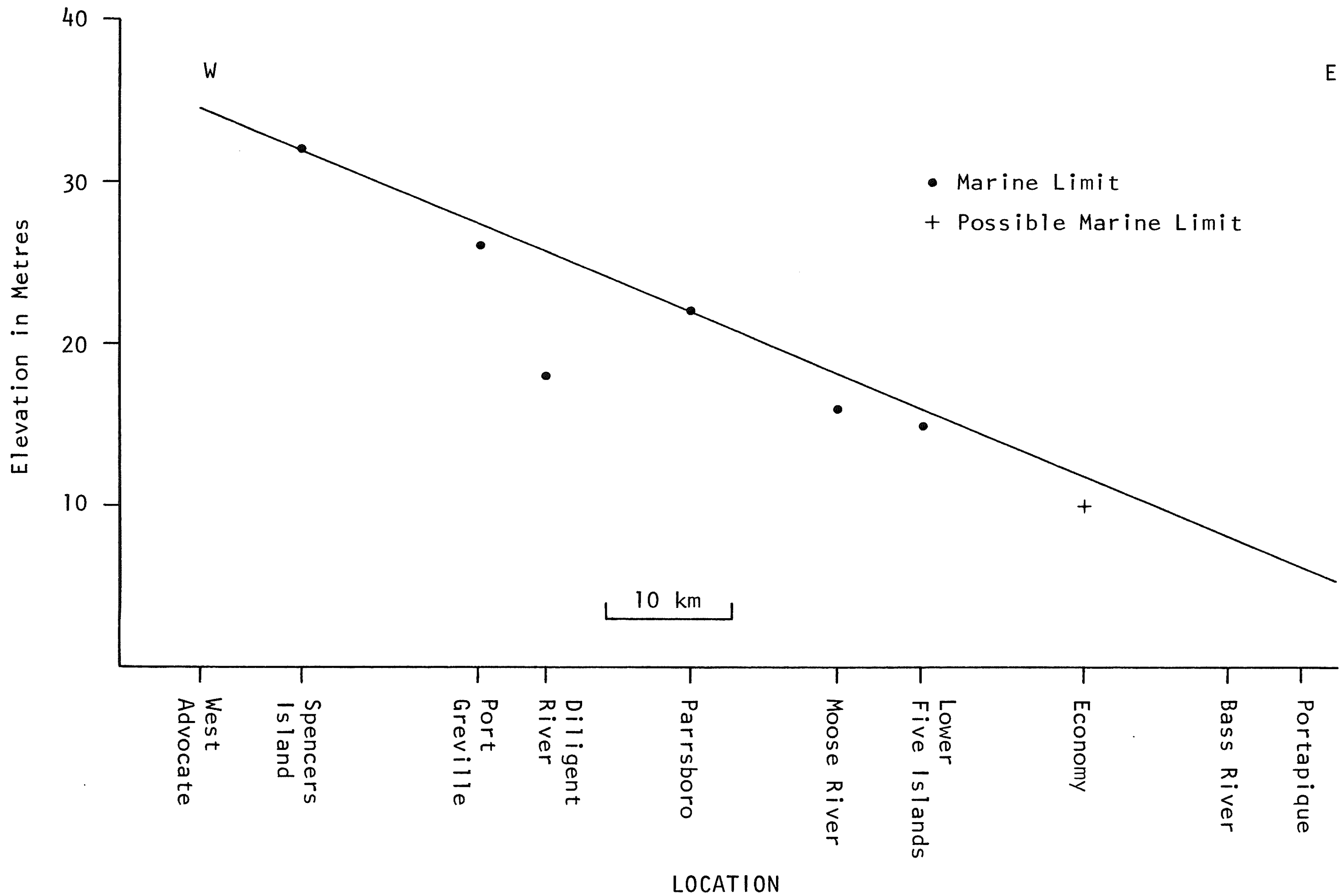


FIGURE 15.1. Emergence data for the deposits in the Minas Terrace.

inland exposure is at least 26 m above sea level and along the coast it is 22 to 19 m in elevation. Continuous exposure is required to establish the true marine limit for each delta. Because the exposures where the marine limits were measured were not from the most proximal parts of the deltas, the line in figure 15.1 is purposely drawn through the two highest points and above the other measured marine limits so as to give the best indication of the probable true marine limit.

The marine limit indicated for West Advocate is almost 35 m, which is higher than the marine limit that would be obtained from the highest raised beach (West Advocate) but could very well be represented by the delta. That the beaches are slightly younger than the delta, and thus lower than the marine limit, has already been suggested in chapter 14. The value for the marine limit at Diligent River seems to be too low but the measurement was taken along a coastal (distal) exposure rather than inland. Furthermore, the exposure was poor and the placement of the contact was uncertain. The relatively high value for the marine limit at Parrsboro suggests that the inland exposure from which the measurement was taken is close to the proximal termination of the foresets. The possible value of 10 m for the marine limit at Economy fits the line quite well and the data predict that no emergent features should exist at Bass River and Portapique, which is the case. The marine limit for Bass River is indicated to be 8 m, which is the same as the level of high tide. Swift and Borns (1967, p. 708) place the marine limit at approximately mean sea level at Bass River but at the present time there is no exposure to verify this. If their marine limit for Bass River is correct, then the line would have to bend downwards at the eastern end of the terrace.

In view of the variability of the elevation of the topset/foreset contact within each delta, the straight line fit of the data in figure 15.1 is surprisingly good. In contrast, Swift and Borns' (1967, p. 709) plot of the marine limit (Advocate Harbour Member) along the terrace forms a concave up curve (Fig. 15.2). The fact that the marine limits used in this thesis were derived from the deltaic topset/foreset contacts while Swift and Borns (1967) used a contact within the topset facies is the probable reason for the discrepancy.

The differential emergence along the Minas terrace could be due to differential rebound (greatest to the west), differential subsidence (greatest to the east) or both. 'Late' ice in the eastern part of the Minas Basin has also been proposed as the cause of the differential emergence but there is no geologic evidence for an eastward deglaciation and Wightman and Cooke (1978, p. 62) believe that the length of time required for the persistence of 'late' ice is unreasonable. Wightman and Cooke (1978, p. 64) conclude that the increase in emergence from east to west along the Minas terrace is related to a centre of uplift in southwestern New Brunswick and this in turn must be related to former ice thicknesses. Because the zone of peripheral subsidence follows the zone of rebound as it migrates towards the centre of uplift, the eastern part of the Minas terrace may well have undergone more subsidence than the western part. Thus, it is probable that differential uplift and differential subsidence have combined to produce the differential emergence of the Minas terrace.

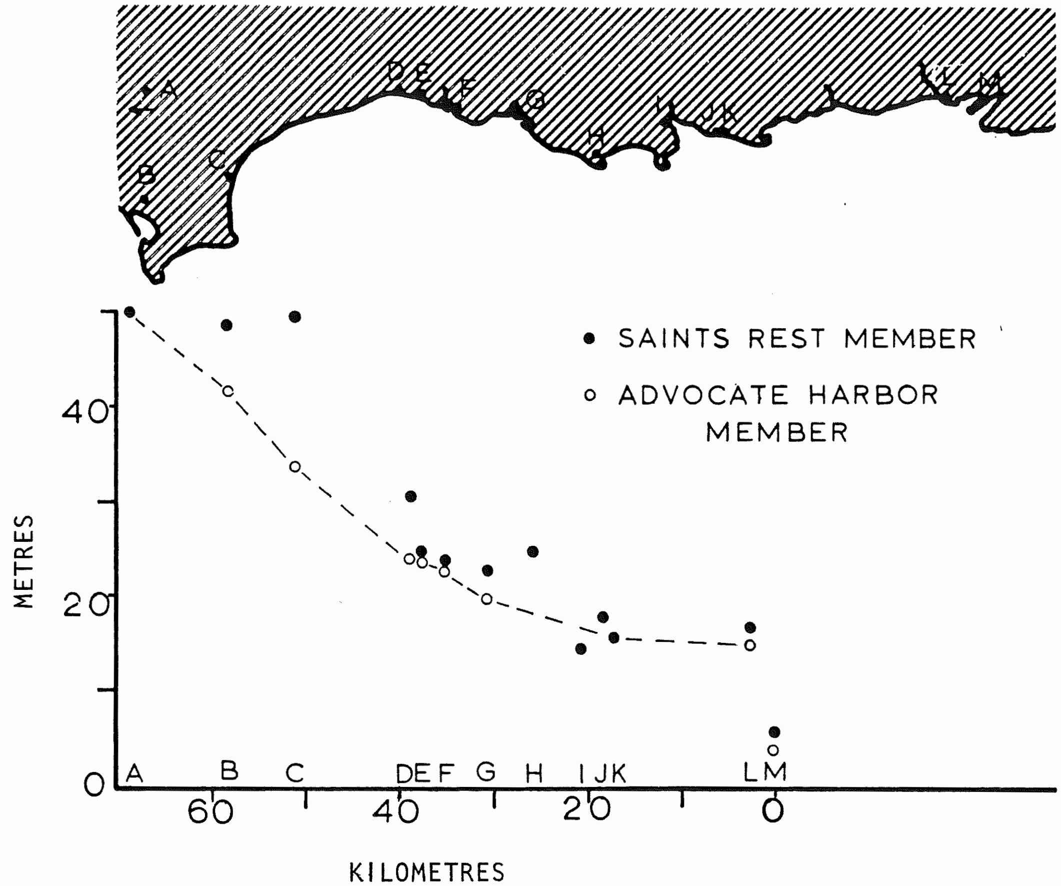


FIGURE 15.2. Plot of terrace elevations from the Minas Basin's north shore, against east-west component of distance between stations. (Modified from Swift and Borns, 1967.)

MARINE LIMITS

It must be emphasized that the deltas in the Minas terrace were the first sediments to be deposited on the shoreline as the ice front receded inland. Raised beaches and other glaciolittoral sediments were formed by sediment eroded mainly from the deltas, but also from till and bedrock, during offlap conditions caused by the uplift of the shoreline. Thus, these features should not be used to obtain the marine limit in this area, although the highest beaches might be close to the marine limit.

Secondly, the best approximation of the marine limit is given by the height of the topset/foreset contact in the proximal part of the deltas. Large errors can result if the heights of the delta plains are used for marine limits, as shown by the delta at Parrsboro. The inland portion of the delta plain, where it may be strictly glaciofluvial, attains a height of over 46 m while the marine limit as indicated by the topset/foreset contact is only 22 m.

CHAPTER 16. DATING AND HISTORY OF DEGLACIATION OF THE BAY OF FUNDY

PREVIOUS WORK

Prest and Grant (1969, p. 3) suggest that at the Late Wisconsinan glacial maximum (about 19,000 yr BP), ice filled the Bay of Fundy and extended across mainland Nova Scotia and on to the Scotian Shelf. During deglaciation, the ice thinned and receded and at about 14,000 yr BP, a marine incursion into the Bay of Fundy was in progress (Prest and Grant, 1969, p. 7). Prest (1969) maps the marine incursion as having reached Cape Chignecto by 13,500 yr BP and by 13,200 yr BP all but the heads of the Minas Basin and Chignecto Bay were ice free.

Earlier work by Lee on the St. John River valley agrees with this picture as a date on marine shells found in postglacial shoreline sediments near St. John (Fig. 16.1) shows that deglaciation was in progress in this area at about 13,325 ± 500 yr BP (I[GSC]-7; Walton *et al.*, 1961, p. 50). Lee interprets the shell bearing sediment as part of an outwash delta that was deposited in close proximity to the melting glacial ice. From his description of the sediment, "Clay lies beneath and intertongues with delta outwash gravels," it would seem likely that the shells were found in the bottomset facies of the delta.

Welsted collected marine shells from postglacial sediment at Sheldon Point, near the site where Lee had collected his material, and they gave a radiocarbon age of 13,200 ± 200 yr BP (GSC-965; Lowdon and Blake, 1970, p. 55). Welsted's description and comments corroborate the interpretation by Lee that the shell bearing sediment is part of an outwash delta formed during deglaciation of the shoreline.

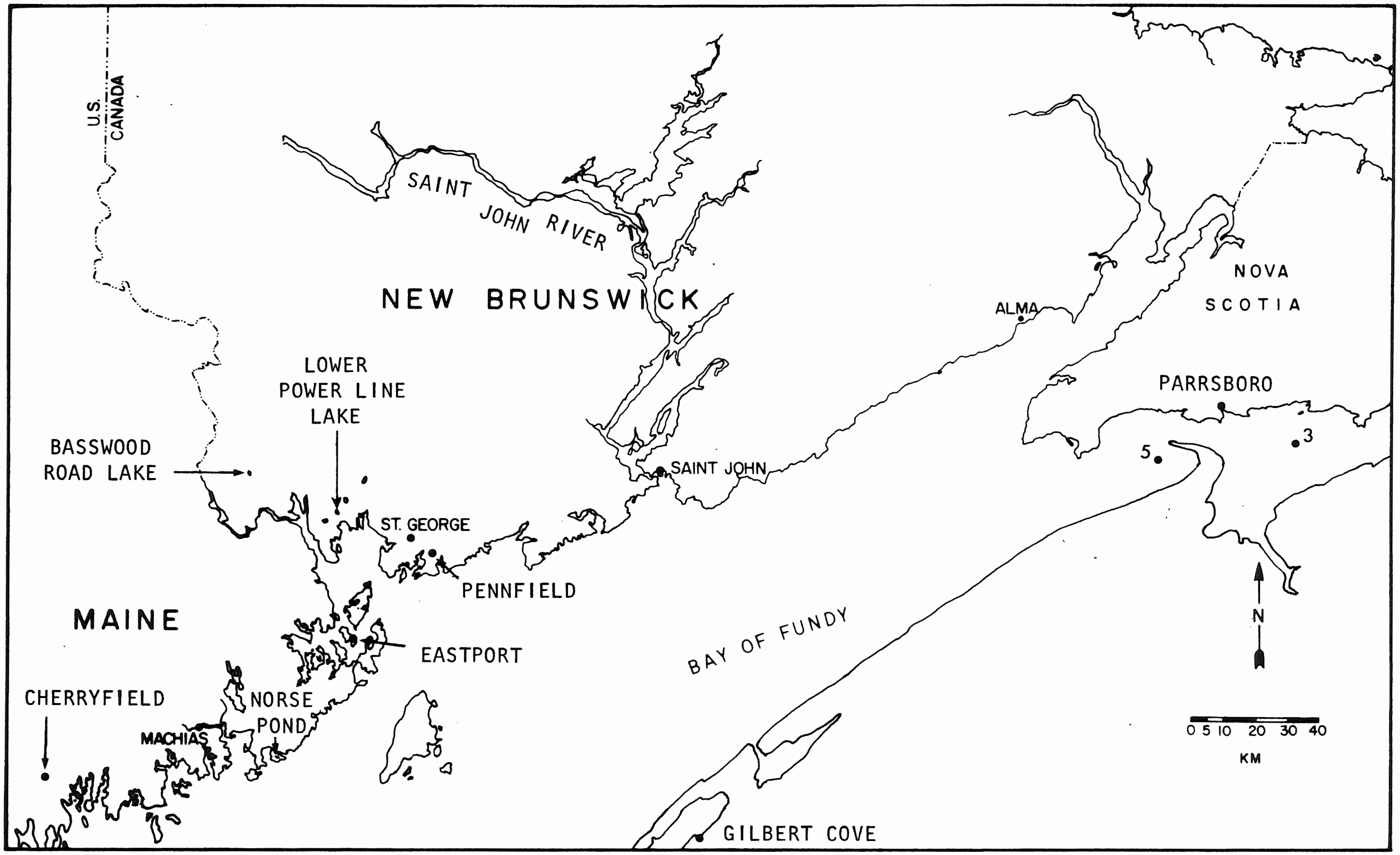


FIGURE 16.1. Location map for dated postglacial deposits in New Brunswick and Maine. Numbers give approximate locations of Amos' (1978) vibrocores in Minas Channel and Minas Basin. (Modified from Scott and Medioli, 1978.)

Gadd (1973) studied the glacial features in the southwestern part of New Brunswick and mapped a series of marine features along the shoreline from St. John to the Maine-New Brunswick border, including the delta described by Lee and Welsted. Gadd's description of the deposit at Sheldon Point is similar to that of Welsted and Lee but he interprets it as a bar or spit developed on a submerged moraine (Lowdon *et al.*, 1971, p. 267) and uses the general term "wave modified moraine" on his map (Gadd, 1973, map 12-1971). Gadd collected marine shells from a level in the clayey part of the deposit higher than that of Lee and Welsted and the shells yielded a date of 13,000 ± 170 yr BP (GSC-1340; Lowdon *et al.*, 1971, p. 267). Thus, the dates on the deposit at Sheldon Point are very consistent within a range from 13,000 to 13,325 yr BP.

Gadd's (1973, p. 15) description and photograph of the deposit at Sheldon Point are consistent with that of a Gilbert type delta but he interprets it as a wave modified kame because in one exposure, the deltaic sediments overlie a ridge of till (Gadd, 1973, p. 26). Based on the descriptions of Lee (Walton *et al.*, 1961, p. 50), Welsted (Lowdon and Blake, 1970, p. 55), Gadd (1973) and the photographs by Gadd (1973, p. 15, 16), the author agrees with Lee and Welsted that the deposit at Sheldon Point is a marine Gilbert type delta. Apparently, in some areas moraine material was deposited by an ice front situated along the shoreline and as the ice front receded inland, the delta prograded over the moraine system. If the ice front was close to the delta during deposition, as suggested by Welsted and Lee and by the ice margin drawn for the "Pennfield Phase" by Gadd (1973, Fig. 12), the dates on the delta give the age of the deglaciation of the shoreline in this area. As

Gadd (1973, p. 26) indicates, the interpretation of the deposit as a reworked moraine means that the dates might not be related to the deglaciation of the shoreline as the reworking might have occurred during the later marine regression caused by postglacial rebound. However, the present author believes that the moraine material is overlain by a delta and the dates are related to deglaciation of the shoreline.

Marine shells collected from the bottomset beds of the delta at Pennfield gave a date of 13,000 ± 240 yr BP (GSC-882), but Gadd (Lowdon and Blake, 1970, p. 55) is unsure whether the shells are related to the deposition or to the later terracing of the delta. Gadd does not state whether the terracing is marine or fluvial but if the terracing is fluvial as it is in the Minas terrace, the marine shells would not be from the terracing phase. Another factor which favours the shells belonging to the constructional phase (versus destructional) of the delta is that the shells were *Portlandia* sp. It has already been stated in the thesis that the optimum habitat for *Portlandia arctica* is off the mouths of meltwater rivers or off glacier fronts where large quantities of silt and clay are being deposited. Thus, it appears that the shells, and therefore the date, are related to the deposition of the delta.

