LATE QUATERNARY HISTORY OF
THE HERMITAGE BAY AREA, NEWFOUNDLAND

By

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LATE QUATERNARY HISTORY OF
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ABSTRACT

Geomorphologic, sedimentologic and stratigraphic investigations have served to provide an interpretation of the late Quaternary history and environments in the Hermitage Bay area, Newfoundland. The area has been affected by at least two major glacial events, the most recent being much less extensive than the earlier one. The earlier event completely inundated the field area, contributing to the molding of large bedrock hills into southwards oriented stoss and lee forms, and eroding broad U-shaped valleys. It is represented by tills at Pass Island Tickle, Seal Cove, and near Salmonier Cove Pond, and a lower till in a two till section at Trout Hole Falls Community Park. Foraminifera recovered from an "old" till in a coastal section at Seal Cove have been assigned a tentative age of ~ 70,000 years.

The most recent glacial event, considered Late Wisconsin in age, was not all encompassing. Differences in till cover, weathering zonations and ice marginal features indicate that a small ice cap centred north and east of the head of Hermitage Bay was separated by the Garrison Hills from a main island icecap to the north. The southern limit of the large island icecap can be traced eastwards to north of the Burin Peninsula where it has been recognized by Tucker (1979). This limit is proposed to represent the maximum extent of the Late Wisconsin glaciation rather than the recessional position of a more extensive advance as suggested by Jenness (1960). Isopleths of Late Wisconsin postglacial emergence show that maximum uplift (30-32 m) occurred in the Hermitage-

Sandyville area.

During deglaciation a tongue of ice feeding down the Little River valley created a large ice-dammed lake in northern Bay d'Espoir which was subsequently infilled by glaciofluvial/lacustrine sediment. Deposits at Conne River suggest that the style of sedimentation was a series of low slope, prograding deltas advancing over glaciolacustrine bottom sediments.

ACKNOWLEDGEMENTS

I would like to thank Dr. S.B. McCann who acted as my supervisor for this project. He suggested a topic which proved to be interesting and stimulating both scientifically and, in the field, culturally. Chris Tucker, coming from and working on the island, provided invaluable advise throughout the project. Peter Hale and Chris Smart were always ready and willing to discuss problems and ideas. Dr. Gustav Vilks, Atlantic Geoscience Centre, Halifax, identified the foraminiferal assemblage and provided a paleoenvironmental interpretation. Mariline and David Leckie were my able-bodied field assistants, although Mariline's steps were somewhat shortened by Serge. Mrs. Phoeboe Savoury, Belleoram, looked after us with characteristic Newfoundland hospitality. Bob Bignell provided photographic services and Cathy Moulder typed the thesis. Financial support was provided by National Research Council Grant No. A5082 and Energy Mines and Resources Research Agreement No. 2239-4-154-77, held by Dr. McCann. A National Research Council Postgraduate Scholarship made sure my family did not have to subsist on beans during the year. Most of all, my appreciation goes to Mariline and Serge; they put up with a lot.

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CHAPTER 1

INTRODUCTION

1.1 Statement of the Problem

This thesis presents the results of a study of glacial phenomena in the Hermitage Bay area, southern Newfoundland. It is concerned with the geomorphologic, sedimentologic and stratigraphic relationships as they relate to the late Quaternary history of the area. Specifically, the study involves the mapping and investigation of glacial erosional and depositional forms, the sedimentologic interpretation of stratigraphic sections, a preliminary analysis of various rock weathering criteria and periglacial phenomena as relative age indicators, and the documentation and interpretation of raised marine features. The topic is of particular interest and relevance because of recent controversy regarding the maximum extent of the Late Wisconsin glaciation on the east coast of North America (e.g., Blake, 1970; Flint, 1971; Grant, 1977a; England and Bradley, 1978; Ives, 1978).

1.2 Purpose and Scope

The purpose of this study is threefold. It is the second and final part of a project concerned with the glacial history of a portion of south coastal Newfoundland. In conjunction with the work of Tucker (1979) on the adjacent Burin Peninsula, it is intended to provide a regional

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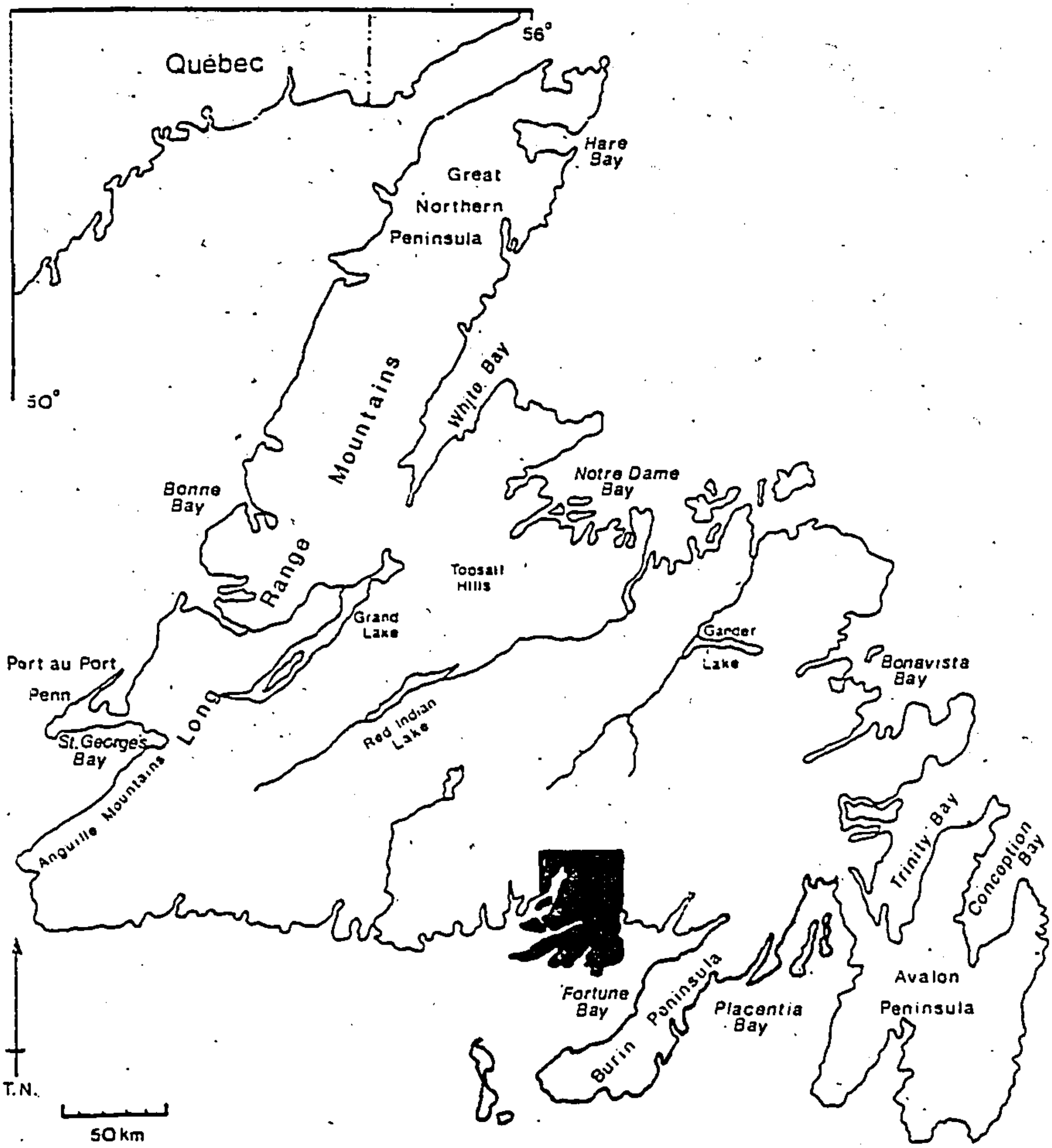


Hermitage Bay area, Newfoundland

interpretation of the Late Quaternary events which affected this part of the island. The second and principle purpose is to determine, on a local scale, the nature and origin of glacial and proglacial deposits, and the sequence of events responsible for their deposition. Finally, in light of new and "rediscovered" evidence regarding the maximum extent of the last glaciation (see Ives, 1978, for review), the third purpose of this study is to determine the limits of Late Wisconsin ice coverage in the Hermitage Bay area.

Previous Quaternary research in Newfoundland is limited and has been carried out on a piecemeal basis. Much of the island remains to be investigated, even in a reconnaissance fashion. This study contributes to the body of information concerning the glaciation and deglaciation of Newfoundland and adds to the larger picture regarding the extent of the Late Wisconsin glaciation in eastern North America. In addition, it provides a local model for glaciolacustrine-deltaic sedimentation which will serve as a refinement to the more general glaciolacustrine model (Church and Gilbert, 1975; Shaw, 1977).

1.3 Description of the Field Area

The field area (Fig. 1.1) is located on the south coast of Newfoundland, west of Bay du Nord and east of Pass Island. It includes the terrain surrounding Bay d'Espoir, generally east of Salmon River and west of Bay du Nord River, and the peninsula projecting into Fortune Bay west of Belle Bay and east of the mouth of Bay d'Espoir. It contains 2396 km^2 of territory (excluding marine waters). At the 1:50,000 scale the area is covered by N.T.S. map sheets 1M/13 (St. Alban's), 1M/12

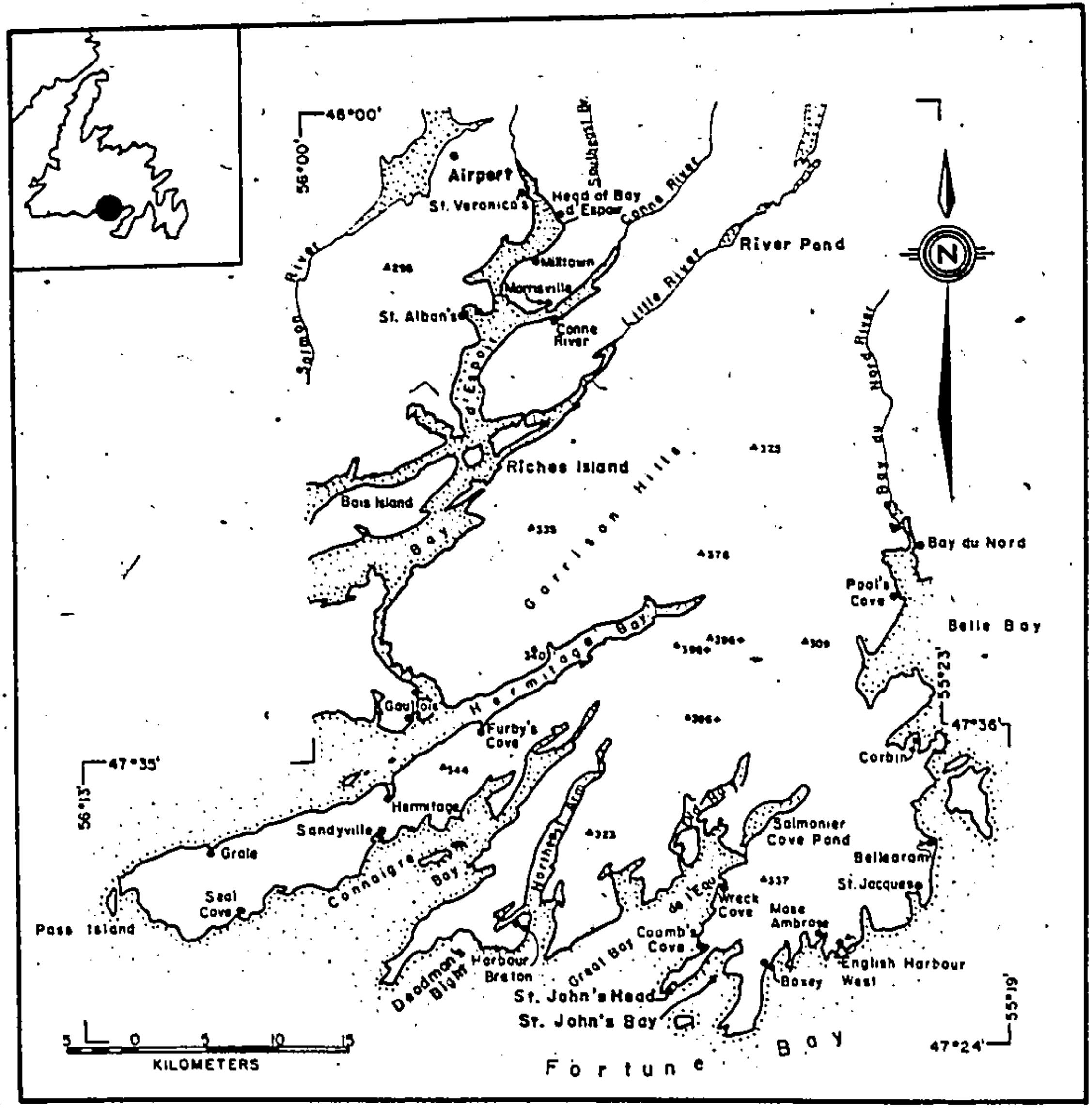


Figure 1.1: Place names of most locations discussed in this thesis.

(Gaultois) and parts of 11P/9 (Facheux Bay), 11P/8 (Pass Island), 1M/5 (Harbour Breton) and 1M/16 (Point Enragee). It covers 1M (Belleoram) and 11P and 11-I (Burgeo) of the N.T.S. 250,000 map sheets. Bounding co-ordinates of the field area are shown in Figure 1.1 (fold-out at back of thesis).

Local bedrock is a highly faulted melange of several rock types including two suites of volcanics, three of granite, granodiorite and granite gneiss, and eight of clastic sedimentaries (see Fig. 2.2 for summary map). More complete bedrock descriptions will be included in various sections of the thesis wherever pertinent to the discussion. Detailed accounts of local geology are to be found in papers by Jewell (1939), Widmer (1950), Anderson (1965), Williams (1967, 1971, 1978) and Coleman-Sadd et al (1979). The large scale topography is characterized by fiords and peninsulas largely oriented northeast-southwest and north-south. Several of the fiords parallel fault lines and are probably of tectonic origin although glacially modified. The height of land in the Hermitage Bay area occurs at about 396 m asl on summits south of the head of Hermitage Bay (Fig. 1.1). The Garrison Hills, north of Hermitage Bay, form a ridge which runs northeast-southwest across the study area. Major drainage routes are the Salmon River (west), Southeast Brook, Conne River, Little River, Salmon River (east) and Bay du Nord River.

1.4 Methods of Investigation

A set of parameters for describing Quaternary sediments, weathering criteria, and raised marine features was established before going into the field. These were based upon descriptions of procedures commonly used in Quaternary investigations (Scott and St. Onge, 1968; Dreimanis, 1971; Andrews, 1971; Gray, 1975).

Texture. Estimates of particle size were determined in the field using the "finger test" by which amounts of sand, silt, and clay can be determined (Folk, 1974). The technique is to moisten and roll a small volume of the sample between the fingers until it reaches the liquid limit. At this point the presence of sand will be indicated by a gritty feeling between the fingers. Silt causes a smooth sensation, as if the fingers were coated with talcum powder, whereas clay causes the fingers to stick. With practice, semi-accurate estimates of sand, silt and clay ratios can be obtained by this method.

Several samples were returned to the laboratory for more rigourous textural analysis. Standard sample preparation and analytical techniques were followed using methods outlined by Folk (1974). The sample was first wet sieved; the sand sized fraction was then determined by dry sieving, and silt-clay portion pipetted. Results were plotted and percentage of sand, silt and clay determined (% sa., % si., % cl.). Size boundaries used are those defined by Wentworth (1922): sand, 2.0-0.62 mm (-1-4Ø(phi)) silt, 0.62-0.0039 mm (4-8 Ø); clay, finer than 0.0039 mm (> 8 Ø). For tills, only the matrix was analyzed, as impractical volumes of samples would be required to obtain proper statistical representation of pebble, cobble, and boulder content.

Color. A visual description of sediment color was recorded, followed by a Munsell color designation of a moist sample (e.g., brownish black, Munsell Color, 10YR:3/2).

Roundness. Roundness refers to the sharpness of corners and the edges of a clast. The degree of particle roundness was determined in the field by visual comparison with a chart subdivided into very angular, angular, subangular, subrounded, rounded and well rounded categories (Powers, 1953).

Fabric. Fabric is the orientation (or lack of it) and imbrication of component rock particles making up a sedimentary deposit. Several studies have shown that stress systems within a glacier will orient elongate particles in either a transverse or parallel position, relative to local ice flow direction, although during deglaciation fabric, especially of supra- and en-glacial material, will become reoriented (Boulton, 1971; Dreimanis, 1976). The purpose of obtaining till fabric data for this project was to provide an indication of ice flow directions at various sites in the study area in order to supplement other directional evidence.

The method was to clear an approximately 1.0 m^2 sampling site in the till and to dig into this, recording the orientation and direction of dip of 25 to 50 clasts using a Brunton compass. Only pebbles greater than 1 cm in length and having a ratio of the long axis to intermediate axis of 2:1 were used (Andrews and Smith, 1970; Andrews, 1971). The data were plotted as two dimensional rose diagrams and the vector mean and other statistical data calculated as outlined by Krumbein (1939). Similar techniques were used to determine the paleocurrent of glaciofluvial gravels (Rust, 1972).

Structure. The structure of particle aggregates in till can be described as platy, blocky, columnar, friable or massive. Marcussen (1975) was able to differentiate supraglacial flow till from lodgement till partially on the basis of structure. Flow till often has a "loose incoherent character" whereas basal till is much more massive and indurated. Fissility is usually an indicator of lodgement till (Dreimanis, 1976).

Sampling. Till samples were obtained on a one site/one sample basis which will suffice when the tills or other sediments are essentially non-bedded. Wherever multiple unit sections were encountered each unit was sampled. As the primary purpose of sampling the tills was to provide a general indication of the nature of the sediment and not strictly process oriented a more regimented sampling plan was considered unnecessary. The sampling of sediments at the Conne River bank was more specific as textural data were required from specific strata and intervals.

Weathering data were collected at random throughout the study area wherever suitable surface outcrops were exposed, although a special effort was made to obtain measurements from summits of hills whenever they were traversed. The height of raised marine features was measured by aneroid levelling wherever access permitted.

Contacts. Contacts between lithologic units are defined as sharp or gradational. A sharp contact is defined by a sudden change in texture, color, or structure. Such a contact may in some instances represent an erosional surface. A gradational contact is not well defined in an outcrop, there being a gradual transition from one unit to the adjacent one. In some units with gradational contacts, material in the underlying unit may have been eroded, incorporated into and deposited as part of the overlying unit. Alternatively there may have been a gradual change in the depositional process or a change in the availability of the sediment.

Diagrams. Reconstructions of stratigraphic sections are drawn to scale from pace and compass bearings, field measurements and photography, topographic maps and air photos.

1.5 Previous Quaternary Investigations in the Hermitage Bay Area.

Recently, there have been several well written summaries and syntheses regarding the late Quaternary history of eastern Canada (Grant, 1976, 1977b; Tucker, 1976, 1979; Ives, 1978) and thus only a brief review is presented here. Ives (1978) reviewed the literature regarding the maximum extent of the Laurentide ice sheet in eastern North America during the "last glaciation" (a period spanning 125,000 to 8,000 years B.P.) and concluded that several coastal areas had never been glaciated during this time. Grant (1976) speculated on the extent of the Late Wisconsin glaciation in the Atlantic provinces of Canada and produced a rough sketch map of the proposed ice limits. The map was subsequently modified and more formally produced in a historical review of Late Quaternary events in Atlantic Canada (Grant, 1977b). In the latter, Grant proposed that, during the Late Wisconsin, Newfoundland supported its own ice cap, or ice cap complex, rather than being overridden by Laurentide ice and, quoting Widmer (1950), added that only local cirque and valley glaciers existed in the Hermitage Hills (sic). Tucker (1976, 1979), restricting his discussion to Newfoundland, provided a comprehensive bibliography and summary of the Quaternary history of the island. Tucker (1979) demonstrated that the maximum extent of the Late Wisconsin glaciation on the Burin Peninsula was to the top of the peninsula and that the southern portions had not been glaciated since the late, mid-Wisconsin by northwards flowing ice from an offshore source. He also found that the zero isobase of postglacial emergence ran through the southeastern tip of the Burin Peninsula, further south than previously reported (Flint, 1940; Jenness, 1960).

The earliest recorded investigation pertaining to the Quaternary geology of the Hermitage Bay area was by Jewell (1939) in a report on the geology and mineral deposits of the Baie (sic) d'Espoir area. He found no evidence to indicate more than one glaciation, suggesting that a large ice cap had completely covered the area and then retreated to an inland position from where it fed valley and fiord glaciers. Faintly varved clays, overlain by sand and gravel at Indian Point and on the Conne River were attributed to several small ice- or moraine-dammed lakes. Jewell commented twice, in his brief section on glaciation, on the lack of striae and grooves on the upland bedrock surfaces, but placed little significance on this. Twenhofel and MacClintock (1940) and MacClintock and Twenhofel (1940) presented an identical scenario to Jewell's for the Baie (sic) d'Espoir area but were more adamant in their contention that a Newfoundland based ice cap, rather than Laurentide ice, was responsible. Twenhofel (1947) stated that the area had recently been completely glaciated and that the upland surfaces had been scraped clean of overburden. Van Alstine (1948) noted granite erratics of Garrison Hills provenance on the northwest side of the Burin Peninsula. He attributed these to an island-based glaciation which passed over the Hermitage Bay area, across Fortune Bay and to beyond the Burin Peninsula. Tucker (pers. comm., 1979) however, believes that their provenance may be from the south and that they were transported by northwards flowing ice.

The most significant piece of work regarding the glaciation of the Hermitage Bay area was a chapter in a doctoral dissertation by Widmer (1950). Widmer believed that the whole of the region had been glaciated during the Wisconsin by a Newfoundland based ice cap which had extended to beyond the Burin Peninsula. The ice cap retreated inland to a position north of the Hermitage Bay area. During "Cary or Mankato times" (15,000 to 8,500 years B.P., Frye and Willman, 1963) "a recurrence of glaciation occurred" whereby the ice cap nourished large valley glaciers feeding into Bay d'Espoir, Hermitage Bay and Connaigre Bay. At this time there was also a phase of local valley glaciation in Harbour Breton Bay, Taylor Bay (Old Bay) and Salmonier Cove Pond. Widmer also levelled coastal strand-lines which he attributed to a series of large offshore proglacial lakes that had occupied Fortune Bay, Placenta Bay and Bay d'Espoir. In support of the proglacial lake hypothesis Widmer cited glaciolacustrine sediments at Conne River, where he counted more than 1600 "varves".

Jenness (1960, 1963), relying on the work of Jewell (1939) and Widmer (1950) for the Hermitage Bay area, proposed that a late Pleistocene glaciation had covered all of eastern Newfoundland as far as the Avalon Peninsula. The ice subsequently retreated rapidly to just inland of the coast where it developed an extensive end moraine system which encircled the eastern half of the island (Fig. 1.2). As deglaciation progressed glaciofluvial deposits radiated from the end moraine and terminated as deltaic deposits at the coast into a higher sea level than present. The end moraine system separated an inner drift zone from a slightly older, outer one. Jenness's inner-outer drift limit passes through the northern portion of the Hermitage Bay area.

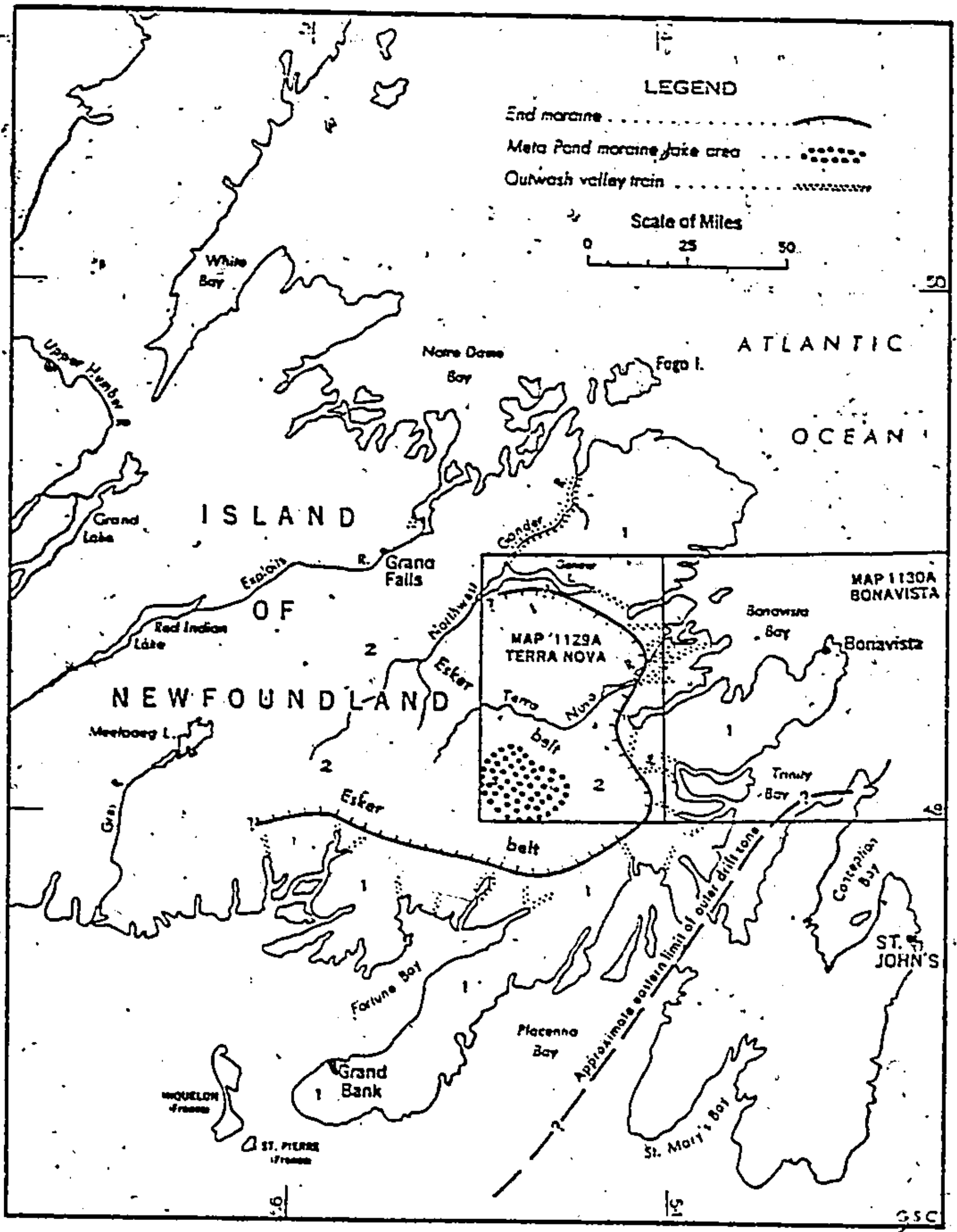


Figure 1.2: General distribution of glacial deposits in east central Newfoundland. Jenness's inner-outer drift zone limit corresponds to the position of the end moraine (Jenness, 1960, 1963)

Grant (1975b), in a brief note, suggested that northwards flowing ice centred in Placentia Bay may have extended across Fortune Bay as far north as Mose Ambrose. He also speculated that late ice in the Garrison Hills may have become isolated from the inland sheet and fed narrow ice tongues terminating in several bays.

Tucker (1979), working on the nearby Burin Peninsula, concluded that the last Newfoundland based ice to have affected the whole of the Burin was the Fortune Bay event, an early Wisconsin, post St. Pierre interstadial glaciation (Fig. 1.3) Before this, an ice mass centered over Placentia Bay flowed northwest probably into Fortune Bay. This event has been assigned a post 38,000 years B.P., or late, mid-Wisconsin age. The final glacial event (Late Wisconsin) to have affected the Burin Peninsula extended to only the northern part of the peninsula.

1.6 Format of Thesis.

Chapter 2 describes the glacial features of the Hermitage Bay area as observed and interpreted in the field and from air photos. Stratigraphic sections and their sedimentologic interpretations, based on facies relationships, are described in Chapters 3 and 4. Chapter 3 considers sections located around the Bay d'Espoir-Conne River area whereas Chapter 4 looks at the more southerly or coastal sections. A local facies model for glaciolacustrine-deltaic sedimentation based on sediments at the Conne River bank is presented in Chapter 5. In Chapter 6, the results of a reconnaissance weathering study are presented and correlations with other weathering zones found in Eastern Canada are proposed. Raised marine features are discussed and interpreted in Chapter 7.

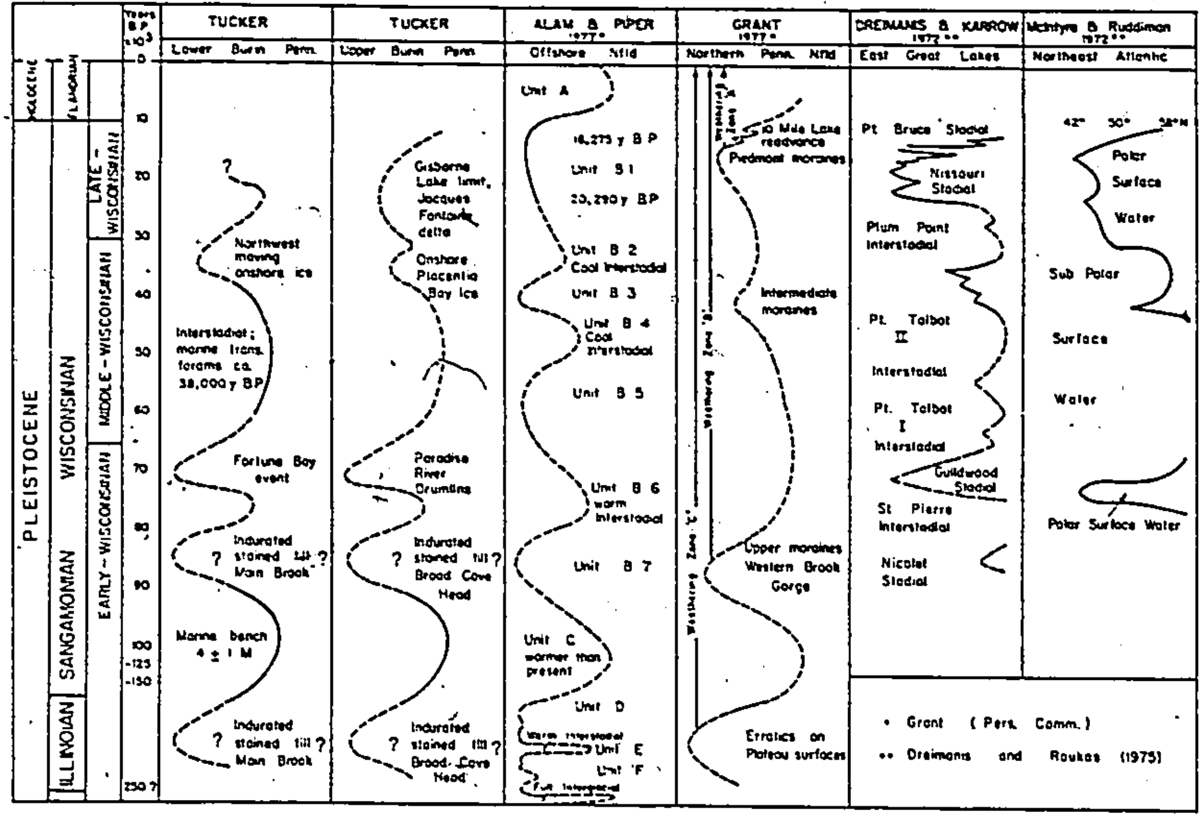


Figure 1.3: Correlation chart of late Quaternary events (Tucker, 1979)

Chapter 8 lists the major conclusions reached in this thesis. Appendix 1 lists the results and locations of striae measurements, bench and terrace elevations and weathering data.

The location of places within the Hermitage Bay area is shown in Figure 1.1, a foldout map at the back of the thesis. The reader is advised to keep the map open for easy referral while reading the thesis. As well, precise location references are given throughout the thesis by using the topographic map sheet number and military grid reference for any given place (e.g., 1M/11: 187652).

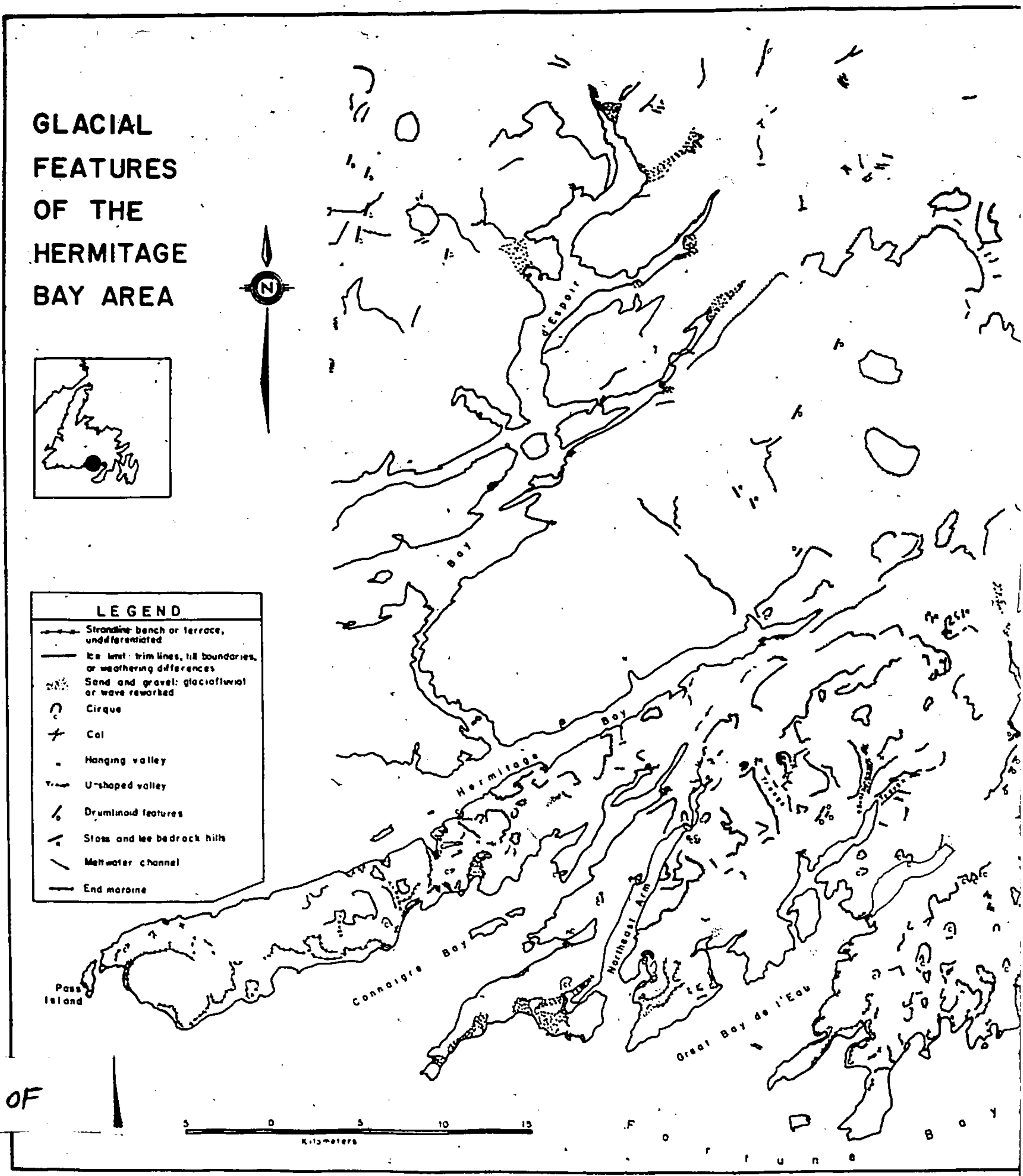
CHAPTER 2

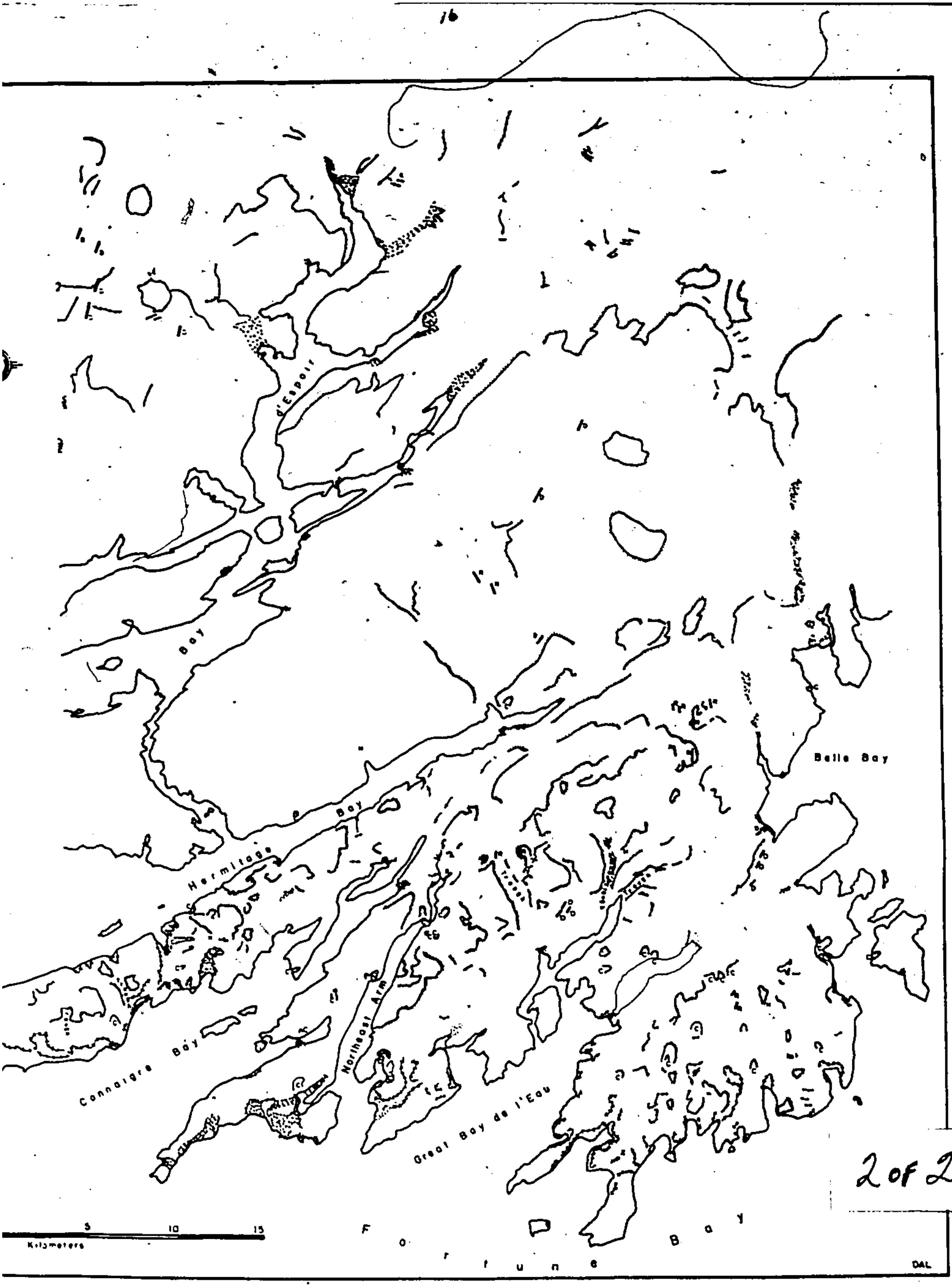
GLACIAL FEATURES

2.1 Introduction

The effects of glacial erosion and deposition can be found in all parts of the Hermitage Bay area. They vary in scale from minute, finely etched striae to large stoss and lee bedrock forms having dimensions of hundreds of meters, and from sparse occurrences of erratic boulders to massive till and outwash deposits tens of meters thick. The following discussion is concerned with these and other glacial features. It is based upon a combination of field observations and air photo interpretation. As a detailed terrain analysis (e.g., Fulton _et al_, 1974; Tucker, 1979) was considered beyond the scope of this project, it was decided to limit mapping to the identification and delimitation of the following glacial features: ice limits, moraine ridges, glaciofluvial outwash, glaciofluvial channels, cirques, glacial troughs and hanging valleys, cols, drumlinoid features, stoss and lee bedrock hills, and abandoned strandlines. The results, shown on Figure 2.1, were identified on 1:50,000 air photos, compiled on 1:50,000 topographic sheets and then photo-reduced.

The term "last glacial event" used below and throughout this thesis refers to the last major glacial advance to have affected most of the Hermitage Bay area, or parts thereof. At this point, no time connotations other than relative are implied. Unless otherwise stated, consider that it is the effects of the last glacial event being discussed.