There are also two younger dates of 12,300 ± 160 yr BP (GSC-795; GSC-886) on marine shells from localities near Pennfield. But as Gadd (Lowdon and Blake, 1970, p. 56) suggests, the dates are probably related to marine offlap conditions that followed the postglacial marine maximum.

In summary, between 13,000 and 13,300 yr BP, an ice front sat just inland of the present Fundy coastline in southwestern New Brunswick and meltwater deposited outwash deltas in the sea. The close proximity of ice during deposition and the structure of the deltas are similar to the deltas in the Minas terrace, but some of the New Brunswick deltas prograded over previously deposited moraines. The moraines suggest that the ice front along the New Brunswick shoreline was more active than the ice front along the north shore of the Minas Basin. The deltas mark the marine limit which, in New Brunswick, attains a maximum of approximately 70 m at Pennfield (Gadd, 1973, p. 16).

The deposits along the Fundy shore of New Brunswick have been referred to as a "moraine system" (Swift and Borns, 1967, p. 703; Mott, 1975, p. 275) but because outwash deltas are an integral part of the system, the term "moraine" is somewhat misleading. A better term might be "delta-moraine system." Swift and Borns (1967, p. 703) and Borns and Hughes (1977, p. 204) correlate the Pineo Ridge (or Cherryfield-Eastport) "moraine" system in Maine with the delta-moraine system in New Brunswick, although Borns and Hughes (1977) give the age of formation of the Pineo Ridge as 12,700 yr BP.

A date of 14,100 ± 200 yr BP (GSC-1259; Grant, 1971, p. 113) on seaweed detritus from Gilbert Cove in southwestern Nova Scotia (Fig. 16.1) corroborates Prest and Grant's (1969) interpretation of a marine incursion into the Bay of Fundy at about 14,000 yr BP. However, the date also suggests a time lag in the deglaciation of the southwestern New Brunswick shoreline.

Grant (1977, p. 352) has a slightly different interpretation of the Late Wisconsinan glaciation of the Bay of Fundy as he believes that the bay was not fully glacierized. Grant (1977, p. 253) believes that glacial ice from New Brunswick flowed southwestward down Chignecto Bay while the main part of the bay, including the Minas Basin, remained essentially marine. The ice margins for the Late Wisconsinan maximum are drawn along, or just seaward of, the present shorelines and the Cobequid Highlands are shown as nunataks. However, Grant (1977, p. 255) retains the interpretation that the outwash deposits fringing the Bay of Fundy mark prominent ice stand positions during deglaciation.

The pebble lithology data presented in chapter 13 show clearly that the ice which melted to form the Minas outwash terrace had crossed the Cobequids, and it had crossed them from the north. It will be shown in the latter part of this chapter that the ice melted at the end of the Late Wisconsinan and the author infers from this that the ice itself was Late Wisconsinan in age. Other deglacial features, such as the esker at an elevation of about 200 m near Sutherland Lake in the Cobequids north of Portapique (Fowler and Dickie, 1978) are therefore also probably Late Wisconsinan in age. The ice which crossed the Cobequids had to come from New Brunswick and therefore had to fill Chignecto Bay.

Fader *et al.* (1976) studied the glacial and postglacial sediments in the Bay of Fundy and interpreted the extensive till sheet as a grounded ice sheet deposit of Late Wisconsinan age. They also state that the till has probably never been subaerially exposed. More recent work has shown that the end moraine complex that lies about 30 km off the Atlantic coast of Nova Scotia was formed by a grounded ice sheet at about 26,000 years ago (King, 1979). As the moraine system extends into the Gulf of Maine at the mouth of the Bay of Fundy, this is confirmatory evidence that the Bay of Fundy was full of ice during the Late Wisconsinan. Thus, the general pattern of deglaciation of the Bay of Fundy presented by Prest and Grant (1969) is preferred to that of Grant (1977).

RESULTS OF THESIS DATING MEASUREMENTS

At present, there is no convenient way to date the deltas in the Minas terrace directly as the carbonate shells of the marine pelecypods have been completely removed by leaching. There is hope of obtaining a date on the deltas in the future as periostracum material from the molds and casts has been collected and advances in the methods of radiocarbon dating now make possible the dating of very small samples (Litherland, 1979, p. 80).

In order to obtain a minimum age for the terrace, freshwater ponds and lakes that are associated with the deltas were cored and the basal organic material was dated. Palynological studies are presently being carried out on the cores by Peta Mudie and J.G. Ogden of Dalhousie University and by R.J. Mott at the Geological Survey of Canada in Ottawa. Preliminary results, furnished by Dr. Mott and Peta Mudie, are available for the lower parts of some cores. The dates and a discussion of the preliminary pollen analyses, where available, are presented in the following pages for each of the freshwater bodies that were cored. A description of the cores and the available palynological analyses are presented in Appendix 3.

Samples were taken from some of the cores for diatom analysis to ensure that the ponds and lakes have been freshwater throughout their postglacial history. Andrew Palmer, presently with Environment Canada, Halifax, identified the diatoms.

LEAK LAKE

Leak Lake was first cored in the summer of 1977 by Dr. J.G. Ogden but glacial sediments were not penetrated. Because of the low organic content of the core (Leak Lake 1), it was decided that several short cores should be taken in the lower part of the Holocene sediment column so that the cores could be matched stratigraphically and combined to give a basal sample for dating. Accordingly, the author took cores Leak Lake 2, 3 and 4, in approximately 13 m of water, and each core started in the lower part of the gyttja and ended in inorganic red sediments (sand, silt and clay). The first date obtained from combining samples from cores Leak Lake 2 and 3 was interesting so another core (Leak Lake 5) was taken that extends from the water/sediment interface to the inorganic red sediment. This is the only core that contains the complete gyttja sediment column from Leak Lake.

Figure 16.2 shows a composite description of the lower part of the Leak Lake cores including the dated intervals. The oldest date (15,900 ± 1200 yr BP; GSC-2880) was obtained by combining the basal 5 cm of gyttja, immediately overlying the inorganic red sediment, from cores 4 and 5. A sample from cores 2 and 3, comprising the basal 12 cm of organic sediments immediately above the inorganic red sediments, was dated at 9580 ± 310 yr BP (DAL-313). A sample of the 5 cm of organic sediment immediately above this from cores 2 and 3 yielded a date of 12,900 ± 160 yr BP

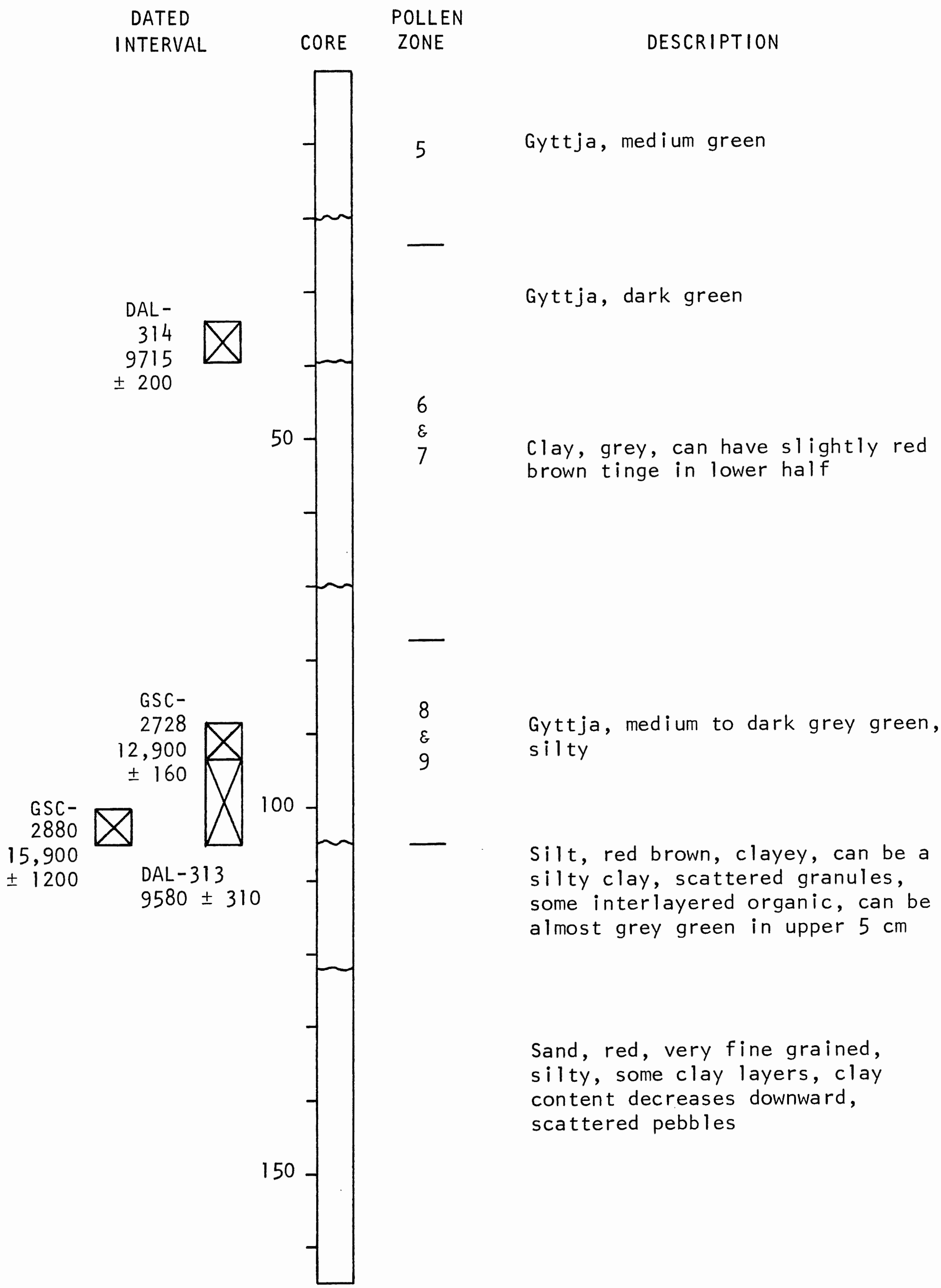


FIGURE 16.2. Composite sketch and description of lower part of Leak Lake cores. Core depth in centimetres.

(GSC-2728). Gyttja immediately overlying a grey clay layer, about 65 cm above the inorganic red sediments, from cores 2 and 3 was dated at 9715 ± 200 yr BP (DAL-314).

The dates from the two laboratories (Dalhousie and the Geological Survey of Canada) are widely discrepant and the two Dalhouse dates (313, 314) are not consistent with their stratigraphic position. The preliminary pollen report on core Leak Lake 4 (GSC Palynological Report No. 79-13, R.J. Mott) sheds some light on the reasonability of the dates. The report will only be discussed briefly here, but it is contained in full in Appendix 3.

The red silt and clay below the gyttja has a very low pollen concentration. This fact, combined with the low organic content (1 to 2%) and the increasing percentage of sand and abundance of scattered clasts with depth, indicate that the red sediment is glacial in origin. It has been interpreted previously in this thesis that Leak Lake was occupied by a large block of ice during the formation of the Parrsboro delta and thus the red sediment probably represents the glacial material that settled out of the water as the lake formed.

The basal gyttja overlying the glacial sediment contains an herbaceous tundra type of pollen assemblage (pollen zones 9 and 8; Mott, 1975), the limit of which is shown in figure 16.2. Above this, spruce and birch tree pollen become prominent (pollen zones 7 and 6) and above the grey clay layer, pollen from white pine trees becomes abundant (zone 5).

Mott (GSC report 79-13) states that the pollen stratigraphy at Leak Lake is amazingly similar to the Basswood Road Lake pollen sequence in southwestern New Brunswick (Mott, 1975). Accordingly, the chronology of the Basswood Road Lake pollen sequence can be used to check the radiocarbon dates at Leak Lake, although the zones may not be of precisely the same age. On this basis, Mott states that the date of 9580 ± 310 yr BP (DAL-313) is far too young for a tundra type vegetation as the zone 9-8 boundary at Basswood Road Lake was dated at 12,600 ± 270 yr BP (GSC-1067). Radiocarbon date GSC-2728 (12,900 ± 160 yr BP) occurs approximately at the zone 9-8 boundary so agrees well with, but is slightly older than, the date at Basswood Road Lake. However, date GSC-2880 (15,900 ± 1200 yr BP) is much older and Mott (personal communication, 1979) feels that it is unreasonably old.

There are also relatively old dates from lake sediment cores taken in southwestern New Brunswick. Marly sediment from the bottom of a core taken in Little Lake, located on the edge of the Pennfield delta in New Brunswick, dated 16,500 ± 370 yr BP (GSC-1063). A date of 14,300 ± 270 yr BP (GSC-1272) was obtained from noncalcareous sediment 29 cm above the sediment used for date GSC-1063. Mott (1975, p. 279) believes that the dates from Little Lake are spuriously old when compared to the Basswood Road Lake dates. Mott's interpretation is corroborated by the fact that shells from the bottomset beds of the Pennfield delta yielded a date of 13,000 ± 240 yr BP (GSC-882). As Little Lake is a kettle lake that occurs on top of the delta, the lake sediments should be younger than the delta by at least the amount of time it took for the ice in the kettle to melt.

Mott (GSC report 79-13) warns that 'old' dates must be viewed with caution in areas where carbon contamination is possible. The Upper Carboniferous rocks of the Cumberland Lowlands north of the Cobequids contain abundant coal seams which are possible sources for such contamination. The possibility of contamination is reinforced by the fact that Paleozoic palynomorphs and coal fragments were identified by Peta Mudie in a clay sample from the bottomset facies of the delta at Spencers Island. Till or other glacial sediment containing fragments of old carbon have possibly contaminated Leak Lake in the past, and perhaps to the present day.

Mott (GSC report 79-13) attaches special significance to the clay layer that occurs in the lower part of the Leak Lake gyttja because a similar clay layer occurs in the lower part of the Basswood Road Lake core. Mott (personal communication, 1979) has also found a similar mineralic layer in the basal part of other cores from New Brunswick. He believes that the mineralic layer may be the result of climatic deterioration between 10,000 and 11,000 yr BP with a reduced vegetation cover that resulted in greater erosion and hence greater sediment input into the lakes. More detailed pollen studies and dates are required to test this hypothesis.

Two samples, one 6 cm below and one 14 cm above the grey clay in the basal gyttja, were taken from core Leak Lake 3 for diatom analysis. The diatoms are a freshwater assemblage and are shown in figures 16.3 and 16.4. The basal gyttja between the red glacial sediment and the grey clay layer has a very high concentration of diatoms.

FIGURE 16.3. Diatom from the base of core Leak Lake 3.
1. *Melosira arenaria ?*

1

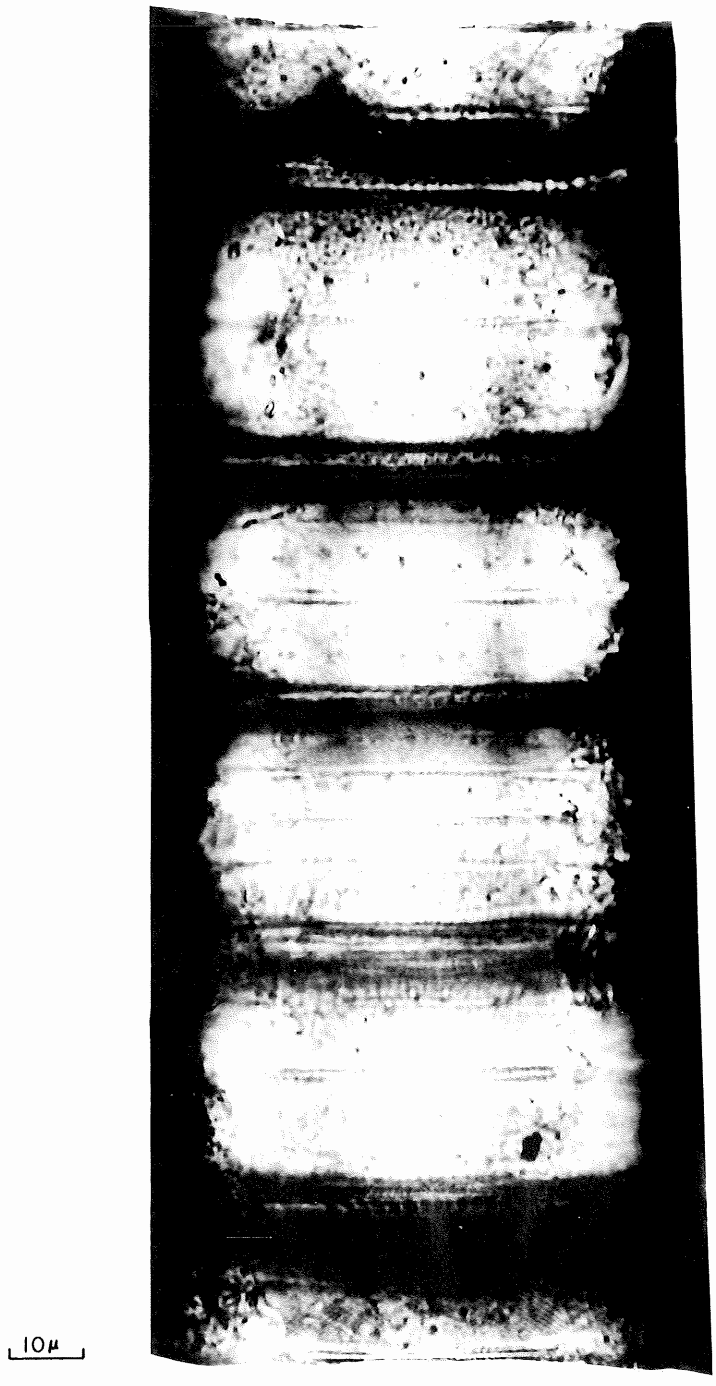
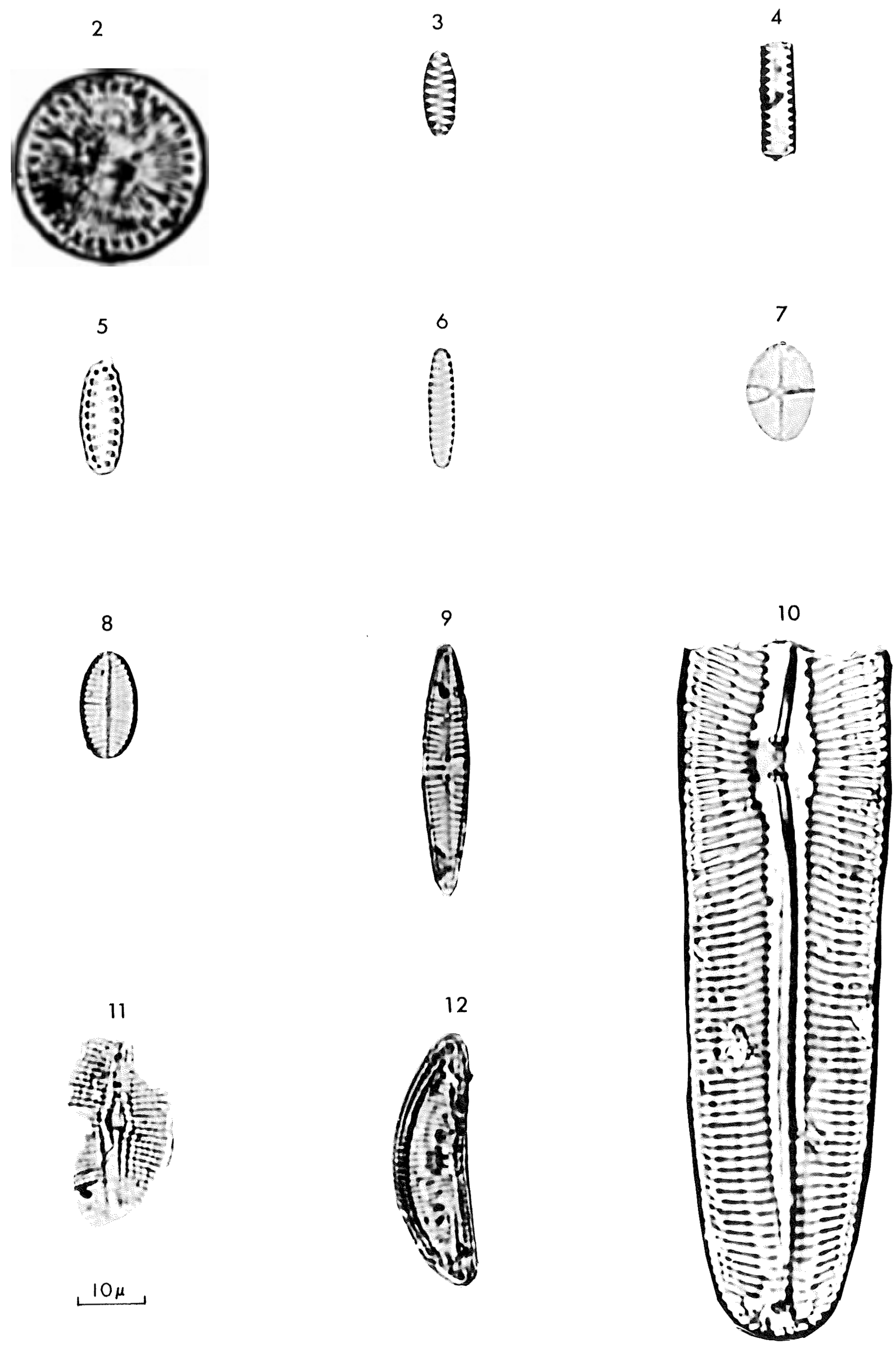


FIGURE 16.4. Diatoms from the base of core Leak Lake 3. 2. *Cyclotella compta*, 3. *Opephora martyi*, 4. *Fragilaria leptostauron*, 5. *Fragilaria leptostauron* var. *dubia*, 6. *Fragilaria pinnata*, 7. *Acnanthes calcar*, 8. *Cocconeis diminuta*, 9. *Rhiocosphenia curvata*, 10. *Pinnularia maior*, 11. *Diploneis elliptica*, 12. *Cymbella hebridica*.


2
3
4
5
6
7
8
9
10
11
12
10μ

DILIGENT RIVER

The westernmost kettle pond on the east side of the Ramshead River estuary was cored and the pond turned out to be surprisingly deep (≃11 m). Two complete cores, Diligent River 2 and Diligent River 3, were taken in the pond as well as a second core sample (Diligent River 2_A) of the basal organics that was taken at the same time as core Diligent River 2. Because the second core sample was taken through the same hole in the ice (coring was done in the winter), it was checked for contamination by the author and by Dr. J.G. Ogden and appeared to be good, so was later used for dating. All of the cores bottomed in an impenetrable, pebbly green clay.

Core Diligent River 2 penetrated 555 cm of gyttja and the basal 7 cm of the core yielded a date of 11,555 ± 230 yr BP (DAL-301) while core sample Diligent River 2_A (basal 17 cm) yielded a date of 10,120 ± 200 yr BP (QU-764). The discrepancy between the dates can be partly explained by the fact that the basal 7 cm were used for date DAL-301 while the basal 7 cm plus the overlying 10 cm were used for date QU-764.

The preliminary pollen report (GSC Palynological Report No. 79-15) on core Diligent River 3 shows that the oldest pollen assemblages of the Leak Lake site are not represented in the Diligent River core. The spruce pollen maximum begins about 5 cm above the impenetrable clay and Mott (GSC report 79-15) feels that a date of 11,555 ± 230 yr BP (DAL-301) is consistent with this pollen assemblage as the spruce pollen maximum is dated about 11,300 yr BP in New Brunswick. The mineral sediment layer is not present in the Diligent River core but Mott suggests that this phenomenon may not occur at every site. Another factor is that the kettle pond at Diligent River does not have a stream feeding into it and thus the mineralic layer, which is dependent upon sediment input from outside the body of water, was precluded from forming.

A sample from the basal part of core Diligent River 2 (at a depth of 533 cm) was taken for diatom analysis and the following diatoms, all with a fresh water affinity, were identified: *Epithemia* sp., *Pinnularia* sp., *Surirella* sp., *Cymbella ehrenbergii(?)* and *Melosira(?)* sp.

WARDS BROOK

The kettle pond on the delta in Wards Brook was cored, although the pond is only about 1.5 m deep. The core (136 cm long) bottomed in a light green pebbly clay and a sample from the organic sediment immediately overlying the clay (117 to 122 cm) dated 3330 ± 178 yr BP (DAL-315) and a sample just above that (107 to 112 cm) dated 4390 ± 70 yr BP (GSC-2730). Thus, the dated sediment accumulated during the Holocene and the impenetrable sediment at the bottom of the core is probably not glacial.

GILBERT LAKE

A 951 cm core was taken in the northern part of Gilbert Lake in about 10 m of water. It did not reach glacial sediment or an impenetrable layer, but the thickness of the gyttja made further coring impossible. A sample of the gyttja from 5 to 15 cm above the base of the core dated 3595 ± 105 yr BP (DAL-309) while a sample immediately above this dated 5640 ± 220 yr BP (QU-765). The relatively young age for this thickness of sediment means either that the sedimentation rate has been extremely high or that there has been slumping of sediments into the deeper parts of the lake. The author believes that it is probably a combination of both.

WEST BROOK

A date of 13,365 ± 420 yr BP (DAL-300) was obtained from a sample from the basal part of a peat layer that underlies fluvial sediments and overlies impenetrable coarse sediments on the West Brook floodplain, the stratigraphy of which has been given in chapter 12. The pollen assemblage of the peat sample suggested sparsely treed tundra conditions. However, in comparison with other pollen assemblages from Nova Scotia, the date is older than expected (Mott, GSC report 78-11). A second sample of the peat yielded a date of 9830 ± 100 yr BP (GSC-2772) but the sample was probably not from the basal part of the layer and the pollen assemblage suggests deposition within the spruce pollen maximum zone, with which the date is compatible (Mott, GSC report 79-2). The pollen assemblages of the two samples suggest that there should be a difference in age, but a difference of about 3000 years is somewhat large. As the coal bearing Carboniferous rocks underlie this area, carbon contamination of the basal part is a possibility. Nevertheless, it seems probable that the basal peat is at least 10,000 years old.

SUMMARY

The dates from Leak Lake obtained from the Geological Survey of Canada radiocarbon dating lab are similar to the dates from Little Lake, which is situated on the outwash delta at Pennfield in New Brunswick, and both groups of dates appear to be anomalously old. The Dalhousie dates on the Leak Lake sediment are spuriously young. The dates from Diligent River are on sediment that was deposited some time after deglaciation as the oldest pollen zones (tundra type vegetation) were not represented in the cores. Therefore, the dating of the lake sediment is inconclusive in establishing a definitive minimum age for the Minas terrace, but combined with the pebble lithology data, it does indicate strongly that the Late Wisconsinan ice crossed the Cobequids from the north, contrary to Grant's (1977) hypothesis.

The pollen sequence and stratigraphy of Leak Lake are amazingly similar to that at Basswood Road Lake and because of this, Mott (personal communication, 1979) believes that the two pollen sequences, and thus the formation of the lakes, are roughly synchronous. As Basswood Road Lake is located near the New Brunswick outwash deposits that have been dated at 13,000 to 13,325 yr BP, a similar time lag might be expected between the Minas terrace and Leak Lake. If this is so, then the Minas terrace was also deposited at about 13,000 to 13,300 yr BP, which is consistent with the interpretations of Borns (1965, p. 1225; 1966, p. 52) and Swift and Borns (1967, p. 703).

Stuiver and Borns (1975, p. 101) found an average time lag of about 800 years for samples from kettles and they suggest that time lags between dates on basal sediment from kettles and dates on shells from the deposits themselves are caused by: 1. the time required for the ice to melt, 2. the amount of time required for plants to invade the area, and 3. a low initial sedimentation rate. The time lag for lakes might be less as the ice is not buried by outwash and hence could melt faster. Thus, the 400 to 700 year time lag for the basal sediment from Basswood Road Lake seems reasonable and could also apply to Leak Lake and the Minas terrace.

The West Brook area must have been deglaciated after the Minas shoreline, as ice retreated northward through the Parrsboro Gap and possibly northeastward along the West Brook valley. Thus, the expected age of deglaciation in this area is likely to be slightly younger than 13,365 ± 420 yr BP (DAL-300) but certainly older than 9830 ± 100 (GSC-2772). Chignecto Peninsula was ice free before 11,000 yr BP as Indians had established campsites at Debert by 10,600 yr BP (MacDonald, 1968).

RECENT WORK IN THE MINAS BASIN

Amos (1978) has studied the recent sediments in the Minas Basin and took several vibrocores (Fig. 16.1) which, along with seismic records, enabled him to establish the postglacial stratigraphy of the basin. The vibrocores bottomed in glacial sediments that are easily distinguished from the overlying Holocene silts and clays. The glacial sediments are red brown, massive clayey silts that are well compacted while the overlying Holocene silts and clays are dark brown, laminated and loosely compacted. The contact between the two units is distinct. Molluscan and foraminiferal faunas show that the glacial sediment and the Holocene sediment were both deposited under marine conditions.

Glacial sediment from the base of core 3, south of Five Islands, yielded a date of 14,180 ± 710 yr BP (GX-4514) while glacial sediment from the base of core 5 in Scots Bay yielded a date of >37,000 yr BP (GX-4522). Apparently Late Wisconsinan glacial and deglacial sediments are absent from Scots Bay. However, the author believes that the date on the glaciomarine sediments within the Minas Basin provides unexpected confirmatory evidence on the age of the Minas terrace.

When the meltwater streams were depositing the deltas along the north shore of the Minas Basin, some of the finer sediment (silts and clays) would have been deposited farther out in the bay. In effect, the zone of massive silty clay deposition seaward of the 'suspension fall out' zone should have extended well beyond the deltas, forming a blanket of glaciomarine sediment on the floor of the bay. Dropstones would also be deposited in this glaciomarine sediment and the deposit should thin away from the meltwater sources. The similarity in structure and texture of the glaciomarine sediments described by Amos (1978) and the massive silty clays below the varves in the deltas suggests that the glaciomarine sediment found by Amos is indeed an offshore equivalent of the Minas terrace. As the date of 14,180 ± 710 yr BP (GX-4514) comes from the top of the glaciomarine sediment, this should represent the last supply of meltwater into the Minas Basin. However, several dates from different localities are needed to substantiate the exact timing of the termination of meltwater inflow. Until more data exist, it seems reasonable to assume a general correlation of the date with the delta building event along the north shore of the Minas Basin.

The date (GX-4514) has a rather large error margin but even when this is taken into consideration, it suggests that the Minas terrace is slightly older than the proglacial deposits in southwestern New Brunswick. Consideration of the deglacial events in the Bay of Fundy support this concept.

It was suggested in chapter 15 that there is no age difference from west to east along the Minas terrace and therefore Prest's (1970) interpretation of 'late' ice in the east part of the basin is not accepted. The author believes that the marine incursion into the Bay of Fundy occurred very rapidly with the head of the bay, including the Minas Basin and Chignecto Bay, becoming ice free shortly after the central part. This separated the ice on mainland Nova Scotia from the ice over New Brunswick and left a small, isolated ice cap on the Chignecto Peninsula. Because of the rapid intrusion of the sea, the ice fronts were oversteepened and the ice thinned by drawdown into the Bay of Fundy. Shoreline outwash deposits would only be formed after the period of outflow (drawdown) was over and a particular ice margin achieved a stable position along the coast. The ice cap over the Chignecto Peninsula was much smaller than the ice cap over New Brunswick, and the ice in New Brunswick was connected to the main Laurentide ice sheet until about 12,800 yr BP (Borns and Hughes, 1977, p. 204). Therefore, the period of ice outflow along the north shore of the Minas Basin was probably shorter than for the southwest coast of New Brunswick, with the result that the Minas terrace may have been formed slightly before such deposits as the Pennfield delta.

This concept of a longer period of ice outflow in southwestern New Brunswick is also partly supported by the date of 14,100 ± 200 yr BP (GSC-1259) on seaweed detritus at Gilbert Cove. Gilbert Cove lies almost directly across the bay from the delta at Pennfield and yet the shoreline was deglaciated about 1000 years before the shoreline at Pennfield. The similarity between the date from the Minas Basin (GX-4514) and the date at Gilbert Cove (GSC-1259) suggest that the entire Bay of Fundy, except for some ice infringed shorelines, may have been marine by about 14,000 yr BP. If this is so, the marine incursion started some time before this.

It must be noted that glaciomarine sedimentation would occur in the Minas Basin from the time of the initial marine incursion and therefore the earliest glaciomarine sediment could be deposited during the period of ice drawdown and outflow. As stated previously, the date of 14,180 ± 710 yr BP (GX-4514) was obtained from the top of the glaciomarine sediments so it should have the same age as the formation of the Minas terrace. A date on the basal glaciomarine sediments should predate the formation of shoreline sediments (such as the Minas terrace) and give the age of the marine incursion into a particular part of the bay.

RATE OF EMERGENCE OF SOUTHWESTERN NEW BRUNSWICK AND THE MINAS TERRACE

David Scott and Franco Medioli of Dalhousie University cored a lake ("Lower Power Line") in southwestern New Brunswick, which has an elevation of 36 m, well below the approximate marine limit of 70 m for the area (Gadd, 1973, p. 16). A sample from the transition zone between marine sediment and the overlying freshwater sediment yielded a date of 12,385 ± 375 yr BP (GX-6053; Scott and Medioli, 1979). They also obtained a core from Norse Pond (elevation 39 m) which is near Passamaquoddy Bay but is in Maine; shell fragments from the top of the marine interval yielded a date of 12,175 ± 120 yr BP (SI-1048; Scott and Medioli, 1978, p. 57). As the nearby outwash deltas in New Brunswick, at elevations near 70 m, have been dated at 13,000 to 13,325 yr BP, almost one half of the total present emergence apparently occurred in only 1000 years.

This gives an emergence rate over that time period of roughly 3.5 m/ century.

While this seems to be exceedingly rapid, Stuiver and Borns (1975, p. 100) suggest that up to 135 m of emergence occurred along the coast of Maine from 13,000 to 12,100 yr BP. Scott and Medioli's (1978; 1979) dating of Norse Pond and Lower Power Line Lake show that emergence was not that rapid in northeastern Maine and southwestern New Brunswick and it is unlikely that it was that rapid over the rest of the Maine coast.

The rate of emergence of the Minas terrace is not known but if it is similar to the emergence rate that apparently existed in southwestern New Brunswick it was very rapid. The amount of emergence of the Minas terrace is only half that of southwestern New Brunswick on the west end of the terrace and emergence decreases until the east end of the terrace is submergent. If the rate of postglacial rebound is dependent upon the total amount of postglacial rebound (Andrews, 1978, p. 42), then the emergence rate would be less for the Minas terrace than for southwestern New Brunswick and it would decrease from west to east along the terrace. If an estimate of 2.0 m/century is taken for the initial emergence rate on the west end of the Minas terrace, the drop in elevation of the topset/ foreset contact from 32 m to 26 m at Spencers Island represents a time period of roughly 300 years. Thus, the portion of the original Spencers Island delta that exists today would have been deposited in less than 500 years. The 7 m drop in the marine limit at Port Greville also shows that the remaining portion of the Port Greville delta would have been deposited in less than 500 years. At Lower Five Islands, the drop in the marine limit along the shoreline exposure is 3 m. Even if the rate of emergence is less at Five Islands, the portion of the delta exposed along the shore would have been deposited in several hundred years.

In order to make more accurate estimates of the rate of emergence for the Minas terrace, the position of the sea level at various elevations would have to be dated, as is being done by Scott and Medioli (1978) in New Brunswick. Unfortunately, there are no lakes along the north shore of the Minas Basin below the marine limit but there is geologic evidence that the uplift of the Minas terrace was indeed rapid.

It has been mentioned previously that the fluvial terracing of the Port Greville delta occurred before blocks of ice buried in the outwash had melted. At Diligent River, there is a terrace of glaciofluvial sediment, closely associated with kames, that must have formed while ice was still melting. The terrace is only 500 m north of the proximal edge of the delta, but it has an elevation of only 26 m while the delta attains an elevation of 41 m. Thus, there had been a relative drop in sea level of 15 m before the ice just up valley from the delta had completely melted. A similar situation exists at Parrsboro. Fluvial terraces cut in kames at an elevation of 29 m occur 1 km north of the delta, which has a maximum elevation of 46 m. The terraces in the kames were also probably cut by meltwater, so there was a relative drop in sea level of 18 m while meltwater was still flowing southward. These examples corroborate the interpretation that at least the upper terrace levels of the deposits were eroded by meltwater.

If the rate of postglacial uplift decreases with time after deglaciation (Andrews, 1968, p. 42), then the rapid initial emergence that apparently occurred in the first 1000 years in New Brunswick must have slowed thereafter. It probably took several thousand years before emergence was completed. Likewise, the rate of emergence of the Minas terrace also decreased with time and probably took several thousand years to complete.