GLACIAL FEATURES OF THE HERMITAGE BAY AREA
N
LEGEND
Strandline bench or terrace, undifferentiated
Ice limit: trim lines, till boundaries, or weathering differences
Sand and gravel: glaciofluvial or wave reworked
Cirque
Col
Hanging valley
U-shaped valley
Drumlinoid features
Stoss and lee bedrock hills
Meltwater channel
End moraine
Bay d'Espoir
Hermitage Bay
Connaigre Bay
Northeast Arm
Great Bay de l'Eau
Fortune Bay
Pass Island
5 0 5 10 15
Kilometers


d'Espoir
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2 of 2

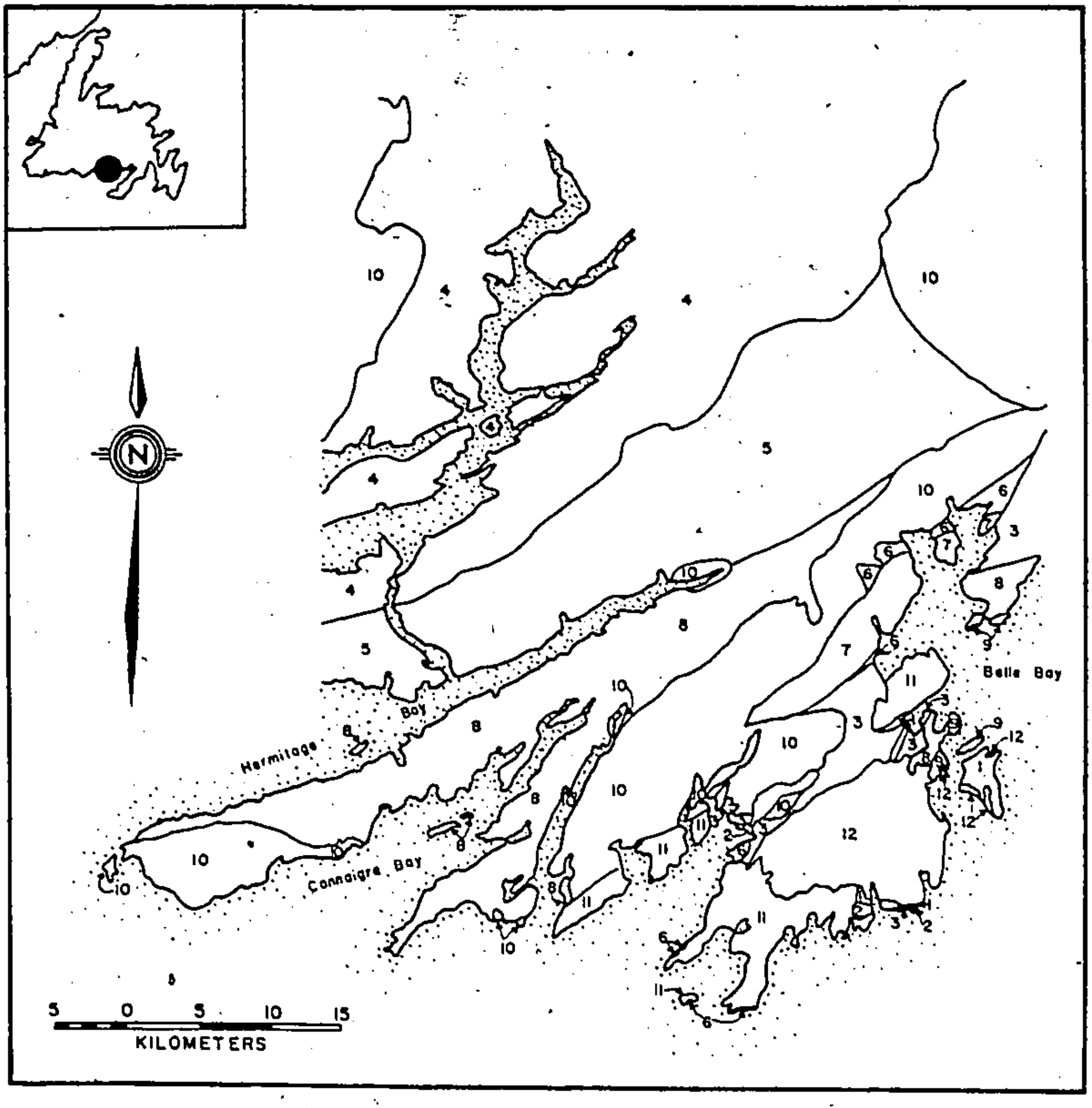
2.2 Drift Cover and Ice Limits

Ice limits are defined as the stationary positions of a glacier margin for a significant period of time. They may represent a still stand that occurred during an overall ice recession or the maximum extent of a given advance or readvance. The limits can be recognized from the location of end moraines and till sheet margins, the proximal positions of meltwater channels and glaciofluvial deposits, variations in the degree of bedrock weathering, differences in the morphologic expression of bedrock (i.e., subdued, molded topography versus very rough craggy, pinnacled terrain), and trim lines. Drift cover in the Hermitage Bay, as mentioned above, varies from isolated, perched erratics to thick till deposits tens of meters thick. Bedrock having a thin, patchy veneer of till is generally the norm. Perched erratics, without exception, occur everywhere.

Bay d'Espoir area. The most extensive and thickest till deposits occur as a till sheet in the northern Bay d'Espoir area and eastwards. The southern limit (Fig. 2.1) is well defined (e.g., see Newfoundland and Labrador Department of Forestry and Agriculture Air Photos 73, 74 NFLD A19832) by a margin that runs northeast, immediately south of and parallel to Little River, until it is south of River Pond where it swings to the east. The limit extends southwards into the lowlands heading the Bay du Nord River and then again continues eastwards. It can be traced intermittently, beyond the study area, inland of the coast to Terrenceville where it has been identified and continued further eastwards by Tucker (1979). West of Bay d'Espoir it is indistinguishable until approximately 8 km southwest of St. Alban's, from which point it can be traced westwards to the edge of the study area.

This till sheet margin outlines an ice limit. South of the limit terrain is characterized by bare rock, perched erratics and thin (less than 1 m thick) isolated patches of till. To the north, there is an olive brown (Munsell color 2.5 Y 4/3) to greyish olive (Munsell color 5 Y 4/2), generally sandy silt till of variable thickness with a high frequency of granite erratics. The till is thickest in the valleys and thins with increased elevation. There is also a marked difference in the overall morphologic expression of the terrain on either side of the limit. On the north side, the countryside has a slightly rounded, molded, almost subdued appearance. To the south, it is much more rugged and craggy, and drainage networks are often deeply entrenched into steepwalled, narrow canyons cutting across local bedrock strike (e.g., the streams feeding into the mouth of the Little River). The southern portion of the Salmon River (west) flows through canyons tens of meters deep and the surrounding countryside is rugged and barren, especially when compared to that north of the limit around Long Pond. Figure 2.2 is an outline of bedrock types in the Hermitage Bay area, presented to demonstrate that bedrock contacts have not been mistakenly identified as ice limits (Fig. 2.1).

On the heights of land northwest of St. Alban's and northwest of the Bay du Nord River, ice limits representing the former surface of the glacier can be recognized at an elevation of 230 to 260 m asl. The summits above the limits would have protruded through the ice as nunataks during the last glacial event. An ice limit on the hill between the Conne and Little Rivers occurs at a lower elevation (~ 152 m asl) probably due to drawdown of the ice into the Bay d'Espoir. A tongue of ice may have fed down the Little River valley as far as Riches Island, which has indications



LEGEND

1 Acid and basic volcanic rocks and associated pyroclastic rocks; sandstone, conglomerate siltstone (Cambrian and earlier,
2 Quartzite, sandstone, shale and silty shale (Cambrian and earlier)
3 Slate, siltstone, sandstone, minor limestones (Cambrian)
4 Slate, phyllite, siltstone, minor quartzite greywacke, argillate (Baie d'Espoir Series)(Middle Ordovician)
5 Granite gneiss including metamorphosed Unit 7 (Carrison Hills Granite)(Ordovician)
6 Arkosic sandstone, minor conglomerate, siltstone limestone (Ordovician)
7 Conglomerate grading upward into pebbly arkose (Ordovician)
8 Felsite, andesite, basalt flows, and associated pyroclastic rocks; siltstone (Ordovician)
9 Volcanic conglomerate, sandstone, slate, siltstone, shale (Silurian)
10 Granite, granodiorite, alaskite (Lower and Middle Devonian)
11 Conglomerate, lenses of slate and sandstone (Great Bay d'l'Eau Formation)(Devonian)
12 Granodiorite, granite (Upper Devonian)

Figure 2.2: Bedrock geology of the Hermitage Bay area (Anderson, 1965)

of a patchy till veneer. The land immediately to the northwest of Riches Island however, is barren. There is also a limit at Milltown and directly across the bay to the northwest. The latter is interpreted as the result of a tongue of ice that fed from the north into the head of the bay. The height of land between Milltown and Morrisville may also have been ice free but this interpretation is tentative and open to question. The western Garrison Hills, Bois Island, and Long Island are barren of till and have very rugged terrain; they were ice free.

Till, 3-5 m thick extends part way down the valley leading to St. Albans and indicates that a tongue of ice fed from an inland ice source down towards the present townsite. This glacier did not however, reach as far as the south side of the town only 5 km distant where, although there are erratics, there is no till. The exact frontal position of the tongue has been obscured by at least 15 m of outwash that covers an area of > 3 square kilometers. The hills immediately west of St. Albans, on which the radio and microwave towers are located, are devoid of till and very rugged (the "Barrens" of Jewell, 1939); they were ice free.

A large bedrock ridge (elevation 230 m asl) ~ 0.5 km south of the St. Albans airport (1 M/13: 855 122) may have acted as a topographic barrier to ice flow, deflecting it north and south into the valleys leading to St. Veronica's and St. Albans. Ice in the vicinity of the airport must have been relatively thin with little erosive power because it was unable to erode several southwest-striking bedrock ridges less than 1 m high. Ice movement, as indicated by southeast oriented striae on the ridges, was transverse to the orientation of the obstructions. On the protected southeast side of the obstruction a trellis drainage network has been preserved, entrenched into the bedrock. The local bedrock is a rotten, very friable slate which has been weathered to a depth of several meters (Fig. 2.3). It is quite different in nature from fresher bedrock of similar lithology exposed further north within the bounds of the ice limits. A period of time longer than that which has elapsed since the last glacial event would have been required to establish the drainage network and permit the extensive bedrock weathering. This area too, was ice free during the last glacial event. Granite erratics which rest on the bedrock would have been emplaced by a previous glaciation.

Head of Hermitage Bay. Inland and on either side of the head of Hermitage Bay there is another series of nunataks (Fig. 2.1) that are virtually devoid of till above 230 to 244 m asl. In some places a faint trim line can be discerned on the ground and from air photos. Bedrock above the trim line has undergone a greater degree of etching and weathering than that below. Traverses to summits of the nunataks (centred at 1M/12: 031848, 1M12:071861 and 1M/12:106865) revealed a marked difference between the tops and bottoms of the hills in terms of the presence or absence of till and degree of bedrock weathering. At a lower elevation and within the bounds of the upper limit there is a second lower limit at ~ 154 m asl (Fig. 2.1). This is thought to represent a lower recessional phase of the same event responsible for the upper limit. Ice which encircled these nunataks flowed southwest into Hermitage Bay, Connaigre Bay and Northeast Arm leaving thick till deposits plastered onto the valley walls. Textural parameters of the till vary from site to site depending on the underlying and immediately up-ice bedrock; granite/granodiorite based tills tend to be coarser grained than volcanic or siltstone based ones. As the ice flowed into Connaigre Bay and Northeast Arm land above ~ 152 m asl at

Figure 2.3: Granite erratic resting on weathered slate bedrock (photo location 1M/13:862056). Bedrock weathered to a depth of more than 2 m.

the heads of the bays remained ice free. At the top of the peninsula separating Connaigre Bay and Northeast Arm (1M/12:948708) there is a very distinct limit showing where the glacier was forced to divide and flow into each of the bays (Fig. 2.4). Figure 2.5 is a trace of Figure 2.4 showing local summits above ~ 152 m asl which would have been nunataks that remained ice free. The ice flowed down Northeast Arm as a fiord glacier into and beyond Harbour Breton and Deadman's Bight. The tills and associated deposits of the Deadman's Bight area are discussed in greater detail in Section 4.1.

South of the head of Hermitage Bay and northeast of Northeast Arm ice flowed around several nunataks (centred at 1M/12:005775, 1M/12:051798, 1M/12:021751, 1M/12:010730, 1M/12:000696, 1M/12:032716, 1M/12:048717, 1M/12:017701, and others) and into a series of deep U-shaped troughs feeding into Old Bay (Fig. 2.4). A very clear trim line and limit has been preserved in the valley which leads to Taylor Bay (1M/12:668024)(Fig. 2.4).

<u>Cinq Islands Bay - Salmonier Cove Pond - Corbin - St. Jacques.</u> Further east, the glacier, possibly somewhat thinner, flowed into Cinq Islands Bay and Salmonier Cove Pond. Northeast-southwest oriented striae and flutings in a lowland trough confirm that ice movement was towards these two bodies of water. Ice in Salmonier Cove Pond, further nourished by cirques on its northern wall (Fig. 2.1) flowed southwest into the Great Bay de l'Eau. South of the outlet, across the bay, there is an exposure (located at 1M/12:064627) of well oxidized, cemented, blocky till. This was the most highly weathered till observed in the whole of the study area and is thought to predate the last glacial event.

Figure 2.4: Airphoto from over the heads of Connaigre Bay and Northeast Arm. Ice limits showing where glacier was forced to divide and flow into Connaigre Bay and Northeast Arm are on left side of photo. Deep U-shaped trough feeds into Old Bay. Lighter toned hill tops (~ 152 m asl) were nunataks during the last glacial event (NAPL photos

Figure 2.5: Trace of Figure 2.4 showing nunataks above 152 m asl of last glacial event. Cross hatched areas were ice free.

Glacial ice also flowed into Corbin Bay but the peak (1M/11:184709) above Corbin remained ice free (Fig. 2.6). The same ice mass that flowed into Corbin and Salmonier Cove Pond may have extended even further south to St. Jacques harbour and the bay immediately to the west of it. Till along the road to these areas is immature being a thin, patchy, very stoney, angular rubble. The ice at this point is assumed to have been relatively thin with little erosive power. Ice limits here, are vague and virtually impossible to recognize although there are obscure traces of small end moraines to the west and southwest of St. Jacques.

At the north end of Belleoram a road cut (1M/11:187652) exposes more than 15 m of very angular, stoney, sandy silt till. Clasts are all local Belleoram granite. The outcrop is very massive and fresh looking, especially when compared to the barren nature of the weathered bedrock summits immediately around it (e.g., Fig. 2.6). Till fabric analysis indicates a seawards ice flow direction (vector mean, 86°N). Its location in the otherwise barren bedrock landscape is problematic, although a cirque origin may be possible. The presence of this fresh till, in juxtaposition with barren weathered uplands, reiterates the fact that in the Hermitage Bay area there are several highlands consistently devoid of till while below a given elevation there are always distinct till deposits.

English Harbour - Mose Ambrose - Boxey. To the west, in the wide U-shaped valleys feeding into Mose Ambrose and Boxey, ice limits are clearly defined by the margins of a well comminuted till on the valley-walls (~ 76-91 m asl at Boxey). The source of the ice was several large inland cirque glaciers situated on the ridge south of Salmonier Cove Pond (Fig. 2.1) which flowed southwards to the sea as valley glaciers. There

Figure 2.6: Weathered bedrock knob above Corbin (1M/11:184709) which remained ice free during last glacial event. Elevation: 168 m asl.

may have been some contribution from the ice mass described above, but this was probably minimal. At Boxey, till (basal (?) overlain by an ablation or wave reworked facies) extends to below sea level. It is a brown (Munsell color 7.5 YR 4/3), platy and well indurated till with a silty matrix containing granite erratics from the north. This till, plus that on the south side of Boxey Harbour Head, indicates that the valley glacier extended out onto the shelf, probably as a piedmont lobe during a period of lower relative sea level. The deposit on Boxey Harbour Head is not "high level kame .... sand and gravel" as stated by Widmer (1950) but rather a dull reddish brown (Munsell color 5 YR 5/4) sandy silt till containing striated conglomerate erratics up to a meter across. The till rests on striated polished bedrock.

At Mose Ambrose, up to 12 m of greyish yellow brown (Munsell color 10 YR 4/2), silty sand till is exposed on either side of the harbour. Only 1 km to the east, at English Harbour West there is no till and bedrock benches are well pronounced. A large bedrock ridge 46 to 61 m asl north of English Harbour West may have acted as a barrier to ice movement into the harbour which would account for the lack of till. The height of the barrier also provides an indication of ice thickness at this location.

Grant (1975b) mentions the presence of northwards-stossed hills at Mose Ambrose, suggesting that they are the result of northwards-flowing ice from the Burin Peninsula. He states that the responsible glacier was unable to reshape previously formed drumlins on the Burin Peninsula but was able to abrade and pluck large bedrock hills some 60 km distant across the 350 m deep Fortune Bay. Grant's "northwards-stossed hills" are more likely structural in origin. His observation of "red rhyolite erratics" having been "derived from the Burin Peninsula" is also open to question, as Widmer (1950, pp 311, 319, 361 and 365) noted orange and orange red colored rhyolite in the underlying Great Bay de l'Eau Conglomerate within the Hermitage Bay area.

Wreck Cove (Tibbos Hill) and Coomb's Cove. The area between Wreck Cove and Coomb's Cove and west of the valley leading to Boxey was situated beyond the limits of the last glacial event. Evidence for this lies in the lack of till, the degree of weathering of conglomerate and intruded sills (Fig. 2.7), and the difference in the degree of dissection of bedrock benches and local bedrock (Fig. 2.8). An isolated 20 m thick pocket of greyish brown (Munsell color 7.5 YR 6/2), blocky, sandy silt till occurs in a depression at St. John's Head. Widmer referred to it as high level kame terrace but this interpretation should be discounted as it is definitely a till. The till was deposited by either a local cirque glacier or, more likely, from ice of a more extensive, pre-last glacial event.

2.3 Glaciofluvial Deposits

Local ice limits can also be identified by the location of till-outwash contacts. At St. Alban's, near Furby's Cove (1M/12:710882), Harbour Breton and east of the study area at the head of Rencontre Lake transitions from till to outwash occur over very short distances denoting stable positions of an ice front from which meltwater and outwash were transported. The proximal upper surface of the outwash is always higher than the adjacent till surface, reflecting the influence of the adjacent

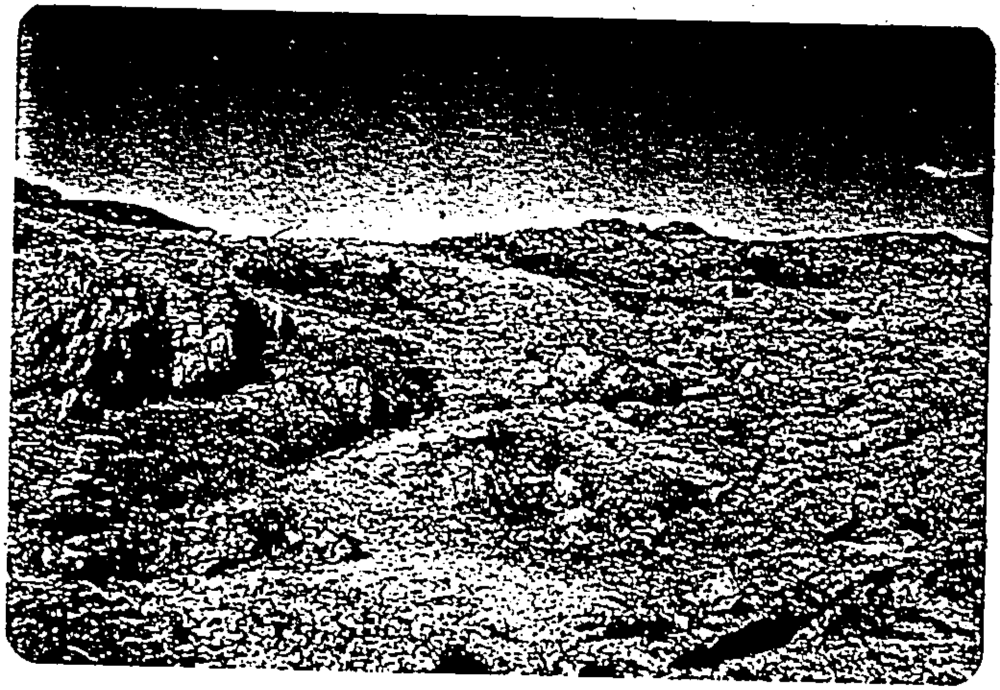


Figure 2.7: Weathered sill in Great Bay de l'Eau conglomerate (photo location 1M/5:045597). Area was not glaciated during last glacial event.

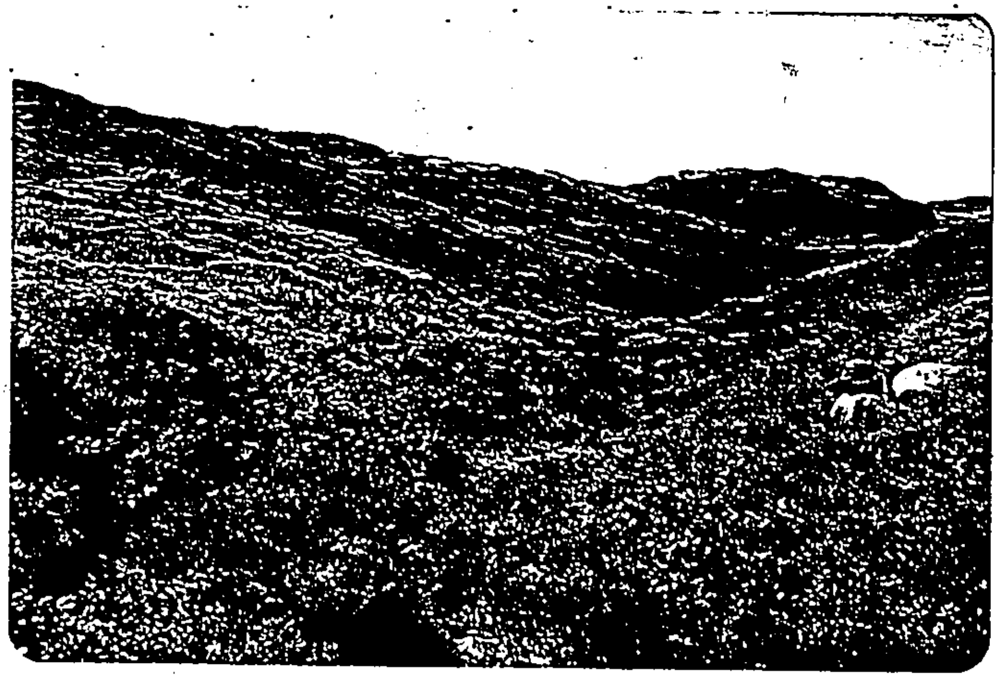


Figure 2.8: Gullied Great Bay de l'Eau conglomerate in the Coomb's Cove area.

ice body. The source of the ice near Furby's Cove was the thick tongue of ice in Hermitage Bay which fed through a topographic low (located at 1M/12:914750) in the valley wall and then down along the spine of the peninsula. When deglaciation occurred this ice would have been severed relatively early from the active ice within Hermitage Bay and down-wasted in situ.

Other occurrences of glaciofluvial deposits are shown in Figure 2.1. The most significant occur at Jersey Harbour, along and at the mouth of the Bay du Nord River, and Southeast Brook.

2.4 Cirques

Cirques are restricted to the southern portion of the Hermitage Bay area with the northernmost being located at Turnip Cove (1M/12:173827), 1 km north of Pools Cove. Cirque widths vary from ~ 150 m to 0.5 km across and lakes of variable sizes occur in several of the basins.

Jenness (1960) stated that cirques in the area were restricted to elevations above 229 m. However, few if any, were identified above this height and some were constructed at or below sea level.

Drowned cirques, formed at a period of lower relative sea level and subsequently submerged are found 0.5 km north of Hermitage (1M/12: 824698) and at Jersey Harbour (1M/5:936600). Several large inland cirques, previously mentioned, were developed on the ridge of Salmonier Cove Pond. The ice of these inland cirques coalesced into wide valley glaciers and flowed south through the valleys leading to Boxey, Mose Ambrose and the harbour east of English Harbour West. Finger lakes now occur in some of the valleys. There may have been restricted ice flow westwards but it did not extend as far as Coomb's Cove and Wreck Cove.

2.5 Ice Flow and Directional Indicators

Indicators of ice flow patterns in the Hermitage Bay area are glacial striae, drumlinoid features, stoss and lee bedrock forms, and clast fabric in till.

Striae. The majority of striae measurements were recorded on freshly exposed bedrock along road cuts or coastal sections. The sense of direction was determined from bevelled surfaces, micro-crag and tail features, and superposition. Field measurements, supplemented by data from local bedrock maps (Widmer, 1950; Williams, 1967, 1969) are summarized in Figure 2.9 (a listing of exact striae locations and orientations are given in Appendix 1). Striae are poorly preserved on the fissile slate bedrock around the head of the Bay d'Espoir and on coarse grained granite further south. They are best preserved on bedrock of volcanic origin, conglomerate and quartz veins. Several of the striae are located beyond the ice limits discussed in Section 2.2. However, as evidence from the Burin Peninsula has shown that striae of pre-Late Wisconsin age can be well preserved under only a thin veneer of till (Tucker, pers. comm.), the striae in the Hermitage Bay area may be multi-generational, representing more than the last glacial event.

The overall regional pattern indicates a generally southwards and/or seawards flow direction onto which topographic control has been superimposed. The local pattern of ice flow northwest of Bay d'Espoir was to the south and southeast with a south-southwesterly component into

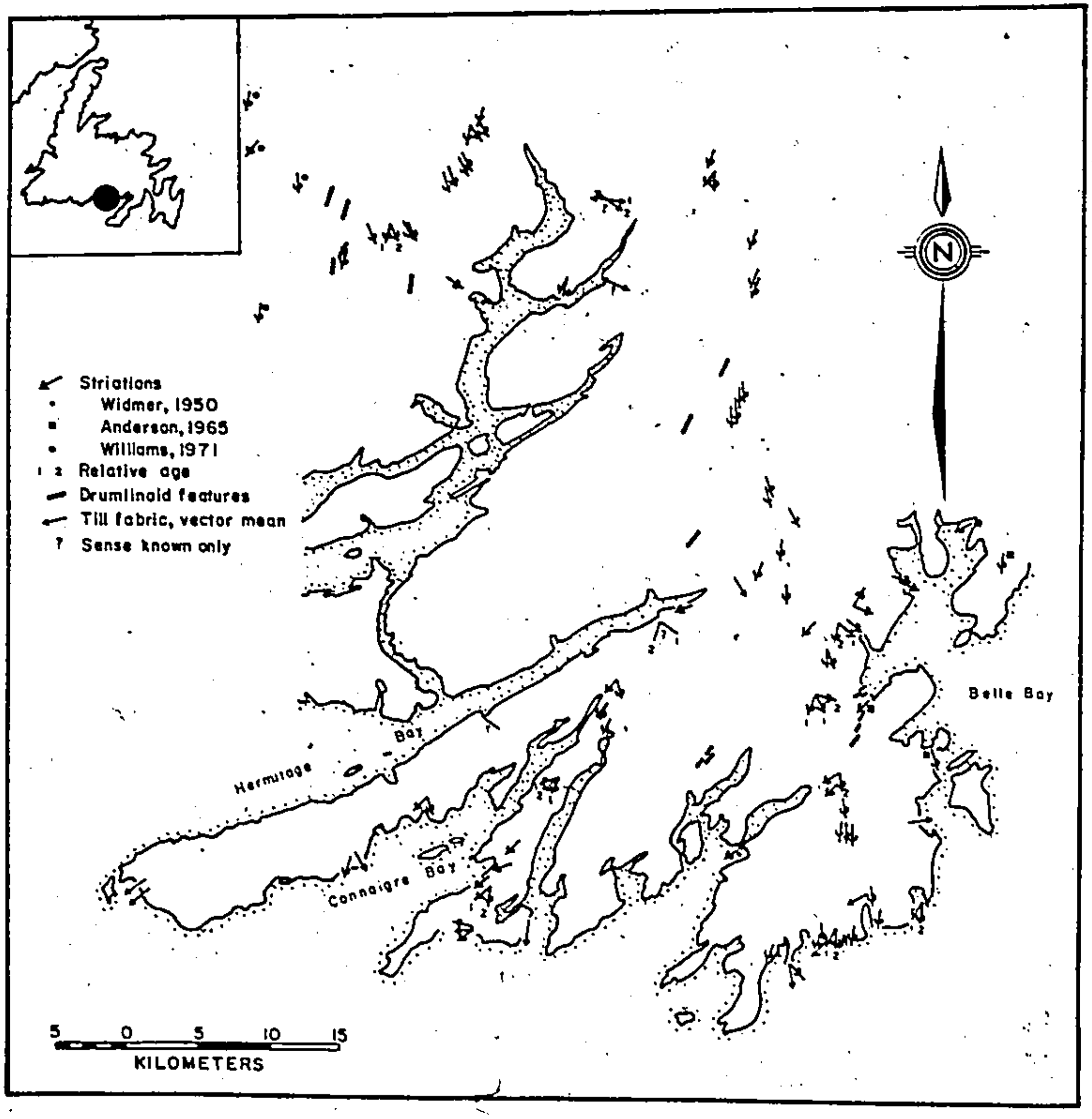


Figure 2.9: Directional indicators: striations, drumlinoid features and till fabric vector mean. Relative ages of multiple striae indicated wherever possible.

the Salmon River valley. East of the head of Bay d'Espoir ice flow was southwestwards, controlled largely by the Conne and Little River valleys, and a northeastwards extension of the Garrison Hills. East of the Garrison Hills ice movement was southerly. At the head of Hermitage Bay it was deflected southwest into the bay and also into Connaigre Bay and Northeast Arm. In a zone of otherwise southeasterly flow, striae in the lowland between Salmonier Cove Pond and Cinq Isles Bay show southwest-northeast orientation. Striae in the valleys between Boxey and St. Jacques all point seaward. At Dog Head Cove (1M/12:778627) and Turnip Cove (1M/12: 173827) the striae are the result of ice from seaward oriented cirques.

Drumlinoid features. Drumlinoid features (ice molded, streamlined till) occur northwest of St. Albans, north of the head of Hermitage Bay, north of Old Bay and in the Corbin Bay-Cinq Islands Bay area (Fig. 2.1 and 2.9). In all instances except the latter they are oriented southwestwards. The features in the Corbin Bay-Cinq Islands Bay area have a slightly steeper southwest side implying ice flow to the northeast into Corbin Bay and not to the southwest as is indicated by a single striae given by Williams (1969). It is not known to which glacial event the drumlinoid features belong. Tucker (1979, Fig. 2.1) illustrated fluted ground occurring beyond the limits of Late Wisconsin glaciation on the Burin Peninsula and it is possible that those in the Hermitage Bay area are of similar age.

Stoss and lee bedrock hills. Several large stoss and lee bedrock hills (with elevations greater than 300 m asl) were observed in the field and together with similar features recognized on air photos are plotted in Figure 2.1. They are characterized by gently stossed northerly slopes and steeper more craggy lee slopes (Fig. 2.10). All stossed slopes that were traversed showed some indication of having been ice molded and perched erratics were always present.

The responsible ice flow, as indicated by bedrock asymmetry, was from the north-northeast to north-northwest which is occasionally at odds with some of the striae and drumlinoid features data. Since considerable thicknesses of ice would have been required to do the molding and plucking that is evident on several of the large hills (e.g., 1M/12:107787) and as the stoss and lee forms occur both within and beyond the ice limits shown in Figure 2.1 it is most probable that these features originated during the penultimate, or an earlier glacial event which was much more extensive than the last.

Till fabric. Andrews (1971, 1975) considered that several one site, one sample collections of till fabric data would be sufficient for an understanding of regional ice flow directions. Consequently, the dip direction of pebbles in till was recorded at 12 locations in the Hermitage Bay area. A sample size of 25 was initially used (as per Andrews and Smith, 1970) but this was later increased to 50 in order to decrease the standard error. The results are plotted two dimensionally in Figure 2.11 and the preferred orientation (vector mean, after Krumbein, 1939) compared with striae trends in Figure 2.9. In Figure 2.9 it can be seen that the preferred orientation generally corresponds fairly well with local striae directions, except for the results at Southeast Brook, Conne River and northeast of English Harbour West which are roughly transverse to local striae trends.

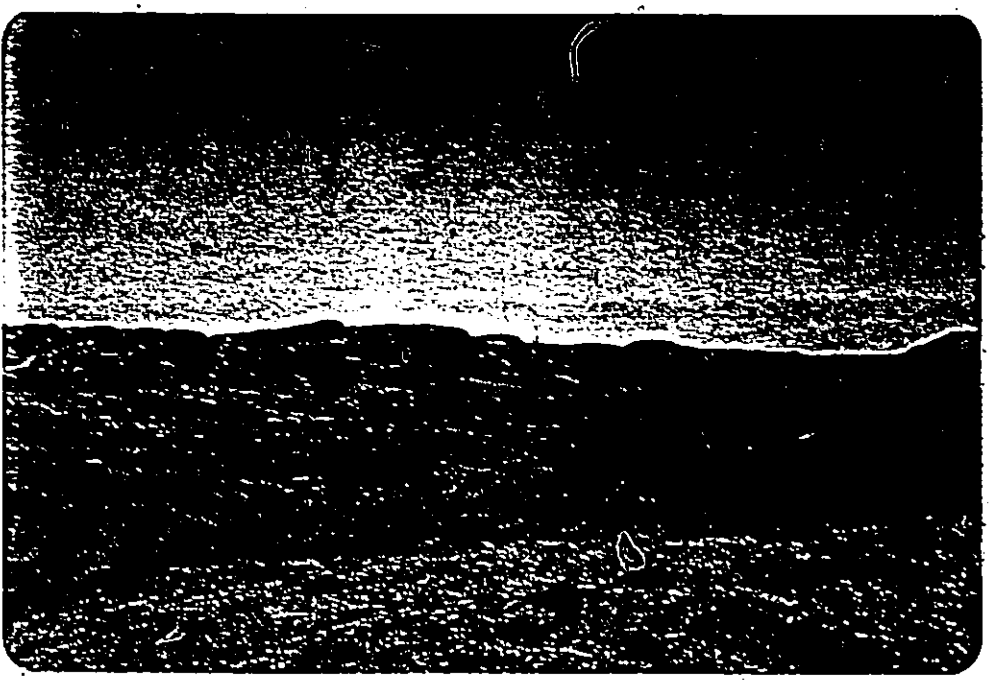


Figure 2.10: Ice molded stoss and lee bedrock hills.

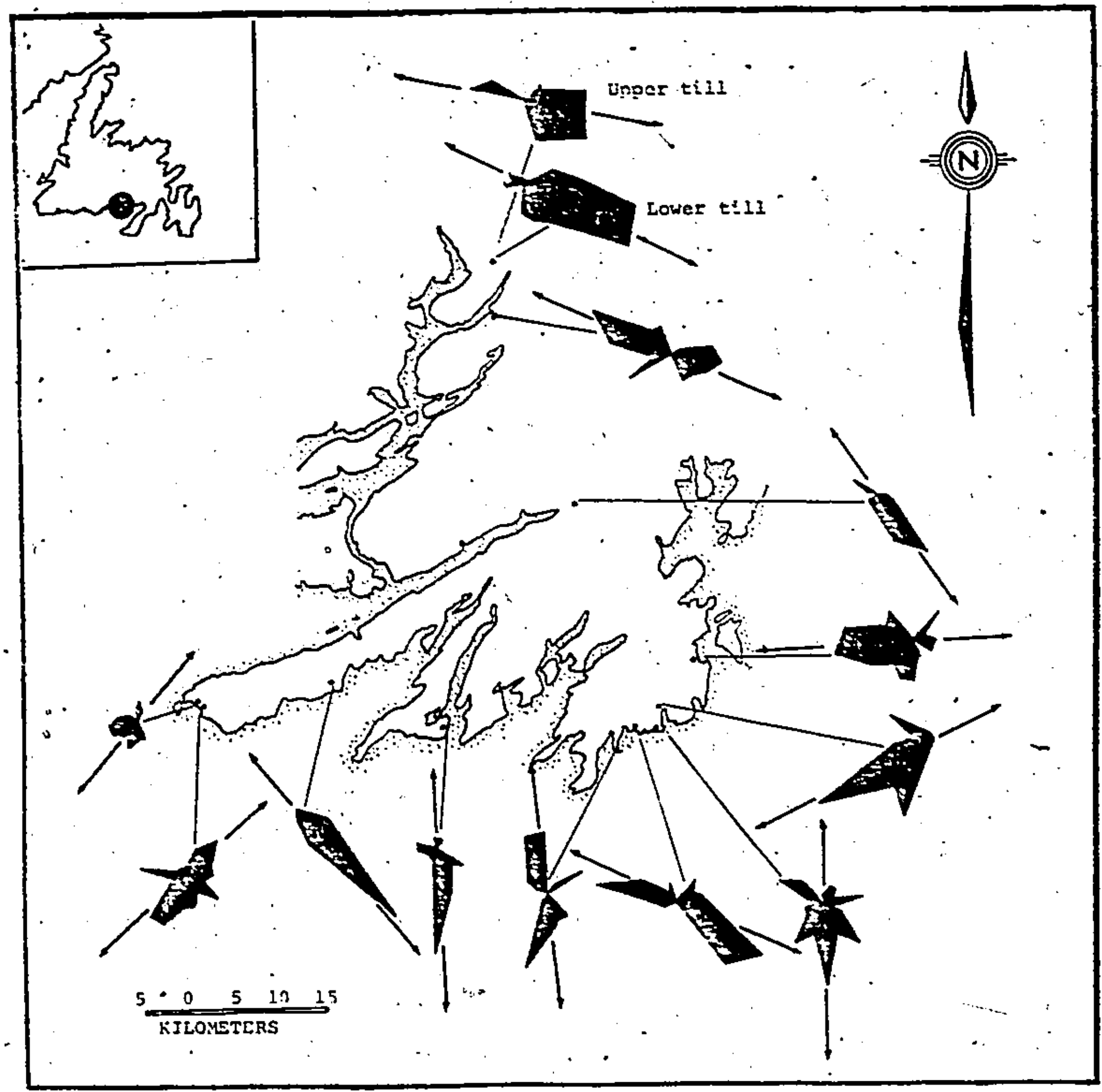


Figure 2.11: Till fabric data and locations, plotted two dimensionally. Arrows indicate preferred orientation (vector mean).

2.6 Marine Planed Bedrock Benches

Although not directly the result of glacial erosion or deposition the planation of marine benches at various levels occurred as a result of glacial advances and retreats. This section is concerned with bedrock benches as they relate to an overall sequence of events. A more detailed account of bench and terrace elevations and their interpretations is presented in Chapter 7.

Grant (1975, Fig. 1) identified three sites where "interglacial marine benches, under till" occur in the Hermitage Bay area. Although Grant does not provide exact locations the general areas he indicated were investigated. At Deadman's Bight (1M/5:867576) there is an ice molded, striated, till covered 9.5 m asl bench that was first described by Widmer (1950). The till cover on its surface indicates that the bench predates at least the last glacial event and is probably of interglacial or interstadial age. This, plus another 4 ± m asl bench in the immediate vicinity were the only drift covered marine platforms recognized in the area during the present study. Another of Grant's examples is east of Pass Island. There is a till covered bedrock ridge (11P/8:625600), but its upper surface is highly weathered, of variable height and has a crude convex upwards profile. This ridge is not a marine cut bench. Grant also shows that there is a till covered bench in the vicinity of Gaultois. However, on a traverse from Gaultois to Piccaire no till was observed, although there were occasional weathered perched erratics. No indication of a bedrock bench could be seen in the field or on air photos. The existence of Grant's "bench" near Gaultois in spite of the vague location given on his map, is somewhat questionable.

Along the coast between Wreck Cove (Tibbos Hill) and Coomb's Cove there is a pair of well-exposed, well-planed benches at 13 to 17 m asl and 28 to 35 m asl (Figs. 2.12 and 2.13). They are significant in that there is a marked difference in the degree of dissection between the upper and lower surfaces. The lower bench, with its several coincident raised sea stacks and caves, is a wide flat surface that has undergone a minimal amount of dissection since its initial planation. (There are occasional pockets of sorted sands and some gravel of possible littoral origin on the lower bench at Coomb's Cove and on eastern St. John's Head Promontory). The upper bench, having a tread of similar width, is not as well defined. Its tread has been highly dissected and gullied, and the riser is indistinct in many places. The degree of dissection of the upper bench is comparable to that of the surrounding countryside of similar bedrock (Upper Devonian conglomerate, Anderson, 1965). The most recent time that planation of the lower bench could have taken place was during the higher relative sea level which occured at the end of the Late Wisconsin glaciation. If this is the case, then the subsequent erosion and dissection of the bench, which is minimal, has occurred since that time. This would then require that the period of time required for the dissection of the upper bench and surrounding terrain have been considerably longer. Therefore, given that the lower bench represents a Late Wisconsin postglacial high sea stand the upper bench must be the product of an earlier interstadial or interglacial raised sea level. Furthermore the lack of any evidence of glaciation on the upper bench implies that the bench has not been overridden by ice since its initial formation. This then suggests that the Coomb's Cove-Wreck Cove area was not ice covered during the last glacial event and that this ice

Figure 2.12: Benches at Wreck Cove (1M/5:057611) Note difference in degree of dissection of the two platforms.