Foraminiferal evidence suggests that there has been a deepening of the Bay of Fundy over the last 8600 years (Amos, 1978, p. 970), but submergence may have begun before this. If emergence was not complete until 9000 to 10,000 yr BP, the lowest terraces had to be eroded without the aid of meltwater streams as glacial ice had disappeared before 11,000 yr BP. Emergence may not have been a linear function as it depends upon the rate of sea level rise as well as on the rate of postglacial uplift. The fluvial terraces may represent slower periods of emergence and the raised beaches in the Advocate Harbour area also represent relative stillstands of the sea. The raised beach in Advocate Harbour is particularly indicative of a stillstand (Wightman, 1976), perhaps coincident with the first or second terrace levels in the other deposits. But there are no beaches in the lower part of the plain and the raised beaches in Advocate Harbour and West Advocate may have been formed in a fairly short period of time. It appears that for the most part emergence was relatively rapid and uninterrupted.

SYNTHESIS

During deglaciation in the Late Wisconsinan, a rapid marine incursion into the Bay of Fundy started about 14,000 years ago and shortly thereafter, the bay, including the Minas Basin and Chignecto Bay, was virtually ice free. This severed the ice on mainland Nova Scotia from the New Brunswick ice and left a small, effectively isolated ice cap over the Chignecto Peninsula. Because of the rapid intrusion of the sea, the ice fronts were oversteepened and the ice thinned by drawdown into the Bay of Fundy. After the period of drawdown and outflow, ice margins began to recede inland and outwash deposits were formed along various parts of the Fundy coastline. Proglacial deposits in southwestern New Brunswick formed at 13,000 to 13,325 yr BP and the Minas terrace formed at about the same time or probably slightly before this as a consequence of the smaller ice mass and hence shorter period of ice outflow into the Bay.

By the time ice outflow had terminated along the north shore of the Minas Basin, ice drawdown and downmelting had thinned the ice over the Chignecto Peninsula until the crest of the Cobequid Highlands was exposed. This separated the ice over the Cumberland Lowlands from the narrow strip of ice that sat predominantly in the valleys on the south side of the Cobequids. As the ice started to recede inland, meltwater deposited glaciofluvial sediment in the valleys and the deposits merged into deltas at the Minas Basin. Postglacial rebound exceeded the rise in sea level so that the deltas were being uplifted even as they were being deposited. Continued ice recession and downwasting supplied meltwater that eroded terraces in the valley trains and in the deltas as the shoreline became emergent. Erosion of the first terrace levels may have occurred less than 1000 years after the intial deposition of the deltas. Contemporary with the delta formation and erosion, kames were deposited inland along the sides of some of the wider, deeper discharge valleys. The supply of meltwater dwindled and then ceased as the dissipating ice receded inland to the upper reaches of the valleys. The lowest terrace levels may have been eroded by nonmeltwater streams, several thousand years after the initial formation of the terrace.

The ice on the north side of the Cobequid Highlands also downwasted but probably lasted rather longer because of its larger size. The ice margin receded to the north of the Cobequids but as it did so in the Parrsboro Gap, it deposited a small recessional moraine that blocked the valley and forms the watershed for the River Hebert and Parrsboro River systems. Gilbert Lake abuts the north side of the moraine. Continued recession of the ice margin may have been towards the northeast so that meltwater flowed southwestward along the West Brook valley and then northward through the River Hebert valley to the Cumberland Basin. An esker along the west side of the River Hebert valley may have formed very early in the deglacial history of the area and flowed southward; alternatively, it may have formed later, after ice separation over the Cobequids, and flowed northward.

Holocene sea level rise, tidal amplification and subsidence have decreased postglacial emergence in the Minas Basin and caused submergence on the east end of the Minas terrace where postglacial rebound was the least. Postdepositional erosion, both fluvial and marine, has removed much of the deposits and at Advocate Harbour, the eroded material forms the bulk of a raised marine plain.

CHAPTER 17. SUMMARY OF SALIENT CONCLUSIONS

A. SEDIMENTOLOGICAL

1. Outwash was deposited by braided streams, 1 to 2 m deep, with sustained velocities of 1 to 4 m/sec.

2. When channels were overtopped and flooding of large parts of the outwash occurred, tabular cross beds were laid down on the lee side of the former interchannel areas. The tabular cross beds were developed only in the finer gravel ($M_Z < \sim -3\phi$).

3. Bed load sediment avalanched down the deltaic foreset slope while the finer, suspended sediment was carried out into the water by a freshwater overflow.

4. Foreset slope avalanches that were large enough to attain true grain flows were infrequent.

5. The amount of seaward transport of the suspended sediment was inversely proportional to grain size. Coarser sand settled out on the foreset slope while finer sand, silt and clay settled out mainly on the bottomset slope. The distance to which coarse silt was transported defined a 'suspension fall out' zone.

6. In the winter, the meltwater supply abruptly ceased and fine silt and clay deposition blanketed the previous deposits, producing varves in the 'suspension fall out' zone. A sharp clay/sand contact combined with a burrowed sand/clay contact and a coarsening up of the summer sand bed produced a coarsening up varve from fall to fall.

7. Beyond the 'suspension fall out' zone, fine silt and clay accumulated to form a massive deposit.

8. Infrequent, large slumps of the foresets initiated small scale turbidites that swept onto the bottomset slope and deposited coarse graded and ungraded sands in the summer sand bed.

9. Ice rafted dropstones were deposited year round but were more abundant in the late fall/early winter.

10. Progradation of the foresets caused undercompaction and instability in the bottomsets especially where the water was deeper (>10 m). Failure of the bottomsets, including faulting, liquefaction and flowage, was significant enough at times to lower the surface of the delta and cause secondary foreset deposition on the submerged surface.

B. CONCLUSIONS DIFFERING FROM THOSE OF SWIFT AND BORNS (1967)

11. There was a relative drop in sea level during delta construction.

12. The Saints Rest Member is part of the overall delta constructional phase, not a separate, later event, and forms part or all of the topset facies where it overlies foreset beds. Accordingly, the Saints Rest Member has been redefined as sediment of glaciofluvial origin whether it be inland outwash plains or the topset beds of deltas. The topset beds may be regarded as the topset facies of the Saints Rest Member. The Advocate Harbour Member comprises the marine portions (foresets and bottomsets) of deltas as well as raised marine littoral gravel.

13. Sediment infilled wedges that occur in the deposits, particularly at Lower Five Islands, are not ice wedge casts and are thus not indicative of permafrost.

14. Postglacial emergence increases steadily from Lower Five Islands towards the west as shown by the elevations of the deltaic topset/foreset contacts which are the best indicators of former sea level.

C. CONCLUSIONS ADDED TO THE DEGLACIAL MODEL

15. During the Late Wisconsinan, ice crossed the Cobequids from the north and the deglacial features on the Chignecto Peninsula were deposited during the waning of that ice sheet.

16. During deglaciation, ice contact deltas were deposited along the north shore of the Minas Basin at least as far east as Portapique and as the ice fronts melted inland, glaciofluvial gravel was deposited in the valleys. There is a 'probable' delta at West Advocate.

17. The deltas were probably formed about 14,000 years ago, perhaps slightly before similar deposits in southwestern New Brunswick that have been dated at 13,000 to 13,325 yr BP. The relatively small size of the Chignecto ice cap, with a shorter period of ice drawdown into the bay, may be the reason for the slightly earlier formation of the Minas terrace.

18. The rock types present in the outwash are closely linked with the bedrock to the north of the deposits and as the northward receding ice fronts crossed lithologic boundaries, different rock types were released.

19. Fluvial terraces were cut into the deltas and inland glacio-fluvial deposits by meltwater while emergence was taking place. The degree of terracing depended upon the amount of meltwater and the length of time over which it was supplied.

20. Contemporaneous with delta formation and erosion, kames were deposited between the glacial ice and the valley walls in some of the deeper, wider valleys. Some of the kames were laid down in ponded water while others were formed by streams that were feeding the deltas.

21. Kame deltas north of the Parrsboro Gap were formed in water ponded along the margins of, and possibly within 'windows' in, stagnating valley ice.

22. Meltwater flow was southwards through the Parrsboro Gap during delta formation and terrace erosion, but it reversed to the north before final ice melt out at the Gilbert Lake 'recessional' moraine. The last ice to melt was on the north side of the Cobequid Highlands.

23. A relative rise in sea level and tidal amplification in the Holocene have decreased the effect of uplift along the north shore of the Minas Basin. Consequently, the east end of the terrace, where rebound was probably least and subsidence was probably greatest, has only glaciofluvial sediments exposed subaerially.

24. Erosion has removed much of the deposits so that their original seaward extent is unknown.

25. Shoreline sediments formed during uplift have been destroyed by subsequent erosion in the Minas Basin proper and are preserved only at Advocate Harbour where they overlie a low rock platform.

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APPENDIX 1

GRAIN SIZE ANALYSIS

Samples with a mean grain size of sand or coarser (Wentworth's scale) were dry sieved from -6.5ϕ to 4.0ϕ at 0.5ϕ intervals, with an additional sieve ($\simeq$ -6.7ϕ) being used for the coarsest samples. A Hewlett-Packard 9100B calculator and 9125B plotter were used to plot cumulative curves and to calculate Folk's (1974) graphic statistics. Samples with a mean grain size finer than sand were wet sieved through a 63μ screen and the silt and clay fraction were pipetted according to Piper (1977). Graphic statistics were also determined for these samples using the calculator.

The following table lists the dry sieving data for each sand and gravel sample. The last table lists the weights from the pipette analyses and weights of 0.0 mean that no sample was taken for that particular grain size. The samples are identified by their distance in centimetres from the bottom of the section measured in 1974. The number of the clay bed is also listed beside the interval.

DRY SIEVE DATA

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* SAMPLE	-6.5	-6.0	-5.5	-5.0	-4.5	-4.0	-3.5	-3.0	-2.5	-2.0	-1.5	-1.0
* 74-191	.000	.000	.000	.000	.000	.000	4.008	3.274	7.152	8.895	11.842	13.068
* 74-197	.000	.000	.000	.000	16.662	23.956	10.201	13.365	15.290	12.385	14.904	23.134
* 74-199	.000	.000	.000	.000	.000	100.686	70.835	112.338	121.823	105.093	121.064	113.872
* 74-203	.000	.000	.000	.000	.000	17.484	4.650	9.120	18.770	15.103	20.707	33.949
* 74-209	.000	.000	.000	.000	80.233	25.375	138.123	323.100	460.760	312.510	70.400	11.520
* 74-210	.000	.000	.000	.000	13.819	26.985	112.535	137.221	276.223	223.263	88.006	12.664
* 74-212	.000	.000	291.900	357.420	298.400	248.680	200.530	29.511	9.201	4.590	6.530	6.450
* 74-213	.000	.000	.000	63.927	30.320	246.840	286.460	113.920	37.480	11.510	7.660	3.630
* 74-215	.000	.000	.000	.000	36.728	63.120	124.480	134.360	138.030	85.640	66.440	50.130
* 74-216	.000	.000	.000	.000	23.813	28.206	52.306	41.516	57.298	70.376	100.915	95.009
* 74-220	.000	.000	172.670	540.100	459.910	148.990	173.205	121.090	122.500	71.730	64.090	40.080
* 74-221	.000	.000	.000	61.635	138.666	250.160	216.280	147.260	139.290	73.550	67.950	61.830
* 74-222	.000	.000	186.690	.000	251.810	117.120	180.060	110.090	101.620	82.170	84.120	82.490
* 74-224	.000	.000	316.670	165.590	160.090	263.270	195.630	65.180	58.020	21.370	10.220	7.210
* 74-225	.000	.000	.000	462.480	534.380	454.540	170.960	76.980	57.690	26.230	26.060	20.270
* 74-250	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.043
* 74-251	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.013
* 74-252	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.034
* 74-253	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
* 74-254	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
* 74-255	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.244
* 74-300	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
* 74-301	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
* 74-302	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.041
* 74-303	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
* 74-304	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.105
* 74-305	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
* 76-11	.000	.000	525.190	389.100	257.420	320.510	229.730	197.870	195.110	152.540	145.760	110.630
* 76-7	.000	.000	216.690	305.310	215.270	197.540	261.150	255.470	322.110	207.520	137.370	48.620
* 77-10	.000	936.920	1421.380	610.100	645.470	729.450	706.000	495.970	338.950	166.040	81.480	46.910
* 77-11	.000	.000	.000	.000	133.288	134.448	222.360	338.390	531.710	883.080	1032.880	1048.560
* 77-12	.000	874.900	930.350	771.870	1121.020	1092.930	686.630	359.960	327.820	215.470	210.210	153.490
* 77-14	.000	361.520	2523.190	1196.960	900.200	499.260	407.550	308.490	252.330	205.510	171.840	176.870
* 77-16	.000	.000	.000	.000	.000	30.060	55.480	145.030	332.320	357.440	377.550	483.170
* 77-17	.000	.000	820.490	1245.240	1427.730	1279.380	862.330	507.240	384.040	251.990	199.670	195.900
* 77-20	.000	.000	642.580	106.206	532.520	539.150	465.030	400.800	412.150	295.210	297.830	253.920
* 77-21	.000	.000	.000	70.417	182.490	339.140	345.150	352.870	455.260	392.020	538.190	568.230
* 77-23	.000	.000	546.710	827.000	514.180	580.800	459.690	446.800	455.720	467.200	531.300	449.540
* 77-25	.000	.000	.000	.000	.000	4.850	.000	.800	4.750	5.810	21.470	41.060
* 77-26	.000	.000	.000	.000	.000	26.300	107.200	114.400	120.140	89.150	26.680	41.440
* 78-1	.000	.000	.000	.000	.000	.000	.000	.000	.000	.894	1.647	1.008
* 78-2	.000	321.700	439.890	357.300	219.370	168.480	112.620	106.050	94.640	65.890	81.940	86.970
* 78-3	2025.520	883.000	570.470	651.100	828.530	395.480	394.070	172.290	164.660	151.170	140.580	125.400

DRY SIEVE DATA continued

* SAMPLE	-0.5	0.0	0.5	1.0	1.5	2.0	2.5	3.0	3.5	4.0	PAN
* 74-191	19.561	41.723	88.096	150.994	317.895	296.229	177.633	96.864	29.192	8.590	8.949
* 74-197	50.323	88.536	148.352	175.193	131.106	24.948	5.124	2.732	1.403	.832	4.184
* 74-199	115.792	99.522	69.862	37.113	19.579	5.885	2.358	1.859	.927	.357	1.390
* 74-203	68.693	60.370	37.084	24.072	14.988	5.516	1.798	1.183	.617	.393	2.158
* 74-209	9.100	7.730	4.980	4.620	7.480	9.080	7.880	11.580	5.310	2.880	4.510
* 74-210	11.901	8.209	3.924	3.549	5.601	5.236	4.900	8.413	4.483	1.447	2.683
* 74-212	6.560	6.440	9.720	20.350	80.120	154.770	121.100	81.570	24.970	5.670	6.030
* 74-213	2.390	3.280	5.720	12.810	39.580	56.970	51.790	58.090	27.700	7.270	7.270
* 74-215	44.140	42.880	53.110	66.370	77.500	41.220	10.660	3.700	1.370	.650	1.930
* 74-216	96.129	100.070	126.007	154.592	228.261	120.371	27.019	7.455	2.036	.731	1.660
* 74-220	32.650	30.850	28.920	27.110	28.690	22.970	17.820	11.400	3.880	1.460	2.590
* 74-221	75.730	85.320	94.460	88.250	88.050	75.150	63.350	45.880	15.190	4.980	9.530
* 74-222	97.390	100.310	108.070	98.090	101.470	82.430	58.660	40.000	13.040	4.040	8.540
* 74-224	5.040	5.250	4.650	5.280	7.690	7.890	10.970	25.490	37.000	24.530	231.570
* 74-225	20.830	22.840	21.350	20.070	23.030	19.210	19.460	31.810	39.780	26.130	305.250
* 74-250	.072	.010	.036	.063	.165	.442	.855	1.399	1.594	.960	1.026
* 74-251	.185	.619	1.112	1.560	1.655	1.249	.790	.445	.172	.057	.077
* 74-252	.029	.072	.312	.741	1.320	1.478	1.235	.924	.427	.141	.225
* 74-253	.010	.035	.087	.320	.575	.813	1.261	1.440	1.139	.620	.775
* 74-254	.000	.014	.218	.779	1.209	.989	.743	.613	.336	.131	.138
* 74-255	.306	.321	.345	.498	.654	.802	1.078	1.073	.740	.323	.541
* 74-300	.010	.004	.013	.033	.114	.335	1.130	2.505	2.354	.993	.790
* 74-301	.012	.028	.010	.045	.126	.391	1.154	1.873	1.473	.690	1.128
* 74-302	.011	.052	.049	.071	.133	.270	.710	1.775	2.010	1.223	1.884
* 74-303	.053	.042	.060	.103	.270	.634	1.577	3.304	2.689	1.332	1.625
* 74-304	.061	.063	.102	.250	.531	.881	1.913	3.112	2.189	1.053	1.398
* 74-305	.054	.081	.099	.274	.634	.969	1.259	1.345	.993	.529	.811
* 76-11	117.370	115.920	128.700	120.210	93.980	37.380	13.850	12.110	8.080	3.630	9.250
* 76-7	29.570	19.150	10.500	7.040	5.540	2.990	1.700	1.490	1.190	.780	3.880
* 77-10	54.810	74.440	127.340	184.890	210.820	118.960	45.080	33.050	17.420	7.860	24.280
* 77-11	1068.170	734.070	415.920	181.170	88.100	25.100	10.460	7.580	4.710	3.140	10.720
* 77-12	112.630	73.650	47.410	35.110	31.540	22.950	13.460	13.630	9.850	6.060	41.700
* 77-14	184.120	196.760	189.330	183.190	181.510	116.240	58.400	41.470	20.640	7.070	23.250
* 77-16	507.050	481.550	548.730	655.150	992.240	878.120	345.180	169.960	65.150	34.400	50.580
* 77-17	226.650	303.340	378.240	310.540	171.700	66.720	26.000	24.690	23.380	16.840	43.010
* 77-20	231.620	214.820	205.450	238.890	350.350	238.720	52.950	20.430	4.980	2.530	7.520
* 77-21	646.600	645.600	531.650	301.070	143.320	51.690	11.920	6.880	2.350	1.170	3.190
* 77-23	369.720	274.740	240.280	170.230	166.320	83.110	32.470	16.100	10.250	5.430	13.490
* 77-25	59.340	56.490	32.480	13.140	7.000	3.150	1.910	2.410	2.920	2.000	5.930
* 77-26	20.200	9.380	4.670	2.540	2.440	2.510	2.150	2.240	1.950	1.410	2.840
* 78-1	1.367	3.975	13.911	41.432	117.304	100.818	29.335	9.434	2.051	.599	.894
* 78-2	120.700	171.640	239.000	201.080	204.590	80.010	15.490	5.020	1.740	.800	2.480
* 78-3	157.960	196.670	243.340	232.120	151.760	38.880	8.660	5.160	2.800	1.760	6.290

PIPETTE ANALYSES

SAMPLE	SAND	4.0	6.0	8.0	9.0	10.0	11.0	12.0
(35) 167.0-168.0	.727	10.160	.000	4.655	.000	.000	.000	.000
(35) 167.5-168.5	1.531	10.290	7.530	5.060	4.000	2.980	1.675	1.305
(35) 168.0-169.0	.167	5.440	.000	2.445	.000	.000	.000	.000
(35) 168.5-169.5	1.817	9.355	6.335	4.400	3.615	.000	2.055	1.180
(39) 189.5-190.5	.911	11.930	8.490	5.805	4.555	3.130	2.205	1.150
(39) 190.5-191.5	2.388	13.845	10.230	7.345	5.910	4.500	3.105	1.905
(44) 222.0-223.0	3.153	8.560	6.725	4.745	3.700	2.800	1.885	1.140
(44) 223.0-224.0	2.405	8.510	6.435	4.435	3.485	2.545	1.775	1.165
(60) 438.5-439.5	.243	4.235	.000	2.210	.000	.000	.000	.000
(60) 439.5-440.5	.641	2.760	.000	1.055	.000	.000	.000	.000
(13) 90-92	.220	13.480	12.700	9.280	7.345	5.615	4.010	2.720
(13) 92-94	2.949	9.880	8.695	6.485	5.235	4.115	2.980	1.635

APPENDIX 2

PEBBLE COUNTS, MINAS TERRACE

The pebble counts are discussed according to deposit, and the locations of the counts, as well as possible bedrock sources for the deposits that were not discussed as such in chapter 13, are given.

After pebbles had been counted at several of the deposits, certain general rock types in which the pebbles could be categorized began to emerge. The rock types are those used for the histograms in chapter 13 and each rock type is used *sensu lato;* that is, the category 'granite' embraces all granitic type pebbles. More precise identification of the pebbles was unwarranted as the sources could not be delineated beyond the general rock types. The pebble count data is listed in a table at the back of the appendix.

PORTAPIQUE, BASS RIVER AND ECONOMY

At Portapique, pebbles were counted in each of the lower, middle and upper parts of DuPaul's pit and one large count of 330 pebbles was taken in the low shoreline exposure just east of the Portapique River estuary. At Bass River, 4 counts (one above the other) were taken along the shoreline exposure on the west side of the Bass River estuary. Three counts were taken in the N.S. Department of Highways borrow pit, from the lower, middle and upper part, and 2 counts were taken along the shoreline exposure at the end of Cove Road.

The Silurian to Devonian Portapique-Parrsboro Succession, the Bass River Metamorphic Suite, and at Bass River and to a lesser extent also at Economy, the Triassic North Mountain Basalt are possible sources for the andesite and diabase (Geology Map of the Cobequid Highlands (East Half). A large intrusion of Devonian to Carboniferous diorite in the north central part of the Cobequids is a possible source for the diorite pebbles. The intrusion does not extend as far west as Economy but another large intrusion of the same diorite occurs north and west of Economy. The two intrusions are separated by about 2.5 km of Portapique-Parrsboro Succession. A lower abundance of diorite at Economy is probably the result of the decrease in the areal extent of diorite in the Cobequids to the north. Andesites and diabases may be associated with these intrusions.

Carboniferous granite intrusions in the Cobequids are the probable source for the granite pebbles. The abundance of granite at Portapique (34%) and Bass River (13%) does not correspond to the limited areal extent of the granites that are mapped in the Cobequids north of the deposits. A rather narrow (up to ≃1.75 km wide), east-west trending granite intrusion that pinches out almost directly north of Bass River occurs on the north side of the Cobequids. It is wider (up to ≃4.5 km) east of Portapique. The dominance of granite (46%) at Economy is more easily accounted for as an east-west trending Carboniferous granite intrusion (up to 3.25 km wide) occurs immediately north of the Cobequid Fault. The intrusion extends from the headwaters at Carr Brook, west of Economy, to north of Bass River. There is also a Devonian to Carboniferous granite intrusion on the north side of the Cobequids, north and west of Economy. Approximately 2 percent of the granite at Economy and Portapique is granite gneiss and this is probably derived from the granite gneiss of the Bass River Metamorphic Suite in the Cobequids north of these particular deposits. The relative proportions of granite and diorite at Bass River and Portapique suggest that some of the intrusion that is mapped as diorite may be granite.

The quartzite and metamorphosed sedimentary rock have two probable bedrock sources in the Cobequids: the Silurian to Devonian Portapique-Parrsboro Succession, which forms a large part of the Cobequids here, and the Carboniferous Londonderry Succession, which extends only as far west as Little Bass River, immediately north of the Cobequid Fault. Rhyolite is also probably derived from the Portapique-Parrsboro Succession. Breccia is a minor component at Portapique and it is interpreted as a fault breccia, derived from one of the many faults in the Cobequids.

The bedrock source for the sedimentary rock is probably the Upper Carboniferous Parrsboro Formation on the south side of the Cobequid Fault but there is also the Upper Carboniferous sedimentary rock north of the Cobequids in the Cumberland Lowlands. All of the outwash at Economy and most of the outwash at Bass River and Portapique overlie the Triassic Blomidon and Wolfville Formations (undifferentiated) but pebbles from these formations are almost nonexistent. About 6 percent of the sedimentary pebbles at Bass River are from the Triassic red beds, but these are in fact poorly lithified clasts and not pebbles. There were no recognizable Triassic sedimentary clasts at Economy or Portapique. Apparently the Triassic red beds were not competent enough to withstand glacial erosion and subsequent fluvial deposition.

The low proportion (2%) of sedimentary rock at Portapique is somewhat puzzling. There is a significant belt (≃2.5 km wide) of Parrsboro Formation adjacent to and south of the Cobequid Fault so source rock is apparently not a problem. The pebble counts were taken at the southern end of the deposit along the shoreline and in a pit beside Highway 2. Bedrock or till were not visible at the base of the exposures where the counts were taken so the sedimentary rock may be present lower in the deposit. Another possibility is that the Carboniferous sedimentary rock was not competent enough to withstand several kilometres of glacial erosion, transportation and fluvial deposition. This possibility would preclude the Cumberland Lowlands as a source area for the sedimentary rocks in the outwash. Sedimentary rock is more abundant in the outwash at Economy but there the Parrsboro Formation lies immediately north of the deposit. At Bass River, the counts were taken at Saints Rest along the shoreline and the Parrsboro Formation underlies part of the deposit immediately north of there in central Bass River. In the Portapique outwash, it is interpreted that the proportion of sedimentary rock increases as the distance to the Parrsboro Formation decreases. Sedimentary rock is probably fairly abundant in the northern part of the deposit that overlies the Parrsboro Formation.

The cleaved sedimentary rock is probably derived from the Londonderry Succession and/or the Parrsboro Formation adjacent to the major faults, including the Cobequid Fault.

LOWER FIVE ISLANDS

At the distal end of the shoreline exposure, 2 pebble counts were taken in the foresets and 1 count was taken in each of the lower, middle and upper part of the topset facies. In the mid part of the exposure, counts were made in the foresets and at the bottoms and tops of units 2 and 3 in the topset facies. At the proximal end of the exposure, 2 counts were taken in the foreset beds as well as counts at the bottoms and tops of units 1, 2 and 3 in the topset beds.

MOOSE RIVER

Pebbles were counted in the foresets and in the bottom, middle and top of the topsets at the shoreline exposure as well as one count at the borrow pit in the inland glaciofluvial gravel.

The sedimentary rock in the outwash at Moose River is probably derived from the Parrsboro Formation that underlies both the delta and the glaciofluvial deposit. As at the deposits to the east, the Upper Carboniferous rocks north of the Cobequids are probably too far away to be an effective source. Cleaved sedimentary rock occurs in the Parrsboro Formation adjacent to the Cobequid Fault. The Silurian to Devonian Portapique-Parrsboro Succession and the North River Succession are sources for metamorphosed sedimentary rock, andesite and quartzite. Rhyolite, andesite and diabase pebbles are also probably derived from the Portapique-Parrsboro Succession. A large Devonian to Carboniferous diorite intrusion occurs northeast of Moose River in the northern part of the Cobequids, from which the diorite pebbles might be derived. The abundance of granite is probably the result of the nearness of a Devonian to Carboniferous granite adjacent to the Cobequid Fault. A similar granite also occurs on the north side of the Cobequids.

PARRSBORO

At Durant's pit, there were 5 counting stations with 1 in the foresets and 4, equally spaced vertically, in the topsets. There were 3 pebble count stations, in the lower, mid and upper parts, at both Gilbert's pit and at Kernohan's pit, while one count was taken at Forkenham's pit.

The abundant quartzite pebbles (41%, 50%) and lesser amounts of metamorphosed sandstone could be derived from several bedrock sources: the Silurian to Devonian Portapique-Parrsboro Succession north of and adjacent to the Cobequid Fault on the east side of the Parrsboro River; the Silurian(?) Advocate Succession on the west side of the river; and the Hadrynian Jeffers Succession. Rhyolite and andesite are also probably derived from the Portapique-Parrsboro Succession and the Advocate Succession. The Upper Carboniferous Parrsboro Formation south of and adjacent to the fault and possibly the Upper Carboniferous Cumberland Group north of the Cobequids are sources for the sedimentary rock. Cleaved sedimentary rocks are probably derived from the Parrsboro Formation adjacent to the Cobequid Fault or the Portapique-Parrsboro and Advocate Successions.

Granite is less abundant (8%, 6%) at Parrsboro than in the deposits to the east of Parrsboro, and granite is also less abundant in the Cobequids north of Parrsboro. A Devonian to Carboniferous granite (≈5 km x 3 km) intrudes the Advocate Succession on the west side of the river and a small (≈0.9 km x 0.4 km) intrusion of the same granite occurs on the east side of the river. The Devonian to Carboniferous diorite intrusion and the Hadrynian Jeffers Brook Pluton on the east side of the Parrsboro River are sources for diorite and possibly for andesite.

DILIGENT RIVER, PORT GREVILLE AND SPENCERS ISLAND

Three pebble counts were taken at the exposure of the Diligent River delta located between the Diligent and Ramshead Rivers; one from the foresets and one from both the lower and upper topsets. As well, one count was taken in the topsets beside the wharf and in the gravel pit near Highway 209. At the proximal (W) end of the shoreline exposure at Port Greville, pebbles were counted in the foresets (lower and upper) and in the topsets (lower, middle, top). On the east end of the exposure counts were taken in the lower and upper foresets and topsets. At Spencers Island, pebbles were counted at two locations in the foresets and one in the topsets near the 1974 exposure while farther to the east where the delta overlies till, one count was taken in the foresets and one in the topsets.

ADVOCATE HARBOUR AREA

Pebbles were counted at 3 locations on the upper surface of the delta at West Advocate; one on the north end in a fire ditch and 2 on the southeastern corner along a road cut. One large count (211) was taken at an excavated face in the raised beach in West Advocate. Pebbles were counted in the foreshore and backshore facies of unit 1 and in the backshore facies of unit 2 in the raised beach at Advocate Harbour.

PEBBLE COUNT DATA

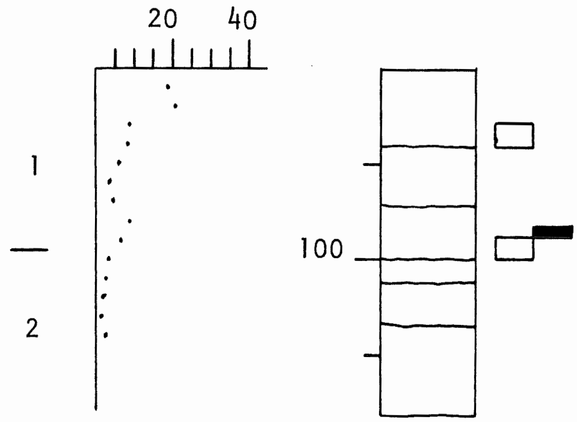
* LOC	SED	CLSD	BREC	MESD	QTZE	GRAN	RHY	AND	DIAB	DIOR	TOT	C:NC
* 1.DELTA,W.ADVOCATE	97	27	5	0	176	13	2	1	0	4	330	71:29
* 2.BEACH,W.ADVOCATE	29	62	1	1	27	47	1	6	0	37	211	86:14
* 3.BEACH,ADVOCATEHARBOUR	127	19	0	10	125	34	2	6	0	11	334	62:38
* 4.TOPSET,S.I.DELTA	89	27	1	6	58	37	4	1	0	1	224	60:40
* 5.FORESET,S.I.DELTA	114	34	0	14	106	57	1	0	0	3	330	65:35
* 6.TOPSET,P.G.DELTA	272	57	0	9	193	9	9	3	0	1	553	51:49
* 7.FORESET,P.G.DELTA	194	31	2	9	184	10	4	3	0	0	437	56:44
* 8.TOPSET,D.R.DELTA	202	19	0	8	197	8	0	0	0	0	434	53:47
* 9.FORESET,D.R.DELTA	64	7	0	0	34	2	0	0	0	1	108	41:59
* 91.TOPSET,PO.DELTA	276	99	1	32	605	80	45	31	0	38	1207	77:23
* 92.FORESET,PO.DELTA	29	18	0	6	46	9	1	3	0	0	112	74:26
* 93.TOPSET,M.R.DELTA	84	18	0	4	48	127	28	17	5	20	351	76:24
* 94.FORESET,M.R.DELTA	64	2	0	1	14	20	8	0	0	0	109	41:59
* 95.GLACFLUV.,M.R.	16	0	0	3	20	51	9	6	2	7	114	86:14
* 96.TOPSET,UNIT1,F.I.DELTA	56	34	0	21	32	31	33	2	0	1	210	73:27
* 97.TOPSET,UNIT2,F.I.DELTA	124	73	2	18	69	35	105	8	0	1	435	71:29
* 98.TOPSET,UNIT3,F.I.DELTA	93	16	0	13	82	393	84	54	13	14	762	88:12
* 99.FORSET,PROX,F.I.DELTA	62	28	0	6	35	35	47	0	0	0	213	71:29
* 991.FORESET,MID,F.I.DELTA	55	11	0	0	22	7	11	7	0	0	113	51:49
* 992.FORSET,DIS,F.I.DELTA	50	17	0	0	57	67	26	9	0	2	228	78:22
* 993.GLACFLUV.,ECONOMY	53	10	0	81	21	258	9	79	0	40	551	90:10
* 994.GLACFLUV.,SAINTSREST	49	10	0	0	103	56	8	110	15	69	420	88:12
* 995.GLACFLUV.,PORTAPIQUE	15	13	8	36	85	222	13	124	33	113	662	98:2

All pebbles > -2.5ϕ were counted
C:NC = Cobequid:NonCobequid Ratio
Disregard numbers in front of locations

APPENDIX 3

LAKE SEDIMENT DATA AND REPORTS

TUBE NUMBER	LOSS ON IGNITION	DEPTH (cm)	LEAK 2	DATED INTERVAL	DESCRIPTION



Gyttja, medium green brown

Clay, grey, infrequent organic laminae

Gyttja, dark green brown

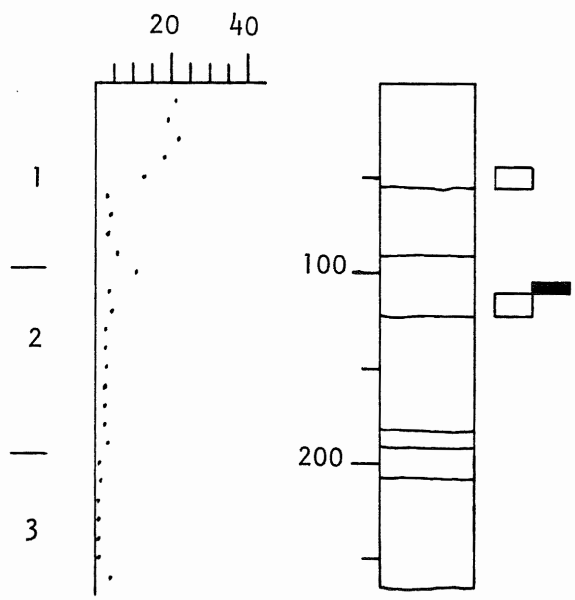
Clay, grey, br., some organic laminae

Clay, red brown granules

Clay, red, granules, some sand and silt laminae

DAL -	313	87.5 - 99.5	9580 ± 310
GSC -	2728	82.5 - 87.5	12,900 ± 160
DAL -	314	29 - 41	9715 ± 200

TUBE NUMBER	LOSS ON IGNITION	DEPTH (cm)	LEAK 3	DATED INTERVAL	DESCRIPTION



Gyttja, medium green brown

Clay, grey, slightly red br. in lower half

Gyttja, medium to dark green brown

Clay, red brown, scattered granules, organic laminae at top

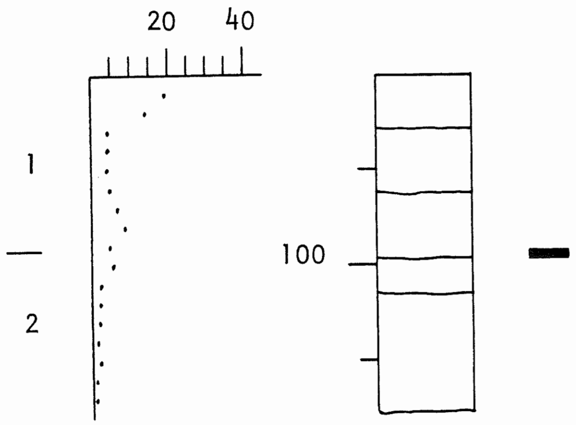
Sand, red brown, clayey

Clay, red brown, granules

Sand, red brown, clayey, some clay laminae

DAL -	313	110.5 - 122.5	9580 ± 310
GSC -	2728	105.5 - 110.5	12,900 ± 160
DAL -	314	44 - 56	9715 ± 200

TUBE NUMBER | LOSS ON IGNITION | DEPTH (cm) | LEAK 4 | DATED INTERVAL | DESCRIPTION



Gyttja, medium green

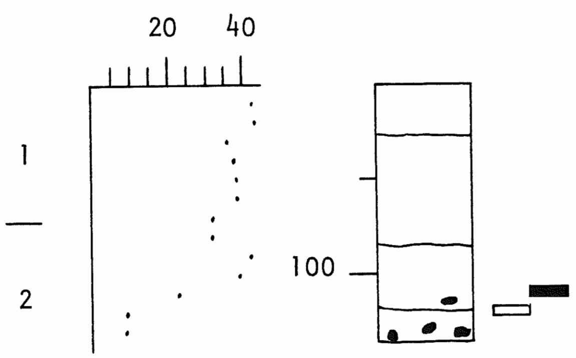
Clay, grey

Gyttja, medium-dark green br.
Silt, red br., clayey, granules, organic laminae at top