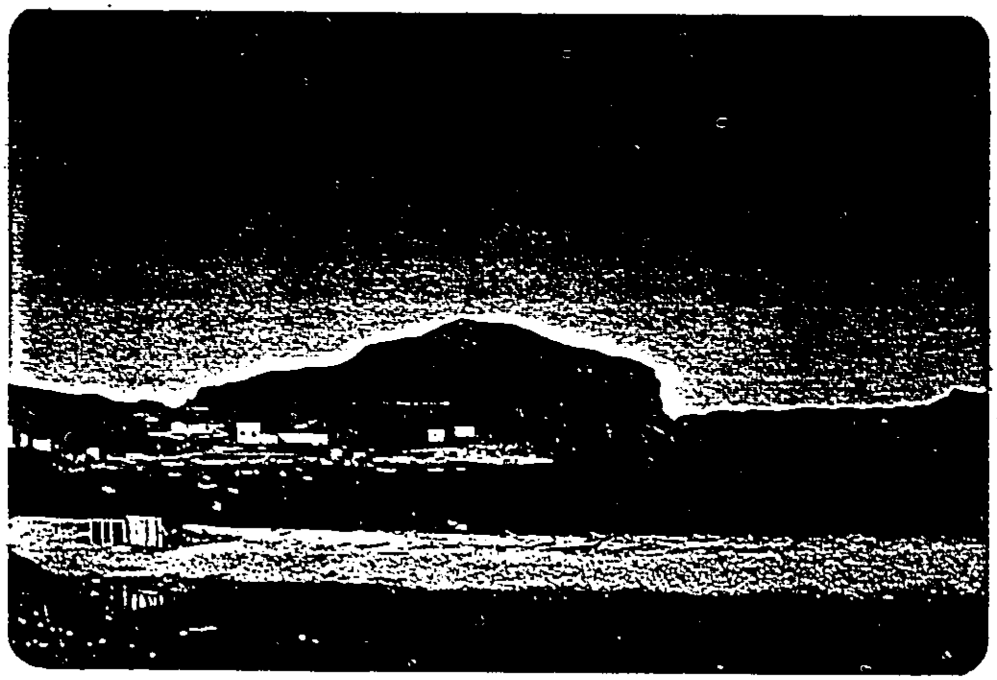


Figure 2.13: Benches at Coomb's Cove (1M/5:033564) Note difference in degree of dissection of the two platforms.

free interval may be extended to at least prior to the last interstadial. Even if it were argued that the lower bench pre-dated the Late Wisconsin post glacial sea level high the area must still have remained ice free during the Late Wisconsin as the well preserved, raised sea stacks (Fig. 2.14) could not have withstood glaciation.

2.7 Fiords

A fiord is a partially submerged, steep walled valley on a glaciated coast (Flint, 1971; Andrews, 1975). By this definition, upper Bay d'Espoir, Hermitage Bay, upper Connaigre Bay, Northeast Arm and Old Bay can be considered fiords. Vertical walls 200 m high and depths greater than 300 m are not uncommon. The Bay d'Espoir and Old Bay fiords appear to be fluvial in origin whereas the others are structural. Hermitage Bay is developed along the Hermitage Fault (Williams, 1978) which extends several kilometers inland. Although glacially modified, its depths to 350 m are probably due to tectonics.

One of the best indicators of a glacially sculpted fiord is a rocky threshold at the mouth (Holtedahl, 1967), although they need not always occur (Andrews, 1975). Longitudinal profiles (Fig. 2.15) of the fiords (constructed from Bathymetric Map 15074-A, Canadian Hydrographic Service, 1977) reveal what may be thresholds at the mouths of Connaigre Bay and Northeast Arm. These two profiles are not unlike those of fiords in Baffin Island (Løken and Hodgson, 1971) or Norway (Holtedahl, 1967). Hermitage Bay does not have a threshold but continues to deepen and widen beyond the limit of the profile. Bay d'Espoir, with its branching pattern is complicated and a distinct threshold is not recognizable.

Figure 2.14: Raised sea stack eroded on 13-17 m asl bench during Late Wisconsin postglacial high sea stand.

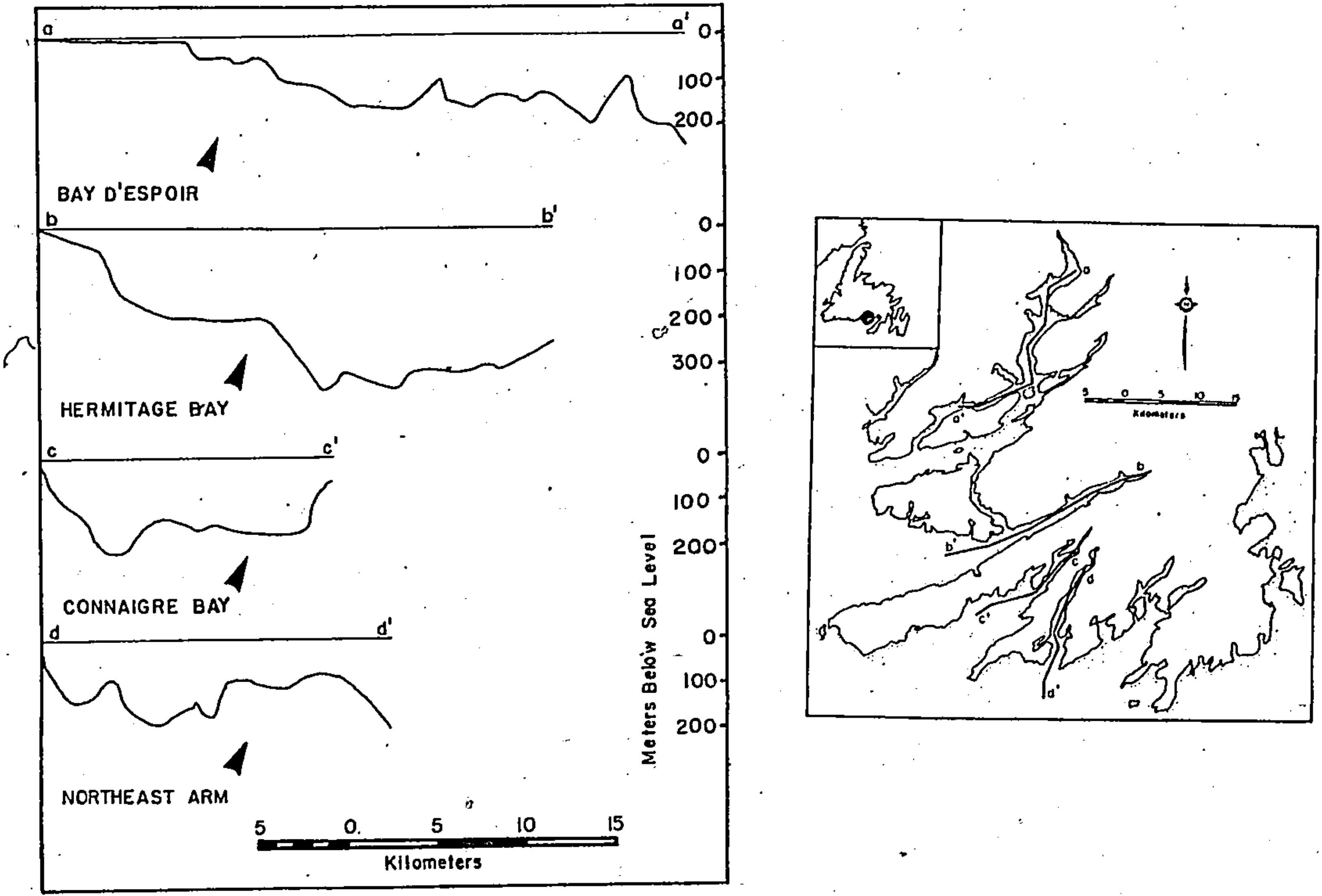


Figure 2.15: Longitudinal bottom profiles of Bay d'Espoir,Hermitage Bay, Connaigre Bay and Northeast Arm.

2.8 Discussion

Jenness (1960) identified an "inner-outer drift zone" in eastern Newfoundland based on the position of end moraines, eskers, and radiating outwash deposits. His location of the limit defining the inner drift zone on the south coast was tentative, based on the earlier observations of Widmer (1950) and Bradley (1954). The present study and that of Tucker (1979) has confirmed the existence and concept of a limit in general and determined its exact location more precisely. The ice limit shown in Figure 2.1, running across the north of the Hermitage Bay area, is the equivalent of the drift limit proposed by Jenness. However, rather than being the recessional position of a more extensive "late Pleistocene" glaciation (Jenness, 1960), the evidence presented above indicates that the limit represents the maximum extent of a Newfoundland based ice cap which existed during the last glacial event. This concurs with conclusions drawn by Tucker (1979) regarding the age of a similar limit extended across to the northern Burin Peninsula.

Further south, a separate and smaller ice cap centered north and east of the head of Hermitage Bay fed ice radially seawards. Profiles showing the general shape of the ice cap have been reconstructed in Figure 2.16 (profile location map on Fig. 2.17) based on the positions and elevations of ice limits shown in Figure 2.1. The ice cap has a convex upwards shape and was thickest around the head of Hermitage Bay. The surface of the ice over Hermitage Bay was depressed due to the draw-down of a large deep outlet glacier which fed out through the bay. The ice cap was also a source for valley glaciers feeding the fiords of Northeast Arm, Connaigre Bay and Old Bay. The northern margins of the cap are indistinct but it did not extend to the southern limit of the main island ice cap

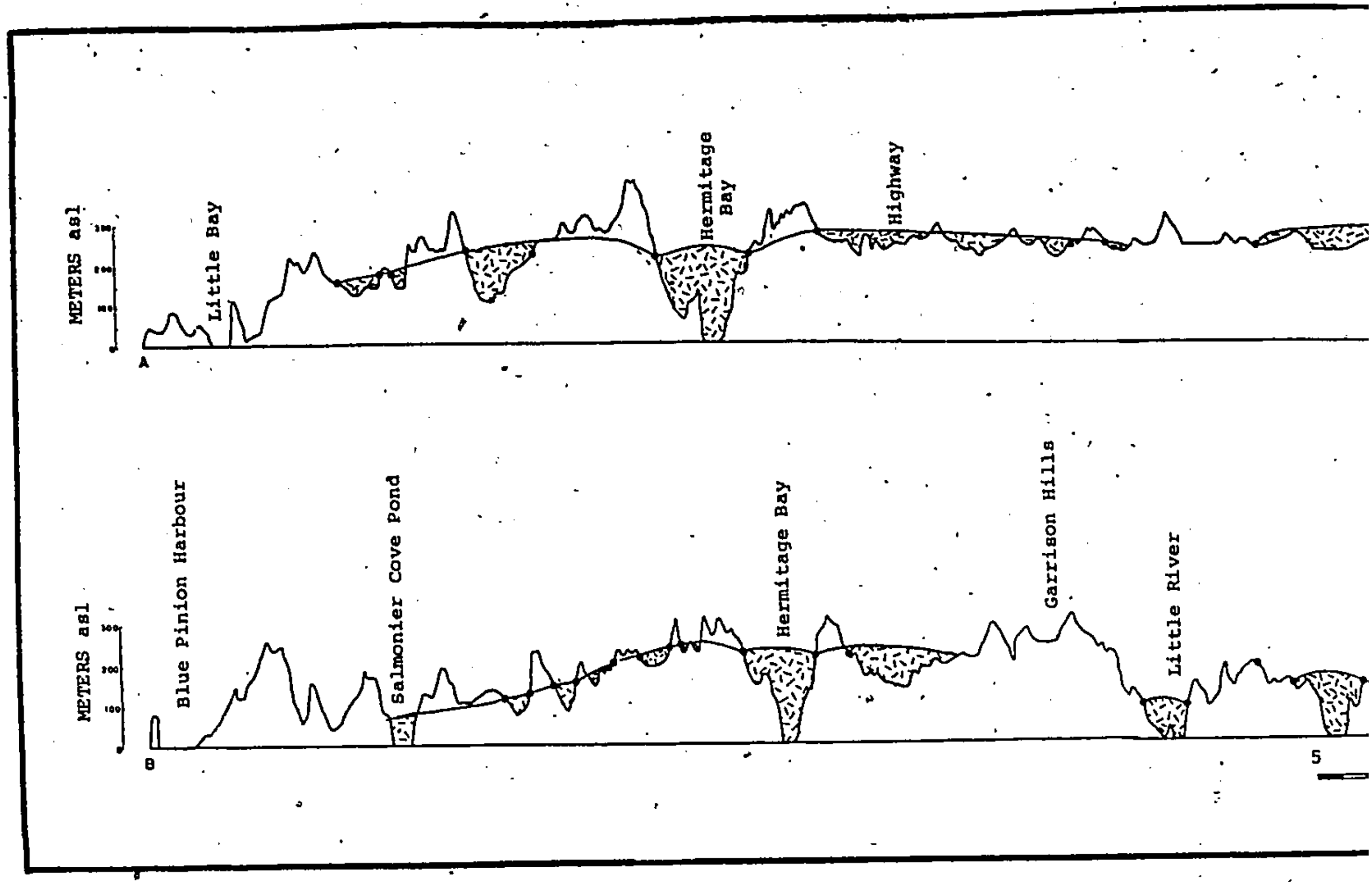
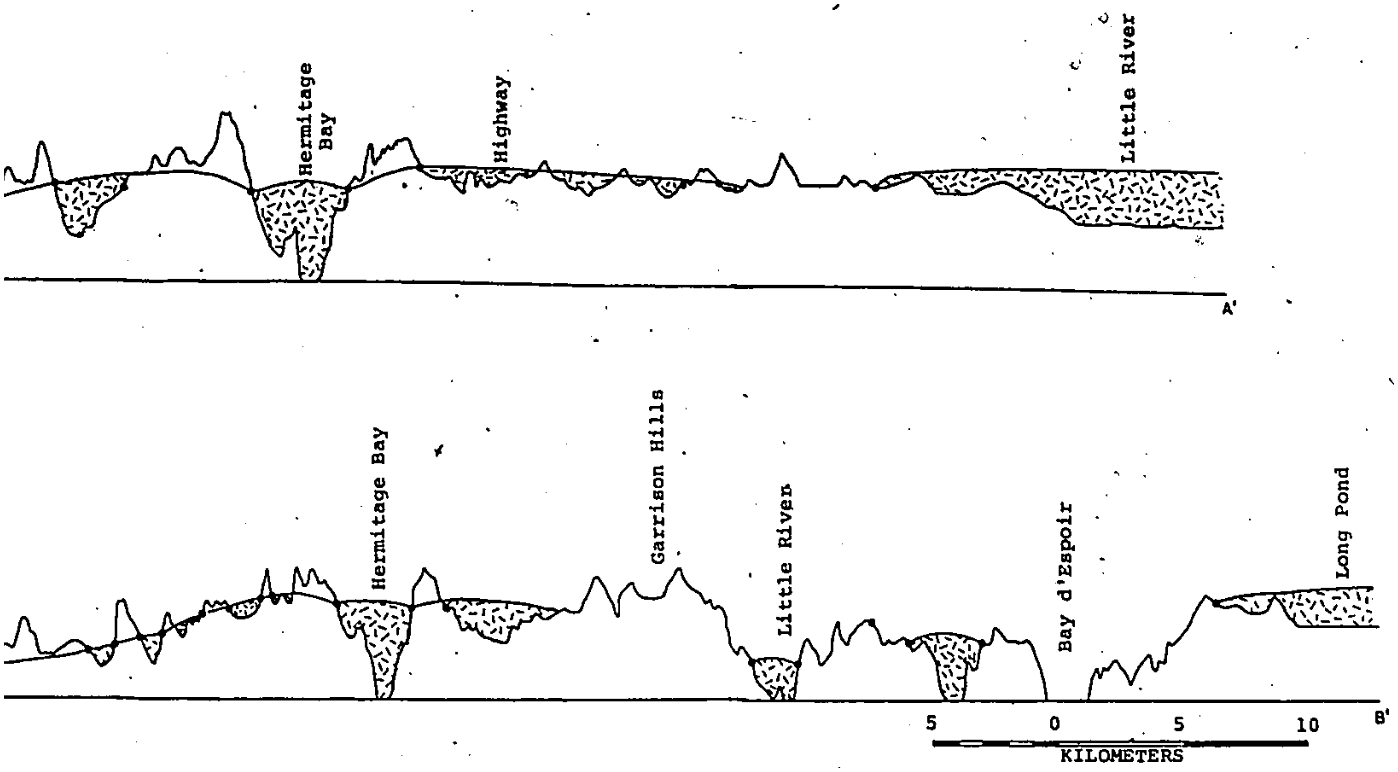


Figure 2.16: Glacier profiles, based on ice limit data. The ice caps have a crude convex upwards shape. Drawdown occurred over Hermitage Bay.



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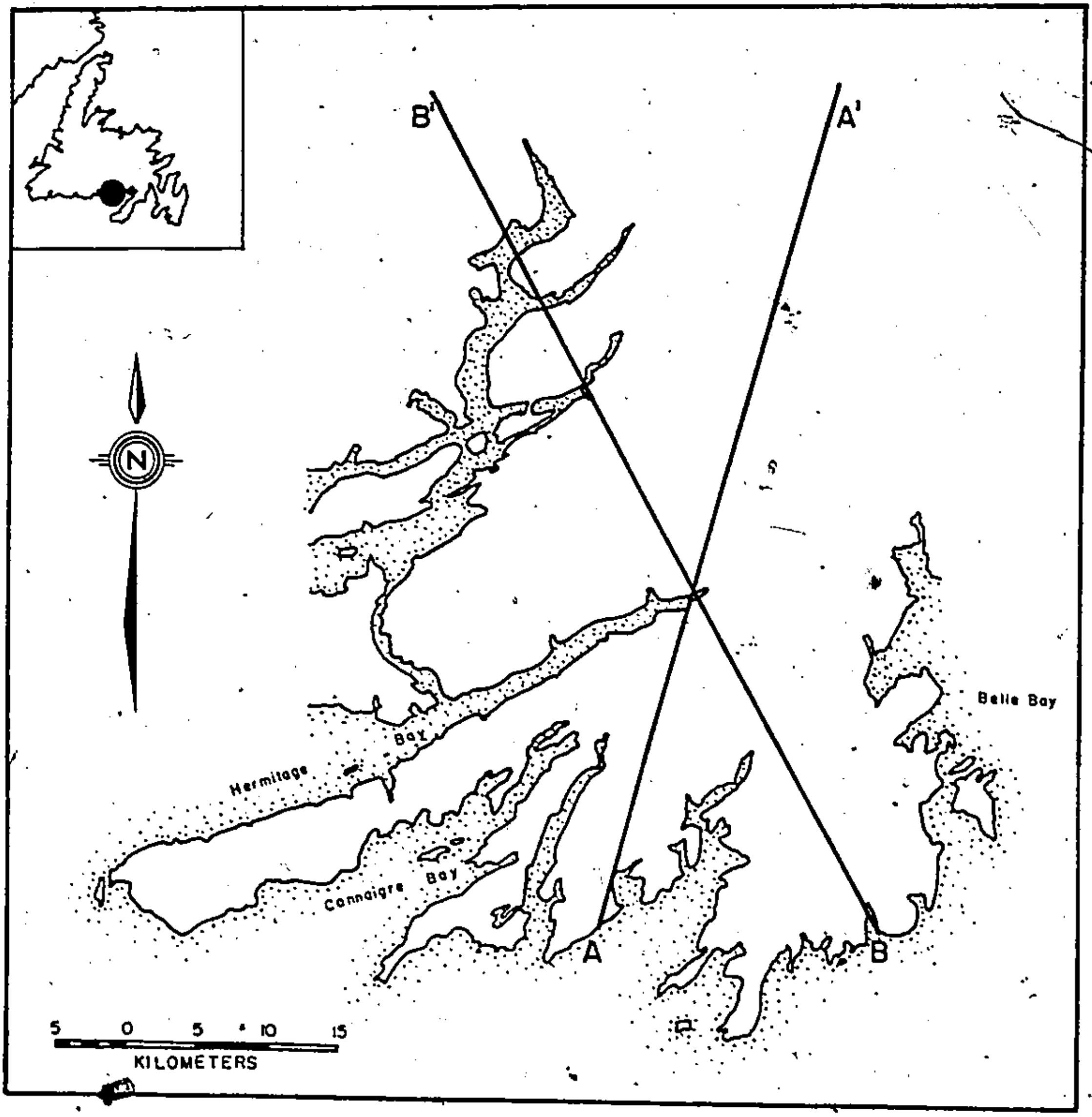


Figure 2.17: Location map of profiles in Figure 2.16.

having been blocked by the Garrison Hills (Fig. 2.16, BB'). The surface of the cap decreased in elevation seawards; Figure 2.16, BB' shows that ice extending as far south as Salmonier Cove Pond was relatively thin, which agrees with earlier observations and conclusions regarding immature tills (Section 2.1). Maximum dimensions of the ice cap were ~ 35 to 40 km across. The profiles also show the valley glaciers which occupied the Little and Conne Rivers, and the southern limits of the large inland ice cap.

Although radiometric control is lacking, the main island ice cap, the smaller one around the head of Hermitage Bay and the several valley and cirque glaciers are tentatively considered as time correlative. They represent the extent of ice coverage in the Hermitage Bay area at the maximum of the last glacial event. At this time, several ice free areas existed, including that between the two ice caps, the nunataks shown in Figures 2.1, 2.4, 2.5 and 2.16, the area west of Boxey valley, the western Garrison Hills, English Harbour West, and possibly the southern Hermitage Peninsula proper (Seal Cove area). Evidence from the Coomb's Cove-Wreck Cove area indicates that ice free areas may have existed throughout the whole of the Late Wisconsin.

Examination of a Landsat image created May 17, 1976 using bands 4, 5 and 7 shows that most of the nunataks outlined in Figure 2.1 are recognizable. They exhibit a light colored tone on an otherwise pink background. The light tone is due to a lack of vegetation in the highlands which is probably due to a lack of till.

CHAPTER 3

STRATIGRAPHY: NORTHERN BAY D'ESPOIR AREA

3.1 Introduction

3.1.1 Preamble

Since the morphologic expression of the earth's surface is generally the result of the most recent processes which have affected an area, any evidence of what has gone on previously is often removed or buried by successive events. As such, stratigraphy provides a means for the documentation of events and processes that have affected an area over time. The purpose of this and the following chapter is to describe and interpret glacial stratigraphic sections found at various coastal and inland sites throughout the field area. Most were discovered on field traverses along river valleys or along the coast. Only one (Deadman's Bight-Harbour Breton) has been described previously in literature. Chapter 3 concentrates on the stratigraphy of the northern Bay d'Espoir area which was located at the fringe of the Late Wisconsin, Newfoundland based icecap centred to the north (Section 2.8). Chapter 4 deals with stratigraphic sections located in the southern Hermitage Bay area which was affected by a small ice cap centred north and east of the head of Hermitage Bay and several small, local cirques and valley glaciers. In both chapters local sequences of events will be provided, based on stratigraphic sections, at a greater level of detail than permitted in Chapter 2.

3.1.2 Northern Bay d'Espoir

Good stratigraphic sections in the northern Bay d'Espoir area are limited in number. In an area encompassing more than 450 km^2 only nine multiple unit sections could be located. Although the problem is in part due to poor access, of even greater significance is the paucity of unconsolidated sediments in a predominantly bedrock and till veneer terrain. The interpretation of the local glacial history of the northern Bay d'Espoir area is based upon a composite suite of seven stratigraphic sections found along coastal, river and road cuts. A description and location of stratigraphic sections is shown in Figure 3.1. The chronology will be relative, based upon cross correlation and stratigraphic position of the units.

Bay d'Espoir is a deep narrow fiord extending ~ 40 km inland from the southern coast. Its depths exceed 160 m bsl (below sea level) (Canadian Hydrographic Service, 1977) and its vertical walls are more than 180 m asl. Local bedrock is Middle Ordovician grey slate, mudstone and vein quartz (Jewell, 1939; Anderson, 1965). Five large valleys feed into the head of Bay d'Espoir: the deep trough-shaped valleys of St. Alban's and St. Veronica's trending northwest-southeast and the smaller valleys of Conne River, Southeast Brook and Little River oriented southwest-northeast, parallel to bedrock strike. The height of land is ~ 295 m asl, occurring northwest of St. Alban's.

3.2 Section I: Trout Hole Falls Community Park

3.2.1 Description

This was one of the few multiple till exposures found in the whole of the Hermitage Bay area. It is especially significant since no others

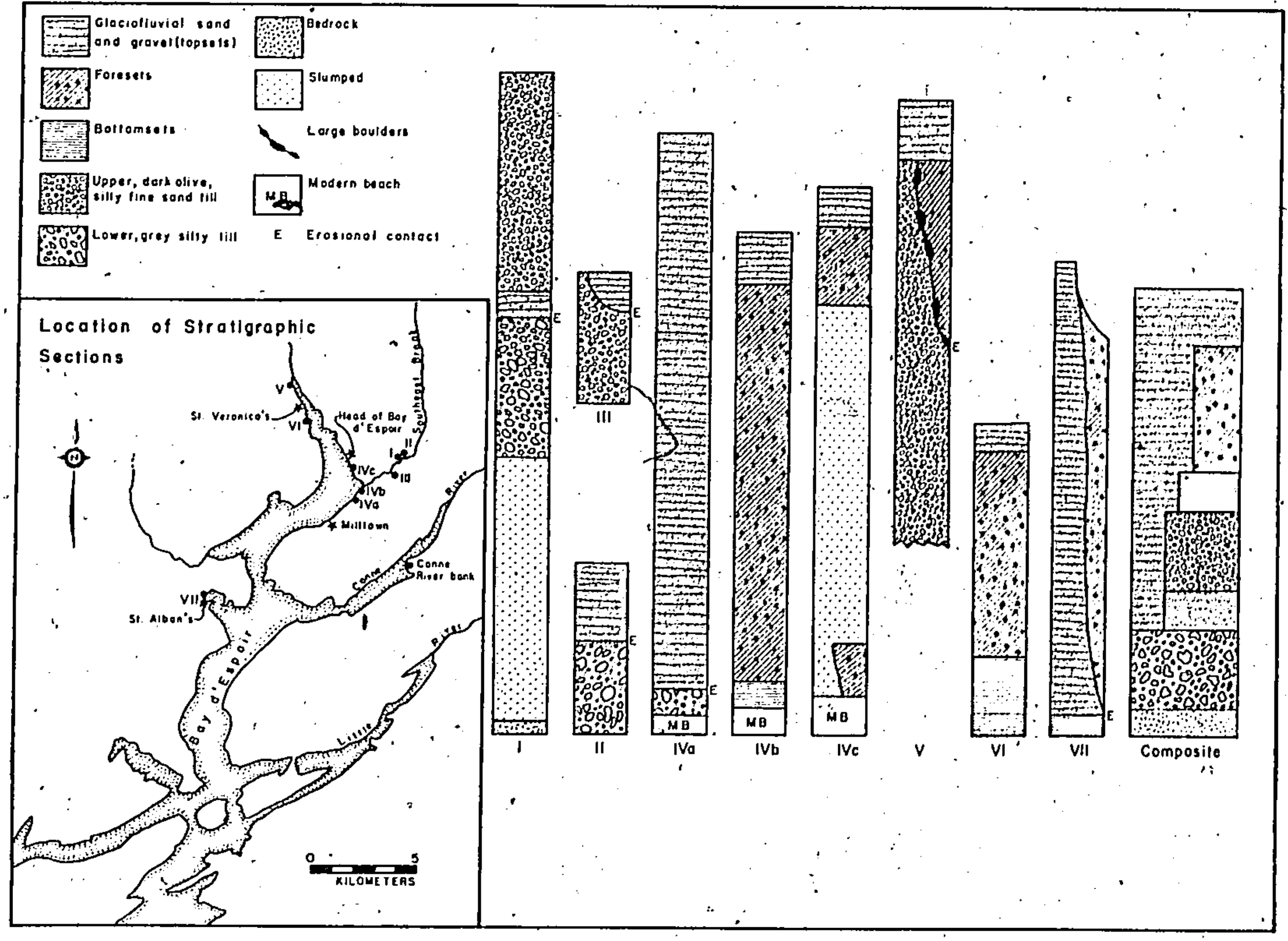


Figure 3.1: Stratigraphic sections in the northern Bay d'Espoir area. Composite section summarizes the overall stratigraphy and shows relationships of units to one another.

have been documented in the literature for this part of southern Newfoundland. The section is exposed on a meander loop of Southeast Brook in Trout Hole Falls Community Park (1M/13:955094)(Fig. 3.1). It is situated beyond the treed cliff in the northeast corner of the campgrounds. The upper portion of the section is well exposed in a recent slump scar, but the bottom nine meters are debris covered.

Unit 1: Grey silty sand till. The lowermost exposed unit is a silty sand (59% sa, 32% si, 9% cl) till with a sharp but irregular upper contact at ~ 4.8 m from the surface (Section I, Fig. 3.1). The most immediately noticeable feature is the grey color (Munsell color N 5/10), markedly different from the overlying and ubiquitious dark olive till characteristic of all local roadside ditches. The grey till is well indurated with a platy structure and few large clasts. Any observed clasts were striated, ranging from angular to subangular. No granite erratics were noted. Till fabric analysis shows the vector mean to be oriented towards 279° N, transverse to the valley axis of 45°-225°. This till is probably continuous to bedrock beneath the slumped material, as it can be observed directly overlying bedrock 0.5 km upstream.

Unit 2: Interbedded sands and gravels. The middle unit is dark olive in color (Munsell color 5Y4/3) and consists of disturbed and contorted interfingering lenses and beds up to 0.3 m thick of massive structureless sands, dirty gravel, and till. The unit is ~ 2 m thick with a sharp basal contact and a gradational upper one. The upper contact suggests an intimate depositional relationship with the overlying unit. Unit 2 was probably deposited from either subglacial melting processes (i.e., basal melt-out till, Dreimanis, 1976) or the deposition of meltwater transported sediments and flow till at the snout of an advancing glacier (Boulton, 1968).

Unit 3: Dark olive silty sand till. The uppermost unit is a silty fine sand (70% sa; 28% si, 2% cl), dark olive (Munsell color 5Y4/3) till with a blocky structure. In contrast to the basal till, granitic erratics up to boulder size are very apparent. At ~ 4.0 m from the surface there is a noticeable increase in sand content as the till grades down into Unit 2. Till fabric analysis showed a vector mean orientation towards 294°, similar that of Unit 1 and again, transverse to valley axis.

3.2.2 Discussion

The two tills exposed in this section were deposited by two separate advances. The fine nature of the matrix and the scarcity of granites in the lower till, deposited during the first advance, suggests local erosion of the underlying slate bedrock, a characteristic of lodgement till (Dreimanis, 1976). A second advance deposited the dark olive till and the underlying interbedded unit. The abundance of granite erratics implies a greater transport distance than for the lower till as the nearest granite outcrops occur 16 km northeast (Anderson, 1965). Although Boulton (1968) observed that till overlying stratified sediments need not always imply a second ice advance, a second advance is proposed in this instance. The color, textural and granite erratic abundance differences are strong arguments for two separate glacial events.

The pebble fabric - transverse to valley axis orientation and local striae is somewhat problematic. However, although fabrics transverse to ice flow directions are less common than parallel fabrics they do occur, due to stress systems at the base of the glacier (Boulton, 1968;

Dreimanis, 1976).

3.3 Section II: Southeast Brook

This is a small section exposed 0.5 km upstream (1M/13:957096) from Section I. It consists of 3 m of glaciofluvial sediments overlying 3.5 m of till resting on bedrock (Section II, Fig. 3.1). The glaciofluvial sediments have crude horizontally stratified sands and gravels with some cross-bedded sands (190° N paleocurrent direction). The till is the same grey till as Unit 1 of Section I.

3.4 Section III: Bay d'Espoir Rod and Gun Club

One kilometer southwest of the Bay d'Espoir Rod and Gun Club (1M/13:947085) is a large clearing on the valley wall revealing an exceptionally high concentration of large and small granite erratics scattered over the surface. Glaciofluvial sediments occur on the lower slopes of the clearing and in the highway ditch. In a small borrow pit a dark olive silty till occurs beneath the glaciofluvial material (Section III, Fig. 3.1). The till outcrops higher on the valley walls above the upper limit of glaciofluvial sediments and is also the same till as Unit 3 of Section I. The high concentration of granite erratics is the result of washing, where the fines were removed leaving only the larger boulders on the upper surface.

3.5 Section IV: Mouth of Southeast Brook

Outcrops of rhythmically laminated fine sand, silt and clay overlain by sands and gravels (Fig. 3.2) occur at several locations

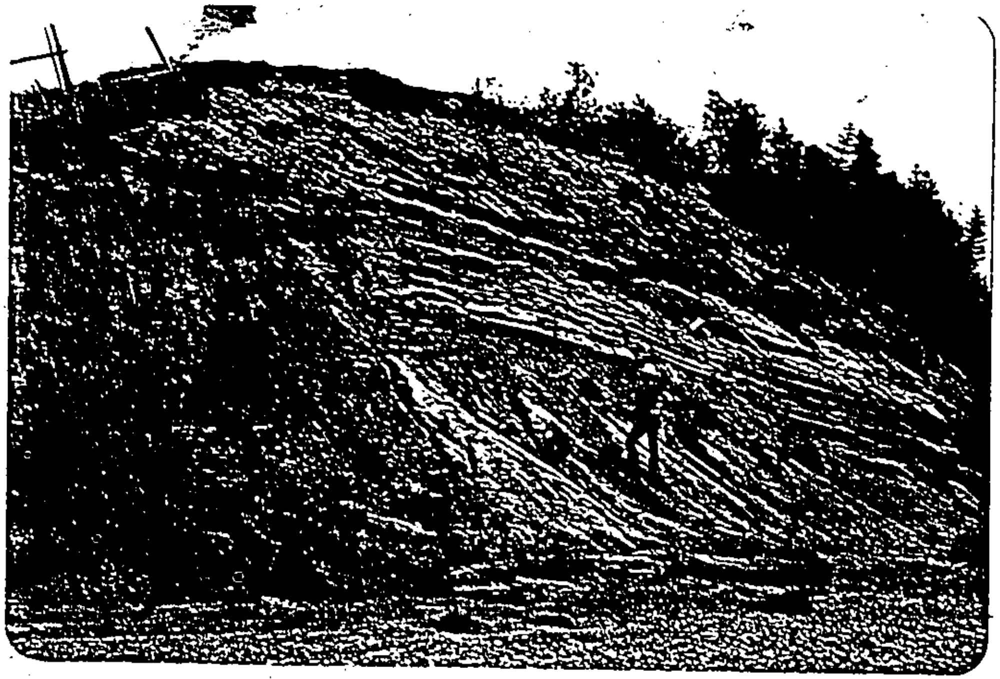


Figure 3.2: Section showing glaciolacustrine/deltaic sediments at St. Veronica's (photo location 1M/13:908097)

around the head of Bay d'Espoir: St. Alban's (1M/13:862025), Long Point (1M/13:909111), St. Veronica's (1M/13:903123), across from Black Duck Hole (1M/13:903123), Reuben Point on the Conne River (1M/13:957042), Bobbett Cove (1M/13:932075) and the mouth of Southeast Brook (1M/13:938081, 1M/13:934078, and 1M/13:932086); rhythmites alone occur at Hen and Cock Cove (1M/13:896075). Three of these outcrops near the present mouth of Southeast Brook are described and interpreted below. The reader is referred to Chapter 5 for a detailed description and interpretation of the sediments at the Conne River bank, considered to be the type section for these deposits.

3.5.1 Section IVa: Head of Bay d'Espoir

South of the bridge where Southeast Brook enters Bay d'Espoir (1M/13:938081) there is 10.7 m of slumped glaciofluvial or deltaic sands and gravels overlying ~ 1 m of grey (Munsell color N 5/10) silty till (Section IVa, Fig. 3.1) which is similar to the lower till of Section I.

3.5.2 Section IVb: Bobbet Cove

One kilometer southwest of the bridge, at Bobbet Cove (1M/13:934077 and 1M/13:932075), three sedimentary units can be distinguished (Section IVb, Fig. 3.1). The bottom unit, less than 1 m thick, consists of alternating layers of coarse to fine sand or silt, and clay. Individual laminae range in thickness from only a few sand grains to 3 to 4 mm thick. Overlying this unit is 14.9 m of seaward (westward) dipping 1 to 9 cm thick sands interbedded with thin silt laminae. The uppermost unit consists of ~ 2 m of glaciofluvial gravels, boulders and cobbles, the top of which has been removed by man.

3.5.3 Section IVc: Head of Bay d'Espoir

In Head of Bay d'Espoir (1M/13:932086) the top of a poorly exposed section (Section IVc, Fig. 3.1) shows northwest dipping beds of sorted and unsorted sands, gravels and cobbles. At 19.2 m asl they are capped by 1.5 m of structureless sands and gravels. Clast imbrication of the upper unit indicates flow to the west-northwest. Fifteen metres north and lower in elevation, there is a small bank of bedded sands and silt also dipping northwestwards.

3.5.4 Discussion

Sections IVa, b, c suggest deltaic sedimentation of glaciofluvial material that was transported down the valley of Southeast Brook into a body of water occupying Bay d'Espoir. Very fine sands, silts and clays settled out of suspension onto the basin floor forming rhythmic bedding which was subsequently buried by prograding deltaic sands and gravels. The coarseness of the foresets in Sections IVa and IVc suggests deposition fairly close to the distributary mouth whereas the finer sands at the base of Sections IVb and IVc may be a more distal facies. The variations in flow directions noted in Sections IVa and IVc would be the result shifting channels as the delta grew in a radial pattern from the river apex. The surface of the water body, as indicated by the top of the foresets of Section IVc, was at least 19.2 m asl.

3.6 Section V: St. Veronica's

North of St. Veronica's there is a small borrow pit (1M/13:903123) which exposes more than 6.0 m of well indurated dark olive (Munsell color 5 Y 4/3), platey, silty sand till (62% sa, 34% si, 4% cl) overlain by 2 m of structureless glaciofluvial sands, gravels and boulders (Section V, Fig. 3.1). Below the glaciofluvial material, cut into the till, is a large anomalous pocket of bedded sands and silts dipping in several directions. Several large boulders ~ 1 m across are located at the till/sand-silt contact giving the impression of a lag deposit, however the variable dip directions of the sands suggest an ice contact origin. Other similar sequences but without the underlying till, 0.5 km south and 1 km north, show seaward dipping sands overlain by glacial outwash. In all instances the gravel/sand-silt contact occurs at 22-23 m asl. The bedded sands are of deltaic foreset origin, whereas the gravels represent glaciofluvial topset beds.

The till at St. Veronica's, in spite of slight textural and color variations was deposited by the same glacial event responsible for Unit 3 of Section I, each till having been deposited by separate tongues of ice feeding from a larger ice mass situated further inland.

3.7 Section VI: Long Point

Coastal sections at Long Point (1M/13:910106) reveal a sequence of glaciolacustrine deposits (Section VI, Fig. 3.1) similar to that at Conne River (Chapter 5). The lowermost unit consists of subhorizontally laminated dark grey clays and silts with minimal amounts of sand. The upper contact occurs at 3 m asl but the base extends below sea level. This unit is overlain by ~ 8.0 m of coarsening upwards olive colored, laminated, seaward dipping silts and sands (true dip of 16° on a 153°

N strike). The section is capped at 10.7 m asl by 1.5 m of crude subhorizontally stratified glaciofluvial sands and gravels. This sequence is interpreted, based on the Conne River type section, as rhythmically deposited, proglacial lake bottom sediments, overlain by deltaic foreset sands. These in turn, are overlain by outwash deposits (the subaerial or topset portion of the delta).

3.8 Section VII: St. Alban's

At low tide along the modern beach at St. Alban's (1M/13:861025) rhythmically laminated grey clays, silts and sands can be observed. These are similar to the lowermost units of Sections IVb and VII, and at Conne River, interpretted as lake bottom sediments. Inland are extensive glaciofluvial deposits 18 to 30 m thick covering an area of ~ 3 km^2. Southeast, on Birchy Point (in Ship Cove) are 14.9 m of steeply dipping Gilbert-type foresets overlain by 1.5 m of crude horizontally bedded topsets. The foresets are constructed of sand, gravel and boulders up to 0.5 m across dipping towards 170°-180° N at angles as high as 29°. The sequence (Section VII, Fig. 3.1) represents a prograding delta advancing out over top of lake bottom sediments.

3.9 Interpretation and Discussion

From the stratigraphy described above it is possible to reconstruct a sequence of events responsible for the deposition of the sediments in the Bay d'Espoir area. The earliest event recorded in the sections is that responsible for the grey silty sand till at the base of Section I, II and IV. This till was deposited by at least the penultimate glacial event, the ice of which occupied the valley of Southeast Brook as either a valley glacier or more massive ice sheet. The washed sediments and flowtill (Unit 2) overlying the lower till of Section I were deposited as either subglacial meltout or at the snout of the advancing glacier responsible for the upper till. The two tills need not be separated by an interglacial or interstadial episode. It is possible that they represent deposition by a fluctuating ice front although the dissimilarities of the tills suggest different sources.

Figure 3.3 shows the proposed reconstruction of the maximum extent of ice coverage during the last glacial event to have affected the northern Bay d'Espoir area. It is based on the location of outcrops of the upper till and evidence presented in Chapter 2. This glaciation is probably the Late Wisconsin maximum and is the equivalent of Jenness's inner-outer drift limit (Section 2.8). Different tongues of ice feeding from the common inland ice mass situated to the north are responsible for the upper till at Trout Hole Falls Community Park (Fig. 10, Section I), and the surficial tills along Southeast Brook, St. Veronica's valley, St. Alban's valley, and the Conne and Little River valleys. This till was not observed in St. Alban's or the uplands on either side of town which were ice free, although the glacier advanced to within 1 km of the present townsite. The height of land northwest of St. Alban's remained ice free, projecting through the glacier as a nunatak. The glacier, unable to surmount the topographic barrier south of the airport, was deflected southwest and northeast into St. Alban's and St. Veronica's. Ice flowing down St. Veronica's valley and that of Southeast Brook coalesced in the Bay but did not extend on land much past Dawson Point

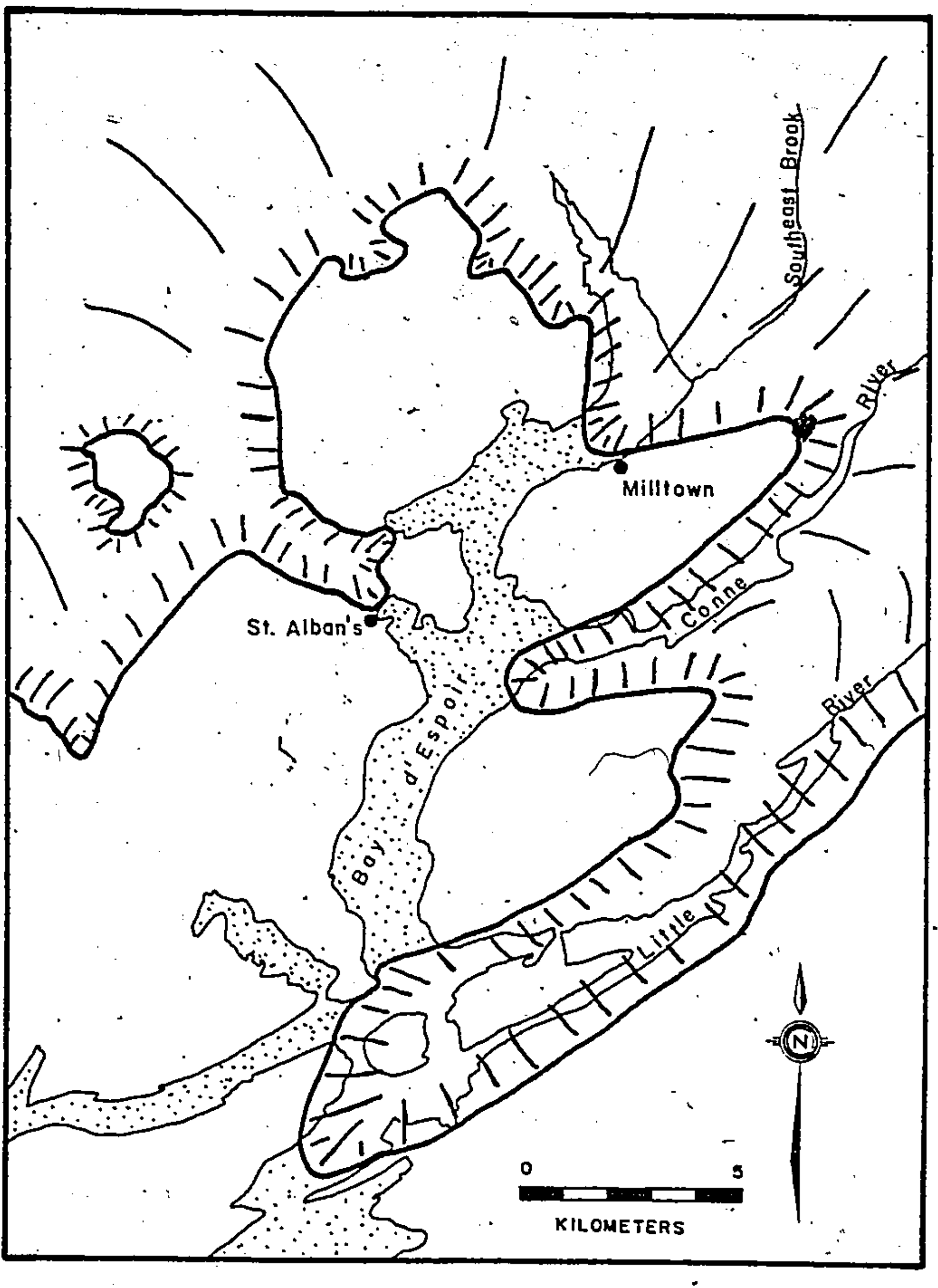


Figure 3.3: Maximum extent of Late Wisconsin glaciation in the northern Bay d'Espoir area. Valley glacier that flowed down Little River valley created large ice dammed lake to the north.

and Milltown. Similarly a tongue of ice flowed down the Conne River Valley to the mouth of the Bay leaving ice free areas between Milltown and Morrisville, and to the south, the upland between the mouths of the Little and Conne Rivers. The hills south and southeast of Little River were ice free as well.

Subsequent to the glacial maxima, deglaciation occurred depositing glaciofluvial, glaciodeltaic and glaciolacustrine sediments into the northern Bay d'Espoir and its valleys. Retreating ice tongues may not have been in phase with one another and consequently ice may have lingered in one of the valleys creating a large ice dammed lake. Modern glacier responses in adjacent valleys of the western North American Cordillera have been shown to be out of phase with one another (Hubley, 1956; Meier and Post, 1962; Post and Lachapelle, 1971, Fig. 37)(i.e., one valley glacier may be actively retreating while simultaneously a nearby one remains stationary, or advances) due to variations in mesoclimate, morphological and lithologic controls, the nature of ice-decay and volume of debris (Shaw, 1972). Air photo analysis showed a series of minor end moraines (Fig. 2.1) at the mouth of Little River and an ice limit on the valley walls suggestive of an ice tongue extending into the bay across Riches Island. The occurrence of lake sediments as far south as St. Alban's and Conne Rivers precludes these valleys as the source of the ice dam and necessitates a more seaward location. The most likely location for the ice dam is at mouth of the Little River valley. Bathymetric charts (Canadian Hydrographic Service, 1977) show a ridge here, which if structural in origin may have played a significant role in causing the ice to linger at this location while other valley glaciers receded (morphologic and lithologic controls of Shaw, 1972).

As the valley glaciers of St. Alban's, St. Veronica's, Southwest Brook and Conne River receded, meltwater and sediment graded to a higher base level flowed into the bay. Sections II and IVa indicate that much of the upper dark olive till in the valleys was eroded by the meltwater and that glaciofluvial deposits were deposited directly on top of the lower grey till. Higher up the valley walls (Section III, Fig. 3.1) and in St. Veronica's valley (Section V, Fig. 3.1) outwash was deposited over top of the dark olive till. As the debris laden meltwater entered the lake its velocity was suddenly checked and bedload transport of sand, gravel and cobbles ceased creating coarse grained deltaic foresets (top of Section IVc, Section VII and possibly IVa). Sand and silt were transported a greater distance before settling out, constructing low angle foresets (Sections IVb, IVc, V and VI). Fines were dispersed throughout the lakes as overflow, interflow and underflow currents before settling to the lake floor. Continuous progradation of the deltas caused the foreset beds to advance over the lake bottom sediments and ultimately for the subaerial glaciofluvial gravels to be deposited over the foresets creating classical deltaic sequences found in Sections IVb, IVc, V, VI and VII. The lake surface as indicated by the top of the foresets in Section VI had a maximum depth of at least 22-23 m asl. The composite section in Figure 3.1 shows the relationships and relative ages of the sediments.

Drainage of the lake would have occurred when the Little River ice dam gave way. Streams were then forced to grade to a new, lower base level (although perhaps not markedly different at the time of drainage due to isostatic depression) and subsequently cut down through previously deposited sediments.

CHAPTER 4

STRATIGRAPHY: SOUTHERN HERMITAGE BAY AREA

4.1 Introduction

This chapter describes and interprets stratigraphic sections discovered in the southern part of the field area. During the last glacial event this area was affected by a body of ice that was distinctly separate from the one which terminated around northern Bay d'Espoir. A coastal section at Deadman's Bight, previously discussed by Widmer (1950), is described in greater detail and reinterpreted. Till found at Pass Island Tickle and near Seal Cove probably predates the last glacial event. Fossiliferous material in till at Seal Cove is tentatively assigned an age of ~ 70,000 years.

4.2 Deadman's Bight-Harbour Breton area

4.2.1 Introduction

Deposits exposed in a 3 km coastal section at Deadman's Bight (1M/5:570860), southwest of Harbour Breton (Fig. 4.1) can be divided into six distinct sedimentologic units as well as two additional, related units exposed away from the shore in nearby borrow pits. Each unit is described below and its mode of deposition interpreted. Following this, five morphologic patterns expressed in unconsolidated surficial material are described and related to the sedimentology. Combined with paleo-sea level indicators a local detailed sequence of

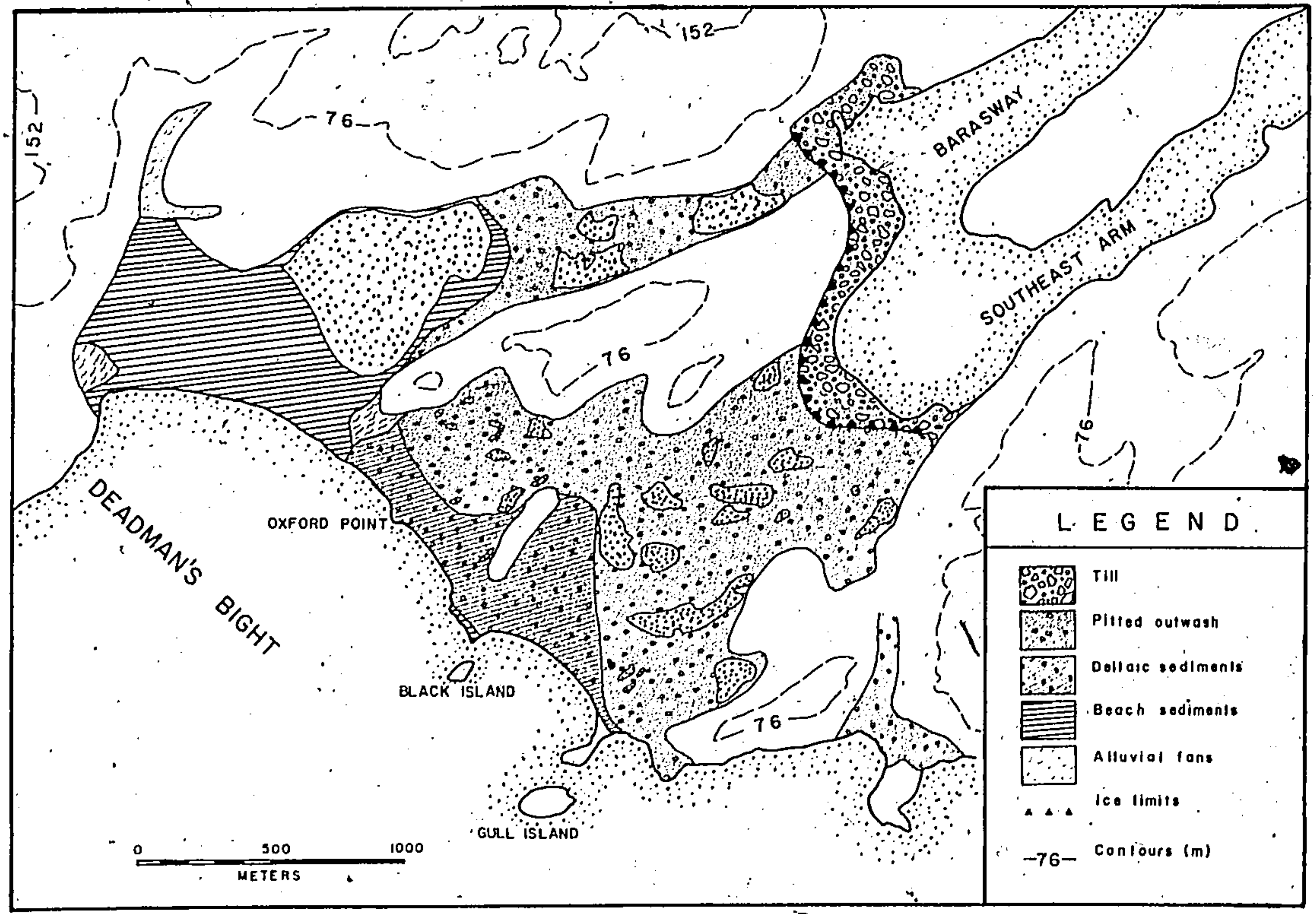


Figure 4.1: Terrain patterns in the Deadman's Bight-Harbour Breton Area.

glacial - early postglacial events is presented. Figures 4.1 and 4.2 should be referred to for the location of sedimentologic units and morphologic patterns. Widmer (1950) described the morphology of Deadman's Bight and recorded strandline elevations explaining that the unconsolidated deposits were a terminal moraine of a valley glacier which were subsequently affected by a fluctuating sea level. He designated a glaciated bedrock bench as Sangamon in age. Grant (1975b) briefly mentioned the area noting "accumulation of ice marginal deposits coeval with wave modification".

Deadman's Bight and Harbour Breton are situated at the south end of narrow peninsula projecting in Fortune Bay. The peninsula is bounded on the west and northwest by Connaigre Bay and on the east by the fiorded Northeast Arm. Deadman's Bight is a shallow offshore platform reaching a depth of only 50 m,3 km from the coast. The peninsula consists primarily of Late Ordovician volcanics along Western Head (Figure 2.2) and unconsolidated Quaternary sediments (Widmer, 1950; Anderson, 1965).

4.2.2 Stratigraphy

4.2.2.1 Unit 1: Till

Unit 1 is a well indurated, silty sand, brown till located above the modern beach immediately west of and flooring the stream northwest of Oxford Point (Figs. 4.1 and 4.2). It has a platy structure and contains angular to subangular clasts. There are occasional interbeds and discontinuous lenses of distorted sands and laminated silts. Granitic erratics are abundant, probably derived from outcrops located

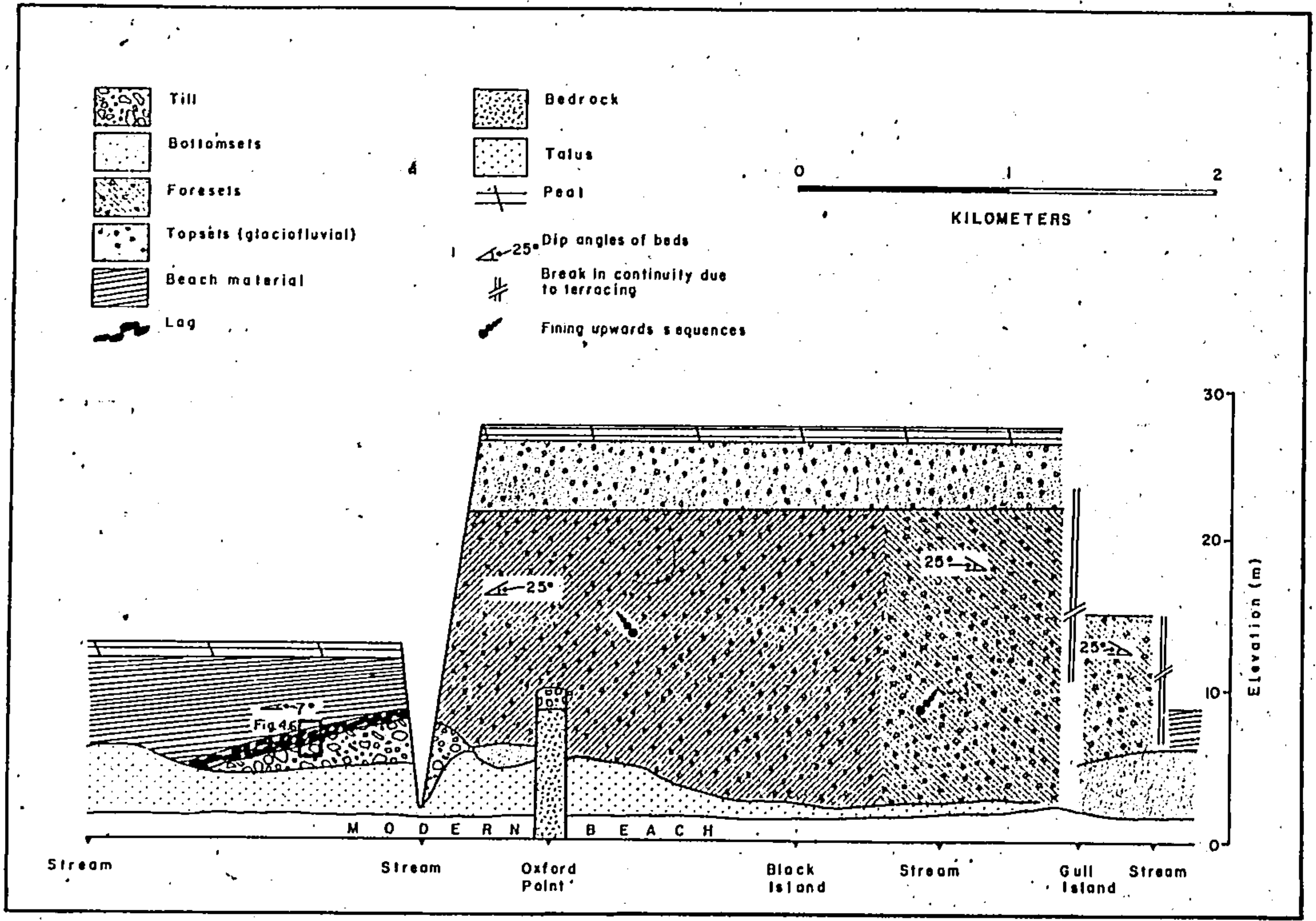


Figure 4.2: Coastal section revealed in Deadman's Bight.

to the northeast in Northeast Arm. Smoothed and striated bedrock is exposed beneath the deposit and on the beach immediately in front of it.

The till was deposited by a valley glacier that flowed down Northeast Arm through Southwest Arm and out onto Deadman's Bight (Section 2.2). Source of the ice was the small ice cap centred around the head of Hermitage Bay. The till is probably an ablation till having been released from the glacier in a watery, unstable state; debris flows, slumping and meltwater washing would account for the interbedded till and sorted sands (Boulton, 1968).

4.2.2.2 Unit 2: Bottomsets

Unit 2 outcrops above the modern beach beginning near Oxford Point and extends intermittently to the southeast limit of the exposure (Fig. 4.2). At one location, west of Oxford Point, it partially overlies Unit 1. The unit consists of low angle, seaward dipping beds of alternating silty fine sand and fine to medium sand. The coarser sand layers range in thickness from 5 to 40 mm and the interbedded silty fine sand laminae are less than 10 mm thick. Occasional pebbles are interspersed throughout the laminae. Other than parallel bedding no other primary sedimentary structures were observed. Secondary deformation is common.

This unit, in stratigraphic association with overlying units 3 and 4, is interpreted as the bottomset beds of a prograding delta, which settled out of suspension onto the sea floor. The alternating silty fine sand and sand layers are rhythmically bedded suggesting periodic episodes of deposition. Pebbles within the strata may be ice rafted. The deformation features are the result of loading by the overlying units.

4.2.2.3 Unit 3: Gilbert-type foresets

Unit 3, overlying Units 1 and 2 (Fig. 4.2), consists of 17 m of steeply dipping strata (dip angles ~25 to 27°) of silt, sand, gravel and boulders (Fig. 4.3). Individual beds, up to 1 m thick, are normally graded perpendicular to dip. There are clast supported cobbles and boulders as large as 0.5 m across at the base grading upwards to gravel, sand and laminated silt at the top. Each bed has sharp upper and lower contacts. The beds dip southwestwards at the west end of the section and south-eastwards at the east end. Striations have been preserved on individual cobbles and boulders.

The sediments of Unit 3 are classical Gilbert-type deltaic foresets. Gilbert-type deltas, especially gravelly ones, are uncommon, rarely being preserved in the geologic record (Jones, 1965; Stanley and Surdam, 1978). Barrel (1912) noted that sediment laden, torrential streams entering deep, quiet water were conducive to foreset bed construction. Jones (1965), studying Lake Bonneville deltas observed that in coarse deltaic systems gravels predominated on the foreset beds whereas fine sands and silts occurred on the bottomsets. From this, it is inferred that the Harbour Breton foresets were deposited into a relatively low energy environment, for there is no indication of erosion, reworking, or winnowing of the sediments as would be expected on a coastline dominated by wave or tidal processes (Coleman and Wright, 1975). Furthermore, preservation of the underlying fine grained bottomsets (Unit 2) also favours quiet water conditions.

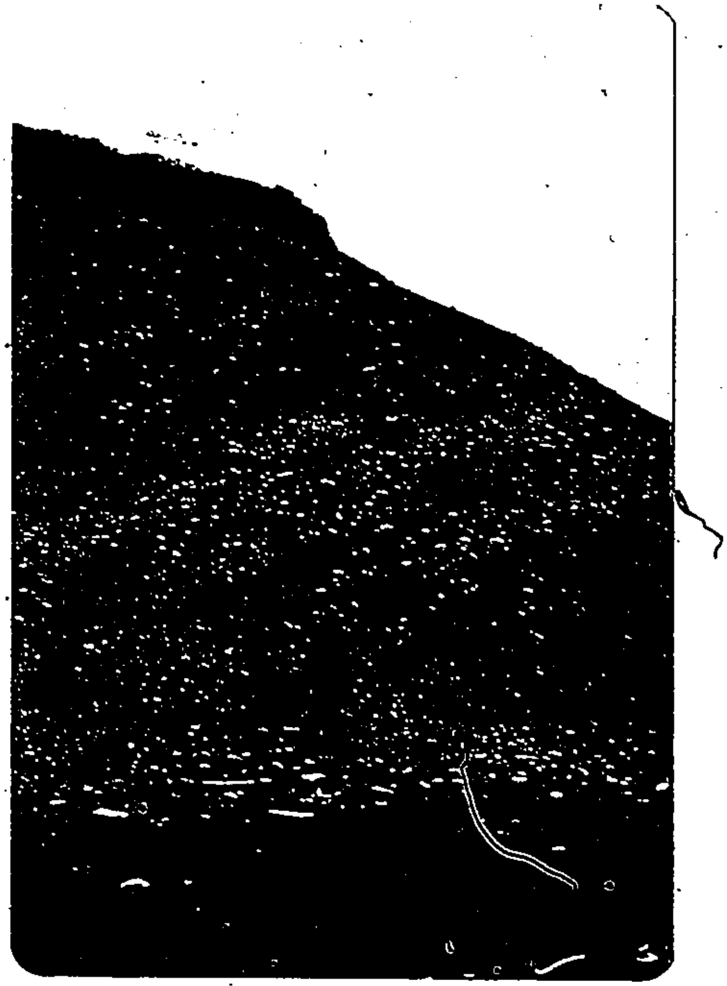


Figure 4.3: Steeply dipping, normally graded, Gilbert-type foresets. Person for scale 1.5 m tall.

Striations preserved on clasts indicate that the source of debris was glacial and suggests minimal abrasion and thus a short bedload transport distance. The normally graded beds with sharp contacts represent rhythmic episodes of flow beginning with extremely high discharges capable of transporting boulders up to 0.5 m across as bedload, subsequently waning to flows where fine sand and silt were able to settle out of suspension. There is no good indication as to whether the period of flow is diurnal, seasonal or annual although discharges capable of repeatedly transporting 0.5 m boulders may suggest spring peak meltwater floods rather than diurnal maxima. The divergence of foreset dip directions at either end of the exposure would be the result of shifting or multiple braided outwash channels on the subaerial portion of the delta.

4.2.2.4a Unit 4a: Topsets (glaciofluvial)

Unit 4a consists of ~ 5.5 m of crude horizontally bedded, poorly sorted, imbricated pebbles and cobbles with occasional lenses of sand and silt. It is capped by 1 m of peat. This unit is interpreted as braided glaciofluvial outwash of proximal origin forming the subaerial, or topset, part of the delta that provided sediment for and prograded over, Unit 3. Sediments with similar internal structure have been observed in longitudinal bars on the proximal reaches of braided glaciofluvial rivers (Rust, 1972a; Boothroyd and Ashley, 1975, Fig. 25).

4.2.2.4b Unit 4b: Pitted (glaciofluvial) outwash

Unit 4b was not observed in the coastal section illustrated in Figure 4.2 but is exposed in small borrow pits bordering the town of Harbour Breton. It has a slightly higher proportion of sand and silt, and horizontal bedding is better defined than in Unit 4a. This unit is also interpreted as braided glaciofluvial outwash. The major feature differentiating it from Unit 4a is the occurrence of fault and slump structures in the stratification (Fig. 4.4). They result from ice blocks which have become buried by outwash and subsequently melted out removing the support for the overlying sediments causing them to shift and slump. Eyles (1977) suggested that in order for the ice blocks to be buried a rapid rate of glaciofluvial deposition is necessary.

4.2.2.5 Unit 5: Beach material

Unit 5, located at the western edge of the exposure has a crude wedge shape and ranges in thickness from 3 to 8.5 m. It consists of low angle strata dipping obliquely seawards with a true dip direction at the west end of the outcrop of 7° on a 227° N strike. Occasional higher apparent dip angles of up to 13° were also recorded. Individual beds, 6 to 40 cm thick, are laterally continuous (Fig. 4.5). One bed could be traced parallel to the cliff face for ~ 250 m with no evidence of channelling, slumping or any other disturbances. Individual beds vary in composition from layers of well sorted sands to matrix free, clast supported cobbles to matrix supported sands and gravels. The majority of the clasts are rounded to subrounded with a maximum observed cobble size of ~ 20 cm. The upper surface of the unit

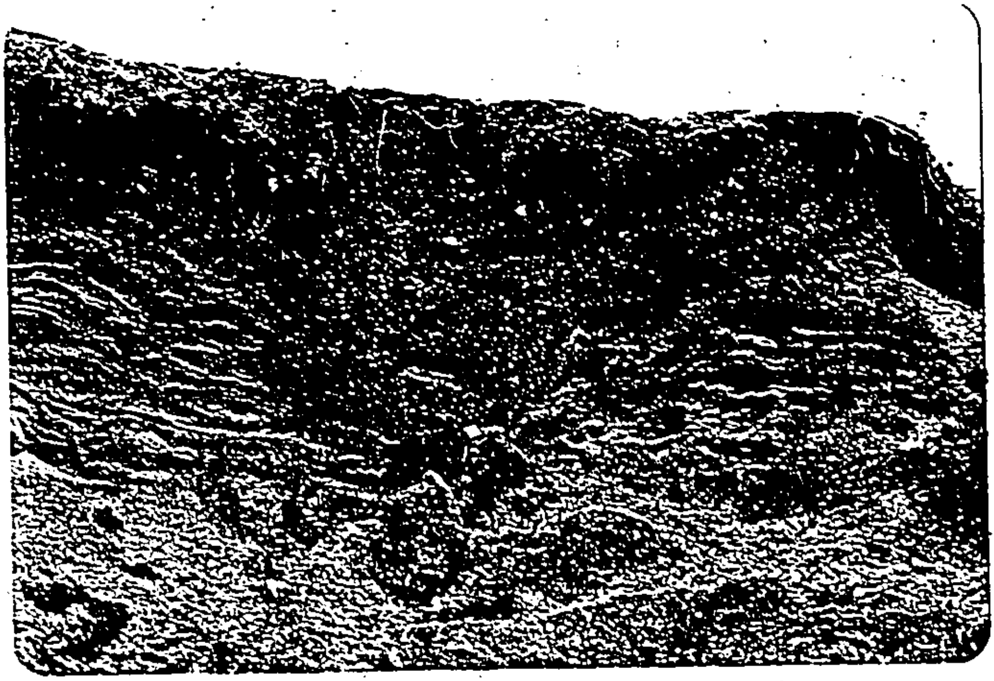


Figure 4.4: Faulted, slumped, pitted glaciofluvial outwash.

Figure 4.5: Low angle, seaward dipping, laterally continuous beds of raised beach. Person 1.5 m tall.

is horizontal and capped by 1 m of peat.

This unit is interpreted as beach sediments deposited in the foreshore, or swash, zone (swash zone describes wave action and currents in the near-shore region whereas foreshore describes part of the beach profile defined by wave swash). Gently seaward dipping, parallel laminations are characteristic of the foreshore zone of a beach (Thompson, 1937; Clifton _et al_, 1971). Whiteman (1975) recorded a dip angle of 5 to 14° on the foreshore beds of a raised Pleistocene beach in Nova Scotia. On the day of observation waves breaking on the modern beach at Deadman's Bight were parallel to the dip direction of the raised beds (i.e., the beds of Unit 5 dip towards 227° N while the waves were approaching the modern beach at a 50°-230° N heading). Variations of grain size and structure are the result of fluctuating wave conditions. Beds having high concentrations of pebbles and cobbles were deposited under high wave energy conditions (storms), the concentration being due to either fines being winnowed out or the coarser material moving landward (Bluck, 1967; Whiteman, 1975). The higher concentrations of sand making up more massive beds were deposited by quieter water.

4.2.2.6 Unit 6: Beach lag deposit

Unit 6 is an elongate lens of structureless sediment dipping southwest, overlying Unit 1 and underlying Unit 6 (Fig. 4.6). Clasts, primarily boulders (some more than 1 m across) and cobbles, are subangular to rounded. There is a noticeably higher proportion of larger and more angular material than occurs in Unit 5, above. The

Figure 4.6: Till overlain by lag deposit overlain by beach material. Location of photo shown in Figure 4.2.

unit is at least 1 m thick but may be thicker as it is talus covered at its western end. It has a dip direction roughly paralleling the beds of Unit 5.

Unit 6 is a lag deposit of cobbles and boulders which exceeded the competency of the processes responsible for the overlying beach sediments. Consequently, this material accumulated as a crude boulder pavement while smaller sediment was winnowed and transported away. The source of the cobbles and boulders may have been from the underlying till (Unit 1) or possibly part of Unit 7, discussed below.

4.2.2.7 Unit 7: Alluvial fan

Unit 7 is not exposed along the coastal section but can be observed in a small borrow pit along the road 1.5 km to the north. It consists of crudely stratified, poorly sorted sands and gravels with angular to subangular clasts. Channelling or lensing is evident and there is a very high concentration of local stones. This unit is higher in elevation than Unit 5 and is in close proximity to a small stream flowing out of the hills to the north.

These sediments are quite similar to glaciofluvial sediments described in Section 4.1.2.4a. However since the morphology of the deposit has a crude fan shape and is near to a break in slope of a stream flowing out of nearby higher gradient hills it is interpreted as an alluvial fan. Alluvial fan deposits are typically "poorly sorted, immature, coarse grained sediments. Usually gravel, cobblestones and boulders predominate .... alluvial fan sediment essentially represents a conglomerate of local provenance" (Reineck and Singh, 1973).

4.2.3 Morphologic Expression

Surficial expression of unconsolidated material in the Harbour Breton area permits the recognition and demarcation of five distinct terrain patterns on 1:50,000 scale air photos (National Air Photo Library, 1966 Lockwood Series, flight line A19832, photos 27, 28, 29). These patterns, outlined in Figure 4.1, can be related to the various units outcropping in the section described above.

Pattern 1 is a narrow strip of northeast dipping land immediately west of Southwest Arm and the Barasway. It is till, believed to be the equivalent of Unit 1 in the coastal section. Pattern 2 has an irregular, highly kettled morphology. It is pitted outwash (Units 4a and 4b) formed where blocks of glacier ice were buried by glaciofluvial sediments and subsequently melted out. The faulting and slumping observed in Unit 4b (Fig. 4.4) is the result of smaller blocks of ice which probably floated down the meltwater streams, become grounded and covered with outwash. The large depressions, now kettle lakes visible on air photos are due to larger ice blocks, detached and isolated in situ from the glacier tongue as it retreated, being buried and subsequently melting out.

Abutting and seaward of Pattern 2 is an area of flat unkettled sediment. This is Pattern 3 which corresponds with the deltaic foresets (Unit 3) overlain by glaciofluvial outwash (Unit 4). The Pattern 2-Pattern 3 contact is interpreted as the proximal limit of the deltaic foresets. It may also represent the limit of the higher relative sea level and marine incursion into which the foresets were deposited.

Pattern 4, a flat lying surface topographically lower than the glacial outwash and deltaic foresets forms a large indentation inland from the sea. ~~It is~~ related to Unit 5, the beach material, and ~~is~~ the result of a marine incursion subsequent to and at a lower level than the one responsible for ~~the~~ foresets. Pattern 5, alluvial fan deposits, is found in three locations (Fig. 4.1). In each it has prograded out over Pattern 4 and is, or was, in close proximity to a stream.

4.2.4 Indicators of Past Sea Levels

The deltaic foreset-topset (Units 3 - 4a) contact represents the highest raised sea level measurement recorded in the Harbour Breton area. Gilbert-type foresets are constructed only into subaqueous environments and thus sea level at the time of their deposition must have been at least to the top of the foresets at 22 m asl (i.e., this value plus unknown depth of stream). A strandline slightly higher than the top of the foresets can be traced paralleling but above the northern and western margins of Pattern 4 (Fig. 4.1). The 22 m asl elevation is indicative of the marine limit in this area and is in close accord with the location of the 20 m isopleth of postglacial emergence illustrated by Whiteman and Cooke (1978, Fig. 1). In detail though, their line should be shifted a short distance southwards.

As sea level fell from its late glacial maximum of ~ 22 m asl terrace levels were cut into the unconsolidated material. The upper surface of the beach sediments (Unit 5, Pattern 4) represents a sea level stillstand at 11.5 m asl where a marine incursion occurred, eroding and cutting back deltaic and outwash sediments. Other less prominent terrace levels occur at 19.5, 17, 10, and 5.5 m asl. The glacially grooved, striated bedrock surface at Oxford Point is a marine bench levelled at 9.5 m asl.

4.2.5 Discussion

A sequence of events for the Harbour Breton area is proposed below based upon the stratigraphic relationships of the various units, surfical morphology and marine strandlines. No dateable organic material was found in the area and consequently the absolute chronology remains speculative. The oldest recognized glacial feature is the raised bedrock bench at Oxford Point. It has been molded and striated by glacier action subsequent to its formation and therefore predates at least the last glacial event. Widmer (1950) and Grant (1975b) assigned a Sangamon age to its planation although an interstadial origin cannot be conclusively ruled out. No deposits predating the last glacial event were recognized. As demonstrated in Chapter 2 the Late Wisconsin glaciation in this area was of limited extent. A valley glacier fed down Northeast Arm from a small ice cap centred north and east of the head of Hermitage Bay. This ice flowed through the Barasway and Southwest Arm of Harbour Breton onto Deadman's Bight at least as far as Connaigre Head Peninsula where there is a large kettle lake. The till of Unit 1 and Pattern 1 was deposited at this time. Subsequent to reaching its maximal position the ice receded to an ice limit defined by the till-outwash contact (Pattern 1 - Pattern 2 contact, Fig. 4.1). Substantial amounts of glaciofluvial sediment (Pattern 2) were transported and deposited, burying large and small blocks of ice which had become detached from the main body. The ice blocks later melted out forming the faulted and pitted outwash (Unit 4b and Pattern 2).

Due to worldwide deglaciation at this time eustatic sea level was rising at a greater rate than local isostatic rebound and relative sea levels were higher than present. Marine limit for this area was at least 22 m asl and it is to this level that the glaciofluvial outwash was graded, and into which the deltaic foresets and bottomsets were deposited. The undisturbed nature of the Gilbert-type foresets and laminated bottomsets suggests deposition into quiet water. In order to account for quiet water conditions it is necessary that a protective structure of some sort have formed across Deadman's Bight, acting as a breakwater. Even had the offshore lake proposed by Widmer (1950) existed it would have had a large fetch and a protective barrier would still have been required.

As relative sea level dropped, the unconsolidated material underwent terracing and the breakwater structure responsible for quiet water was destroyed. A significant stillstand (where isostatic uplift equalled eustatic sea level rise for a period of time) occurred when sea level was at ~ 11.5 m asl and was responsible for a major marine incursion which deposited the beach sediments (Unit 5, Pattern 4). Adjacent previously deposited deltaic material and till would have provided an ideal source of supply for beach construction. Bedrock outcrops along the modern beach and at Oxford Point were protective and prevented all the deltaic sediment and till from being reworked and eroded. The beach environment was a high energy one similar to the modern beach capable of transporting and reworking large cobbles. Cobbles and boulders exceeding the competency of the beach processes accumulated as a lag under the beach sediment (Unit 6, Fig. 4.5). Source of the boulders for the lag was the deltaic sediments, the underlying till, and/or alluvial fan material (Unit 7). The alluvial fan sediments were transported seaward during the incursion and also subsequent to it, as relative sea level fell, depositing the sand and gravel (Unit 7, Pattern 5) out over top of the beach material.

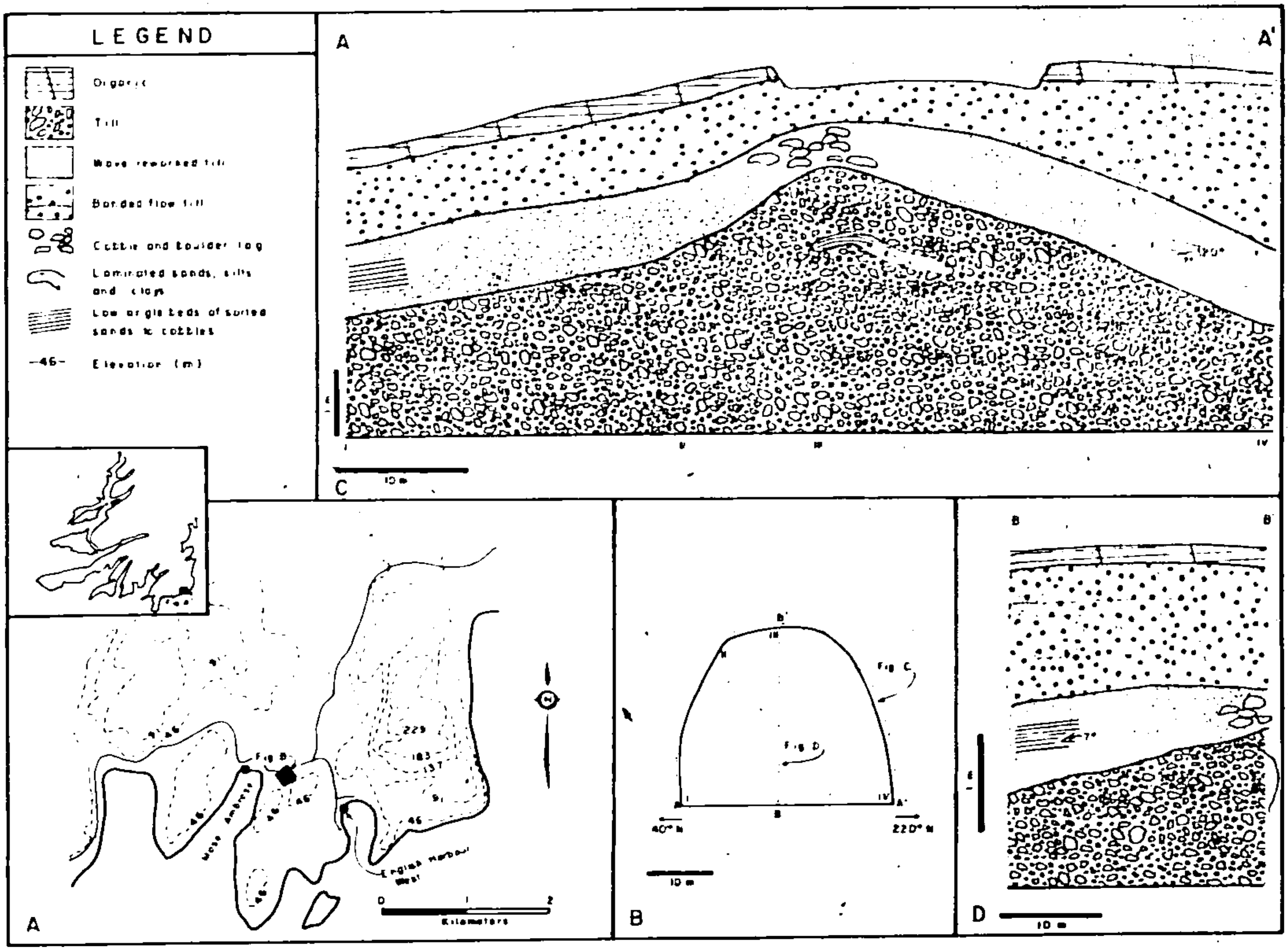
The lack of dissection of the unconsolidated material in this area suggests that only a relatively short period of time has elapsed since its deposition. This is an admittedly weak criterion but it reinforces previously made conclusions that the deposits are Late Wisconsin in age (Section 2.8).