Sand, red brown, very fine grained, silty, some clay laminae, clay content decreases downwards

GSC - 2880	93.0 - 98.0	15,900 ± 1200

TUBE NUMBER | LOSS ON IGNITION | DEPTH (cm) | WARDS BROOK I | DATED INTERVAL | DESCRIPTION

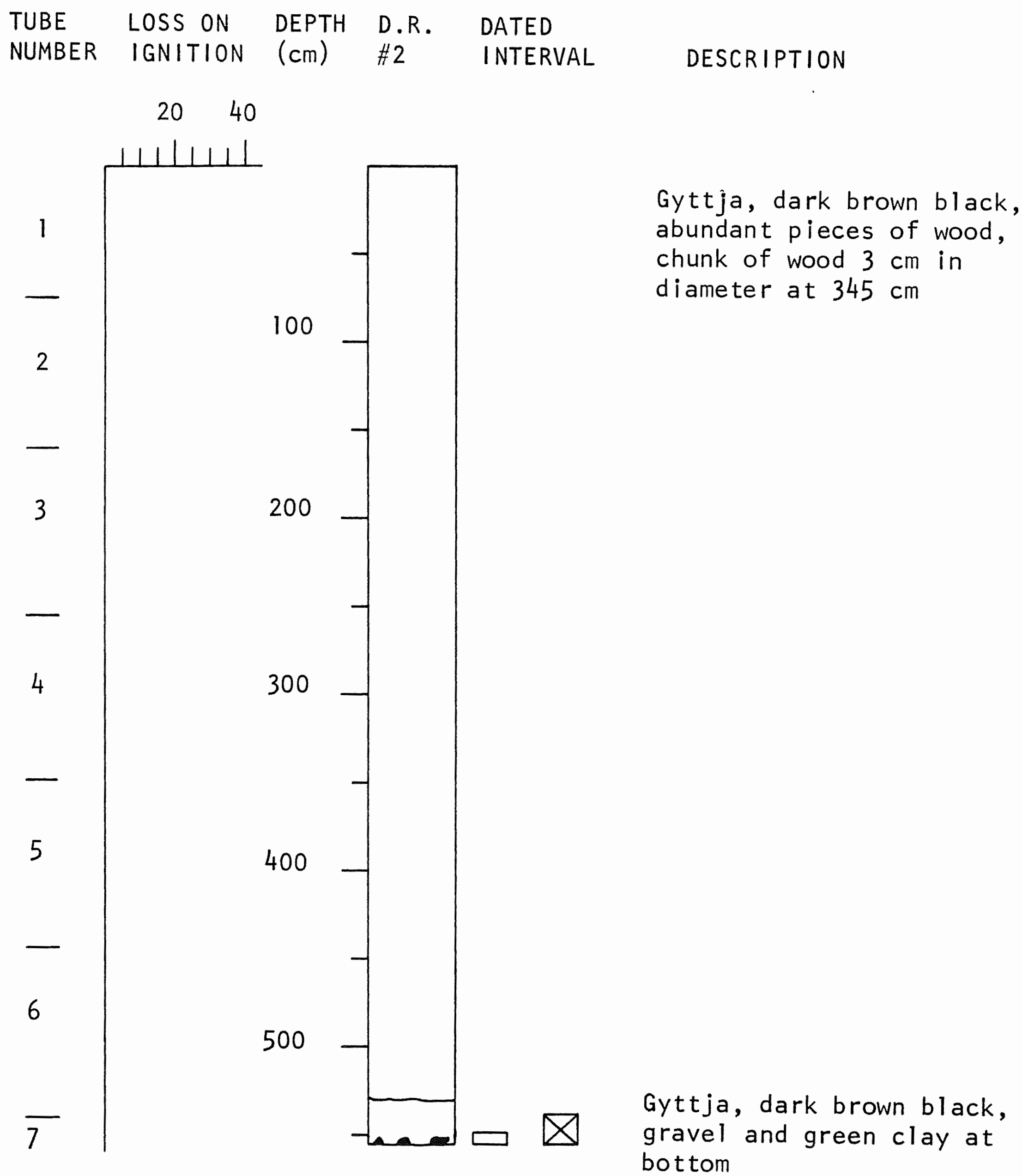


Gyttja, dark green black

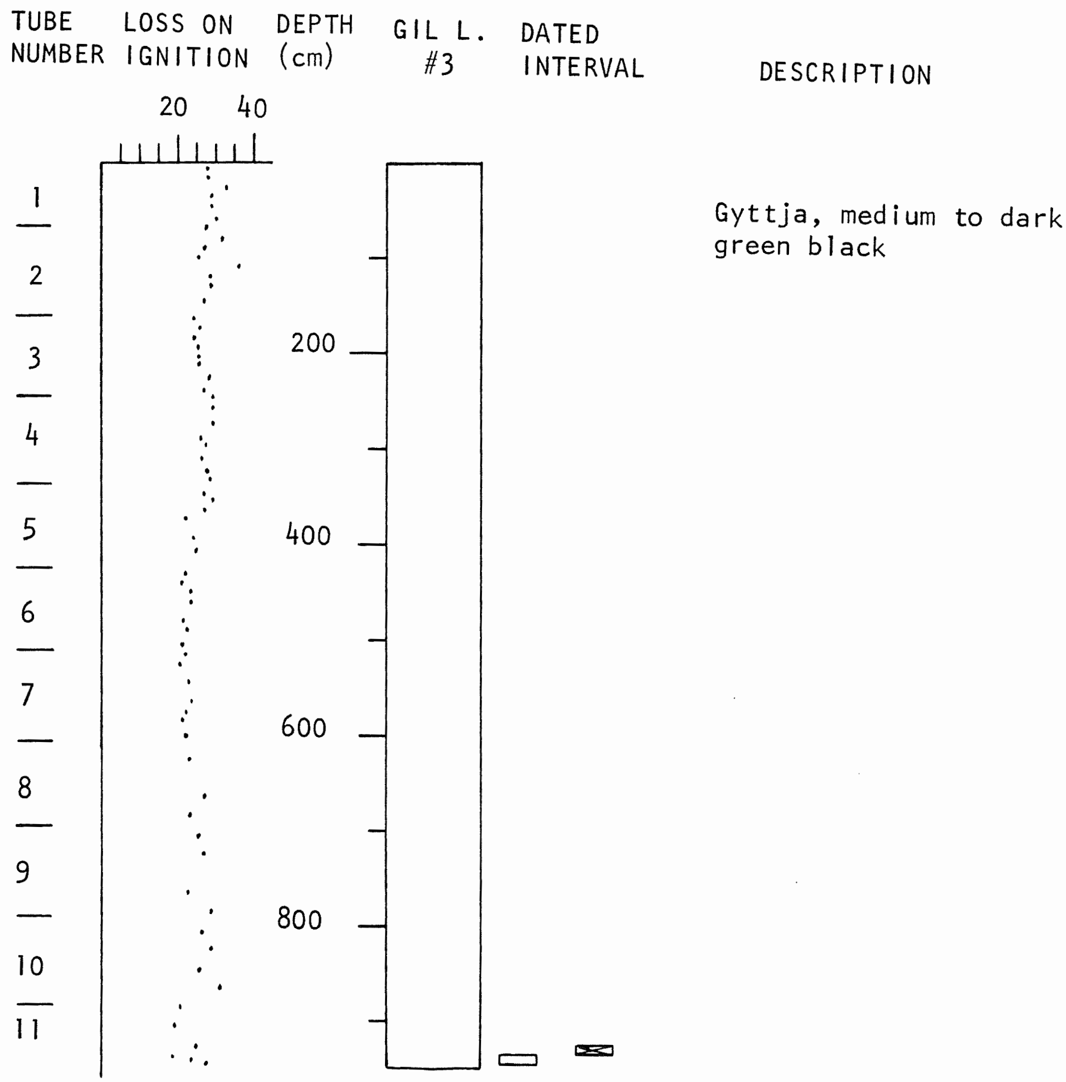
Gyttja, light-medium green black

Gyttja, dk. green black, pebbly and clay laminae at bottom
Clay, lt. green, pebbly, organic laminae towards top, wood at 120-122 cm

GSC - 2730	107 - 112	4390 ± 70
DAL - 315	117 - 122	3330 ± 178



DAL - 301	548 - 555	11,555 ± 230
QU - 764	538 - 555	10,120 ± 200



DAL - 309	936 - 946	3595 ± 105
QU - 765	926 - 936	5640 ± 220

PALYNOLOGICAL REPORT NO. 78-11

Date: Oct. 6, 1978

Locality: ½ km east of West Brook (settlement), Nova Scotia

Lat.: 45°33.5' N Long.: 64°17.5' W

NTS: 21 H/9W

Submitted by: W. Blake Jr. for Daryl Wightman

Field No.: DW-76-2

Lab No.: PL-78-48

Description of Sample: Peat

Results and Interpretation:

Trees	%	Herbs	%
Picea	16.5	Ericaceae	3.8
Pinus sp.	1.9	Gramineae	4.7
P. strobus	0.9	Tubuliflorae	0.9
P. banksiana/resinosa	14.4	*Artemisia*	1.4
Abies balsamea	0.2	Rosaceae	0.7
Betula	30.0	*Rubus chamaemorus*	0.2
Quercus	0.7	Caryophyllaceae	0.2
Fagus	0.2	*Polygonum*	0.4
		Lycopodium clavatum	0.5
Shrubs		*L. obscurum*	0.5
Alnus	1.7	*L. complanatum* type	0.7
Salix	3.8	*L. annotinum*	0.9
Myrica gale	7.6	Pteridophyta	1.2
		Polypodiaceae	0.7
Aquatics		*Selaginella selaginoides*	2.4
Cyperaceae	283.7	*Botrychium*	0.2
Myriophyllum	0.2	Unidentified	2.4
Sphagnum	; 8.0		

The assemblage noted above is similar in many ways to those foun at or near the base of several profiles from Nova Scotia (Railton, 1972; Hadden, 1975; Livingstone, 1968) and to assemblages contained in other buried postglacial organic deposits (Gill, Internal Palynological Report, 1974-4).

The preponderance of sedge (Cyperaceae) pollen accompanied by *Sphagnum* spores, heath pollen types (Ericaceae), grasses (Gramineae), cloudberry (Rubus chamaemorus) and Selaginella selaginoides are characteristic of bog or fen environments indicating such conditions prevailed at the time of deposition. Some of the clubmosses (Lycopodium annotinum; L. obscurum) are also indicative of bog margins or wet woods. This wet or boggy area or its surroundings also supported such shrubs as sweet gale (Myrica gale), willow (Salix) and alder (Alnus).

Among the arboreal pollen types birch (Betula) is the most abundant but many of these may relate to the shrub birch (Betula glandulosa). Spruce (Picea) pollen is less abundant but some spruce trees may have been present in the area or on the bog itself. Pine pollen is about as abundant as spruce with most of the pine pollen of the jack/red pine (Pinus banksiana/resinosa) type. The former was the most likely tree involved but it may not have been growing in the immediate area. Only a few grains of white pine (Pinus strobus) are present. Open treeless or sparsely treed tundra conditions are suggested by this pollen assemblage.

Comparison with other assemblages from Nova Scotia suggest an age of between 9,500 and 11,000 years B.P. or possibly as old as 11,700 years B.P.

The occurrence of white pine wood (see Wood Report No. 78-35) associated with this deposit is an enigma. The few grains of white pine pollen are probably attributable to long distance transport and white pine trees were not present until a much later time. Possibly the wood is stratigraphically higher in the section or may have fallen into an older peat deposit.

Hadden, K.A. 1975. A pollen diagram from a postglacial peat bog in Hants County, Nova Scotia. Can. J. Bot., 53, 1, 39-47.

Livingstone, D.A. 1960. Some interstadial and postglacial pollen diagrams from eastern Canada. Ecol. Mon., 38, 2, 87-125.

Railton, J.B. 1972. Vegetational and climatic history of southwestern Nova Scotia in relation to South Mountain Ice cap. Dalhousie U., Ph.D. thesis, 146 p.

R. J. Mott

R. J. Mott
Quaternary Paleoecology Laboratory

PALYNOLOGICAL REPORT NO. 79-2

Date: Jan. 30, 1979

Locality: ½ km east of West Brook (settlement), Nova Scotia.

Lat.: 45°33.5' N Long.: 64°17.5' W

NTS: 21 H/9W

Submitted by: W. Blake Jr. for Daryl Wightman

Field No.: Peat block 77-

Lab No.: PL-78-48

Description of Sample: Peat

Results and Interpretation:

The pollen spectra on the accompanying sheet comprise the results obtained from analysis of samples removed from a peat block having a vertical thickness of 7.5 cm. Comparison of these spectra with the assemblage reported in Internal Palynological Report No. 78-11, purported to be from a block of peat of similar age, shows some significant differences in content. Notable are the extreme values for sedge (Cyperaceae) in the latter along with higher values for birch (*Betula*) and pine (*Pinus*) pollen and the much higher value for willow (*Salix*) and spruce (*Picea*) in the spectra included herein.

Without doubt, deposition of the peaty layer took place during the spruce pollen maximum, a zone that is somewhat variable in age throughout the Maritimes. In southwestern New Brunswick the maximum ranges in age between 10,500 and 11,500 years B.P. (Mott, 1975), whereas in Prince Edward Island, it is younger than 9,500 years B.P. (Anderson, personal communication). In Nova Scotia the spruce pollen zone is not as well defined in most cases but is probably less than 11,700 years B.P. in central Nova Scotia (Railton, 1972) and 10,300 years B.P. or less on Cape Breton Island (Livingstone and Livingstone, 1958). One site from northern Nova Scotia (Livingstone, 1968) shows the spruce zone to be older than 10,700 years B.P. but this seems to be anomalous. An age of between 9,500 and 10,500 years B.P. for the spruce pollen zone in this area would be expected. In light of these estimates the date of 9,830 ± 100 years B.P. (GSC-2772) for the base of the peat block seems realistic. The peat sample reported previously should be slightly older than this according to the pollen assemblage obtained but the recovery of white pine wood from this peat (Wood Report No. 78-35) still remains an enigma and the interpretation of the pollen assemblage a problem.

Other components of the spectra are typically those associated with the spruce zone. Pine values are low as are most other tree pollen types although birch (Betula), aspen or poplar (Populus) and oak (Quercus) pollen are present in small amounts. Among the shrubs, willow (Salix) is extremely abundant and Myrica is plentiful. Herbaceous pollen types are represented by a number of taxa especially grass (Gramineae), sedge (Cyperaceae), and the Tubuliflorae group of the Composites. Clubmosses (Lycopodiaceae) and ferns (Polypodiaceae) are also represented.

The preponderance of spruce pollen indicates the presence of spruce trees in the area if not as a closed boreal forest, at least a spruce woodland. White spruce cones (Wood Report No. 79-8) recovered from the basal peat confirms the presence of this species in the area. Minor amounts of poplar or aspen, birch and oak may also have been present. Pine was probably not represented locally. Willow must have been extremely abundant locally and Myrica plentiful as well, probably as a border of vegetation around a wet sedgy area.

More complete studies in this area are required before the pollen stratigraphy and chronology can be worked out in detail.

Hadden, K.A. 1975. A pollen diagram from a postglacial peat bog in Hants County, Nova Scotia. Can. J. Bot., 53, 1, 39-47.

Livingstone, D.A. 1968. Some interstadial and postglacial pollen diagrams from eastern Canada. Ecol. Mon., 38, 2, 87-125.

Livingstone, D.A. and Livingstone, G.R. 1958. Late-glacial and postglacial vegetation from Gillis Lake in Richmond County, Cape Breton Island, Nova Scotia. Amer. J. Sci., 256, 341-359.

Mott, R.J. 1975. Palynological studies of lake sediment profiles from southwestern New Brunswick. Can. J. Earth Sci., 12, 2, 273-288.

R. J. Mott

	Depth From Top Of Block %			
TREES	0-1 cm	2-3 cm	4-5 cm	6.5-7.5cm
Picea	35.2	23.8	21.6	37.7
Pinus	0.7	1.7	0.6	0.8
P. strobus	0.7	1.1	1.6	0.9
P. banksiana/resinosa	1.2	3.3	3.8	1.9
Tsuga canadensis	0.5			
Larix laricina	0.2		0.3	
Betula	7.2	5.0	3.8	3.2
Populus	0.7	1.1	6.9	4.6
Quercus	0.7	0.9		1.2
Ulmus	0.2			
Acer		0.3		+
Fraxinus	0.2			
Carpinus/Ostrya	0.5	0.3	0.3	
SHRUBS				
Juniperus/Thuja				0.2
Alnus crispa	0.5	1.4	0.6	0.3
A. rugosa				0.3
Salix	32.8	39.8	42.2	26.4
Corylus				0.2
Myrica	7.7	3.6	1.6	10.2
Cornus stolonifera	0.2			
HERBS				
Ericaceae		0.6		
Gramineae	1.2	2.2	0.9	0.5
Tubuliflorae	1.2	2.2	3.1	1.6
Artemisia		0.3		0.3
Rosaceae	0.2			
Ranunculaceae	0.2			
Umbelliferae			0.3	
Leguminosae			0.3	
Lycopodium annotinum	0.5	1.7	1.6	1.6
L. clavatum	0.7	1.1	0.3	1.1
L. complanatum type	0.7	0.6	0.6	1.8
Equisetum	0.2	0.3		
Botrychium		0.6		
Polypodiaceae	5.0	7.2	8.4	5.2
Unidentified	0.5	1.1	0.3	
AQUATICS				
Cyperaceae	1.4	2.5	2.8	6.2
Typha	0.2	0.6		
Potamogeton	0.2		0.9	
Sphagnum	0.2	0.3		0.3

Percentages are based on total pollen excluding aquatics equal to 100.
+ = present

PALYNOLOGICAL REPORT NO. 79-13

Date: June 11, 1979

Locality: Leak Lake, 2.5 km north of Parrsboro, Nova Scotia.