4.3 Mose Ambrose

4.3.1 Description

A small borrow pit at Mose Ambrose (inset, Fig. 4.7A) exhibits an unusual set of deposits. It is located on the south side of the road (1M/5:124577) as the town is entered from the west at an elevation of less than 18 m asl. The shape of the pit was a half oval (Fig. 4.7B) with good exposures all round. It is floored by till which is also the lowermost unit of the section (Fig. 4.7C). The brownish-grey till (Munsell color 7.5 YR 5/1) has a silty, fine sand texture and a fissile structure. Clasts are striated, subrounded to angular, and are up to boulder size. The unit is thickest at the south-east side of the pit (Location III on Figs. 4.7B and C) and thinnest

Figure 4.7: Mose Ambrose borrow pit: A. Location of borrow pit; B. Plan view of the pit showing orientation, reference locations (I, II, III, IV) and positions of cross sections AA' and BB'; C. Cross section AA' of the pit showing units. Imagine the cross section to have been wrapped around the pit in Figure 4.7B; D. Proposed reconstruction of the deposit as it would have occurred in cross section BB', Figure 14B, prior to removal of the sediments. The elevation of the boulder concentration at B' coincides with the terrace levels in Mose Ambrose and represents the upper wash limit formed during terracing.


LEGEND
Organic
Till
Wave reworked till
Banded flow till
Cobble and boulder lag
Laminated sands, silts and clays
Low angle beds of sorted sands to cobbles
-46- Elevation (m)
A
A'
I
II
III
IV
1 m
10 m
C
B
B'
Fig C
Fig D
40° N
220° N
D
7°
A
Fig B
Moss Ambrose
English Harbour West
229
183
137
91
46
0
1
2
Kilometers
N

at its northwestern edges (Locations I and IV on Figs. 4.7B and C), there being 2.2 vertical meters difference over a 26 m horizontal distance on a 120^{o}-300^{o} strike (transect BB' in Figs. 4.7B and D). Near location III in Figure 4.7C there is a 8 m long, 20 cm thick lens of well sorted laminated sands, silts and clays within the till. The lens is well preserved in spite of a post-depositional deformation. Individual laminae can be traced the length of the deposit. It is interpreted as supraglacial in origin perhaps having settled out of suspension into a pond at the surface of the ice. Deformation occurred when the underlying ice melted out. The fissile nature of the surrounding till does not negate supraglacial origin,as Dreimanis (1976) noted that fissility occasionally occurs in supraglacial material as well as lodgement tills.

Sitting erosively on the till is a unit of crudely stratified, in places almost structureless, reddish coloured sands and gravels. Individual clasts are subangular to angular with angular predominating. This unit is thickest,and the underlying till surface lowest in elevation,at locations I and IV, Figures 4.7B and C. Here the sediments are better sorted, smaller (sands to cobbles) and have gently dipping beds (a dip direction of 7^{o} on a 320^{o} strike at Location IV) and occasional pinch outs. The unit thins and increases in elevation towards the southeast wall of the pit where at location III, it is 0.7 m thick. There is also a notable change in its structure at Location III where structureless clast supported gravels, cobbles and boulders (some greater than 0.5 m across) predominate. In effect, where the unit is at its highest elevation it is also thinnest, has the highest proportion of coarse material and is most highly disorganized. Its upper surface at Location III is 15.2 m asl which is the same as nearby terraces of wave reworked till at Mose Ambrose (1M/5:122573). The internal structure, low dip angle of the beds and elevation coincident with a nearby terrace suggest this unit is wave reworked till (beach sediments). The structureless large boulders may be washed till where the fines have been removed at the upper limit of wave action.

Overlying the sands and gravels are up to 2.2 m of unsorted, subtly stratified till-like material. Stratification is defined by bands ~ 10 to 35 cm thick having slight variations in colour, or by crude subhorizontal rows of pebbles and cobbles within and at the base of individual bands. The material is somewhat coarser gravel than the lower till and there are boulders up to 0.3 m across. It is interpreted as flow till deposited at or near the snout of the glacier. Boulton (1971) described how stones and boulders will settle to the bottom of a "mobile, liquid flow" of till at the surface of a glacier. This would account for the lineations of clasts. Boulton also noted that liquid flow till is often washed of its fines which would account for the slightly coarser nature of the till. The multiple banding may be the result of repeated flows as the till, in a semi-viscous state moved downslope (Boulton, 1971). The slight variations in colour cannot be satisfactorily explained.

4.3.2 Discussion

The sediments exposed in this borrow pit were deposited from a valley glacier which flowed south from an area of large coalescent inland cirques (Section 2.2). By the time the ice reached the Mose Ambrose area it was relatively thin and of limited extent as shown by a faint trim line or ice limit at 60 to 76 m asl immediately north and west of the townsite. Ice was prevented from entering English Harbour West by a ~ 41-61 m bedrock barrier. When deglaciation occurred and higher relative sea levels prevailed, previously unconsolidated material underwent terracing at the 15 m asl mark. Figure 4.7D is a reconstruction of the stratigraphy that would have been present along transect BB' of Figure 4.7B prior to the anthropogenic removal of sediment in the pit. The level of the upper cobbles and boulders at B' is coincident with the terrace levels on either side of Mose Ambrose and the internal structure, profile, and thicknesses of wave reworked till are very similar. The concentration of boulders at B' is a washing limit or lag (Synge, 1977) signifying that coastal processes were sufficient to remove fines but not the larger gravels, cobbles and boulders. Had the sands and gravels not been overlain by the flow till, this pit would have been a continuous part of the terrace extending from the harbour.

The banded flow till may have formed initially as melt-out till in the terminal zone of the glacier and subsequently moved downslope covering the underlying beach deposits. In order that the flow deposits have covered the wave reworked sands and gravels it is implicit that the glacier snout have been in the immediate vicinity of the borrow pit at the time of deposition, possibly having undergone a minor readvance from a more recessional position. The readvance would not have reached Mose Ambrose as the terrace in the Harbour is well preserved and not overlain by subsequent flow till deposits.

4.4 Pass Island Tickle

4.4.1 Description

The Pass Island Tickle coastal section (11P/8:627604), at the southwest tip of the Hermitage Peninsula directly east of Pass Island (Fig. 4.8), consists of about a kilometer of well exposed sea cliffs. Most of the local bedrock although weathered and generally devoid of till does have a distinct ice molded appearance. There is a small southwest oriented cirque and col located ~ 1.5 km north of the outcrop (Fig. 2.1). Hermitage Bay, to the north, separates the peninsula from the island by a greater than 300 m deep channel. Fortune Bay, to the south, is relatively shallow only reaching 100 m deep, 6 km out to sea. Widmer (1950) acknowledged the existence of Quaternary sediments off Pass Island and Grant (1975b, Fig. 1) speculated on an interglacial (?) marine bench under till.

4.4.1.1 Unit 1: Periglacial rubble

The bedrock at the base of this section has been included and described as a Quaternary deposit because of some of its unusual features. There is no indication of a till covered marine bench as suggested by Grant (1975b)(Section 2.6) but, rather, it merely has the appearance of being till covered bedrock. The bedrock surface is not smoothed and striated as would be expected underneath a till cover. Instead, there is a rubbly broken surface of fractured bedrock which grades upwards into till accompanied by an increase of the matrix content. It is similar to "head" described by Eyles and Slatt (1977) on the Avalon

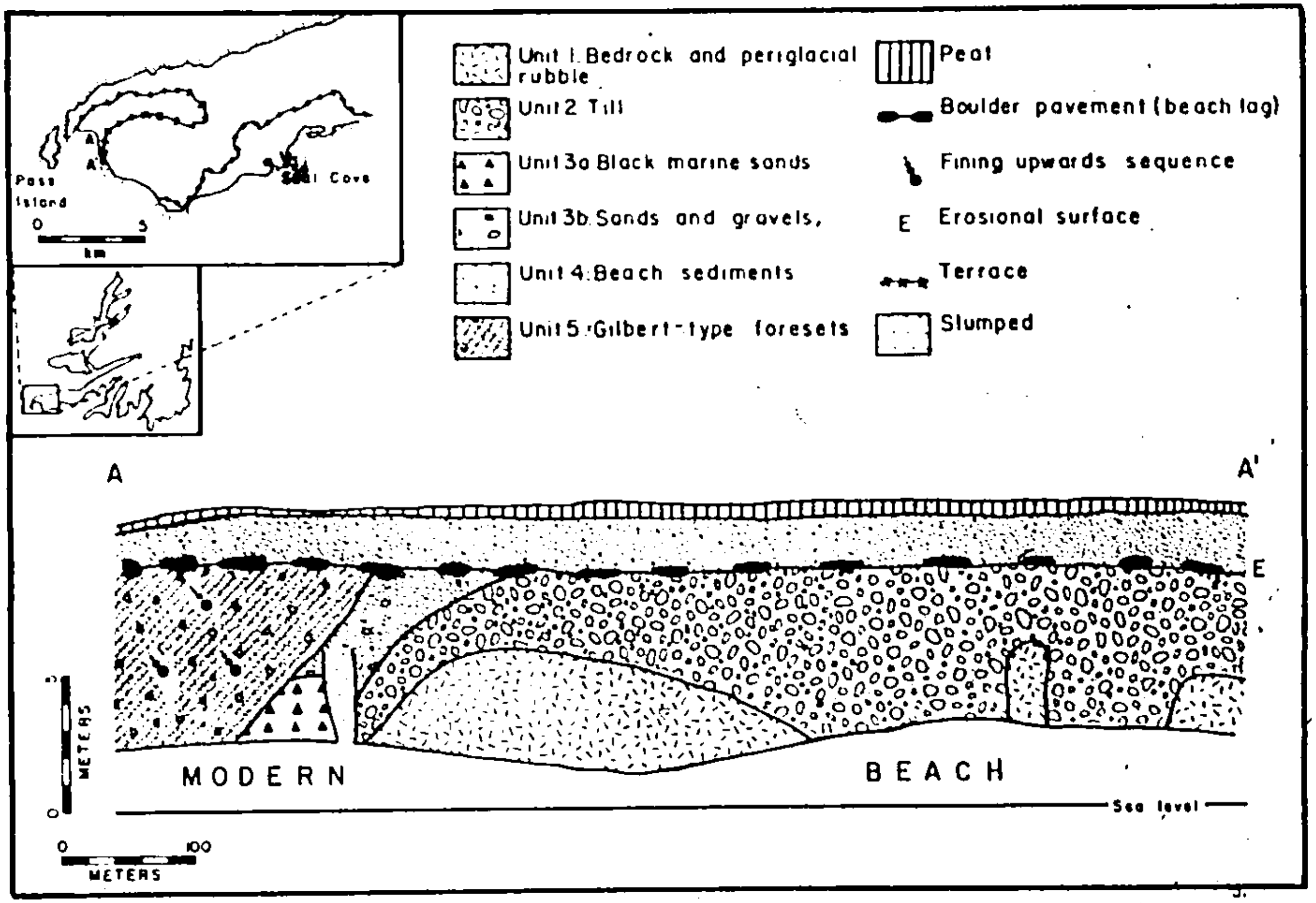


Figure 4.8: Pass Island Tickle coastal section.

Peninsula and recorded at many locations in Britain (e.g., Bowen, 1973). Eyles and Slatt (1977) defined head as a periglacial rubble resulting from the breakup of bedrock during a period of climatic deterioration preceding glaciation. As such it would signify a period of periglacial weathering previous to the deposition of the overlying till.

4.4.1.2 Unit 2: Till

The overlying till (Fig. 4.8) is stoney with a platy structure and sand matrix (78% sa, 22% si). Clasts range from subrounded to very angular and are as large as 3 m across. The colour is greyish brown (Munsell color: 5 YR 5/2) although it appears to have a faint pinkish tinge due to the high concentration of coarse local granite. There are erratics of green volcanic rock which outcrop immediately to the north and northeast (Williams, 1967). Two slate erratics were found which broke apart into thin plates upon handling. Similar slate bedrock was observed in outcrop (Bay d'Espoir Series Slate, Anderson, 1965) in the northern Bay d'Espoir area, although closer outcrops occur 20 km north in northern Long Island. Till fabric analysis showed a vector mean having a strong 46-226° orientation with a minor 110°-290° transverse mode.

4.4.1.2 Unit 3: Marine sands

Unit 3 (Fig. 4.9) is at the same stratigraphic level as, but north of, Units 1 and 2 (Fig. 4.8). The lower third of the unit consists of massive, somewhat consolidated, very dark grey to black sands (78% sa, 22% si). There are no large clasts and few small ones. Whatever stones are present are angular to subangular. Pebble fabric analysis showed a strong vector mean orientation of 121-131° (dipping towards 121°)(Fig. 2.9). Above the dark grey/black sands there is ~ 5 m of yellow brown, oxidized, interbedded, thin sandy silt (39% sa, 56% si, 5% cl) and thicker coarser sands (100% sa). Internal bedding has been highly distorted and the original structure is difficult to discern. Angular pebbles and cobbles, although not abundant, are found throughout the sands. The contact between the two coloured sands was obscured by a slump and gully.

The origin of Unit 3 is problematic. Although there is a slight possibility that it is a sandy marine till, a more likely interpretation is that it is a marine sand deposited in the nearshore during a period of higher sea level.

4.4.1.3 Unit 4: Gilbert-type foresets

Partially overlying and north of Unit 3 (Fig. 4.8) are a series of Gilbert-type foresets (Fig. 4.9) having dip angles up to 24 to 26° towards a general 340° to 20° N direction. Individual sets are 0.7 to 1.0 m thick, clast supported at the base and normally graded (perpendicular to foreset dip) with clast size ranging from 0.5 m boulders to coarse sands and gravels. Upper and lower contacts of individual sets are sharp. Top of the foresets occurs at ~ 9 m asl, but this value is of little use as an indication of past sea level as the upper part of the unit has been eroded and there is no indication of the original height. The base of the foresets extends to below the modern beach.

Figure 4.9: Black marine sands overlain by Gilbert-type foresets which in turn are overlain by wave reworked (beach) material. Cliff face is ~ 9 m high.

The exact origin of this deposit is not known although it can be assumed that it was related to a higher sea level where high energy, north flowing, sediment laden currents encountered deeper water, lost their competence and deposited their load as Gilbert-type foresets. Deposition was episodic, beginning with flow capable of transporting the large boulders and then waning to a point where only sand could be moved. Fluvioglacial origin should be ruled out as there is no evidence of an ice mass situated off the south shore.

4.4.1.4 Unit 5: Wave reworked (beach) material

The uppermost unit, ~ 2 m thick, extends the length of the section (Fig. 4.8) overlying Units 2, 3 and 4. It consists of gently dipping beds (4 to 6° in a seawards direction) of sand, gravels, cobbles and boulders. Cobbles and gravels are usually clast supported with occasional isolated pockets of sand. The base of the unit is erosive and defined by a continuous layer,at 6-7 m asl,of boulders 0.8-1.0 m across (Fig. 4.10). The boulders appear to be slightly smaller over the foresets (Unit 4.)

The gently dipping internal structure of the seaward dipping beds is characteristic of beach sediments (Thompson, 1937; Clifton _et al_,. 1971) and are similar to deposits at Deadman's Bight (Section 4.1). The underlying till and foresets would have provided an ideal sediment source for the beach. Any material exceding the competency of the wave processes remained in place as finer sands and gravels were winnowed out leaving a concentration of large boulders as a pavement that would eventually provide a protective mechanism for the underlying till. Similar raised boulder pavements have been recorded and described

Figure 4.10: Wave reworked material overlying till. Note concentration of boulders which eventually forms a protective pavement at base of wave reworked material.

in Wales by John (1971, from Bowen, 1973). A good modern analogue for this deposit can be found in the beach immediately in front of the outcrop (Fig. 4.11). The surface has a low seaward angle of dip. There is a base of large boulders which remains stationary in spite of high energy wave conditions (observed on the day of observation) while sands, gravels and cobbles are continuously moved about with each incoming wave swash. Clast supported gravels and isolated patchs of sorted sands are present, which when buried would resemble the gently dipping beds of the raised beach sediments.

4.4.2 Discussion

Pass Island Tickle was situated beyond the proposed limit of the Late Wisconsin glaciation in the Hermitage Bay area (Section 2.8, Figure 2.1). A small cirque north of the outcrop (Fig. 2.1) could not have generated sufficient ice to account for all the till in a valley of this size. Bay d'Espoir Series slate erratics in the till indicate regional ice flow from the north, across Hermitage Bay to the Pass Island area. Local bedrock above the deposits has undergone a considerable amount of weathering and the occurrence of tafoni above nearby Seal Cove may indicate a greater period of subaerial exposure than has occurred since the Late Wisconsin. On these grounds the till is assigned a pre-Late Wisconsin age.

Deposition of the sands (Unit 3) and Gilbert-type foresets (Unit 4) occurred into a higher relative sea level than the present but it is not known when, other than after emplacement of the till. The final event recorded in the section is that of a higher relative

Figure 4.11: Modern beach immediately in front of outcrop. It provides a modern analogue for the depositional environment of beach material and lag in Figure 4.10.

sea level which reworked and planed off all previously deposited consolidated sediments depositing the beach deposits and creating a 19.5 m asl terrace around Beak Bay. The terrace is continuous inland along the road to Seal Cove (Fig. 2.1). It represents the marine limit of Late Wisconsin postglacial sea level rise, when the southern tip of the peninsula may have been an isolated island, perhaps connected by a narrow isthmus near Seal Cove. There is also a 22.5 to 23.5 m asl bedrock bench but it may be of any age.

4.5 Little Barasway, Seal Cove

4.5.1 Description

One of the most important coastal sections in the study area occurs at Little Barasway (11P/8: 744609), east of Seal Cove on Connaigre Bay, where a silty till, underlying wave reworked sediments, was found to contain foraminifera. The till, forming the lower 3 to 4 m of the section, has a silty matrix with a blocky structure and contains angular to subangular pebbles. It is olive grey in colour, although in its lower portions there is a faint pinkish tinge reflecting underlying granite content. Overlying the till is less than 1 m of structureless, sandy (no clay or silt), very stoney material thought to be an ablation or wave reworked facies. The top 2.8 m of the section consists of stratified, seaward dipping, low angle (~ 8°) beds of coarse sand and sandy gravel with occasional boulders up to 0.5 m across. This unit is situated below the Late Wisconsin marine limit (Section 7.3) and is a wave reworked (beach) modification of the underlying till.

A sample of the silty till returned to the laboratory for grain size analysis was found to contain foraminifera. The sample was washed through a 200 mesh wire screen and then dried. Foraminifera were separated from material retained on the sieve by floatation in carbon tetrachloride. Identification of the specimens was carried out by Dr. G. Vilks, Atlantic Geoscience Centre, Dartmouth, Nova Scotia. The foraminiferal species and their numbers in the sample are presented in Table 4.1.

Table 4.1

Elphidium excavatum (clavatum)	2,156
Protelphidium orbiculare	65
Elphidium subarcticum	5
Islandiella islandiea	75
Bucella frigida	19
Astrononion gallowayi	1
Nonionella auriculata	1
Pseudoplymorphina novangliae	4
Virgulina schreibersiana	1
Globigerina bulloides	1

4.5.2 Discussion

This section is especially significant as it provides the only source of fossiliferous material found in the whole of the study area. The foraminifera and shell were incorporated in what is clearly a till, and indicate therefore, a marine source before being redeposited onto the present shoreline in the Seal Cove area. G. Vilks (pers. comm) provided the following paleoenvironmental interpretation for the foraminiferal assemblage.

> "*E. excavatum* is the same species that I have been referring to as *E. clavatum* in my core samples from Beaufort Sea, Labrador and Scotian shelves. I interpret its presence as indicating marginal marine environment with lower salinities ($25\text{-}30\ {}^{o}/_{oo}$). Its dominance in your sample would suggest a nearshore-estuarine environment and possibly seasonal sea ice. The one well-preserved specimen of *G. bulloides* seems to be out of place, because the present day coastal waters off Newfoundland contain *Neogloboquadrina pachyderms* and *Globigerinita uvula* rather than this offshore North Atlantic planktonic species. The presence of *G. bulloides* could indicate a situation where warmer offshore waters are not too distant; a model recently proposed by Ruddiman and McIntyre, Science, Vol. 204, p.173, where during the period of glacial growth warm waters came close to Newfoundland. If this is so, the age of the sediments could be approximately 70,000 years."

Vilks also adds the proviso that " .... the paleo-environmental synthesis is reasonably reliable, the suggested age of the sediments is pure speculation."

The most logical source for these foraminifera and therefore the till is from inner Connaigre Bay, having been incorporated into glacial ice flowing from the north and northeast down into the bay. According to the model of Ruddiman and McIntyre (1979), during periods of rapid glacial growth warm marine waters would have flowed directly adjacent to the ice covered coast of Newfoundland. As *G. bulloides* may have been introduced into the area by warm North Atlantic currents abutting an advancing Newfoundland based ice cap at this time it is quite probable that the foraminifera were almost immediately eroded and then later redeposited by the growing ice cap. The age assignment of 70,000 years by Ruddiman and McIntyre (1979) for the rapid onset of a major glacial event is pertinent to a glacial chronology of southern Newfoundland as it is similar to age proposed independently by Tucker (1979)(Fig. 1.3) for the Fortune Bay event on the Burin Peninsula. The next most likely alternative for glacial erosion of the marine sediment is 30 to 40,000 years later by a major late, mid-Wisconsin advance recognized by Tucker (1979). It could be argued that the glaciation responsible for eroding the marine sediments may have been Late Wisconsin. However, as the outcrop is located beyond the proposed limits of the Late Wisconsin glaciation and there is only one till in the outcrop the age assignment of pre-Late Wisconsin for the deposition of this till is quite tangible.

CHAPTER 5

GLACIOLACUSTRINE/DELTAIC SEDIMENTATION AT CONNE RIVER

5.1 Introduction

At the mouth of the Conne River there is a well exposed bank of rhythmically-bedded fine sands, silts and clays overlain by coarse sands interbedded with clayey silt and capped by poorly sorted gravels. The outcrop is ~ 0.6 km in length and 15.4 m high. It lies between the Conne River and a smaller stream to the south (Fig. 5.1) approximately 3.5 km northeast of the settlement of Conne River. Water laps directly at the foot of the cliff at high tide and this, when combined with storm activity, causes periodic undercutting and slumping providing new, well exposed faces. The Conne River sediments were first described by Jewell (1939) who noted "faintly varved" finely laminated grey clays overlain by "fluvial gravels". The clays were proposed to have been deposited into a lake formed behind the apex of glaciers flowing down the Conne River and Southeast Brook valleys. Widmer (1950) counted more than 1600 "varves" in the clays, suggesting they had been deposited into a freshwater proglacial lake which stood at 19.8 m asl.

Initial inspection of the sediments for this study revealed that the outcrop was more complex than earlier descriptions had indicated and that a detailed investigation would provide a good paleoenvironmental interpretation. In the discussion that follows, the sediments and

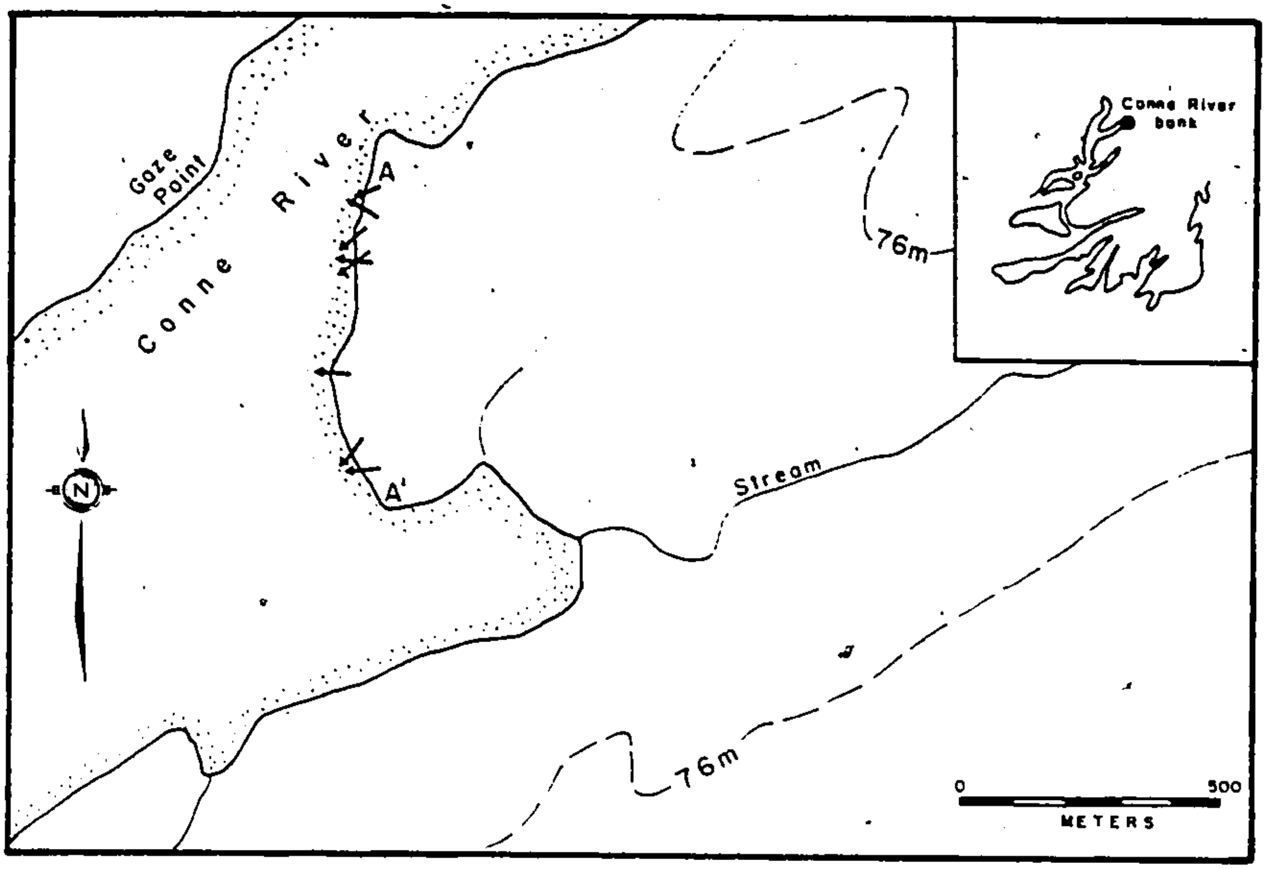


Figure 5.1: Conne River bank. Paleocurrent (arrows) indicate west to southwestwards flow.

sedimentary structures of the Conne River deposits are systematically described, mechanisms responsible for their deposition are proposed and finally, a model taking into account the inferred processes will be developed.

5.2 The Sediments

Sediments in the Conne River bank can be separated into three distinct lithologic units (Fig. 5.2, and 5.3), each characterized by variations in texture and sedimentary structure: a lower unit of thin, parallel bedded, very fine sands, silts and clays, an intermediate unit of ripple cross-laminated and massive sands interbedded with clayey silts, and an upper unit of poorly-sorted, structureless and crude horizontally stratified gravels.

5.2.1 Unit 1: Thinly bedded and laminated fine sands, silts and clays.

Description

The lowermost and thickest unit consists of thin, gently dipping and sub-horizontally stratified very fine sands, silts and clays. Average mean grain size of several random samples is 6.8 Ø. The base of the unit rests on bedrock at one location but otherwise is not visible as it is continuous to below sea level. The upper contact is gradational over a very short vertical distance into the overlying unit and varies in elevation from 3.9 to 11 m asl (Fig. 5.2). An unusual and deceptive feature of the unit is the marked color difference between the lower (dark bluish grey, Munsell color 10BG 3/1) and upper (greyish olive, Munsell color 5Y 5/3) portion. In places the greyish olive sediments extend downwards into the bluish grey sediments as what appear to be

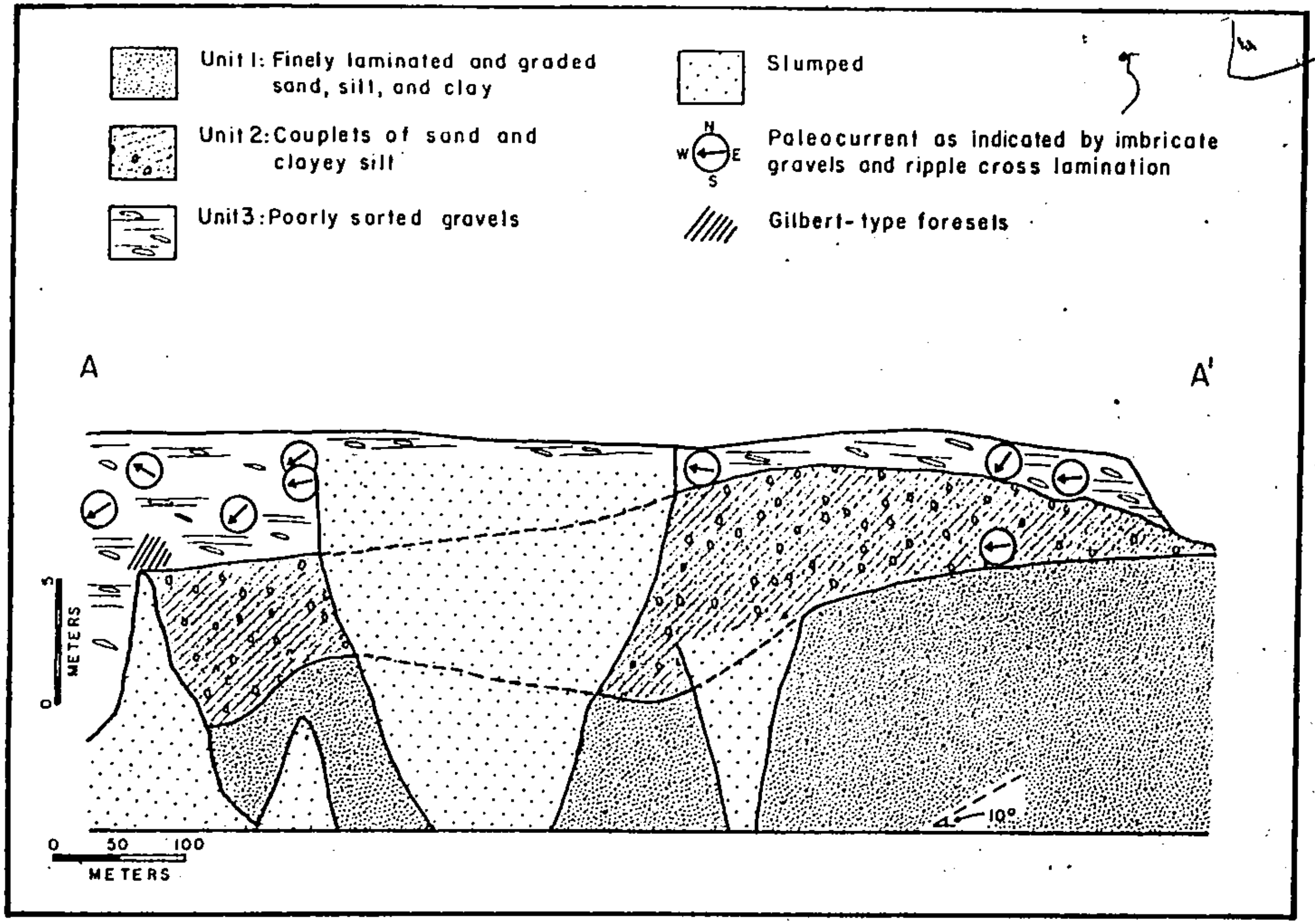
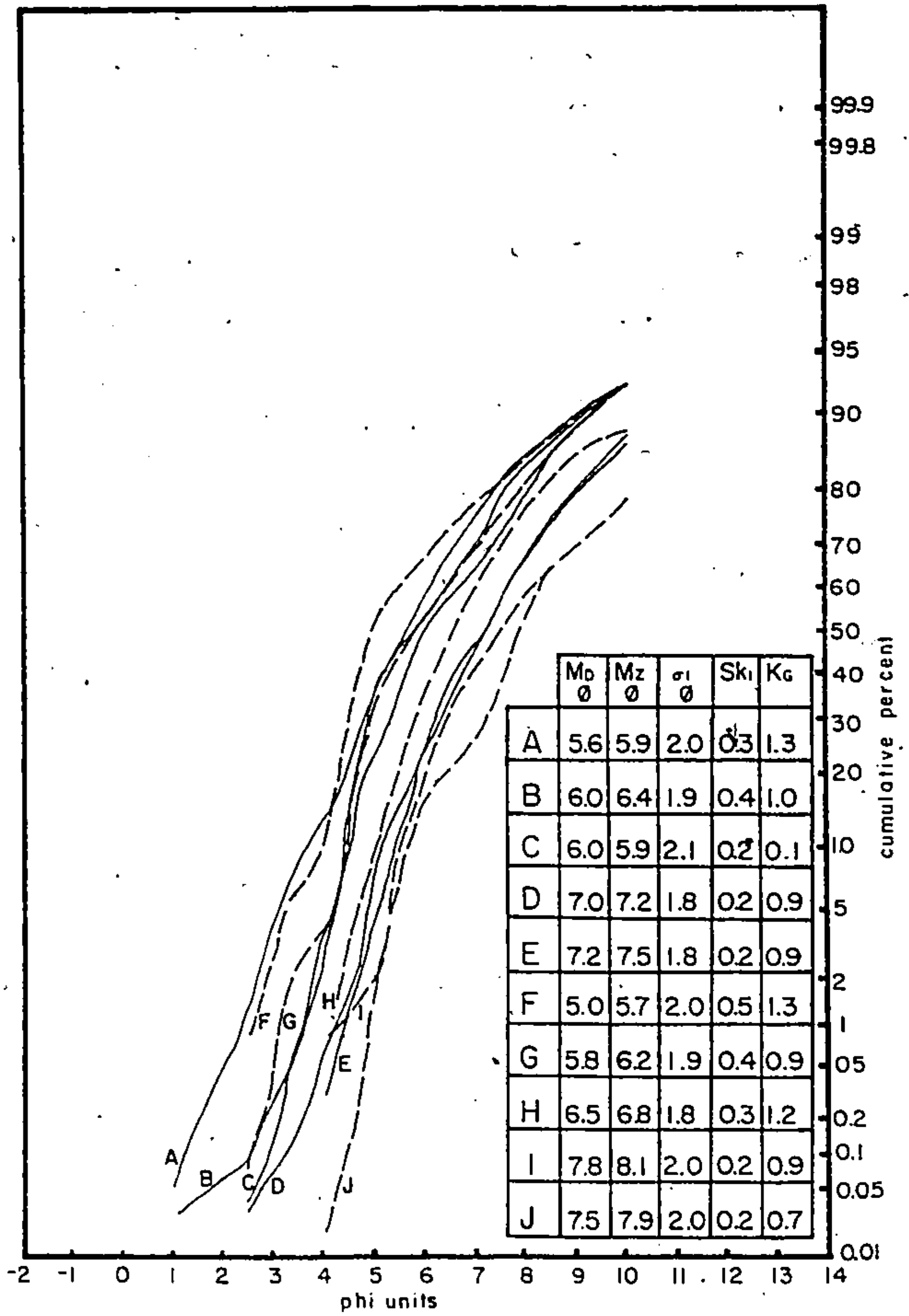


Figure 5.2: Cross section of the Conne River bank. Location of AA' is shown in Figure 5.1. Mean tide level used as datum. Dashed lines represent inferred contacts hidden by slumping. Dash line in lower right corner shows location of slump scar in Unit 1.

Figure 5.3: Sediments in the Conne River bank. Lower two thirds of section are glaciolacustrine bottom sediments which are overlain by "varved", rippled sediments of the low slope prograding delta. Arcuate shape at contact of the two units is not a channel. Glaciofluvial sands and gravels cap the section. The unconformity to the left of the tree is a lake bottom slump scar.

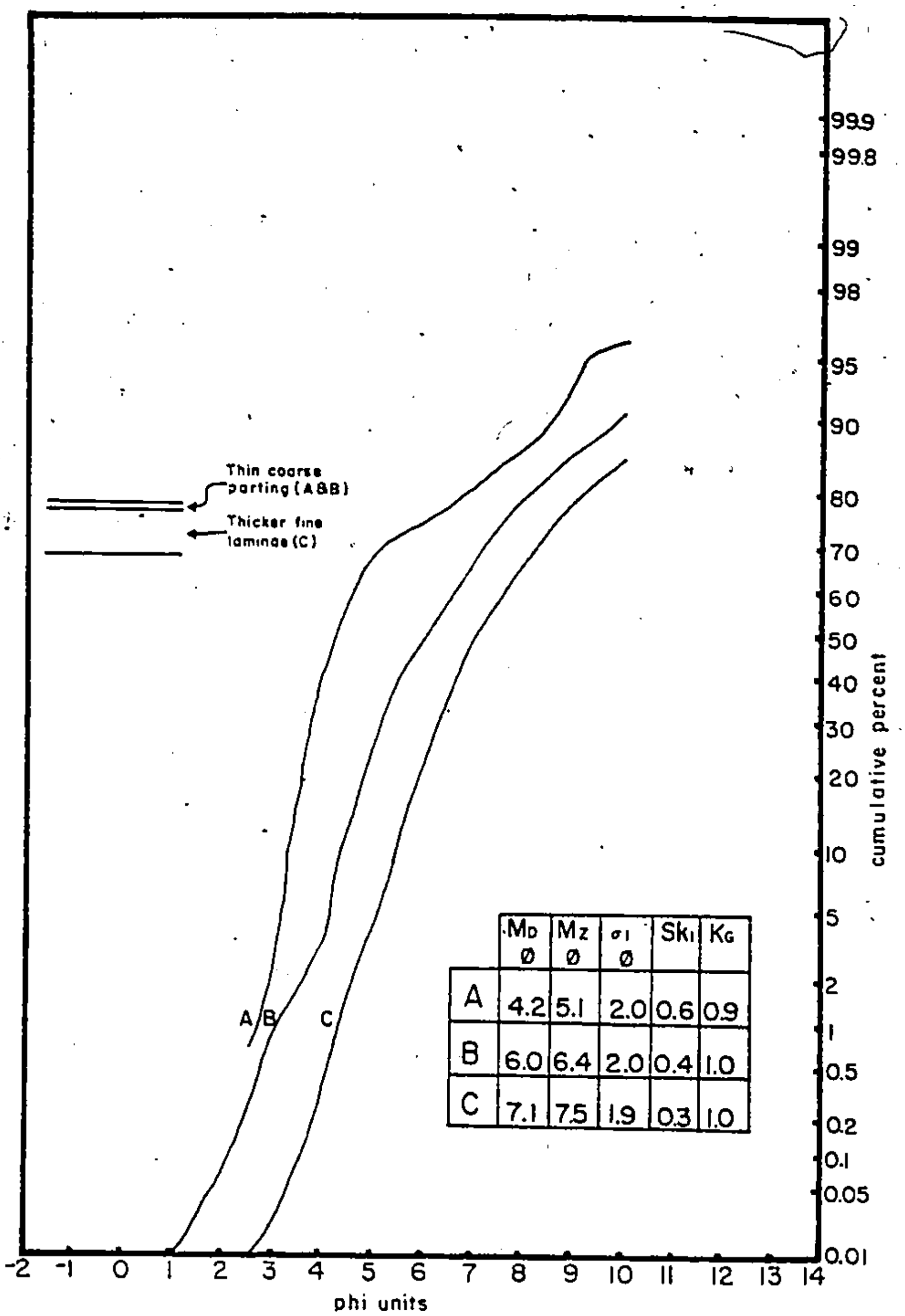
channels entrenched into the lower material. Close examination, however, shows that individual continuous laminae can be seen in every instance crossing the apparently erosive contact between the two different colored materials. Textural analyses of five samples from each color zone reveal no significant difference (Fig. 5.4) in grain size. Average mean phi size for the dark grey sediments is 6.9 Ø and that of the dark olive is 6.6 Ø. The color change is probably due to oxidation of the upper sediments or alternatively, is diagenetic in origin, or related to groundwater seepage from the overlying coarser sandy unit.

Within the unit as a whole (both colors) there are two primary bedding types: finely laminated and graded bedding. The former is more common but both occur arbitrarily throughout the unit. The finely bedded material consists of dark clayey silt strata separated by thin (0.5 to 2 mm) partings of slightly coarser silts and very fine sands. The clayey silt beds vary in thicknesses from a fraction of a millimeter to more than 12 cm. Individual beds, defined by the coarser partings, may contain several micro-laminae. One 12.3 cm thick bed was comprised of more than 70 microlaminae, each showing slight variations in tone and/or texture (unmeasured). Alternatively, there are also relatively thick (e.g. 2 cm) beds having a massive structure in which no microlaminae can be distinguished. Textural analysis of a single laminated couplet (Fig. 5.5) gave a mean grain size of 7.5 Ø for the clayey-silt layer and 5.1 and 6.4 Ø for the two bounding silt partings. Microscopically, some of the thicker sand-silt partings can be seen to have a sharp contact. The upper contact grades over a relatively short vertical distance into the finer overlying clayey silts.



	Mᴅ Ø	Mz Ø	σI Ø	SkI	KG
A	5.6	5.9	2.0	0.3	1.3
B	6.0	6.4	1.9	0.4	1.0
C	6.0	5.9	2.1	0.2	0.1
D	7.0	7.2	1.8	0.2	0.9
E	7.2	7.5	1.8	0.2	0.9
F	5.0	5.7	2.0	0.5	1.3
G	5.8	6.2	1.9	0.4	0.9
H	6.5	6.8	1.8	0.3	1.2
I	7.8	8.1	2.0	0.2	0.9
J	7.5	7.9	2.0	0.2	0.7

Figure 5.4: Textural analysis of glaciolacustrine bottom sediments. Average mean phi size for the dark olive sediments (A to E, solid line) is very similar to that for the dark grey sediments (F to J, dashed lines).
Folk (1974) statistical parameters are shown in box: median in 0 units (M_D), mean in 0 units (M_Z), sorting in 0 units (σ_I), skewness (SK_I) and kurtosis (K_G).



	M_D ø	M_Z ø	σ_I ø	Sk_I	K_G
A	4.2	5.1	2.0	0.6	0.9
B	6.0	6.4	2.0	0.4	1.0
C	7.1	7.5	1.9	0.3	1.0

Figure 5.5: Textural analysis of a single laminated couplet: the thicker silt layer (C) and two bounding silt partings (A and B). Folk (1974) statistical parameters are shown in box.

Micro-faulting (Fig. 5.6) is common in the laminated sediment, usually as normal faults and fault stepping with occasional micro-graben. Some of the laminae terminate abruptly as if broken or pulled apart (Fig. 5.7) and are replaced by thoroughly mixed and disturbed sediment. Minute fragments of the original laminae can be distinguished in the otherwise now homogeneous mixture.

The second major type of primary structure consists of a series of well graded, sharp based beds varying in thickness from 0.5 to 2.0 cm (Fig. 5.8). Textural analyses of the upper and lower parts of a single graded bed (Fig. 5.9) show that the latter contained coarser grained sediment for all percentiles than the former. The mean grain size difference between the upper and lower portions is one phi unit. Occasionally the coarse portion of a graded bed will be interupted by a 1 to 2 mm thick laminae of fine silt and clay followed by coarser material, again grading up into the fine.

Small scale load structures, deformation features and erosional surfaces abound in the graded bedding (e.g., Fig. 5.8) although none were observed in the finely laminated sediments. Undisturbed oriented monoliths from Unit 1 were returned to the laboratory for detailed inspection and micro-photography. A sample of graded bedding was dried and pried apart along its bedding planes. At the base of one plane was a series of several parallel, coarse grained ridges, 1 to 2 mm deep, 1 to 2 mm wide and more than 7 cm long protruding into the underlying clayey silts. The grooves were oriented northwest-southeast. The lower portion of another one graded bed contained several small pebbles up to 6 mm across. This was the only noted occurrence of material larger than sand size in this unit. The pebbles were not ice rafted as there was no indication

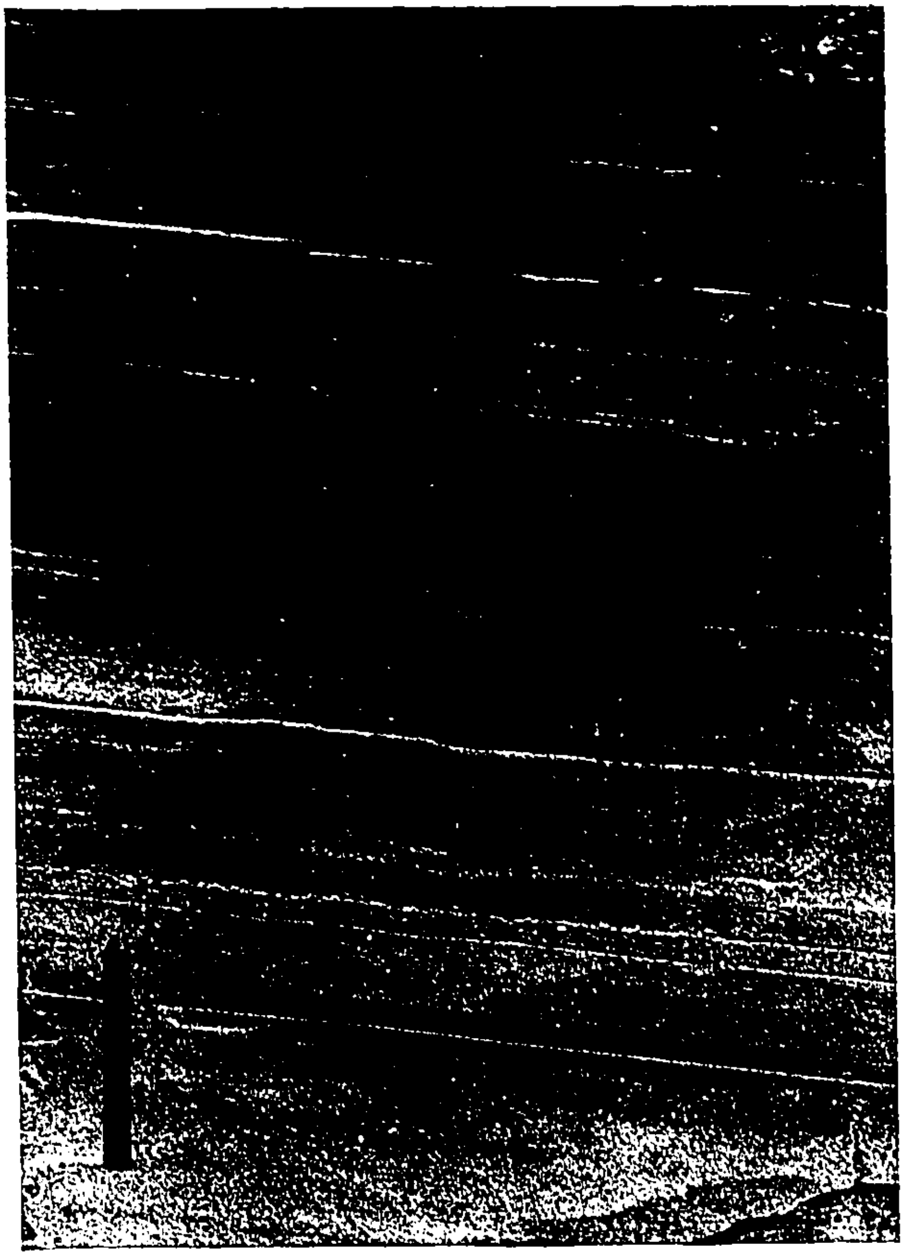


Figure 5.6: Blowup of finely laminated lake bottom of Unit 1 showing microfaulting. Bar represents 1 cm.

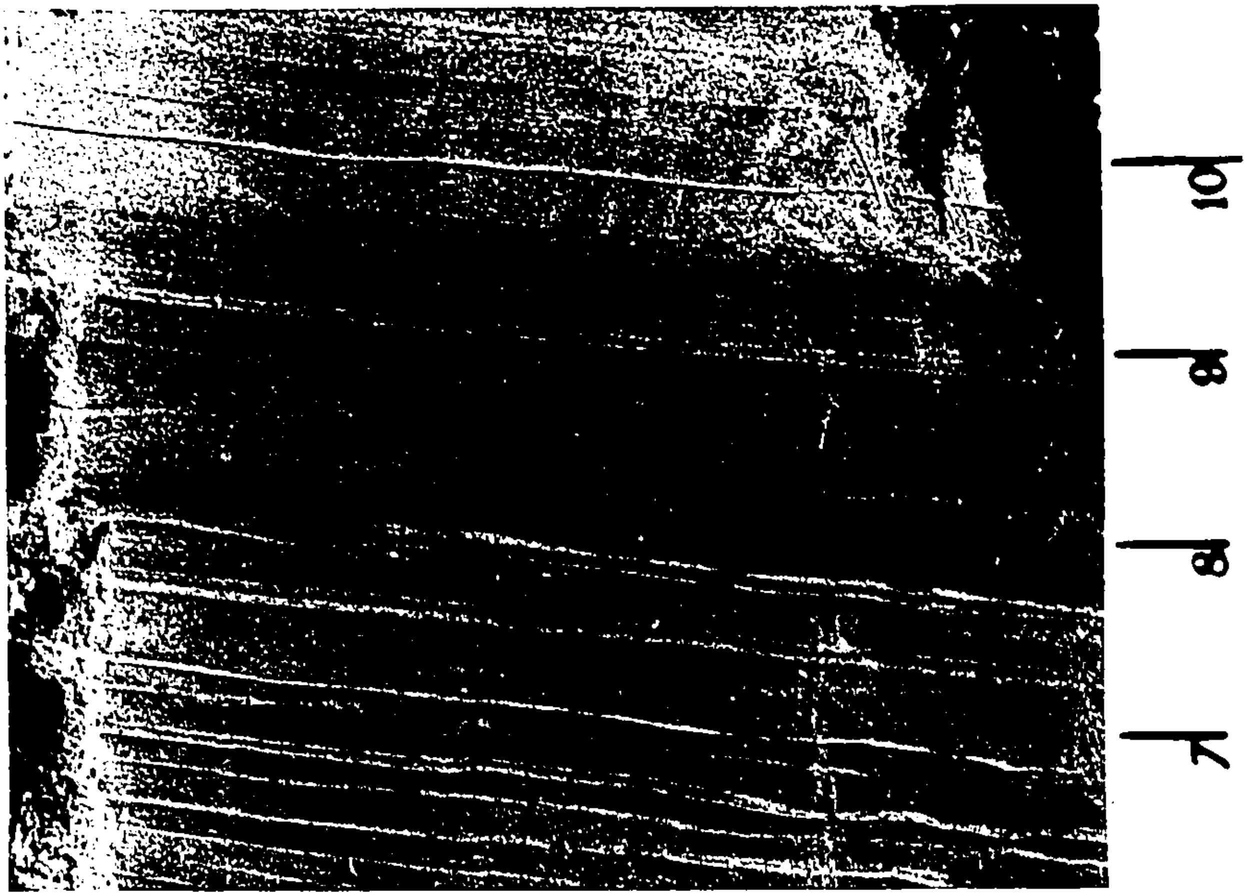


Figure 5.7: Blowup of finely laminated lake bottom sediment of Unit 1. In the 1 cm thick massive bed in middle of photo the laminae suddenly terminate. Bed(s) probably underwent tension and or liquefaction. Fragments of the original bedding are still preserved in the massive mixture. Scale in centimeters

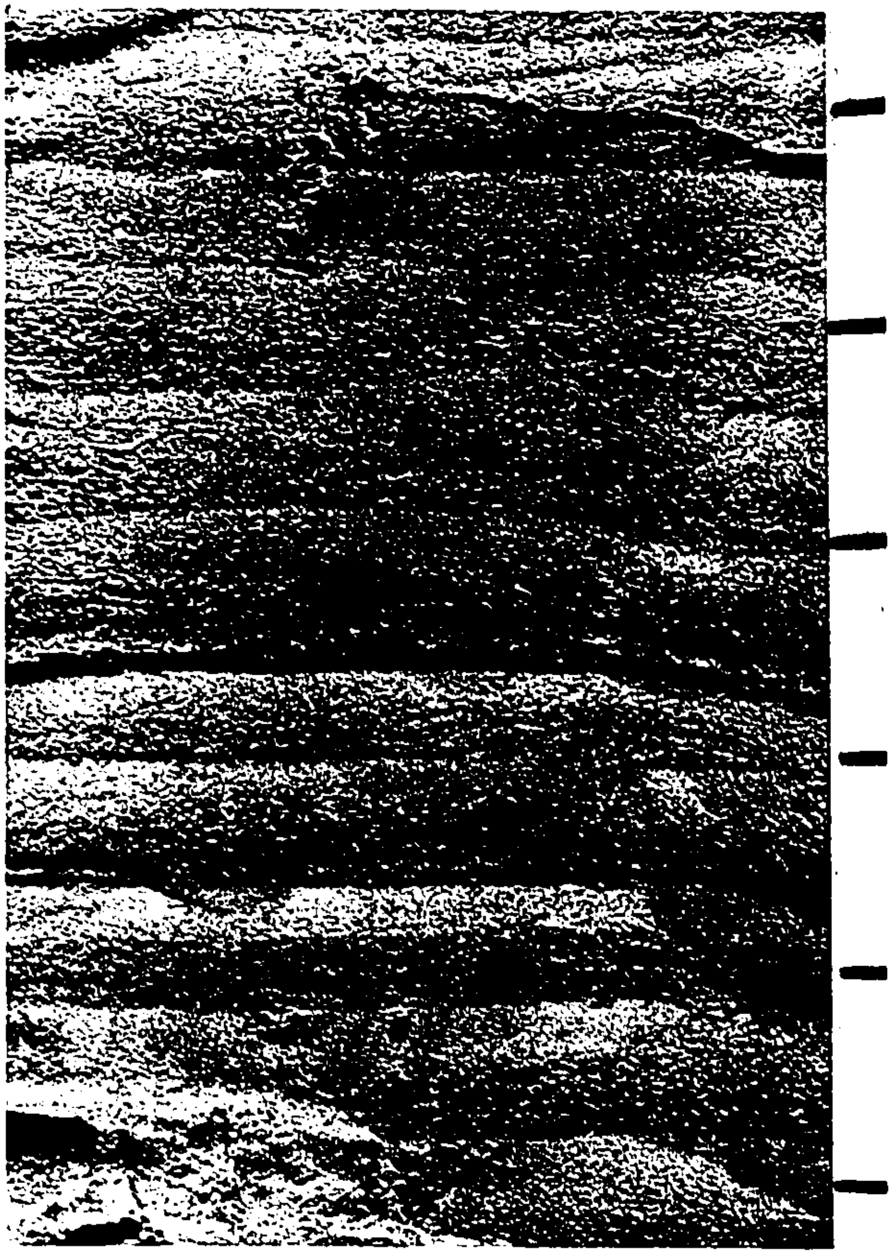
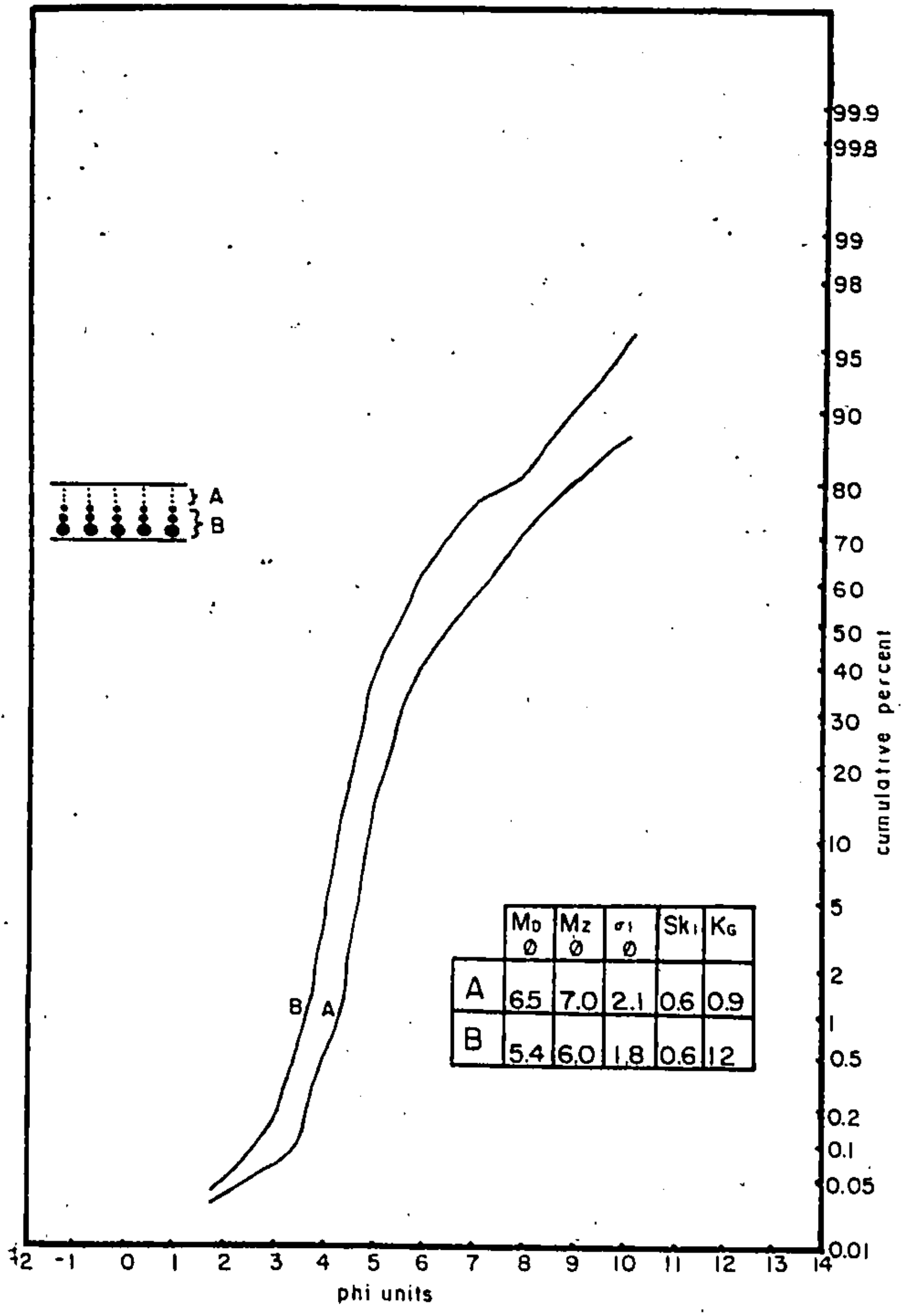


Figure 5.8: Blowup of sharp based, graded beds. Note micro-erosional structures and silt rip-ups. Difference in grain size between upper and lower parts a single graded bed shown in Figure 5.9. Scale in centimeters.



	M_D ϕ	M_z ϕ	σ_I ϕ	Sk_I	K_G
A	65	7.0	2.1	0.6	0.9
B	5.4	6.0	1.8	0.6	12

Figure 5.9: Textural analysis of the upper (A) and lower (B) parts of a single graded bed. The lower portion contains coarser grained sediment for all percentiles than occurred in the upper part. Folk (1974) statistical parameters are shown in box.

of disturbance or warping by the pebbles of overlying and underlying beds.

There are several instances of continuous 4 cm to ~ 2 m thick deformed bedding overlain and underlain by undisturbed parallel strata (Figs. 5.10 and 5.11). In a few instances the tops of convolutions have been truncated by overlying horizontal beds. One particular ~ 20 cm thick convoluted zone, dipping along strike to the north, could be followed for more than 150 m. The folds vary from open anticlinal and synclinal forms to very tight extremely involuted features. Faulting and minor thrusting is common in extremely contorted structures. The frequency and magnitude of disturbance is greatest near the central portion of the outcrop where Unit 1 is also thinnest. With the exception of very small zones of convolute bedding there is very little secondary deformation at the southern end of the exposure.

Large scale erosional surfaces, although not abundant, do occur. At one location a broad shallow channel cutting down through several centimeters of laminae could be seen.

Interpretation

The occurrence of two distinct types of bedding (finely laminated and graded) indicates that two independent processes contributed to the depostion of Unit 1. Recent process studies of glaciolacustrine sedimentation have demonstrated the existence (Gustavson, 1975a,b; Gilbert, 1975) and predominance (Smith, 1978) of overflow and interflow currents. Although the density of the inflowing water may be insufficient to create an underflow current there may be sufficient sediment in suspension so that visible laminations will be constructed (Smith, 1978). Harms (1974) theorized that coarse to fine silt introduced into a density stratified

Figure 5.10: Deformed (convolute) bedding overlying and underlying undisturbed parallel strata of Unit 1.

Figure 5.11: Deformed (convolute) bedding overlying and underlying undisturbed parallel strata in Unit 1.

basin as an interflow current would settle out of suspension creating thin parallel strata, less than 1 to 2 mm thick, of alternating light colored coarse silt laminae and dark, finer sediments (in Harms' model, density stratification was the result of differences in salinity and temperature between inflowing and basin water; the lake studied by Smith was thermally stratified). Theakstone (1976) and Shaw (1977) interpreted very finely laminated silts with a base of "minor sand laminae" as having been deposited out of suspension in glacial lakes. Thus, by analogy with modern and ancient studies, the finely laminated sediments of Unit 1 probably settled out of suspension having been introduced into the lake as interflow and overflow currents. The consistent lack of flow structures or erosional features strongly confirms that current action did not play a significant role. Some of the thicker, slightly coarser sand partings (up to 2 mm) however, may have been transported by weak, sluggish underflow currents which stagnated in the immediate area, producing the overlying massive non-laminated clayey silt beds. Although the micro-laminae undoubtedly represent periodic variations in sediment supply or rate of sedimentation, the time cycle involved (diurnal to annual) cannot be distinguished.

The second major bedding type has several features typical of deposition by current action. Repetitive graded bedding is generally accepted as being the result of turbidity currents (Dott and Howard, 1962; Dott, 1963; Harrison, 1975; Lambert et al, 1976) formed as progressively finer sediment was deposited by a decelerating flow. The sharp contact at the base of the graded beds is a feature common to turbidites (Walker, 1967). Channelling and truncation of the tops of contorted bedding requires that currents have flowed on the bed. The parallel grooves found at the base of a graded bed are sole marks formed by turbidity currents. They indicate that the local paleoflow direction of the current was oriented northwest-southeast. The small pebbles found within the coarse unit of a graded bed may have been transported as bedload within the turbidity current. They represent the type of tools responsible for the sole marks. Although the pebbles may have been ice rafted, this is doubtful because of their relatively uniform small size, being scattered solely along a single horizon and not visibly having disturbed any overlying or underlying strata.

A turbidity current origin is proposed to account for the graded bedding and related features. The idea is not new to glaciolacustrine sedimentation, having been suggested as a (partial) mechanism of "varve" formation by several authors in the past three decades (e.g., Kuenan, 1951; Smith, 1959; Jopling and Walker, 1968; Banerjee, 1973; Ashly, 1975; Gustavson, 1975b; Harrison, 1975b; Shaw, 1977). The best documented mechanism for generating lacustrine turbidity currents is one which sediment laden incoming water, having a higher density than ambient lake water, sinks and flows as a density underflow current capable of erosion and deposition (Houbolt and Jonker, 1968; Gilbert, 1975; Gustavson, 1975a,b; Lambert _et al_, 1976). The result is a series of graded beds formed by a pulsating turbidity current (representing diurnal, subseasonal, seasonal and/or possibly annual cycles). Another generating mechanism is subaqueous slumping of the prodeltaic slope caused by seismic activity (calving of ice bergs) or overloading and deepening of unconsolidated material. As slumping occurs, unconsolidated sediment often mixes with water to become a slurry which in turn develops into a turbidity current (Morgenstern, 1967; Middleton, 1970). Each slump event would cause a single, normally graded bed although a series of regressive slumps might lead to a short series of beds. Slump generated graded beds will probably be indistinguishable from those deposited from density underflows. When slumping does not create turbidity currents the displaced sediment often accumulates at the foot of the delta slope as slump mounds (Mathews, 1956; Fulton and Pullen, 1969; Gilbert, 1972, 1975; Gustavson, 1975b; Smith, 1978). Slumping may also account for some of the deformation features with minimal displacement.

The coarse portion of some of the graded beds are occasionally interupted by a thin laminae (~ 1 mm) of clayey silt. This was probably due to a temporary cessation of the density underflow which permitted fines to settle out of suspension. Gilbert (1975) has shown that underflow currents need not be continuous throughout the whole melt season. A drop in temperature can temporarily reduce ice melt and thus also water and sediment discharge, which will affect the density of meltwater entering the lake as an underflow current.

The convoluted bedding (Fig. 5.10 and 5.11) is most likely of subaqueous slump origin although other processes cannot be completely ruled out (i.e., plastic deformation at the time of deposition, or shearing of the sediment as a result of an overriding turbidity current). Slumping can occur on lake bottom slopes as low as 1° (Morgenstern, 1967), and Shaw (1977) has presented an argument for slumps occuring on very low slope angles as the result of compaction of lake bottom sediment. Sediment shifting a short distance over a bed of detachment will maintain its plasticity, contorting and deforming the displaced bedding. Faulting occurred when the plasticity of the sediment was exceeded. After the disturbance, deposition continues as it did before. Passage of a turbidity current would plane off any irregularities.

The diagonal break in the sediments shown in Figures 5.2 and 5.3 is probably a slump scar of a disturbance which occurred in the floor of the lake. It must have happened while the lake floor was being sedimented as only the lower portion of Unit 1 has been disrupted. The upper strata of the unit are continuous, gently curving over the top of the scar (Figs. 5.2 and 5.3) The micro-faulting of the finely laminated material (Fig. 5.6) occurred as the result of consolidation and dewatering. Consolidation may also have been responsible for the suddenly terminated laminae in Figure 5.7. As compaction and consolidation progressed water would have been released from the finer grained laminae increasing pore water pressure in the slightly coarser material and decreasing its shear strength to the point where liquifaction occurred, creating a homogeneous medium with occasional preserved fragments of the original bedding.

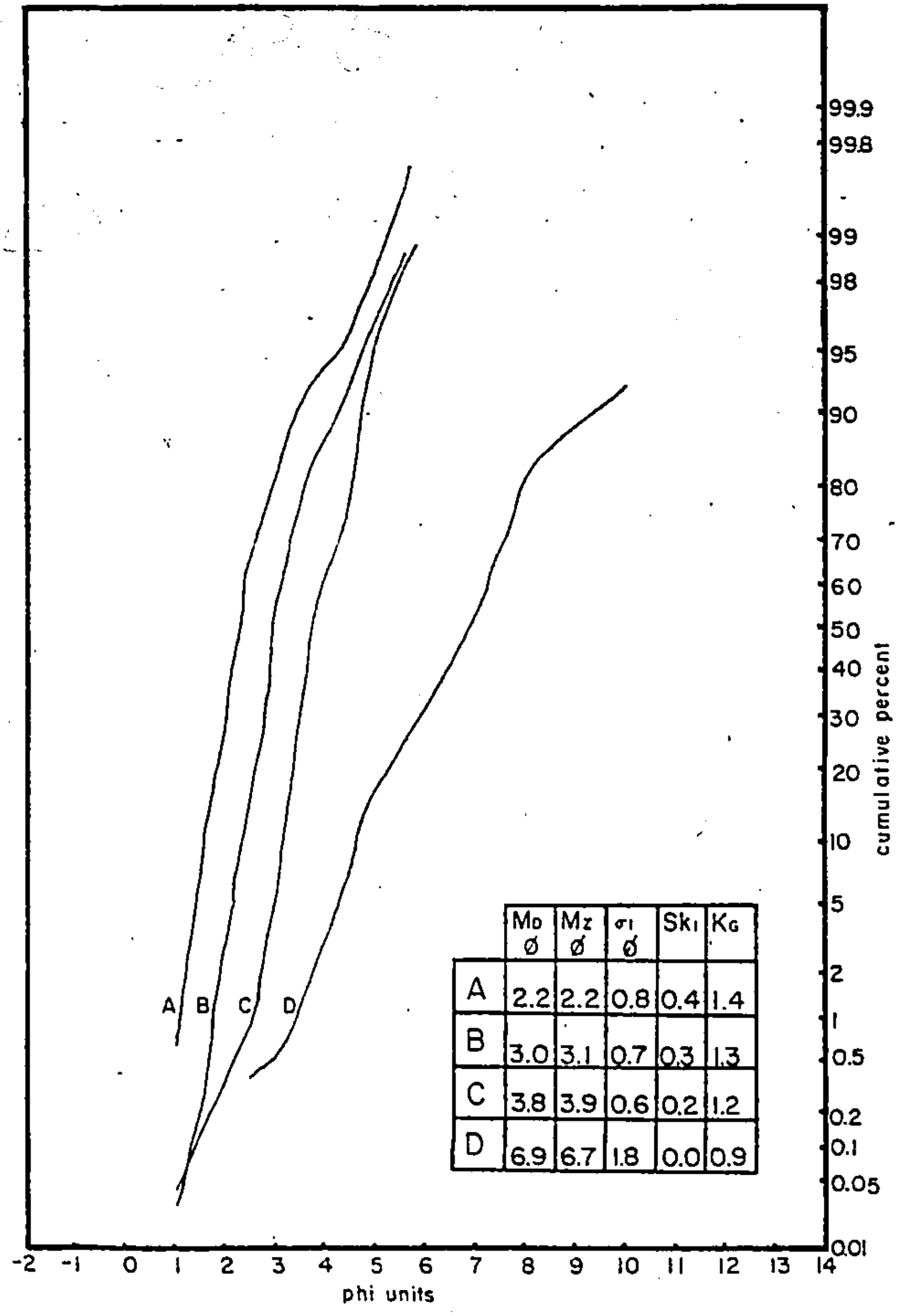
In summary, the finely laminated beds resulted from deposition from suspension of material introduced into the lake as interflow and overflow currents, whereas the repeated graded bedding was deposited by turbidity currents. The lack of cross-bedding, commonly found in sediments of silty lake bottom beds (Gustavson, 1975b), suggests that all the deposition in the turbidity currents was from suspension, there being no, or little, bedload transport. Turbidity currents were generated by density underflows, slumping of the deltaic slope and/or slumping of the lake bed. An annual period of deposition has been shown for finely laminated couplets settling out of suspension from inter- and overflows (Smith, 1978) and from density underflow generated graded bedding (Gilbert, 1975; Gustavson et al, 1975; Gustavson, 1975b) but in Unit 1 at the Conne River bank there is no good evidence of an annual periodicity. A distinct clay drape would be expected (Agterberg and Banerjee, 1969; Ashley, 1975) but was not observed, nor was an over-abundance of clay noted.

The uniform dark grey color of Unit 1 indicates deposition of the sediments occurred under anaerobic reducing conditions suggesting stagnant bottom water having a low oxygen content (Theakstone, 1976).

5.2.2 Unit 2: Rythmically bedded ripple cross-laminated sands and clayey silts

Description

Unit 2 consists of rythmically bedded medium to fine grained sands and clayey silts overlying Unit 1. It varies in thickness from 1 to 8 m (Fig. 5.2). The lower contact is gradational into Unit 1, the change being defined by an increased thickening and coarsening of sand layers between alternating clayey silts. The upper contact is erosive having been scoured by the overlying unit. The sands are usually small scale, cross-laminated (cross bedding at one location indicated flow towards 255°; Fig. 5.1). Sand in any one bed is moderately well to well sorted and mean grain size varies slightly from bed to bed (Fig. 5.12). Thicknesses of sand beds range from 1 cm to greater than 15 cm. Occasionally there will be a relatively thick, massive structureless bed of sand with a wavy, loaded lower contact and a planar upper contact, markedly different in appearance from regular sand beds. In one instance a massive sand bed overlay a regular sand bed without the intervening



	Mᴅ Ø	Mz Ø	σ₁ Ø	Sk₁	Kɢ
A	2.2	2.2	0.8	0.4	1.4
B	3.0	3.1	0.7	0.3	1.3
C	3.8	3.9	0.6	0.2	1.2
D	6.9	6.7	1.8	0.0	0.9

Figure 5.12: Textural analysis of 3 rippled summer sand beds (A,B,C) and one overlying winter fine layer (D). Folk (1975) statistical parameters shown in box.

clayey silt layer (possibly erosively removed).

Alternating with the sands are beds of poorly sorted clayey silts (Fig. 5.12) having relatively uniform thicknesses (1 to 2 cm) throughout. The contacts between the sand and overlying silt were either sharp or transitional, the latter especially in the lower portion of the unit. The lower contact of the sand is always sharp. Couplets of sand and clayey silt near the base of the unit have an overall higher proportion of silt to sand than those higher up towards the middle or at the top. There were several instances where sequences of couplets would show thinning and fining upwards tendencies over a vertical distance of 0.3 to 0.5 m (i.e., within a given sequence the thickness of the sand portion of a couplet decreases upwards although the silt-clay remained relatively constant).

Load, flow and flame structures are common in several sets of beds. Normal faults also occur, although infrequently.

Interpretation

The alternation of sand and clayey silt beds in this unit are typical of deposition in proximal glaciolacustrine environments (Gustavson et al, 1975; Gustavson, 1975). The cross-laminated sands were transported primarily as bedload by density underflow currents on gently sloped, deltaic foresets during the summer melt season whereas the overlying clayey silts settled out of suspension in quiet water during the following winter. Although it was not possible to measure the foreset slope angle it is envisaged as having a dip of ~ 5°, based on similarities with a kame delta measured by Jopling and Walker (1968). Any given couplet of sand and clay-silt is a varve representing a year's deposition. The lower couplets having a high silt to sand ratio represent deposition on the distal prodelta slope immediately overlying the lake bottom sediments; couplets higher in the unit having a much lower silt to sand ratio are a more proximal prograded facies. The thinning and fining upwards of some sequences of couplets is due to the thinning of the summer sand layer. This may be due to channel abandonment as distributaries on the delta top gradually shifted in position from year to year. Alternatively, increasing distance from the ice front may be responsible but this is highly improbable as the sequences are recurrent and appear to have occurred over short periods of time.

The beds of structureless sands, having a planar upper and loaded lower contact, are similar to features described by Shaw (1977) as grain flows. Grain flows can occur on delta fronts as the result of drawdown of the lake late in the melt season. They occur on slopes as low as 3 to 6° and may be accelerated to a point where transformation into a low concentration turbidity current occurs.

The deformation features (flame, flow and load structures) occur as the result of relatively heavy, low porosity sands being deposited onto a high porosity somewhat plastic clayey silt layer which is very malleable and easily deformed. The features are characteristic of several environments and not diagnostic by themselves (Potter and Pettijohn, 1963).

5.2.3 Unit 3: Poorly-sorted sand and gravel

Description

The uppermost unit consists of poorly-sorted, structureless and crude horizontally stratified gravels with occasional pockets and pinch-outs of better sorted pebbles and sands. Individual clasts are generally less than 8 cm long although the largest clast observed had a long axis of 20 cm. The base of the unit is sharp and erosive, with occasional small channels ~ 1 m across and less than 0.5 m deep incised into the underlying unit. Its thickness varies; at one location it changed from 1.8 to 3.6 m thick over a horizontal distance of 9 m.

The gravels are moderately imbricated. Results of pebble fabric analysis (resultant vector mean) at 8 locations in the unit are shown in Figures 5.1 and 5.2. Flows responsible for the gravel transport and deposition, as indicated by the vector mean of pebble dip, ranged from between 275° to 230°. At the north end of the exposure there are recurring southwest oriented paleoflow directions. The gravels are also thickest at the north end of the section and at one location (Fig. 5.1) there is 1.5 m of poorly exposed angle of repose gravel foresets dipping to the south-southwest (the foresets are probably more extensive but the face is badly slumped.

Interpretation

These sands and gravels are glaciofluvial in origin (i.e., braided outwash). The imbricate clasts, crude horizontal stratification and general lack of cross bedding suggest deposition at a location fairly proximal to the source (Boothroyd and Ashley, 1975, Fig. 25). This material formed the delta surface over which sediment and meltwater were transported to the prodelta slope from the glacier. It appears, based on paleocurrent data (Figs. 5.1 and 5.2), that two meltwater systems may have been operative, occupying the Conne River valley and the valley of the smaller stream to the south. The southwest oriented paleoflow directions, especially at the north end of the section, the southwest dipping Gilbert-type gravelly foresets and the much thicker gravels of what may have been a major channel distributary represent flow coming out of the Conne River valley.

5.3 Summary and Discussion

On the basis of sedimentary structures and textures the sequence of deposits exposed at the Conne River bank is interpreted as a low slope (~ 5°), prograding glaciofluvial delta which advanced over glaciolacustrine bottom sediments. The mean grain size of the sediments increases upwards through the three units from silts and clays to alternating sands and silts to sands and gravels. A similar coarsening trend of sediments has been observed in samples from a modern lacustrine environment where a fluvial supplied delta is prograding out onto the lake floor (Gilbert, 1972, Figure 6).

Unit 1 was deposited on the glacial lake floor by two different processes. The finely laminated strata settled out of suspension from sediments introduced into the lake as interflow or underflow currents and the series of multiple graded beds are the result of turbidity current deposition. The best understood mechanism for generating lacustrine turbidity currents is density underflows where density differences between incoming meltwater having a relatively high silt/clay content and ambient lake water cause the inflow to plunge to the lake bottom as a continuous turbidity current (Gustavson, 1975a,b; Gilbert, 1972; Lambert _et al_, 1976). Significance of the density underflow varies from year to year as they may occur for only a few hours one or two times during the meltseason one year, whereas the next year they may occur for at least half the season

(Church and Gilbert, 1975). Contorted bedding (Unit 1), faulting (Unit 1) and grain flows (Unit 2) indicate that slumping occurred on the lake floor and delta front and may have been a second mechanism for generating the turbidity currents. Deposition on the lake floor apparently alternated between episodes of turbidity current generation and sedimentation by suspension from above. Deposition from suspension of the interflow and overflow introduced material occurred throughout the meltseason but was only preserved when turbidity currents were not active. It is not possible to assign any periodicity to the bedding in Unit 1. Agterberg and Banerjee (1969) and Banerjee (1973a, Fig. 2) suggested that the upper portion of a single turbidite, or the top of a sequence of turbidities will be capped by a winter clay layer. This could not be discerned in the graded bedding, nor anything similar in the laminated material.

Unit 2, made of alternating cross-laminated sands and clayey silts, was deposited as a low slope, prograding delta. Episodic slumping caused massive grain flows which occasionally developed into turbidity currents. The repetitive couplets of sands and clayey silts represent the summer melt and winter freezeup, respectively. Why seasonal effects can be seen in Unit 2 but not in Unit 1 cannot be explained at this point. The upper unit represents the glaciofluvial, or topset, portion of the delta. The sands and gravels were probably deposited in a fairly proximal location relative to the ice front. Paleocurrent data and other evidence suggests the influence and convergence of two outwash systems, that of the Conne River valley and, of less importance, the smaller stream to the south.

A model demonstrating the proposed depositional environments for the sediments in the Conne River bank is shown in Figure 5.13. Figure 5.14 is an enlargement of the inset in Figure 5.13 showing the processes responsible.

Creation of the lake responsible for these deposits was discussed in detail in Chapter 4. The surface of the lake, as indicated by the maximum height of the topset-foreset bed contact (Unit 3 - Unit 2), had a minimum elevation of ~ 15 m asl (a minimum elevation is given as it is not known how deep the glaciofluvial channels were). Widmer (1950) suggested an elevation of 19.8 m based on observations at the head of Bay d'Espoir which is similar to data presented elsewhere (Section 7.4) regarding the maximum height of the lake surface. Widmer counted more than 1600 varves in the Conne River sediments. On the basis of the observations and conclusions made above it is suggested that Widmer's varve count was in error, possibly due to misinterpretation of couplets in Unit 1 that represent a period of deposition less than annual.

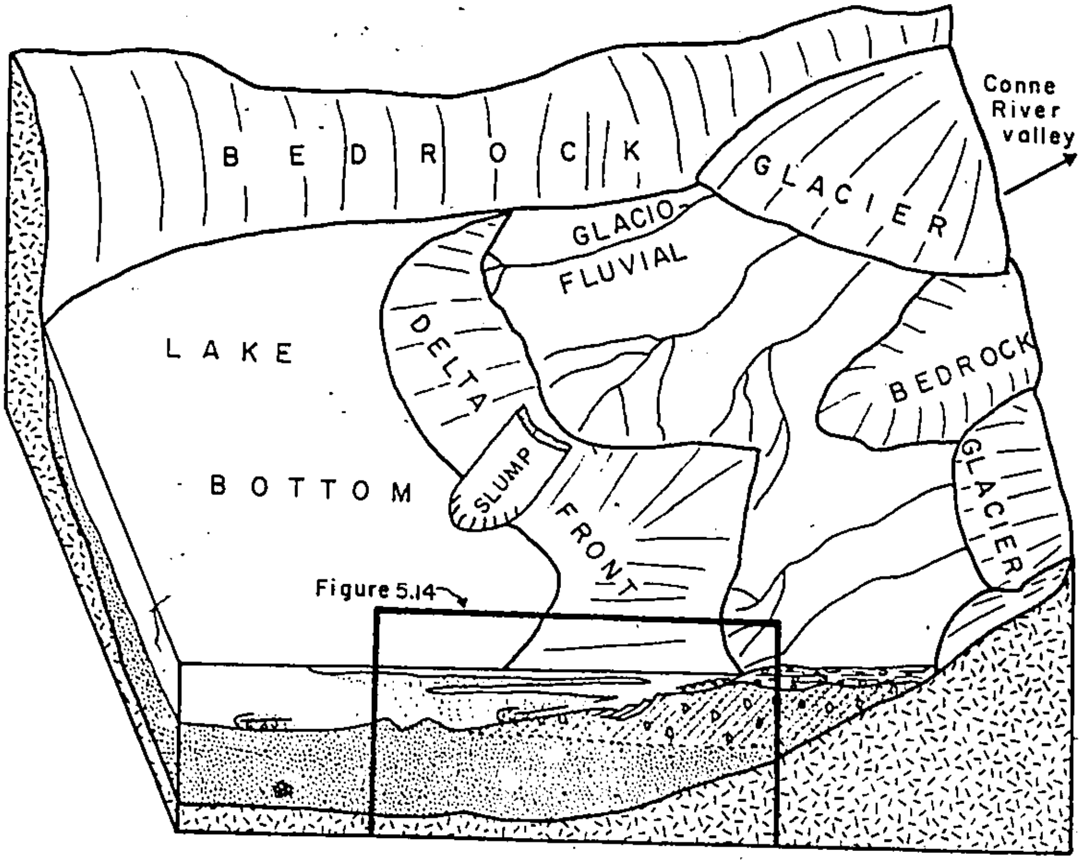


Figure 5.13: Model demonstrating the proposed environment of deposition for sediments of the Conne River bank. Box is enlarged in Figure 5.14 to show the processes also.