Lat.: 45°26.2' N **Long.:** 64°21' W

NTS: 21 H/8

Submitted by: Daryl Wightman

Field No.: Leak Lake Core 4

Lab No.: PL-78-71

Description of Sample: Lake sediment core

Results and Interpretation:

A preliminary analysis of Leak Lake Core # 4 involving 12 samples yielded the accompanying pollen profiles: a relative frequency and a pollen concentration diagram. The latter includes only the most abundant taxa. The purpose of the exercise was to assess the reliability of the 3 radiocarbon dates obtained on the core as shown to the left of the pollen diagram on the stratigraphic column.

The basal red silt and clay has a very low pollen concentration with an assemblage dominated by diploxylon type pine (*Pinus banksiana/resinosa*) pollen with smaller percentages of other tree pollen. Willow (*Salix*) pollen increases in abundance toward the upper boundary of this sediment.

In the overlying grey-green silty gyttja the pollen concentration is greater and pine pollen is less abundant. Willow and sedge (Cyperaceae) are abundant. Birch (*Betula*) pollen is present as are a variety of herbaceous types. Towards the top of this layer, birch and then spruce (*Picea*) pollen reach maximum values and shrub and herb pollen values decline.

Pollen concentration values decline again in the grey clay unit. Spruce and birch pollen percentages are lower whereas pine pollen increases slightly. Aspen or poplar (*Populus*) reaches a low maximum as do some of the clubmosses (*Lycopodium*) especially *Lycopodium annotinum*. Sedge pollen is more abundant and fern spores (Polypodiaceae) attain peak values. Most of the fern spores are oak-fern (*Dryopteris disjuncta*).

Spruce pollen increases again just above the base of the more organic upper clayey gyttja as does birch. Shrub and herb percentages decline to very low values. However, spruce is replaced by pine pollen of the haploxylon white pine (*Pinus strobus*) type accompanied by oak pollen. Pollen concentration values are very high in this lake sediment.

Comparison with other pollen diagrams from the Maritime Provinces reveals many similarities. This is especially true for the Basswood Lake profile from southwestern New Brunswick (Mott, 1975). The sediment stratigraphy of the two sites is also very similar. An herbaceous/shrub pollen zone is evident in the basal part of both diagrams with successive maxima in willow, *Artemisia* and sedge and declining amounts of jack pine type pollen. Very low pollen concentrations suggest an herbaceous tundra environment with concentrations of willow at suitable sites. Birch, probably shrub birch (*Betula glandulosa*) then became more plentiful, and pollen production in general increased. A prominent *Populus* pollen peak is not present at the Leak Lake site as it is at Basswood Lake. These pollen assemblages are delimited as Zones 9 and 8 in the New Brunswick profile.

Spruce trees then began to invade the area to some degree as evidenced by the increase in spruce pollen, but although pollen values increased generally, not enough is present to indicate closed forest conditions. An abrupt change in lithology with deposition of grey clay of low organic content then occurred at both sites and pollen values, especially spruce, decline considerably. The pollen assemblage in this clay unit is characterized by a dominance of sedge pollen and fern spores. Unfortunately, the fern involved at the New Brunswick site was not identified to species but it was a polypodiaceous type spore as is the case at Leak Lake. Above the clay layer spruce shows a resurgence especially at Leak Lake as does birch which at this time was probably paper birch (*Betula papyrifera*). White pine then began to invade the area surrounding both sites. Therefore, pollen zones 7, 6 and 5 of the Basswood Lake profile also appear to be present at Leak Lake.

Since the pollen sequence and stratigraphy at Leak Lake are amazingly similar to the Basswood Road Lake site, the chronology at the latter can be used to assess the reliability of the Leak Lake radiocarbon dates, although it does not necessarily follow that pollen zones are of equivalent age at different sites. The radiocarbon date of 9715 ± 200 (DAL-314) for the base of the upper gyttja appears reasonable when compared with the date of 9460 ± 220 (GSC-1643) for the zone 5-6 boundary in New Brunswick. The date DAL-313, 9580 ± 310 for the base of the lower silty gyttja horizon containing a tundra type pollen assemblage is far too young for this level and is considered spuriously young. If the correlation between sites is valid, according to the New Brunswick profile an age of about 11,300 would be expected for the top of the lower silty gyttja horizon at the spruce pollen maximum. An age of 12,600 years B.P. was obtained at Basswood Lake for the zone 9-8 boundary which, if the correlation is correct, also occurs at Leak Lake at about the depth of the 12,900 ± 160 (GSC-2728) year B.P. date. Therefore, given the statistical errors involved in the dates, they are equivalent. However, caution should be exercised in accepting these old dates at face value, especially in areas of Carboniferous age bedrock where the possibility of contamination by old carbon is great.

The significance of the grey clay layer is of interest because of the occurrence of such a layer at two widely separate sites. The clay layer at Basswood Lake was originally interpreted as a local phenomenon resulting from local erosion and washing of clay into the lake (Mott, 1975). A similar stratigraphic occurrence at Leak Lake and at other sites in New Brunswick (pollen not yet analyzed) suggests a regional phenomenon with climatic change implications. The well known Younger Dryas Stadial (10,950 to 10,000 years B.P.) of northwestern Europe (Mörner, 1970) and Britain (Coope et al, 1971) is a possible correlative judging by the dates obtained thus far. A climatic deterioration just as spruce trees began moving into the area with a consequent decline in vegetation cover and hence greater erosion, would cause deposition of a clay layer. When the climate warmed again, spruce continued to advance into the area followed eventually by more thermophilous trees.

More detailed pollen work and radiocarbon dating are required on this and other cores before the detailed fluctuations, if such did in fact occur, are worked out.

Coope, G.R., Morgan, Anne, and Osborne, P.J. 1971. Fossil Coleoptera as indicators of climatic fluctuations during the Last Glaciation in Britain. Palaeogeogr., Palaeoclimatol., Palaeoecol., 10: 87-101.

Mörner, Nils-Axel. 1973. Climatic changes during the last 35,000 years as indicated by land, sea, and air data. Boreas, 2: 33-54.

Mott, R.J. 1975. Palynological studies of lake sediment profiles from southwestern New Brunswick. Can. J. Earth Sci., 12: 273-288.

RJMott

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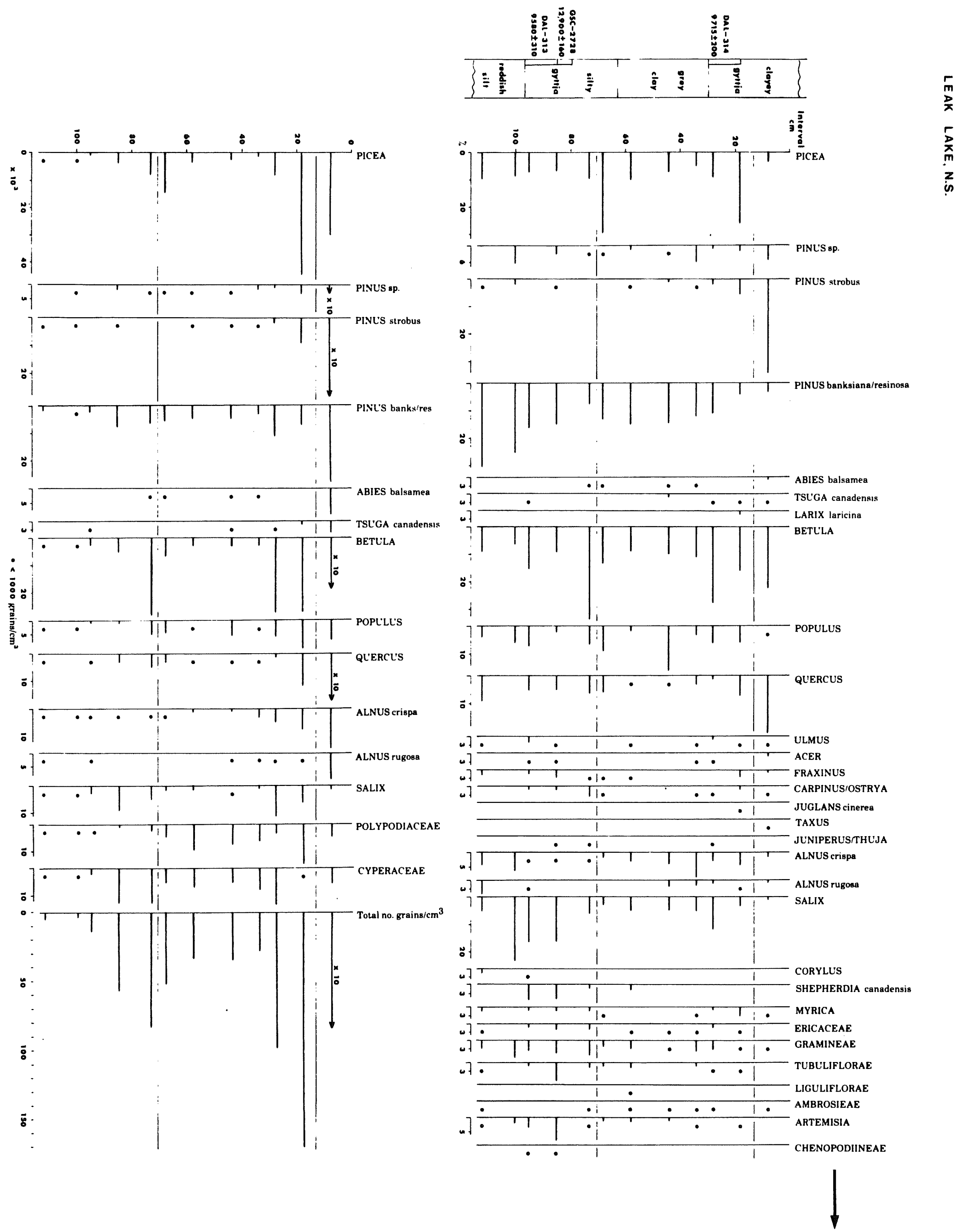
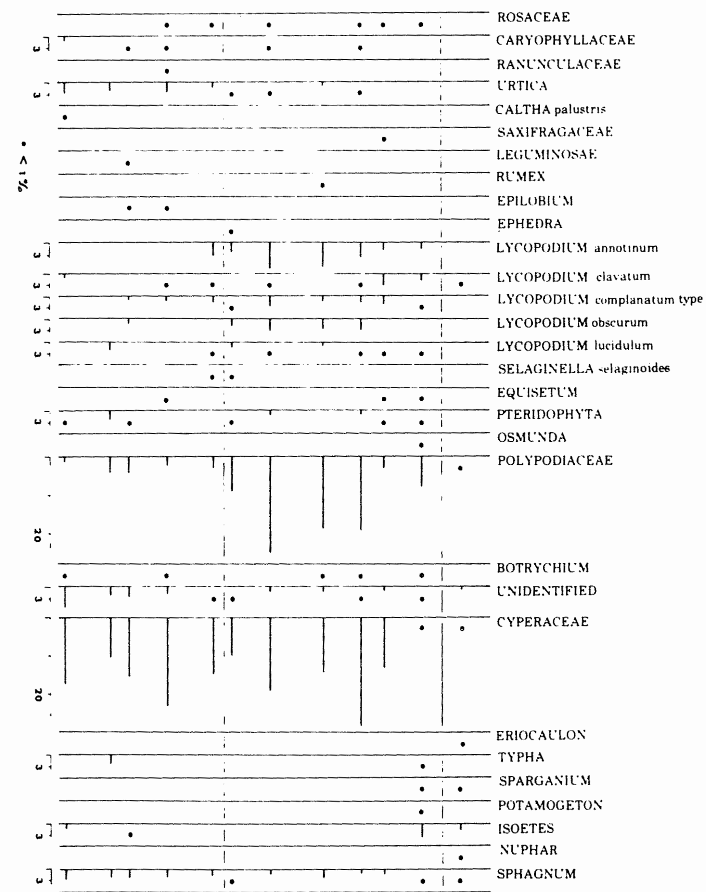


LEAK LAKE, N.S.
OSC-2728 12,900±160
DAL-313 9380±310
DAL-314 9715±200
reddish silt
gyttja
silty
clay
grey
gyttja
clayey
Interval cm
PICEA
PINUS sp.
PINUS strobus
PINUS banks/res
PINUS banksiana/resinosa
ABIES balsamea
TSUGA canadensis
LARIX laricina
BETULA
POPULUS
QUERCUS
ALNUS crispa
ALNUS rugosa
SALIX
POLYPODIACEAE
CYPERACEAE
Total no. grains/cm^3
ULMUS
ACER
FRAXINUS
CARPINUS/OSTRYA
JUGLANS cinerea
TAXUS
JUNIPERUS/THUJA
CORYLUS
SHEPHERDIA canadensis
MYRICA
ERICACEAE
GRAMINEAE
TUBULIFLORAE
LIGULIFLORAE
AMBROSIEAE
ARTEMISIA
CHENOPODIINEAE
x 10^3
• < 1000 grains/cm^3
x 10
%


ROSACEAE
CARYOPHYLLACEAE
RANUNCULACEAE
URTICA
CALTHA palustris
SAXIFRAGACEAE
LEGUMINOSAE
RUMEX
EPILOBIUM
EPHEDRA
LYCOPODIUM annotinum
LYCOPODIUM clavatum
LYCOPODIUM complanatum type
LYCOPODIUM obscurum
LYCOPODIUM lucidulum
SELAGINELLA selaginoides
EQUISETUM
PTERIDOPHYTA
OSMUNDA
POLYPODIACEAE
BOTRYCHIUM
UNIDENTIFIED
CYPERACEAE
ERIOCAULON
TYPHA
SPARGANIUM
POTAMOGETON
ISOETES
NUPHAR
SPHAGNUM
• < 1%
5
20

PALYNOLOGICAL REPORT NO. 79-15

Date: July 23, 1979.

Locality: Diligent River, Nova Scotia.

Lat.: 45°24' N (approx.) **Long.:** 64°26' W (approx.)

NTS: 21 H

Submitted by: Darly Wightman

Field No.: Diligent River Core 3

Lab No.: PL-79-32

Description of Sample: Lake sediment core

Results and Interpretation:

Preliminary analysis of 8 samples from the basal part of the core from Diligent River, Nova Scotia, yielded the results compiled in the accompanying diagrams. A percentage diagram and an abbreviated pollen concentration diagram are attached. Pollen was abundant even in the basal pebbly clay as can be seen from the pollen concentrations. A basal date of 11,555 $\pm$ 230 (DAL-301) years B.P. was obtained from a core collected previously. Presumably the increment dated corresponds to the basal organic sediment increment immediately above the pebbly clay of this core, although the depth below the sediment/water interface was not equivalent in both cores.

Three pollen zones can be distinguished and are delineated on the diagrams by horizontal lines. The basal zone is dominated by birch (*Betula*), grass (Gramineae) and sedge (Cyperaceae) pollen and only small percentages of tree pollen. Spruce (*Picea*) pollen increases abruptly in the overlying zone, and birch, although still abundant, declines considerably. Grass and sedge and herbs in general are less abundant. Spruce decreases upward as birch increases again. Poplar/aspen (*Populus*) pollen forms a small peak and other tree pollen is slightly more abundant. Clubmoss (*Lycopodium ssp.*) and ferns (Polypodiaceae) are also more plentiful. Total concentration declines slightly at first but then increases again upward. White pine (*Pinus strobus*) and oak (*Quercus*) pollen increase and spruce and birch decline in the upper zone of the diagram.

The above sequence of changes in the pollen assemblages is similar to that obtained from Leak Lake (see Palynological Report No. 79-13) and from other sites in Nova Scotia (Hadden, 1975; Livingstone, 1968) and New Brunswick (Mott, 1975). From the pollen data obtained, it is obvious that the older assemblages of the Leak Lake site are not represented at Diligent River, and therefore, a younger basal radiocarbon date would be expected. The date of 11,555 ± 230 (DAL-301) is not unreasonable for a pollen assemblage pre-dating the spruce pollen maximum. In New Brunswick the spruce pollen maximum dated about 11,300 years B.P. and at nearby Folly Bog the post spruce pollen maximum dated 10,764 ± 101 years B.P. (Livingstone, 1968).

A mineral sediment layer above the basal organic sediment similar to that occurring at Leak Lake is not present at Diligent River, but this phenomenon may be exaggerated at some sites and not others. A slight decline in total pollen concentration does occur at the spruce maximum at Diligent River which may correspond to the prominent decline at Leak Lake. The evidence at Diligent River for a climatic deterioration as postulated for Leak Lake is very weak. However, circumstances at various sites may have been different enough to cause a climatic deterioration to be reflected differently at these sites. An alternative hypothesis is that the basal radiocarbon date is slightly too old, and in fact, the basal herb-shrub zone occurred during the colder recession and spruce became abundant after the climate began to warm again. More work is required throughout the maritimes before this critical time can be delineated correctly.

REFERENCES

Hadden, K.A. 1975. A pollen diagram from a postglacial peat bog in Hants County, Nova Scotia. Can. J. Bot., 53/1, 39-47.