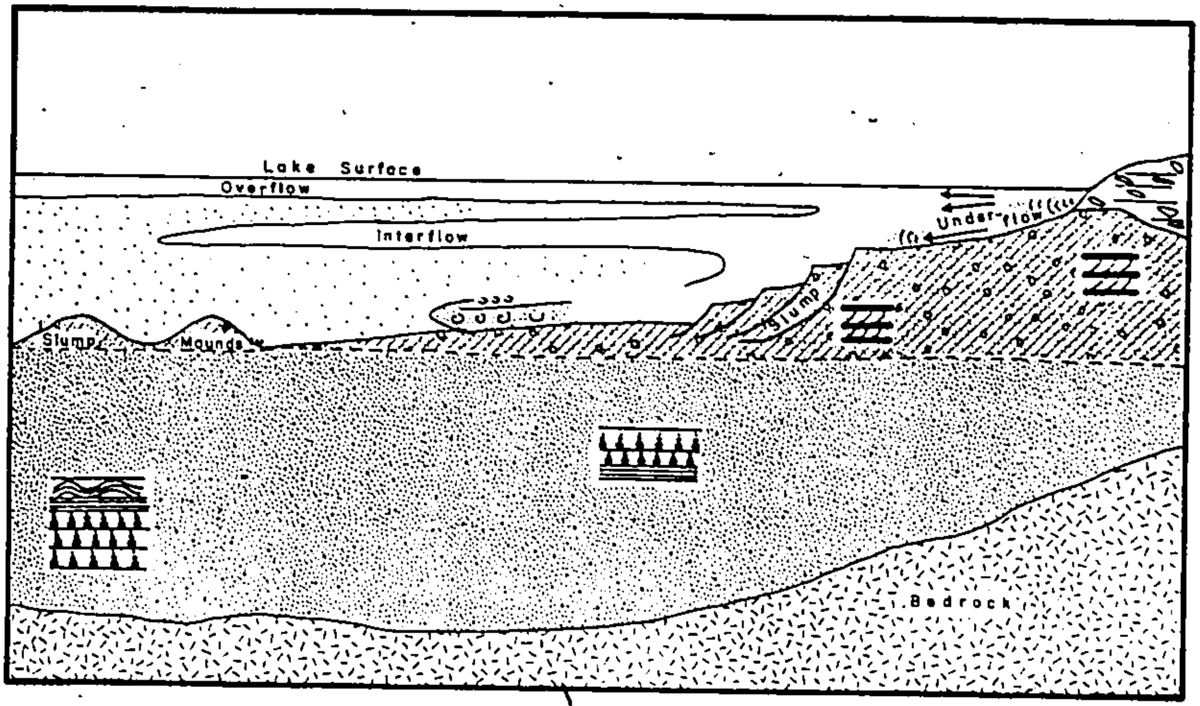


Figure 5.14: Processes responsible for the sediments in the Conne River bank. Lowermost unit is lake bottom sediments, showing graded bedding, fine laminations and convolute bedding. Overlying unit represents low slope prograding delta with alternating couplets of clayey silt and rippled or massive sand. Top unit is crude horizontally bedded, imbricated glaciofluvial (topset) gravels. Symbol downslope from slump represents a turbidity current. Dots indicate sediment in suspension.

CHAPTER 6

WEATHERING AND PERIGLACIAL PHENOMENA

6.1 Introduction

In areas having poor radiometric control it is often necessary to establish a glacial chronology based primarily on surficial rock weathering criteria. Relative age dating techniques involving differential rock weathering features have been used successfully to date late Quaternary glaciations in the Rocky Mountains (e.g., Birkeland, 1973; Carroll, 1974) and to establish multiple weathering zones in the Canadian Arctic (Pheasant and Andrews, 1972; Boyer and Pheasant, 1974; Dyke, 1978). The rationale, summarized by Dyke (1978), is that "degree of weathering, just as degree of soil development, is a function of climate, parent material (structure, lithology), topography (drainage), organisms, and time. If (sample) sites are selected in such a way that the first four variables are held roughly constant, then degree of weathering is a function of the length of time that the parent material, be it a bedrock outcrop or till clast, has been exposed to the weathering processes." The occurrence of relict periglacial phenomena has also been utilized in some studies in attempts to assess relative ages and as an indicator of extra-glacial areas (e.g., Tucker, 1979). Boyer and Pheasant (1974) considered that weathering zones could be used as regional stratigraphic markers on which Quaternary interpretations could be based.

In Newfoundland, "old" highly weathered "unglaciated" surfaces were recognized on the summits of the Long Range Mountains over 50 years ago (Fernald, 1925; Coleman, 1926). More recently, Brookes (1977; 1978) and Grant (1977a) identified three weathering zones on the west coast of Newfoundland. Their youngest weathering zone (Brookes', 1977, Zone 1; Grant's, 1977a, Zone C) is known to be Late Wisconsin in age and is correlated with the Saglek Zone of Ives (1978) and Zone III of Boyer and Pheasant (1974). The next oldest zone (Brookes', 1977, Zone 2 and Grant's, 1977a, Zone B) has been called early to pre-Wisconsin by Brookes and an unknown age (10 to 100,000 years) by Grant. Brookes' age assignment for Zone 2 is crudely correlative with the Koroksoak Zone of Ives (1978) and Zone II of Boyer and Pheasant (1974). Grant's Zone B is difficult to correlate, although it may be the equivalent of Brookes' Zone 2a. Correlations and age assignments of the oldest weathering zones are even more difficult and tenuous at best. Tucker (1979) recognized two weathering zones on the Burin Peninsula, but was unable to accurately depict their limits from weathering data alone. His zones represented terrain covered by Late Wisconsin ice (Brookes' Zone I and Grant's Zone A) and an earlier zone (Brookes' Zone 2a and Grant's Zone B).

With these facts in mind, weathering data was collected on a reconnaissance basis in the Hermitage Bay area to serve as a test for the ice limits previously identified from other criteria. If differences in the degree of weathering occur within and beyond the bounds interpreted as ice limits it is most probable that they result from differences in the duration of exposure and, as such, represent weathering zones which can be correlated with zones of other areas. The indicators of duration of exposure to weathering processes used in this study were thickness of oxidized weathering rinds, height of quartz vein protrusions, depth of weathering pits, and the presence of tors and felsenmeer.

6.2 Weathering Criteria

6.2.1 Weathering rinds

The oxidation of iron bearing minerals on rock surfaces exposed to the atmosphere leads to the development of a discolored weathering rind which becomes progressively thicker with time. Although the rate of increase in rind thickness is nonlinear (Carroll, 1974), comparisons of the absolute thickness of weathering rinds on similar rocks in different locations may provide some indication of the relative duration of the weathering episodes involved. For the present study, erratics and bedrock outcrops were broken exposing a fresh surface from which the thickness of the weathering rind was measured to the nearest millimeter. At all sample sites a minimum of four measurements was taken, from which only the thickest rind was recorded (as per Birkeland, 1973; Carroll, 1974); the thickest rind is considered to be more representative of the weathering period involved. The results, plotted on Figure 6.1 (precise location, rock type and rind thickness are listed in Appendix 1), reveal that the summits northeast of the head of Hermitage Bay, previously identified as nunataks during the last glacial event, have bedrock rind thicknesses of 24 and 18 mm. These values are, in general, thicker by a factor of two than those found at other locations within the limits of the last glacial event. Figure 6.2 shows rind thickness plotted against elevation. From the data shown there does not appear to be any trends or separate populations which would indicate more than one weathering zone.

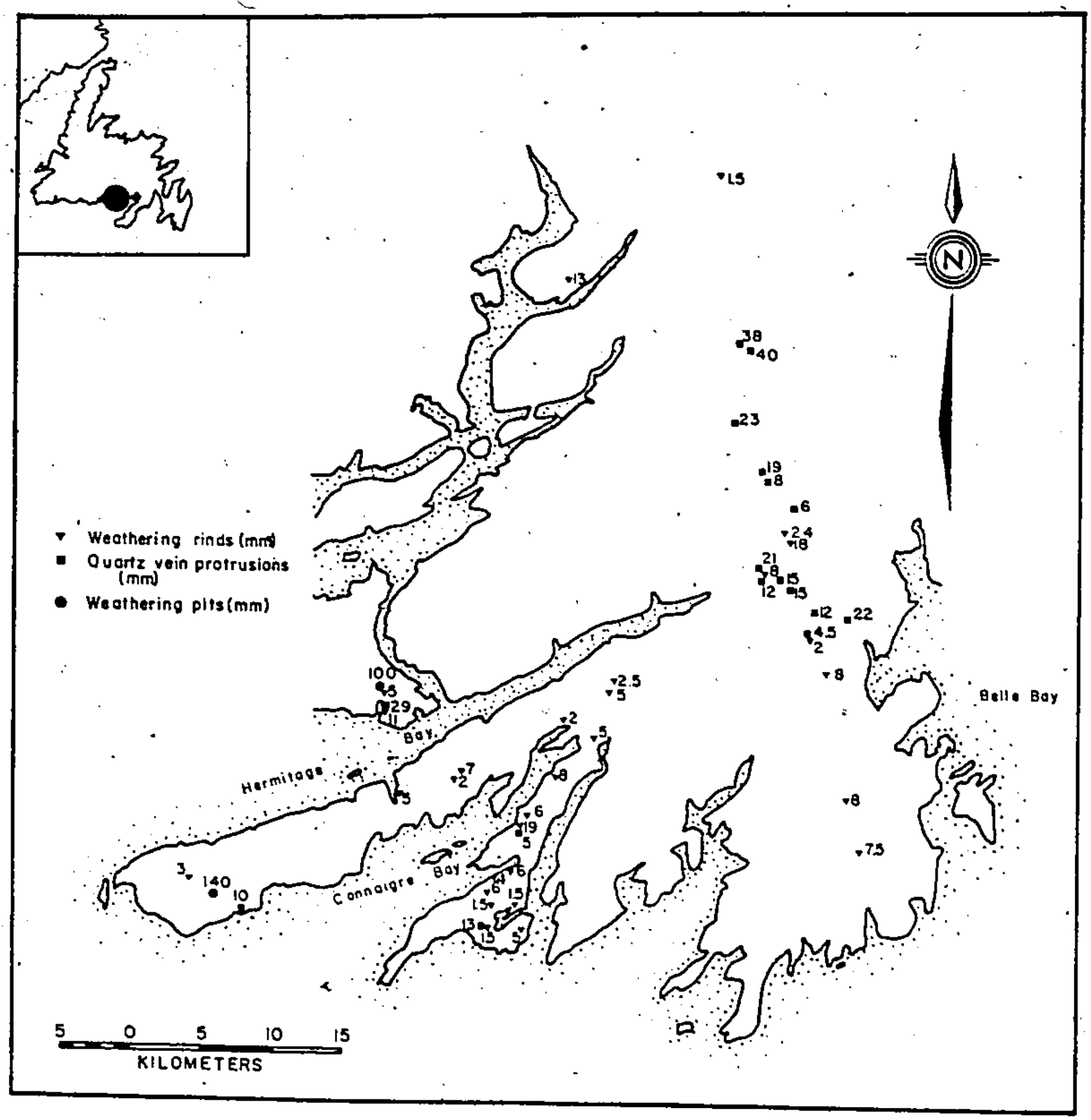


Figure 6.1: Location and thickness of weathering rinds, quartz vein protrusion, and weathering pits (taforis). Maximum thicknesses correspond well with nunataks and extra-glacial areas shown in Figures 2.1 and 8.1.

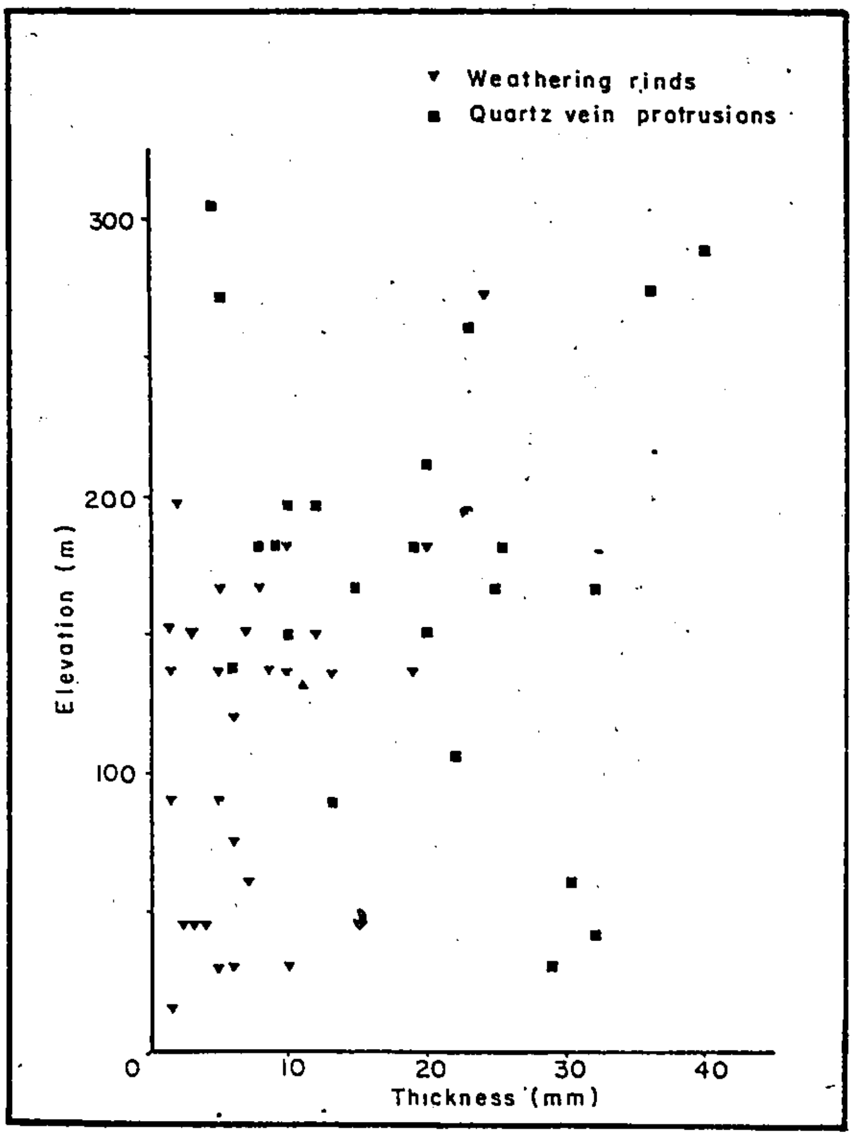


Figure 6.2: Weathering rinds and quartz vein protrusions plotted against elevation. There does not appear to be any trends or separate populations which would indicate more than one weathering zone.

6.2.2 Quartz vein protrusions

The differential weathering and grain by grain disintegration of less resistant rock on either side of quartz veins leaves the veins protruding above the local bedrock surface. For similar bedrock types, the higher the quartz vein protrusion, the greater the duration of exposure to weathering. At all sample sites, only the quartz veins having the greatest relief were recorded. The results, plotted on Figure 6.1 are more conclusive than the data collected for weathering rinds. Although there are exceptions, the maximum vein protrusions tend to occur on the heights of land that were above and beyond the limits of the last glacial event. The values obtained at Gaultois (29 mm) and east of Little River (19, 23, 38 and 40 mm) are comparable to those obtained by Tucker (pers. comm.) beyond the limits of the Late Wisconsin glaciation in the Burin Peninsula. Brookes (1978) recorded quartz vein protrusions of 100 mm in his Zone I (Zone 3 of Brookes, 1977) which may never have been glaciated. Figure 6.2 shows again that there is no apparent trend or separate populations of quartz vein protrusions which would indicate more than one weathering zone.

Figure 6.3 is a photo of a well rounded granite erratic taken on the ridge south of Salmonier Cove Pond. Considerable grain by grain disintegration of the erratic has occurred over time and left the rock in its present form. The location is above the limit of the last glacial event (Fig. 2.1).

6.2.3 Weathering pits

Weathering pits occur on the surface of medium to coarse grained granite and granitic gneisses as a result of chemical and/or mechanical weathering. Results from the Antarctic (Calkin and Cailleux, 1962) have

Figure 5.3: Well rounded, weathered granite erratic found near the mouth of Salmonier Cove Pond (photo location 1M/12:069638). The location is above the limit of the last glacial event.

shown that the degree of weathering pit (tafoni) development increases with the length of subaerial exposure. This agrees with conclusions drawn from observations of weathering pits in Labrador (Ives, 1966). Dahl (1966), however, believes that weathering pits in Norway have developed since the Würm glaciation and are of limited value as a relative age indicator. Only two locations of significant weathering pit development were found in the study area (Fig. 6.1): Seal Cove (140 mm) and Piccaire (100 mm). Both locations were beyond the limit of the last glacial event. The depths of the pits are similar to values noted by Brookes (1978) in his unglaciated zone, although in the present study a similar interpretation is not implied.

6.2.4 Incipient tors and felsenmeer

Pheasant and Andrews (1973) proposed that "old" weathering zones will be characterized by the presence of characteristic morphological features such as tors. Within the Hermitage Bay area, some of the summits depicted as nunataks during the last glacial event have been frost shattered to such a degree that they can be considered as tors and felsenmeer. However, since the features are not as well developed as many Arctic examples (e.g., Dyke, 1976) the modifier "incipient" is added to imply an embryonic stage of development.

Weathered bedrock, with incipient tors, occurs on the summits northwest of St. Albans (Fig. 6.4), at the Head of Hermitage Bay (Fig. 6.5), south of Salmonier Cove Pond (6.6) and in the Seal Cove-Pass Island area. The incipient tors consist of <u>in situ</u>, or partially loosened bedrock blocks. All are highly jointed and the degree of roundness of individual

Figure 5.4: Weathered bedrock on summits northwest of St. Albans (photo location 1M/13:806056). Weathering rind thickness of 28 to 30 mm and quartz vein protrusion of 36 mm. Elevation: 290 m asl.

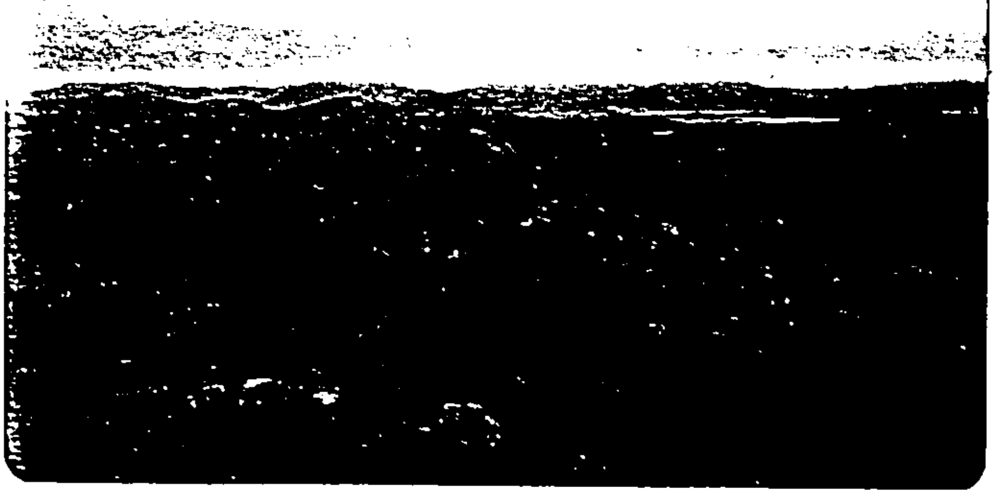


Figure 6.5: Weathered bedrock at the head of Hermitage Bay (photo location 1M/12:088861). Elevation: 274 m asl.

Figure 6.6: Weathered bedrock south of Salmonier Cove Pond (photo location 1M/12:069638). Elevation: 112 m. Note well rounded, weathered granite in lower left hand corner of photo.

blocks varies from angular to round. Grus development is minimal with the maximum depth being ~ 6 cm.

Frost-shattered bedrock detritus, or felsenmeer, occurs at the top of the hill above English Harbour West (Fig. 6.7), east of Little River (uplands centred at M/13:072965 and M/13072924)(Fig. 6.8), around the head of Hermitage Bay (Fig. 6.9 and 6.10) and inland between Cinq Isles Bay and Old Bay (Fig. 6.11). The incipient tors and felsenmeer all occur beyond the limits of the last glacial event (Fig. 2.1). Erratics are always present, usually rounded and weathered, indicating that the whole area has been glaciated at some period of time in the past.

6.2.5 Discussion

Initial observations of differences in weathering characteristics (i.e., quartz vein protusions, weathering rinds, weathering pit development, and morphological features related to weathering) between the uplands and lower elevations in the Hermitage Bay area indicate that two weathering zones may exist. Although it was not possible to define the limits of the zones solely from the weathering data alone, there is a general spatial correlation of areas exhibiting a higher degree of weathering and those depicted as having remained beyond and above the limits of the last glacial event (Section 2.2, Fig. 2.1). Terrain which has been recently glaciated is characterized by abundant well preserved glacial striae, grooves and polish, minimal development of weathering rinds and quartz vein protrusions, and "fresh" till and glaciofluvial deposits.

The older weathering zone has, in general, thicker weathering rinds, thicker quartz vein protrusions, deep weathering pits and a more

Figure 6.7: Frost shattered bedrock detritus, or felsenmeer, at top of hill above English Harbour West (photo location 1M/6:140580). Elevation: 213 m asl.

Figure 6.8: Frost shattered bedrock detritus, or felsenmeer, east of Little River (photo location 1M/13:071924). Elevation: 324 m asl.

Figure 6.9: *In situ* weathered bedrock on uplands around the head of Hermitage Bay.

Figure 6.10: Weathered bedrock and well rounded erratics around the head of Hermitage Bay (photo location 1M/12:121770). Elevation: 228 m asl.

Figure 6.11: Weathered bedrock between Cinq Isles Bay and Old Bay (photo location 1M/12:107786). Elevation: 309 m asl.

highly degraded bedrock surface. Quantitative evidence of this weathering zone is not overly conclusive from the data presented in Figure 6.2. More data is needed, especially from the ice free summits identified in Figure 2.1 and greater consideration should be made of the many varied bedrock types (Burke and Birkeland, 1979). However, in spite of these shortcomings the author is convinced that the second, older weathering zone exists in the region. The limited occurrence of grus and weathering pits, the incipient tors and felsenmeer, and the paucity of striae all indicate an older degraded surface. Figure 6.12 is a location map of photographs in this thesis which illustrate advanced weathering features. In all instances the locations were nunataks or lay beyond the limits of the last glacial event.

It is proposed that the lower zone is correlative with Brookes' (1977) Zone 1 and Grant's (1977a) Zone A, both of which were ice covered during the Late Wisconsin glaciations. Areas located above and beyond the limits of the ice cover have been exposed to weathering processes for a greater period of time and are comparable to Zone 2a of Brookes (1977), Zone B of Grant (1977a) and, more importantly, to an unnamed zone on the Burin Peninsula which has remained unglaciated since the late, mid-Wisconsin (Tucker, 1979).

Dyke (1978) was able to assign absolute values to the duration of weathering accomplished on granitic gneiss of various morphostratigraphic units in southeastern Baffin Island. The characteristics of his weathering zone A4 (Dyke, 1978, Table 2), to which he assigns an age of 20 to 40,000 years, are comparable with those of the older zone, developed on similar bedrock, in the Hermitage Bay area: "sufficient crystal removal to destroy all striae and grooves, either major disruption by frost

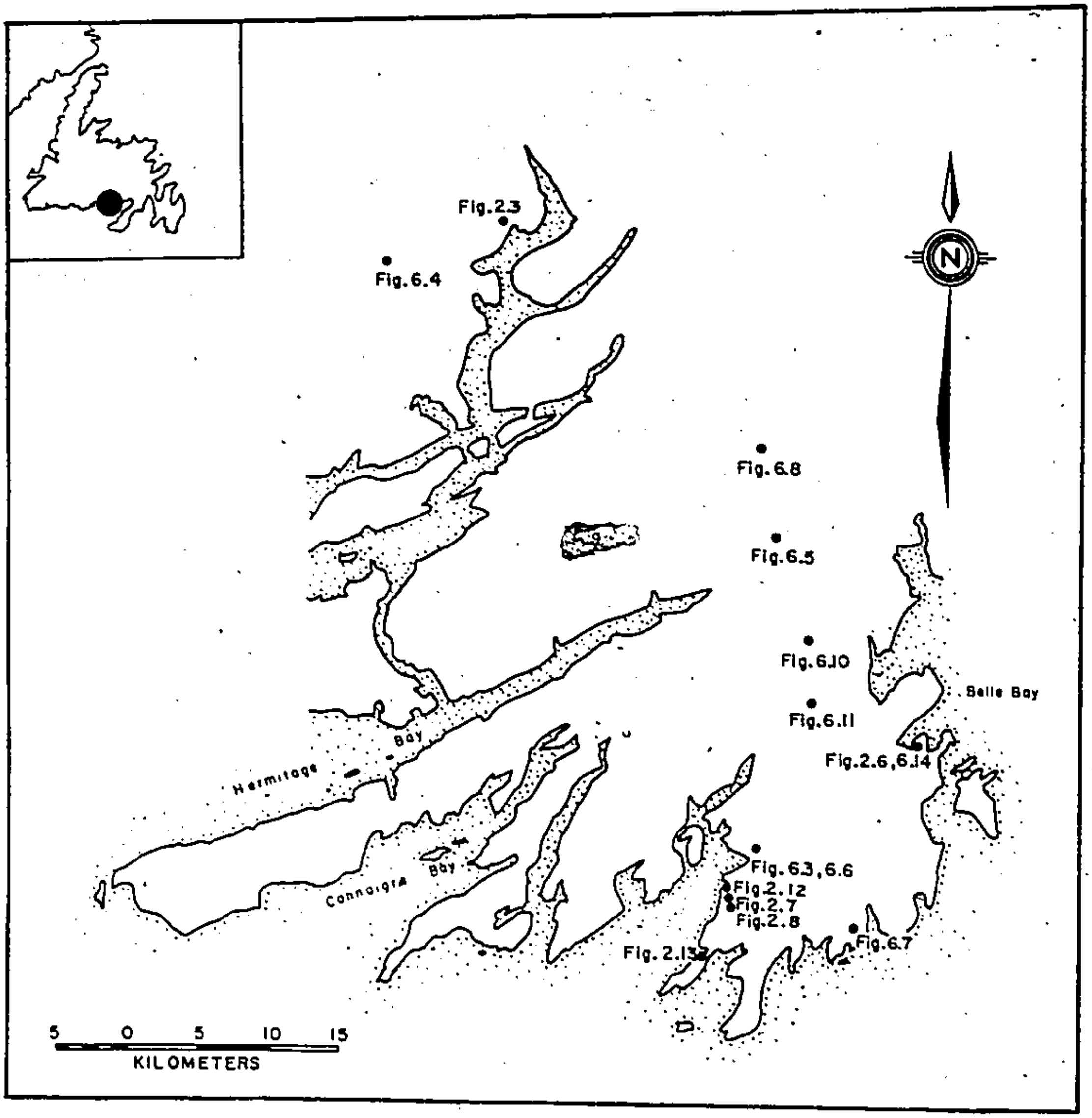


Figure 6.12: Location map of photographs in this thesis which illustrate advanced weathering features. In all instances the locations were nunataks or lay beyond the limits of the last glacial event.

heaving or felsenmeer (not as pronounced in the Hermitage Bay area), and first appearance of macropits and surface grusification of bedrock." Dyke's age assignment is significant in that it agrees remarkably well with the late, mid-Wisconsin designation (~ 30 to 40,000 years B.P.) for a comparable zone on the Burin Peninsula (Tucker, 1979). On these grounds, it is tempting to conclude that the older weathering zone in the Hermitage Bay area has not been glaciated for approximately 20 to 40,000 years and that the younger less weathered zone represents the maximum extent of the Late Wisconsin glaciation.

6.3 Periglacial Phenomena

Periglacial phenomena in the Hermitage Bay area are largely restricted to the occurrence of small scale, active patterned ground (circles and stripes), frost jacking and talus.

Small scale sorted circles and stripes. Active circles on a scale of centimeters across (Fig. 6.13) occur on bedrock rubble throughout the southern half of the study area. None were observed in the Bay d'Espoir-Little River region. Individual circles consist of an outer perimeter of coarse angular stones containing no fines, typically as large as 8 X 6 X 5 cm, although the median size is generally about half that. The inner portion consists of angular stone chips as small as 1 mm with silts and clays occurring immediately below the surface. The circles merge to form networks 0.2 to 2 m in width that may be bounded by larger stones which in turn form a larger less distinct circular pattern. The circles are usually constructed in small bedrock depressions in which detritus has accumulated, often to depth of no more than 6 cm. Stones are both local bedrock and more rounded, erratic pebbles and cobbles. Bedrock lithology

Figure 6.13: Active sorted circles at Harbour Breton (photo location 1M/5: 882579).

is not a factor in circle formation as they occur in all bedrock types. No circles were observed in till.

Small scale sorted stripes consisting of alternating rows of coarse and fine rock fragments form on slopes having angles as low as 3°. Their general characteristics are similar to those of sorted circles. In several instances a transition from sorted circles to elongate sorted circles to sorted stripes is evident wherever there is a slight change in slope angle from a horizontal surface to a gently dipping one. Most of the circles and stripes are currently active as indicated by the freshly broken, non oxidized stones on their surface. A set of stripes located between Wreck and Coomb's Coves is situated on a road cut and thus must have been formed within the last 20 to 30 years.

Inspection of meteorologic data for Grand Bank (Marine Sciences Directorate, 1974, p.65) on the Burin Peninsula, 45 km from Harbour Breton, provides some indication of why the active patterned ground exists at this latitude. Mean daily temperatures fluctuate on either side of the freezing point throughout the winter permitting repeated freeze-thaw cycles. Moisture content in the form of fog, rain and snow is high, providing a source of water for freezing. Snow falls can be heavy but frequent rain and heavy winds prevent significant accumulations. Reconnaissance work on the Burin Peninsula showed that in February, 1978 there was less than 3 cm of snow on the ground. The thin wet snow has poor insulative properties and permits maximum frost penetration when temperatures drop below freezing. No permafrost was observed and it is not necessary for the formation of these features.

Relict patterned ground. The only occurrence of relict patterned ground was at the summit of the hill above Corbin (~ 168 m asl). It

Figure 6.14: Relict patterned ground (large sorted circle) on hill above Corbin (photo location 1M/11:184709). Elevation: 168 m asl.

consisted of large scale partially vegetated, polygonal ground (Fig. 6.14) with large angular blocks up to 0.5 m across and small material in the centre. Individual forms are 2 to 3 m across. The central portions are currently active in some polygons. These features differ from large sorted polygons in till on the Burin Peninsula (Tucker, 1979) and Avalon Peninsula (Henderson, 1968) in that there is no till present and they were constructed from frost-shattered hill top detritus. It was not possible to ascertain the age of the polygons, but the site is located beyond the extent of the last glacial event (Fig. 2.1) and may represent a period of intense periglacial activity such as is likely to occur beyond the perimeter of a glacier. The active central portion of small material undoubtedly reflect recent meteorological conditions.

Frost heave. Several examples of frost heave were observed. Bedrock blocks were typically jacked up as high as 0.7 m or forced laterally up to 0.5 m apart by the repeated freezing and expansion of water in joints and cracks.

Talus. Talus occurs at the foot of many large steep cliffs, especially in the southern region. The talus slopes are as high as 30 to 40 m and individual blocks 2 to 3 m across were noted. The screes are still active as occasional rock falls were observed.

With the exception of the large scale relict polygons, none of the observed periglacial phenomena are indicative of extreme conditions which would have occurred during the last glacial event. All show some evidence of current activity, although the large stones of the relict polygons may have attained their original pattern during intense periglacial conditions of the last glacial event.

6.4 Conclusions

This reconnaissance investigation of weathering features suggests that two weathering zones exist in the study area. Any single weathering criterion by itself is non-conclusive, but when combined they indicate that surficial weathering characteristics tend to be more enhanced above and beyond the limits of the last glacial event (Section 2.2). Correlation with other weathering zones indicates that the older weathering zone has not been glaciated since the late, mid Wisconsin (Dyke, 1978; Tucker, 1979) and that the younger zone represents the extent of Late Wisconsin glaciation in the area. Periglacial phenomena, in general, are active and are not indicative of glacial limits.

CHAPTER 7

RAISED SHORELINES

7.1 Introduction

Relict shoreline features provide evidence of marine action when land-sea level relationships were different from the present. In glaciated areas the changes of relative levels of land and sea resulted from a combination of glacio- and hydro-isostatic depression of the earth's crust, and of eustatic sea level fluctuations that accompanied deglaciation. Raised shoreline features are well preserved along the coasts of the Hermitage Bay area taking the form of raised beaches and terraces formed in unconsolidated material, the upper limit of wave washed boulders, bedrock benches with and without surficial cover, and the upper contact between deltaic foreset and topset beds. They commonly occur at specific localities separated by several kilometers of coast showing no raised shoreline evidence. One problem is to correlate the phenomena from place to place and reconstruct the raised strandlines. In the discussion that follows, the term "benches" refers to platforms carved into bedrock while "terraces" refers to the erosion of unconsolidated sediments.

The only previous record of strandline elevations in the area was provided by Widmer (1950). He identified, at Deadman's Bight southwest of Harbour Breton, an 8.8 m asl striated, till covered bench overlain by terraced glaciofluvial deposits. The bench was interpreted as

Sangamon in age. At other localities he did not differentiate between benches and terraces, but rather considered their planation to have been coincident with 5 sets of strandlines occurring at 21.3, 17.6, 10.7, 8.8 and 7.0 m asl levels. As the Hermitage Bay area was then thought to lie south of the zero isobase (Flint, 1940), Widmer ascribed the origin of the strandlines to a series of large proglacial lakes of "Cary or Mankato age" on the south coast. Widmer also speculated on the existence of a submerged bench at 20 fathoms.

Grant (1975b, Fig. 1) noted the bench at Deadman's Bight, calling it "interglacial(?)" and also two other purported (Section 2.6, this study) benches under till. Whiteman and Cooke (1978) constructed a map showing isopleths of net postglacial emergence in Atlantic Canada. Their map, based on widely spaced data points, suggests that 20 to 30 m of emergence has occurred within the present study area in the last 13,500 ± 500 years B.P. Tucker (1979), with much better control, was able to determine more precisely the extent of marine overlap on the Burin Peninsula and showed that the line of zero postglacial emergence is located on the southeast tip of the peninsula, further south than previously reported (Flint, 1940; Jenness, 1960, 1963). Tucker also identified a 4 ± m asl Sangamon platform around the Burin Peninsula which is somewhat at odds with the level determined by Widmer at Deadman's Bight.

A model developed by Clark _et al_ (1978) shows Newfoundland to be located in a transition zone beyond the Laurentide ice sheet but within a zone of submergence due to a collapsing forebulge. The sea level response to deglaciation in this transitional zone is one of initial emergence followed by submergence. The maximum emergence of a strandline formed 18,000 years ago will be ~30 m asl but the value will decrease with increased distance from the ice sheet. The amount of submergence increases, with distance from the ice sheet, to a maximum of 22 m asl.

7.2 Methodology

Strandline elevations were measured using a Wallace and Tiernan barometric altimeter, a standard method typically used for similar work in the Arctic (e.g., Blake, 1975; Løken, 1978). Base level was taken as the still water mark at the time of measurement and then later corrected to mean tide level using local tide tables (Fisheries and Marine Service, 1978). Elevations are to the nearest half meter. The heights of wavecut benches and terraces were recorded at the base of the platform immediately before the landward rise (see Gray, 1975). Outwash terraces were measured at their leading edges. If bedding was visible the former lake or sea level was considered to be represented by the angular unconformity between topset and foreset beds. Wash limits, being concentrations of boulders from which fines have been removed were also levelled.

7.3 Raised Marine Shorelines

7.3.1 Benches

Elevations of bedrock benches are plotted in Figure 7.1 and listed with locations in Appendix 1. Only two instances of till covered benches were noted in the area. A 9.5 m asl bench at Oxford Point in Deadman's Bight, discussed previously (Sections 2.6 and 4.2), is overlain by till and deltaic sediments. Benches of similar elevation, but lacking surficial cover, also occur on nearby Black and Gull Islands. The

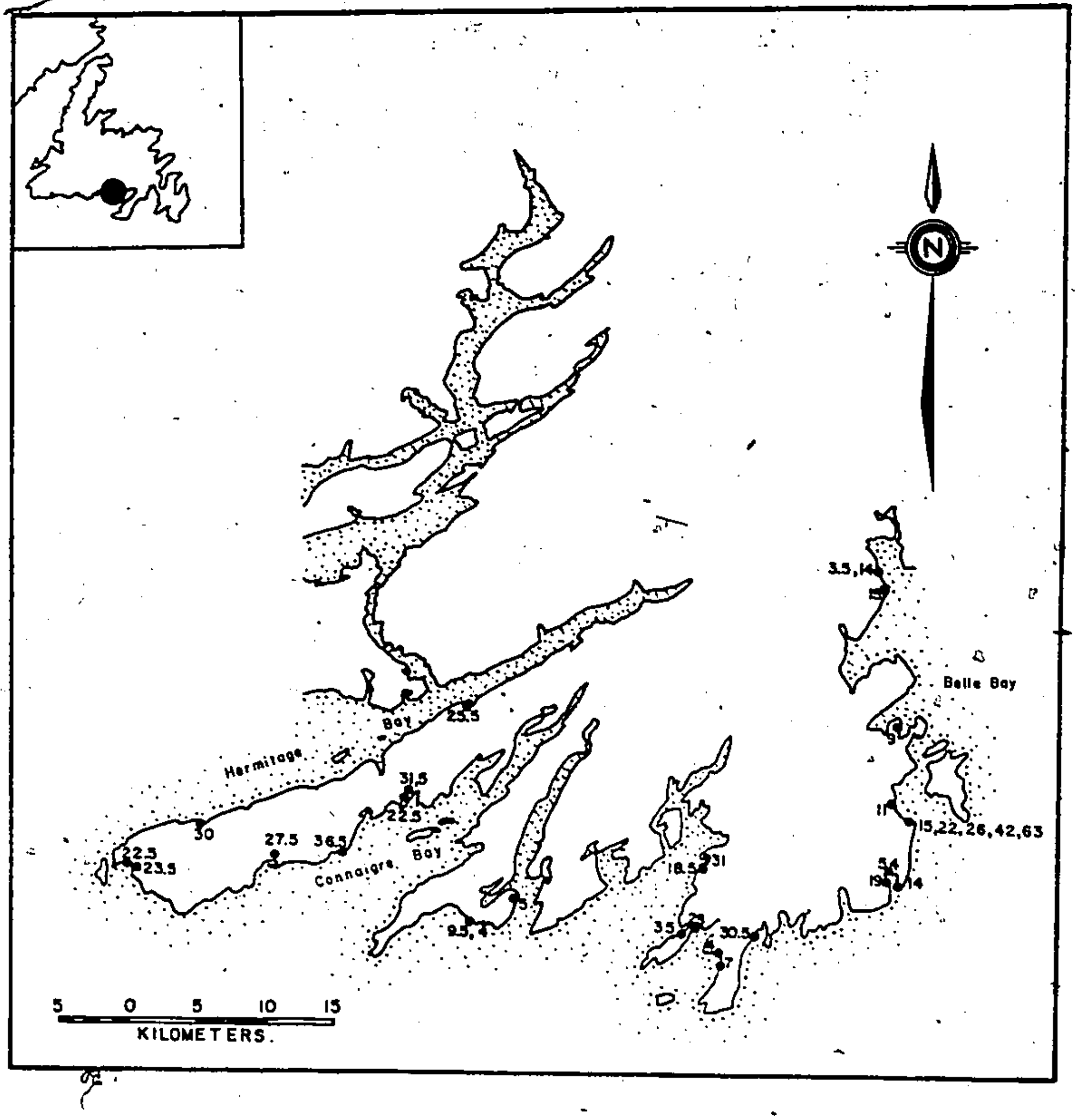


Figure 7.1: Elevations (in meters) of bedrock benches.

unconsolidated material overlying the bench has been terraced subsequent to its deposition and thus relates to a higher relative sea level which occurred after the last glacial event. The underlying rock platform antedates the deposition of the till and the episode of terracing. Widmer (1950) and Grant (1975b) have suggested that the bench is of Sangamon age.