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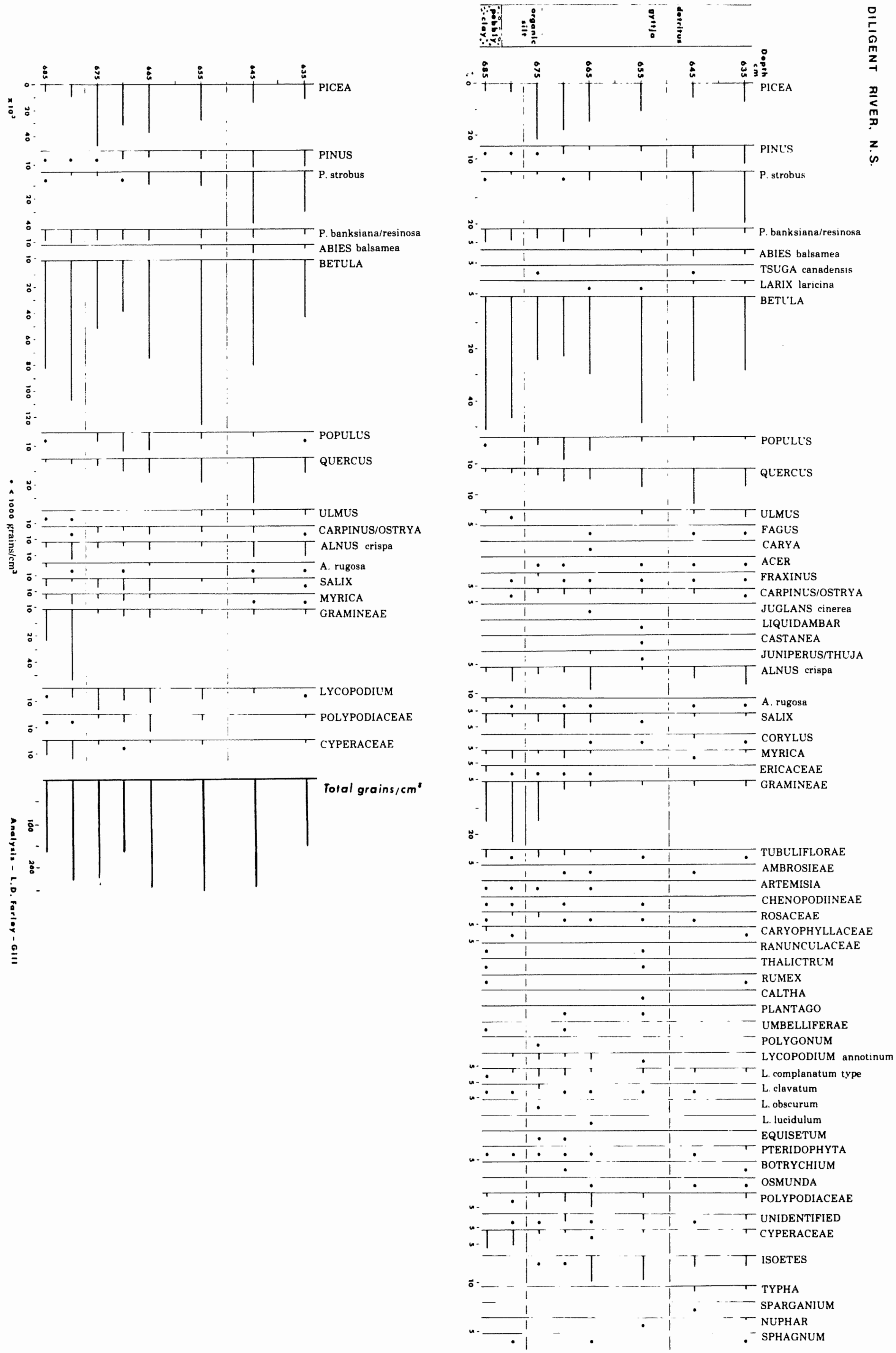
RJ Mott.

R. J. Mott,
Quaternary Paleoecology Lab.

DILIGENT RIVER, N.S.



NOVA SCOTIA DEPARTMENT OF MINES AND CANADA DEPARTMENT OF REGIONAL ECONOMIC EXPANSION

LEGEND FOR THE GEOLOGICAL MAP OF THE COBEQUID HIGHLANDS, NOVA SCOTIA

EAST AND WEST SHEETS

1978

STRATIFIED ROCKS ADJACENT TO COBEQUID HIGHLANDS

MESOZOIC

TRIASSIC

FUNDY GROUP

- **TSB** — *SCOTS BAY FORMATION: red and green siltstone, calcareous siltstone*
- **TNM** — *NORTH MOUNTAIN BASALT: basalt flows*
- **TB** — *BLOMIDON FORMATION: red siltstone, wacke*
- **TW** — *WOLFVILLE FORMATION: red quartz and lithic wacke, siltstone, polymictic conglomerate*
- **TBW** — *BLOMIDON AND WOLFVILLE FORMATIONS: undifferentiated*

PALEOZOIC

UPPER CARBONIFEROUS

(Westphalian B to D)

PICTOU GROUP

- **CP** — *Undivided; grey, red, siltstone, wacke*

CUMBERLAND GROUP

- **CC-U** — *UPPER PART: red and grey siltstone, wacke, polymictic conglomerate*
- **CC-L** — *LOWER PART: red and grey siltstone, lithic wacke, polymictic conglomerate;* PC-LC *coal bearing, grey siltstone, wacke*
- **PCP** — *PICTOU AND CUMBERLAND GROUPS: undifferentiated*

(Namurian to Westphalian A)

No stratigraphic order implied between areas.

Eastern Area

RIVERSDALE GROUP

- **CBP** — *BOSS POINT FORMATION: grey to red lithic wacke, siltstone, shale*
- **CM** — *MILLSVILLE CONGLOMERATE: brown-red polymictic conglomerate, lithic wacke*

North Central Area

RIVERSDALE GROUP

- **CBP** — *BOSS POINT FORMATION: grey to red lithic wacke, siltstone, shale*
- **CCl** — *CLAREMONT FORMATION: brown red arkosic conglomerate; minor red wacke, siltstone*
- **CMb** — *MIDDLEBOROUGH FORMATION: dusky-red siltstone, shale, wacke, minor conglomerate*

Southern Area

RIVERSDALE GROUP

- **CPa** — *PARRSBORO FORMATION: red and grey siltstone, shale, wacke; red polymictic conglomerate at base*
- **CWB** — *WEST BAY FORMATION: red, greenish-grey siltstone, wacke, conglomerate*

(Namurian)

CANSO GROUP

- **CCa** — *Undivided; grey wacke, siltstone, some calcareous siltstone, minor red siltstone*

LOWER CARBONIFEROUS

(Viséan)

WINDSOR GROUP

- **CW** — *Undivided; grey calcareous siltstone, grey limestone, gypsum*

(Tournaisian and (?) older)

HORTON GROUP

- **CH** — *Undivided; grey, reddish-brown shale, siltstone, wacke*

STRATIFIED ROCKS OF THE COBEQUID HIGHLANDS

PALEOZOIC

(Late Visean to Namurian)

- **CLo** — *LONDONDERRY SUCCESSION: grey, green quartz wacke; red, purple siltstone*

DEVONIAN TO LOWER CARBONIFEROUS

- **DCN** — *NUTTBY SUCCESSION: grey, red siltstone, wacke, red polymictic conglomerate, rhyolitic volcanic rock*

DEVONIAN

(Eifelian to Famennian)

RIVER JOHN GROUP

- **DRJ-U** — *UPPER UNIT: red polymictic conglomerate, lithic wacke*
- **DRJ-L** — *LOWER UNIT: basalt, grey to red lithic wacke, polymictic conglomerate*

SILURIAN TO DEVONIAN

(Llandoverian to Pridolian and younger)

- **SDPP** — *PORTAPIQUE - PARRSBORO SUCCESSION: greyish-green quartz wacke, siltstone, red siltstone, rhyolitic and andesitic volcanic rock*

SILURIAN

(Llandoverian to Pridolian)

- **SE**, **SE-A**, **SE-B** — *EARLTOWN SUCCESSION: (In ascending order)* SE-1 *grey siltstone, mudstone;* SE-2 *rhyolitic, dacitic and some andesitic flows, tuff,* SE-3 *rhyolitic, dacitic flows;* SE-4 *blue-grey wacke siltstone, tuff*
  SE-A *rhyolitic and dacitic flows, tuff;* SE-B *intermediate tuff; volcanic wacke*

SILURIAN (?) (May be in part younger)

- **SA** — *ADVOCATE SUCCESSION: grey quartz wacke, siltstone, intermediate tuff, basalt, rhyolite, minor quartz arenite; granitic clast conglomerate*

AGE OF ROCKS UNKNOWN

- **PHE** — *HARRINGTON - EAST RIVER SUCCESSION grey quartz wacke, siltstone, granitic clast conglomerate, tan quartz arenite, shale; minor tuff*
- **PNR** — *NORTH RIVER SUCCESSION: andesitic tuff, quartz wacke, siltstone, minor limestone*

PROTEROZOIC OR PALEOZOIC

No stratagraphic order implied

BASS RIVER METAMORPHIC SUITE
(No stratagraphic order implied)

- **PPB** — PPB-1 *andesitic meta-volcanic rock;* PPB-2 *quartzite, biotite schist;* PPB-3 *hornblende and feldspar-chlorite gneiss;* PPB *undifferentiated*

MOUNT THOM COMPLEX:

- **PPT** — *Biotite-muscovite ± garnet schist, amphibolite, granite gneiss*

PROTEROZOIC

HADRYNIAN

- **HJ** — *JEFFERS SUCCESSION: chlorite-rich tuff, green tuffaceous wacke, siltstone*

IGNEOUS ROCKS OF THE COBEQUID HIGHLANDS

Letter designation identifies pluton

CARBONIFEROUS

- **Cg** — *Hornblende granite;* O *Salmon River,* P *West North River,* Q *Gain Brook,* R *Chain Lakes,* S *Byers Lake,* T *Hart Lake,* U *Pleasant Hills,* V *Chignecto*

DEVONIAN TO CARBONIFEROUS

- **DCg** — *Hornblende granite;* I *Shatter Brook,* J *North River,* K *West Moose River,* L *Gilbert Mountain,* M *Hanna Brook,* N *Apple River*
- **DCd** — *Hornblende and/or pyroxene diorite;* A *Gully Brook,* B *Cranberry Brook,* C *Chiganois River,* D *McElman Brook,* E *Gleason Brook,* F *Wyvern,* G *New Prospect,* H *Soldier's Brook*
- **PPgg** — *Granite Gneiss*

HADRYNIAN

- **HJB** — *JEFFERS BROOK PLUTON: hornblende quartz diorite*

SYMBOLS

Bedding (tops known, overturned; tops unknown inclined, vertical) 90 50 50

Flow layering in volcanic rock (inclined, vertical) . . . 80

Structures in the Cobequid Highlands

Slaty or fracture cleavage associated with F_1 folds . . . 50

Anticline, F_1, upright / *Syncline, F_1, upright, overturned* — *arrow indicates plunge direction* . . .

Local cataclastic fabric in plutonic rocks 40

Cleavage and foliation in Jeffers Succession 60

Schistosity or metamorphic foliation associated with T_A, B_A, F_X folds 60

Antiform F_Y, arrow indicates plunge direction . . . . .

Structures north and south of the Cobequid Highlands

Anticline, upright / *Syncline, upright* / *overturned* — *arrow indicates plunge direction* . . . . .

Faults (defined, approximate, assumed; arrows indicate relative movement; solid circle indicates down thrown side; single arrow represents dip direction) . . . . .

Geological boundaries (defined, approximate, assumed). . . . .

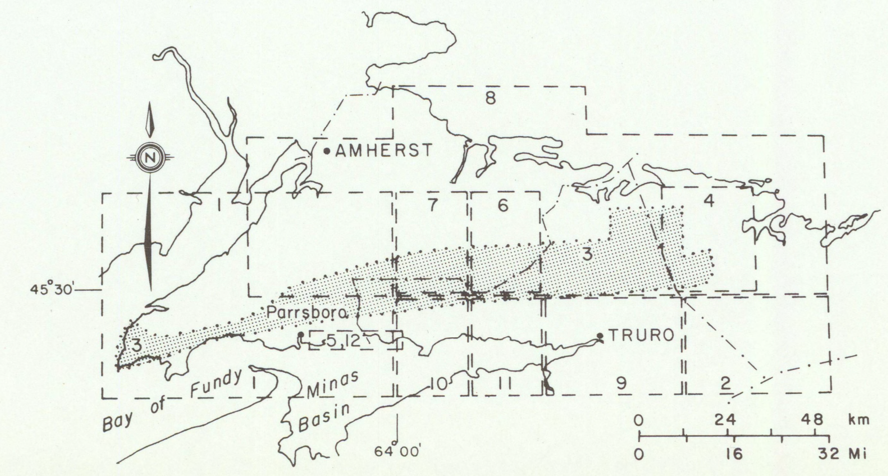
Fossils: (flora or fauna; spores) . . . . . Ⓕ, Ⓢ

Mineral Occurrences . . . . . × Cu
Cu-copper, *Fe*-iron, *Pb*-lead, *Zn*-zinc, *As*-arsenic, *U*-uranium, *Ba*-barite, *Mn*-manganese, *C*-coal

ABBREVIATIONS

CF:	*Cobequid Fault*	LVF:	*Loganville Fault*
LF:	*Londonderry Fault*	BBF:	*Balmoral Brook Fault*
PF:	*Portapique Fault*	NRF:	*North River Fault*
RF:	*Riversdale Fault*	AF :	*Alma Fault*
EF:	*Economy Fault*		

SOURCES OF GEOLOGICAL INFORMATION



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2 *Benson D.G., 1967, Geological Map of Hopewell Area, Nova Scotia; Geol. Surv. Canada Map 1215A.*

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6 *Norman, G.W.H. and Bell, W.A., 1938, Geology of the Oxford Sheet (East Half), Cumberland and Colchester Counties, Nova Scotia; Geol. Surv. Canada Map 409A.*

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11 *Weeks, L.J., 1946, Geology of Londonderry Area, Colchester and Hants Counties, Nova Scotia; Geol. Survey Canada Map 874A.*

12 *Olsen, P. and Donohoe, H.V., 1977, unpublished detailed stratigraphy of the Triassic in the Five Islands region.*

PALEONTOLOGICAL DATA

SPORES:

Identification of spores by M.S. Barss and D.C. McGregor, Geol. Surv. Can., from samples submitted by E.S. Belt, J.W. Gillis, and D.J. Kelly.

Identification of spores by M.S. Barss, Geol. Surv. Can., from samples submitted by H.V. Donohoe.

FOSSILS: STRATIFIED ROCKS OF THE COBEQUID HIGHLANDS

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Harper, C.W., 1973, Brachipods of the Arisaig Group (Silurian-Lower Devonian) of Nova Scotia; Geol. Surv. Can., Bull. 215, 103p.

FOSSILS AND FLORA: STRATIFIED ROCKS ADJACENT TO THE COBEQUID HIGHLANDS

See sources of geological information.

RADIOMETRIC DATA

Rb/Sr whole rock isochrons by R. Cormier and J. Stirling (unpublished, 1977)

Chignecto Pluton (V) *339 ± 22 m.y.* $^{87}Sr/^{86}Sr = 0.7064 \pm 0.0073$

Hart Lake Pluton (T) *331 ± 17 m.y.* $^{87}Sr/^{86}Sr = 0.7076 \pm 0.0095$

Byers Lake Pluton (S) *331 ± 27 m.y.* $^{87}Sr/^{86}Sr = 0.7097 \pm 0.0093$

I.U.G.S. Subcomission on Geochronology decay constants used (Steiger, R.H. and Jager, E., 1977, Earth and Planetary Sci. Letters, vol. 36, p. 359-362).

This is a preliminary map, subject to modification

GEOLOGY MAP OF THE COBEQUID HIGHLANDS (EAST HALF)

by

Howard V. Donohoe Jr. and Peter I. Wallace

1978

JOINT PROJECT

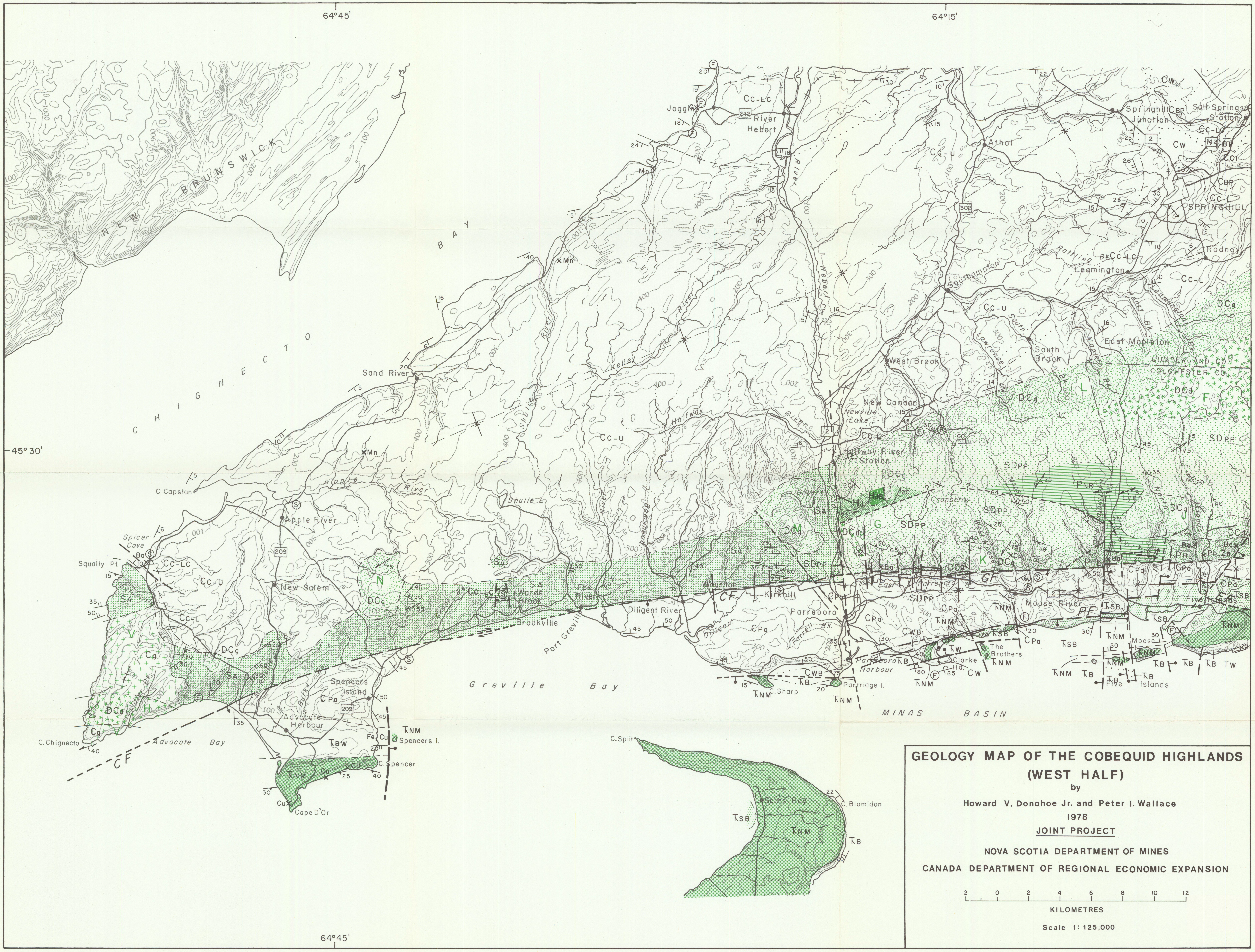
NOVA SCOTIA DEPARTMENT OF MINES

CANADA DEPARTMENT OF REGIONAL ECONOMIC EXPANSION

2 0 2 4 6 8 10 12

KILOMETRES

Scale 1: 125,000



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