In the immediate area there is also a 4 ± 1 m asl platform exposed at two locations, one of which is till covered. Opposite Gull Island, 1 km to the southeast of Oxford Point, there is a 4 m asl bench overlain by Gilbert-type foresets. At Rocky Harbour, 1 km east of Harbour Breton a 5 m asl bench is overlain by till and wave reworked material. Other occurrences of a 4 ± 1 m asl bench (Fig. 7.1) occur in St. John's Bay and Pool's Cove. The significance of this bench level is that Tucker (1979) has interpretted a discontinuous, but recurrent, 4 ± 1 m asl bench around the whole of the Burin Peninsula as Sangamon in age. Although Widmer (1950) speculated that there were two Sangamon benches of different elevations (levels unspecified) in the Hermitage Bay area, it is doubtful that both the 9.5 m asl and 4 ± 1 m asl benches represent Sangamon marine planation. That they predate the last glacial event is undeniable, but further age resolution is impossible without absolute dates and good stratigraphy.

In the Wreck Cove-Coomb's Cove area there are two sets of bench levels. The upper bench is discontinuous and highly dissected (Figs. 2.12 and 2.13); its elevation varies from 28 to 35 m asl. A lower bench, at 15 ± 2 m asl, is much more continuous and has suffered considerably less dissection than the upper surface. It is characterized by several well preserved, sometimes delicate sea stacks, caves, and notches. The lower bench level was probably cut during the Late Wisconsin postglacial high sea level whereas the upper bench represents a much earlier, interstadial or interglacial episode of planation (Section 2.6). At the entrance to Wreck Cove a modern platform tens of meters wide has been cut by recent coastal erosion.

The discordant nature of the remainder of the bench levels in the Hermitage Bay area makes further interpretations difficult. Most are carved in competent bedrock and thus would have required a greater period of sea level stability for planation than would have occurred during the postglacial. They may be of any age, some possibly predating the Sangamon.

7.3.2 Raised marine features in unconsolidated sediments

Figure 7.2 is a plot of elevations of raised marine beaches, wash limits, deltaic topset/foreset bed contacts and postglacial benches (see also Appendix 1). Marine limits, representing the highest level reached by the sea on the glacio-isostatically depressed coast during the late glacial and postglacial time are also shown. Levels in the northern Bay d'Espoir represent the strandlines of a large proglacial lake and will be discussed separately (Section 7.4).

In the Hermitage-Sandyville area a continuous 31-32 m asl terrace extends from Hermitage Bay to Connaigre Bay. There is wave reworked sediment overlying undisturbed dark grey till to this elevation at Hermitage. This was the highest observed strandline feature and may represent the maximum amount of postglacial emergence in the area. The lowlands on the peninsula southwest of Hermitage are girdled by a continuous 20-23 m asl strandline, sometimes more than a kilometer inland from the present coast (Fig. 2.1). Till is evident in several coastal

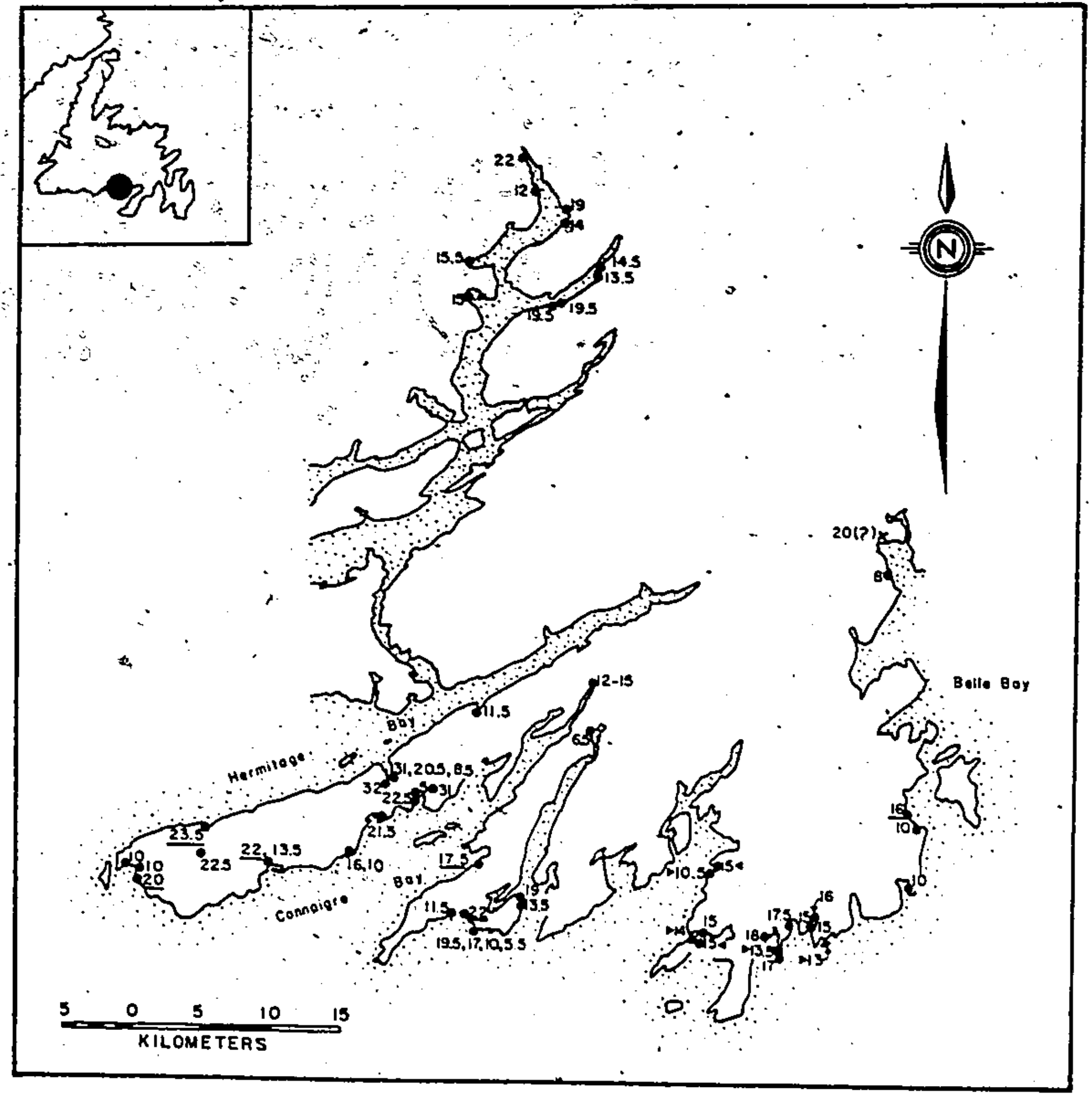


Figure 7.2: Elevation (in meters) of raised marine beaches, wash limits, deltaic topset/foreset bed contacts and post-glacial benches. Underlined values indicate marine limit. Triangles represent benches cut into Great Bay de l'Eau Conglomerate concurrent with terraces in unconsolidated material. "X" is a level attained by White (1939).

sections (Chapter 4) but is always overlain by as much as 2 to 3 m of wave reworked sediment. At Pass Island Tickle (Section 4.4) the base of the wave reworked sediments is sharply outlined by a lag deposit of large boulders resting on till. Strandlines on the Hermitage Bay side of the peninsula have been preserved only in the coves and inlets such as Grole (marine limit at 23.5 m asl).

At Deadman's Bight the marine limit, defined by a deltaic topset/foreset contact, occurs at 22 m asl. A lower strandline at 11.5 m asl represents a period when relative sea level remained constant for an extended interval of time (isostatic uplift equalling eustatic sea level rise) and a large marine incursion occurred reworking previously deposited deltaic sediments (Section 4.2). Further north, in Great Harbour Bight, a deltaic topset/foreset contact at 17.5 m asl may represent the marine limit there.

The limit of postglacial submergence along the Wreck Cove-Coomb's Cove-Boxey-English Harbour West coast is represented by a generally continuous bench at 15 ± 2 m asl. As noted earlier, the platform has several notches, caves and stacks which could not have survived glaciation. This plus the lack of dissection of the poorly lithified bedrock indicates a recent episode of planation. For this reason these benches are grouped with postglacial terrace levels. At Mose Ambrose, terraces cut into till, capped by up to 2 m of wave reworked sediment, on either side of the harbour occur at 14.5 to 15.5 m asl. From these terraces a faint strandline can be traced inland for ~ 2 km (Fig. 2.1) up the valley. At Belleoram, wave reworked sand and gravel over till form a terrace at 16 m asl.

Outwash/deltaic (?) sands and gravels greater than 15 m asl in elevation at the mouth of the Bay du Nord River can be seen from Pool's Cove. As it was not possible to obtain access to the deposit, a value of 20 m, recorded by White (1939) has been included in Figure 7.2. In spite of possible differences in levelling techniques this value is considered more representative of the maximum amount of submergence in this area than an 8 m value obtained for a terrace at the mouth of a cirque in Turnip Cove.

7.3.3 Discussion

An attempt to accurately determine and correlate postglacial marine limits in the study area has been severely restricted by the lack of datable organic material. Andrews (1975) has stated that "on the outer coast of glaciated regions which were deglaciated early and had only a limited load (Hermitage Bay area), the postglacial rise of sea level might have transgressed over the original marine limit and formed a higher and younger sea level." To this, can be added the complication of a migrating collapsing forebulge (Clark et al, 1978) in which the pattern is one of emergence and then subsequent submergence. It may also be possible that several of the terraces relate to separate glacial events.

Figure 7.3 is a construction of postglacial emergence based on what the author believes to be the Late Wisconsin marine limits. The isopleths were derived from a greater number of data points than used by Whiteman and Cooke (1978) for the area and are considered to be more realistic. The major difference is that the 30 m asl isopleth has been shifted southwards from the northern Bay d'Espoir (Whiteman and Cooke) to Hermitage (this study). The lines correspond in general, with isopleths constructed by Tucker (1979) on the Burin Peninsula.

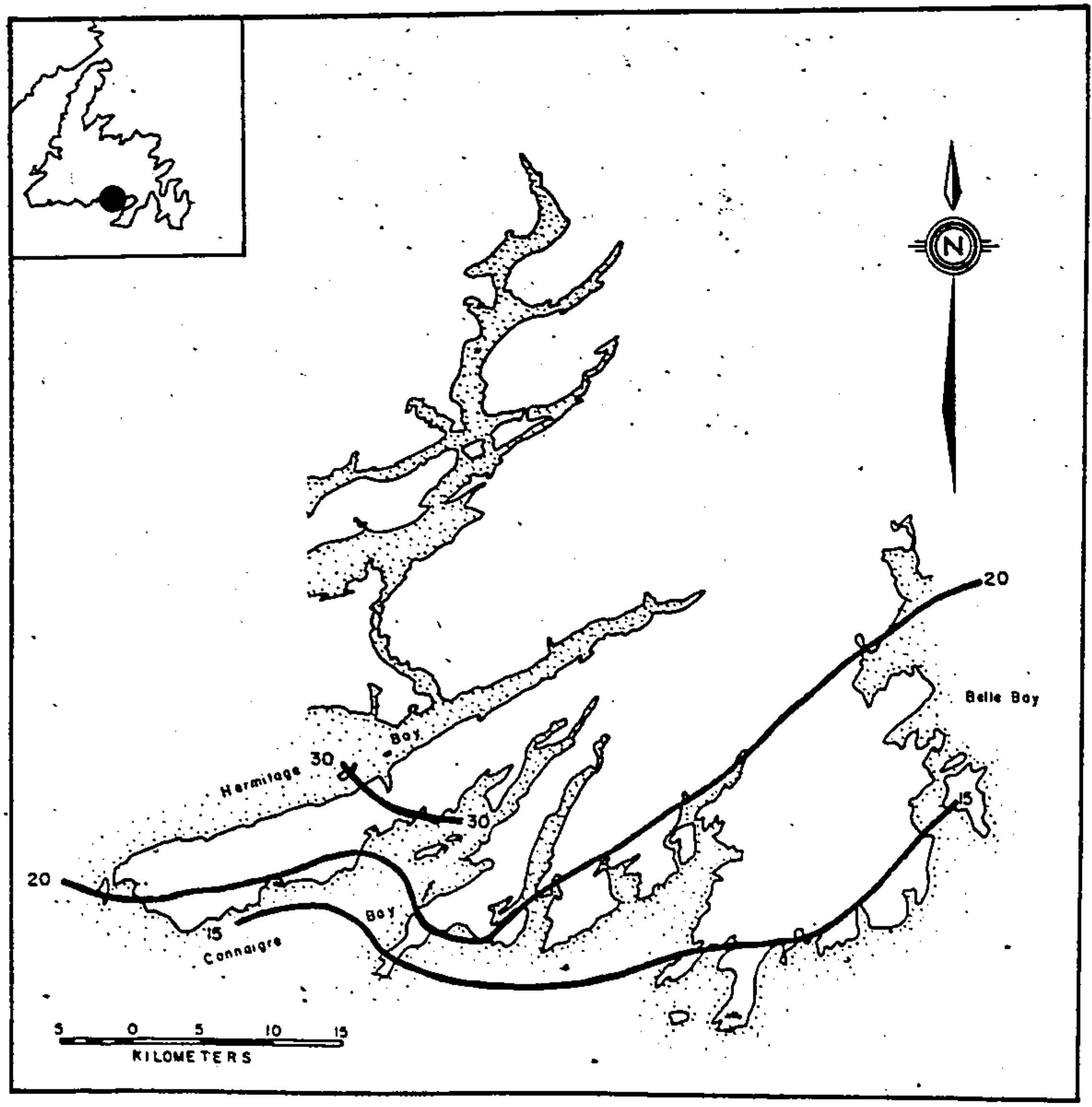


Figure 7.3: Isopleths of postglacial emergence.

Southward deflection of the isopleths reflects the presence of an ice mass in the Hermitage Bay area. The southwards dip and the close spacing of the 15 and 20 m asl isopleth and the location of the 30 m asl segment may represent significant thick amounts of ice over Hermitage Bay, Connaigre Bay and Northeast Arm. A lack of ice or thin ice may be responsible for the wider spread of the isopleths to the southeast, which agrees with conclusions made earlier (Sections 2.2 and 2.8). The 11.5 m marine limit at Furby's Cove, 12-15 m asl terrace at the Head of Connaigre Bay, 6.5 m asl terrace at Northeast Arm and the 17.5 m asl marine limit in Great Harbour Bight are all located further north than the isopleths would indicate. This suggests that late ice may have remained in the fiords and they may not have become ice free until a later date after several meters of uplift had already occurred. This is further confirmed by the general lack of terraces north of Furby's Cove especially when compared to their relative abundance along the coast south of Hermitage. From this, it can also be concluded that the glacier which fed down Hermitage Bay extended to a location between Hermitage and Furby's Cove. The lack of recent glacial evidence in Gaultois further restricts the position of the ice tongue terminus to north of Gaultois. Hermitage Bay widens by a factor of greater than three past Hermitage and had the glacier extended past Gaultois it would have spread laterally and thinned causing it to become buoyant.

Figure 7.4 shows the extent of postglacial marine overlap in the Hermitage Bay area. The amount of submergence varies depending on the amount of isostatic depression and the time of local deglaciation.

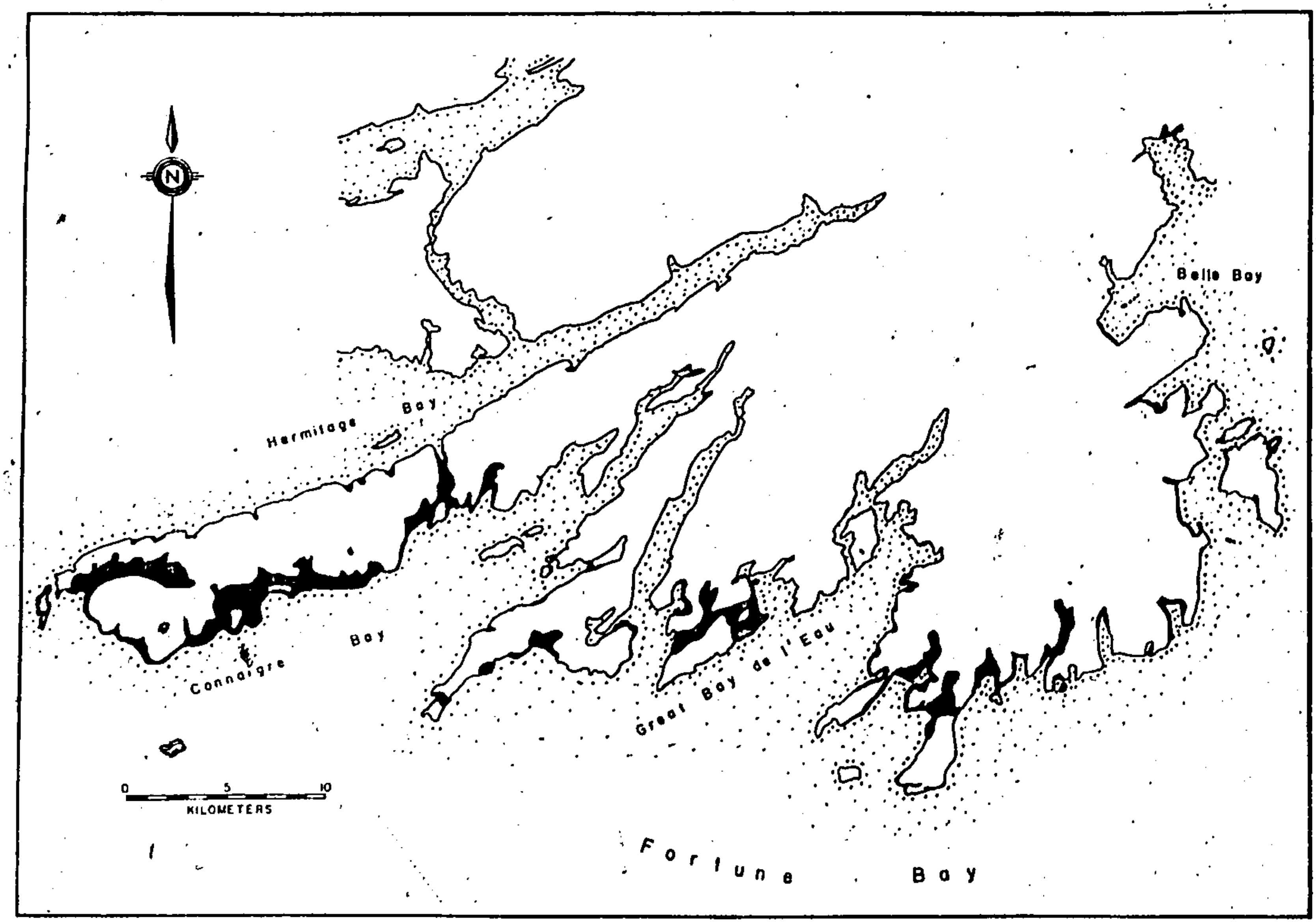


Figure 7.4: Extent of postglacial marine overlap.

7.4 Ice Dammed Lake Levels in Northern Bay d'Espoir

Evidence presented earlier (Section 3.9) has shown that northern Bay d'Espoir was occupied by a large ice dammed lake. Proglacial deltas were constructed into the lake at St. Albans, St. Veronica's Head of Bay d'Espoir and Conne River (Chapter 5). The elevation of deltaic topset/foreset contacts are shown in Fig. 7.2 (note however, that the two 19.5 m asl elevations taken at the Conne River townsite were from sands and gravels of unknown origin). The elevations can be grouped into two separate sets (~ 14.5 and ~ 20 m asl) which suggest that there may have been two stages of the lake at different levels. The maximum height of the lake surface was 22 m asl as represented by littoral sands and topset/foreset contacts near St. Veronica's. There is insufficient data to recognize any tilt to the lake surface.

CHAPTER 8

CONCLUSIONS

The research described in the preceding chapters has served to provide an interpretation of late Quaternary history and environments of the Hermitage Bay area which is summarized below:

1. The area has been affected by at least two major glacial events, the most recent being much less extensive than the earlier one. Two dissimilar tills separated by a sandy gravel unit at Trout Hole Falls Community Park (Section 3.2) demonstrate that at least two advances occurred here. Otherwise stratigraphic evidence of multiple till sections is absent. However, very clear evidence is provided by the persistent recurrence throughout the field area of "fresh" unoxidized till at low elevations, juxtaposed against barren, weathered bedrock surfaces directly upslope, indicating that the more recent glacial event occupied only the lower ground (Section 2.2). Furthermore, differences in the degree of bedrock weathering between uplands and lowlands indicate that two weathering zones may exist representing a younger, Late Wisconsin glaciated surface an an older surface which remained ice free during the Late Wisconsin (Section 6.2).

2. The earlier glacial event completely inundated the Hermitage Bay area. Its maximum extent is not known but the ice was sufficiently thick to contribute to the molding of large bedrock hills into stoss and lee forms (Section 2.5) and erode broad U-shaped valleys leading into Boxey, St. Jacques and Mose Ambrose. Ice flow, as determined by bedrock asymmetry, was largely due south. This glaciation accounts for the ubiquitous occurrence of erratics and emplaced the "old" tills found at Pass Island Tickle (Section 4.4), Seal Cove (Section 4.5) and near the mouth of Salmonier Cove Pond (Section 2.2).

3. The most recent glacial event was not all encompassing. Several hill tops remained ice free as nunataks, as well as several other extra-glacial areas. The maximum extent of this glaciation is illustrated in Figure 8.1, based on the compilation of ice limit data presented in the preceding chapters. A small ice cap centred north and east of the head of Hermitage

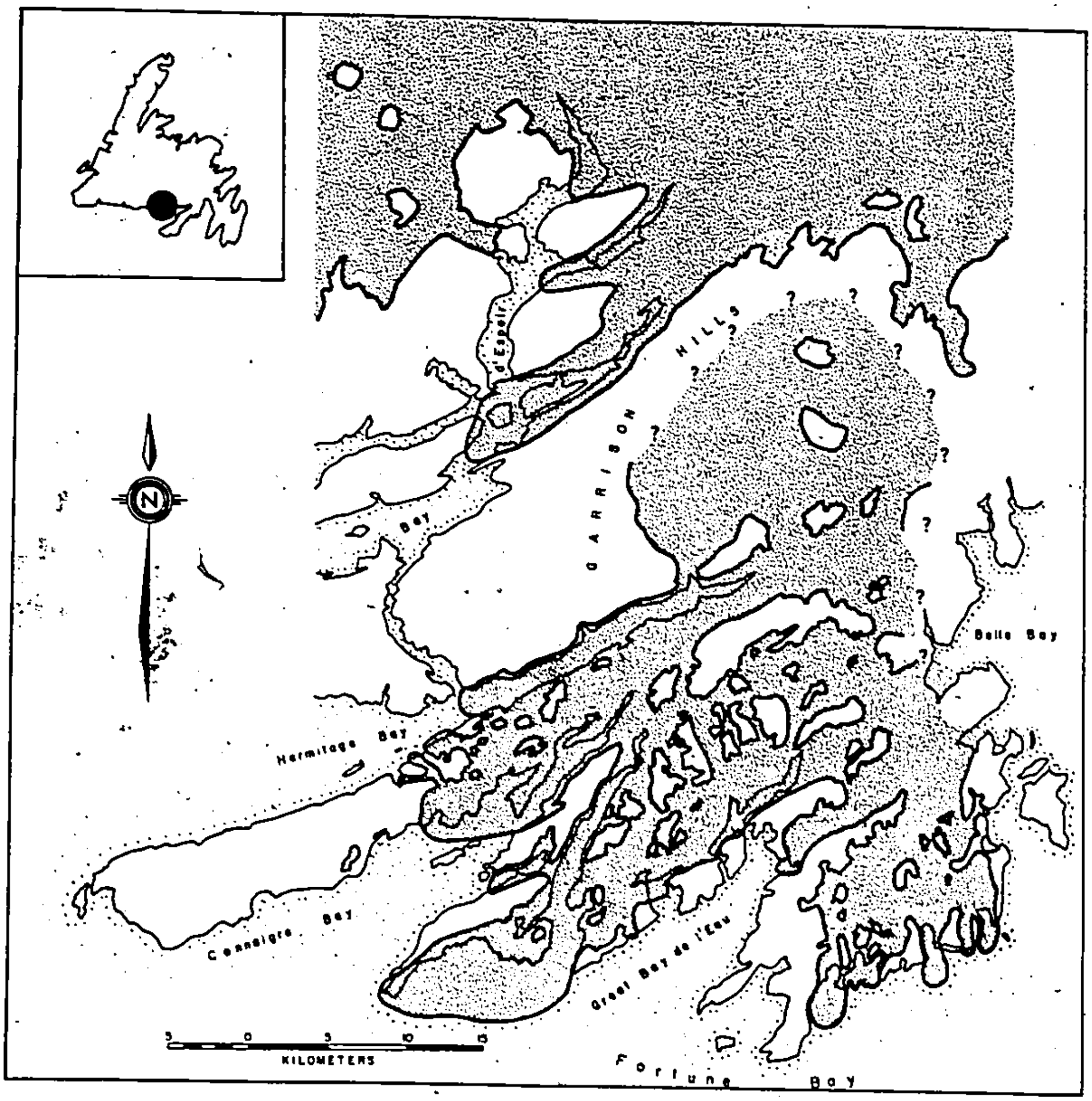


Figure 8.1: Maximum extent of the Late Wisconsin glaciation in the Hermitage Bay area (stippled). "?" indicates where margin is indistinct, possible due to thin ice.

Bay was distinctly separate from a large, main island based ice cap centred to the north. The Garrison Hills acted as a topographic barrier to southwards flow of this northern ice cap and northwards flow of the smaller cap. The southern boundary of the northern cap is a well defined margin extending across the top of the field area from west of St. Albans to north of Bay du Nord. It can be traced intermittently beyond the field area to north and west of Terrenceville where it connects with a similar limit recognized by Tucker (1979) on the Burin Peninsula. Many coastal areas remained ice free at this time (Sections 2.2, 2.6, and 6.2). Several large, coalescent, inland cirques south of Salmonier Cove Pond contributed ice to the smaller ice cap feeding down the valleys leading to Boxey and Mose Ambrose. Isopleth curves of postglacial emergence (Section 7.3.3), ice cap profiles (Section 2.8) and immature till (Section 2.2) indicate that ice in the English Harbour West, St. Jacques and Belleoram area was relatively thin and had little erosive power. There is no evidence for north flowing offshore ice affecting the southern part of the area, as proposed by Grant (1975b). During deglaciation, as the area underwent uplift, late ice may have remained in Hermitage Bay, Northeast Arm and Great Harbour Bight until after several meters of uplift had already occurred (Section 7.3.3).

4. The rate of deglaciation of tongues of ice feeding into northern Bay d'Espoir was unequal. A valley glacier which occupied the Little River valley extended across Bay d'Espoir (Fig. 8.1) and created a large ice dammed proglacial lake (Section 3.9) which was subsequently infilled with sediment. Maximum height of the lake was 22 m asl. A model for a low slope, prograding delta advancing over glaciolacustrine sediments (Figs. 5.13 and 5.14) is developed based on a type section at Conne River (Chapter 5). Three distinct lithological units are present. A lower unit consists of finely laminated silts deposited out of suspension from sediments introduced into the lake by inter- and overflow currents, and sharp based, graded beds deposited on the lake bottom by turbidity currents. Ripple cross-laminated and massive sands interbedded with clayey silts in the middle unit were deposited on the front of a low slope, prograding delta. A coarse upper unit of poorly sorted, structureless sands and gravels with crude horizontal stratification represents the proximal glaciofluvial topset beds of the advancing delta.

5. Based on correlation with other weathering zones (Brookes, 1977, 1978; Grant, 1977a; Dyke, 1978; and Tucker, 1979) it is tempting to conclude that the older weathering zone (areas lying beyond and above the margins of the last glacial event) has not been glaciated for 20 to 40,000

years and the younger, less weathered zone (within the limits of the last glacial event) represents the maximum extent of the Late Wisconsin glaciation (Section 6.2). This age assessment is further supported by differences in the degree of dissection of bedrock benches at Coomb's Cove and Wreck Cove which have remained unglaciated since at least the last interstadial (Section 2.6). The southern margin of the northern ice cap is interpreted as representing the maximum extent of this ice body during the Late Wisconsin glaciation and not the recessional position of a more extensive glaciation as proposed by Jenness (1960). Independent research by Tucker (1979) led him to assign a similar age for a continuation of this same limit over the Burin Peninsula. Marine foraminifera in till, tentatively assigned an age of ~ 70,000 years, were deposited by a glaciation more extensive than the last (Section 4.5).

6. Isopleths of postglacial emergence show that at least 30 m of uplift occurred since the Late Wisconsin deglaciation (Section 7.3.3). The maximum amount of uplift (32 m) occurred in the Hermitage-Sandyville area. Till covered bedrock benches (4 ± 1 and 9 m asl) in the Deadman's Bight-Harbour Breton area predate the last glacial event and may possibly represent Sangamon or earlier marine planation.

These results, in conjunction with those of Tucker (1979), should provide a broader understanding of the style of glaciation on southern coastal Newfoundland during the Late Wisconsin. At this time the south coast was not totally inundated. A Newfoundland based ice cap extended southwards, terminating against topographic barriers just inland of the coast, leaving several coastal areas and nunataks ice free. The southern Hermitage Bay area supported isolated cirques and a small icecap from which valley glaciers fed seawards; the southern Burin Peninsula remained largely ice free. The fact that several ice free areas existed at this time supports the general hypothesis put forward by Grant (1976, 1977b) and Ives (1978) that eastern North America was not totally glaciated during the Late Wisconsin.

Any future research in the area should make a concerted effort to obtain dateable organic material and thus provide absolute dates for the events discussed above. To this end the marine till in the outcrop at Seal Cove (Section 4.5) should be further investigated.

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Appendix 1a

Striations: Location and Azimuth

Note: Number in parenthesis after multiple striation indicates relative age of striation i.e., (1) is older than (2); (1)(2)(1) means one striation (2) is younger than the other two (1)(1) with further age differentiation not possible.

Map Location	Azimuth (degrees)
1M/5 128577	206(1), 180(2)
1M/5 122573 or 121568	210
1M/5 118573	195
1M/5 093553	138
1M/5 083567	207
1M/5 865576	284
1M/5 867374	227, 260
1M/5 882598	170
1M/5 879606	230(1), 190(2)
1M/5 878608	193
1M/5 873600	266 N
1M/11 132657	187
1M/11 136653	183
1M/11 138651	182
1M/11 193618	115

Map Location	Azimuth (degrees)
1M/12 777623	200
1M/12 005802	245
1M/12 988780	225
1M/12 833668	221, 159
1M/12 090853	04
1M/12 094879	149
1M/12 969754	345(1), 224(2)
1M/12 963736	218
1M/12 963726	206
1M/12 878611	230
1M/12 893621	253
1M/12 926681	251(1), 208(2), 232(1)
1M/12 927683	208(1), 124(2)
1M/12 079891	165
1M/13 073095	149
1M/13 074902	160
1M/13 948051	178
1M/13 052953	172
1M/13 805076	154(2), 195(1)
1M/13 788078	156
1M/13 768061	194, 201
1M/13 772063	195
1M/13 846038	112
1M/13 037117	42

Appendix 1b

Weathering Features: Differential Weathering, Rind Thicknesses, Map Locations, Elevation

Map Location	Rind Thickness (R) Differential Weathering (D)	Lithology	Thickness (millimeters)	Approximate Elevation (meters)
11P/9 663614	R	sandstone	3.0	15
11P/8 701587	D		10	
11P/8 686597			140	
1M/12 072838	D	granite gneiss	21	213
1M/12 026806	D		32	168
1M/12 936726	R	sandstone	7	610
1M/12 856692	R	volcanic	10	137
1M/12 851686	R	volcanic	2	198
1M/12 087847	R & D	pink granite	0	92
1M/12 090853	R		20	183
1M/12 088061	R	granite	24	274
1M/12 094879	D		6	137
1M/13 076898	D		8	183
1M/12 761610				15
1M/12 810671	R	metamorphosed volcanic	5	
1M/12 801732	R		11	30
1M/12 801732	D	very coarse grained granite	29	30
1M/12 802737	R	medium grained granite	6	76.2
1M/12 969754	R	volcanic	2.5	45.7
1M/12 964740	R	volcanic	5	91.9
1M/5 897597	R		3	45.7
1M/5 899597	R		1.5	15.24
1M/5 878581	R	siltstone (volcanic?)	1.5	91.4
1M/5 878581	D	"	13	91.4
1M/5 882569	R		1.5	91.4
1M/5 897606	R		6	122
1M/12 882617	R	volcanic	4	45.7
1M/12 893621	R	volcanic	6	30.5
	R		19	137.2
1M/12 906662	R	basalt?	6	137.2
1M/12 926681	R	slate	5	137.2
1M/12 926681	R	sandstones coarse & fine	8.5	137.2
1M/12 942703	R		5	168
1M/5 889584	R	granite	5.0	30.5
1M/12 076836	R	granite	8	152
1M/12 076836	R	volcanic	.5	152
1M/12 076836	R	volcanic	7	152
1M/12 076836	D	volcanic	12	152

Weathering Features: Differential Weathering, Rind Thicknesses, Map Locations, Elevation (continued)

Map Location	Rind Thickness (R) Differential Weathering (D)	Lithology	Thickness (millimeters)	Approximate Elevation (meters)
1M/12 086828	D		14	183
1M/12 092820	D		15	168
1M/12 107786	R	volcanic	2.0	305
1M/12 107786	D		4.5	305
1M/12 110804	D		12	198
1M/11 132800	D		20	107
1M/12 119763	R	conglomerate	8	167
1M/11 133674	R	siltstone	1.5	152
1M/22 143634	R	granite	7.5	122
1M/5 052598	D	conglomerate	30	61
1M/13 074902	D	granite	10-19	183
1M/13 051939	D		23	260
1M/13 059984	D		36-40	290
1M/13 056994	D	slate	36	274
1M/13 037117	R	sandstone	1.5	152
1M/13 019026	R	granite	13	137
1M/12 072838	Tafoni	granite	140	91
1M/12 802737	Tafoni	granite	100	76

Appendix 1c

Benches and Terraces: Map Location and Elevation

Map Location	Bench (B) Terrace (T) Delta Foresets (D)	Elevation (meters)
1M/12:813677	T	31, 20.5, 8.5
1M/5:906585	B	4
1M/5:905589	T	19
1M/12:958712	T	6.5
1M/11:173826	B	3.5, 14
1M/11:173826	T	8
1M/11:180812	B	15
1M/6:181602	B	19
1M/6:183608	B	54
1M/11:200644	B	63, 42, 26, 22, 15
1M/11:198646	T	10
1M/5:123559	B*	13
1M/5:128559	B*	13
1M/5:131559	B*	13
1M/5:190602	B	14
1M/5:122573	T	16
1M/5:121568	T	15
1M/5:118573	T	15
1M/5:119579	B*	16
1M/5:119579	T	14
1M/5:083558	B	30.5
1M/5:084566	T	18
1M/5:058548	B	3.8
1M/5:059538	B	7.2
1M/5:054570	B	7.1
1M/5:038565	B*	15
1M/5:038565	B	28
1M/5:032563	B*	14
1M/5:032563	B	35
1M/5:046611	B	18.5
1M/5:046611	B*	10.5
1M/5:104577	T	17.5
1M/11:188651	T	16
1M/11:188653	B	11
1M/5:052613	B*	15
1M/5:052613	B	31
1M/11:189716	B	9
1M/6:094558	B*	13.5
1M/5:092555	T	17
1M/13:958042	D	13.5
1M/13:957044	D	14.5
1M/13:932017	T	19.5

Map Location	Bench (B) Terrace (T) Delta Foresets (D)	Elevation (meters)
1M/13:926017	T	19.5
1M/13:934077	D	14
1M/13:932096	D	19
1M/13:903123	D	22
1M/13:863146	T	15.5
1M/15:833668	T	22.5
11P/9:719615	T	22, 13.5
11P/9:728615	B	27.5
11P/8:618608	B	22.5
11P/8:618608	T	10
1M/12:803647	T	21.5
1M/12:802642	B	22.3
11P/9:661633	B	30
11P/9:661633	T	23.5
11P/9:719615	T	13.5
1M/12:777617	T	16,10
1M/12:777617	B	36.5
1M/12:873725	T	11.5
1M/12:873725	B	25.5
1M/12:830666	T	5
1M/12:830666	B	31, 5
1M/12:806674	T	32
1M/5:856580	T	11.5
1M/5:865576	D	22
1M/5:875567	T	19.5, 10, 5.5
1M/5:875567	D	17
1M/5:907588	B	5
1M/5:907588	T	13.5

Datum take as mean tide level

* Great Bay de l'Eau Conglomerate

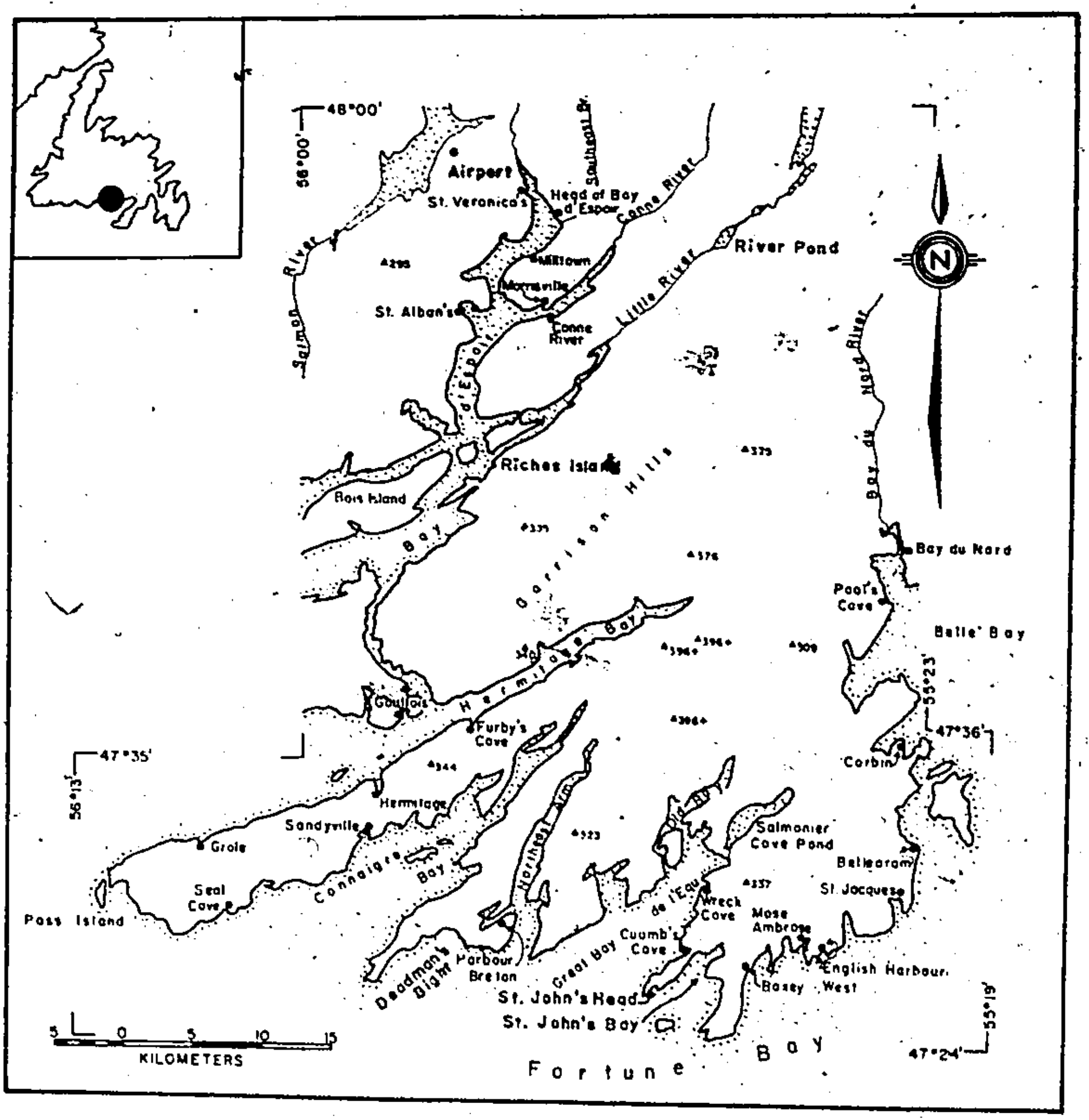

Figure 1.1: Place names of most locations discussed in this thesis.