THE QUATERNARY HISTORY OF THE CARMANVILLE (NTS 2E/8) AREA, NORTHEAST NEWFOUNDLAND

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BY

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ABSTRACT

The Carmanville (NTS 2E/8) area is located in northeast Newfoundland. It is marked by a complex assemblage of sediments and landforms. These are in order of diminishing areal importance: till which has a variety of geomorphic expressions such as veneers, hummocks, ridges, and lineations; organic deposits; glaciofluvial sand and gravel (also meltwater channels); colluvial sediment; and fluvial deposits.

Four ice flows are detected in the area: successively eastward, northeastward, northwestward, and northwestward. These are all argued to be Late Wisconsinan in age. Retreat and down-wasting of the third ice-flow, northwestward, was associated with large quantities of meltwater and the formation of the Ragged Harbour Moraine. This formed shortly after 12,000 BP. The final ice-flow event (also northwestward) may be correlative with the Younger Dryas cooling event dated to between approximately 11,000 and 10,000 BP.

Till deposits are mainly formed by northwestward moving ice and are mostly fine-textured with high percentages of local clasts. They display only minor regional textural and lithological variations. All tills observed and analyzed are basal, and most are interpreted as melt-out or lodgment tills.

The Ragged Harbour Moraine in the northeast of the study area is approximately 15km long and is composed of the most diverse sedimentary assemblage in the region. High angle-planar cross-bedded gravels, extensively deformed silts and sands, rhythmically and ripple bedded silts and sands, and diamictons are present within this complex and occur directly adjacent to each other. This sedimentary assemblage is interpreted as a sequence of subaqueous

fans which formed proglacially under rapid sedimentation rates, while ice was stagnating. Rogen *moraines* formed contemporaneously with the Ragged Harbour Moraine. The moraine formation does not represent a lengthy stillstand of glacial ice.

The marine limit in the area is at least 57m asl (above sea level) based on emerged marine sediment at Noggin Hill and Wing's Point. Five further major sea level stands occurred at 52m, 38m, 34m, 17m, and 11m asl indicated by the presence of emerged beaches. Two terraces at 5m and 2m asl are evident around most of the coastline but may represent major storm events rather than being representative of relative sea-level change. These sea-level stands all occurred between 12,500 and 10,000 BP. Sea levels dropped to below present during the mid-Holocene and then continued to rise until present times.

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CHAPTER 1
INTRODUCTION

1.1 - INTRODUCTION

Limited Quaternary work has been undertaken in northeast Newfoundland, and what has been done has concentrated primarily on offshore sediments. Quaternary research in other parts of Newfoundland has been relatively more widespread. For instance, the Quaternary history of regions including the Northern Peninsula, the Baie Verte Peninsula, Bonavista Peninsula, parts of the Avalon Peninsula, the Southern Shore, and parts of the interior, such as Gander and Buchans, is moderately or well understood.

The Carmanville area is one where little work has been done. Detailed regional analysis of this area will aid in understanding the Quaternary history of northeast Newfoundland, and of the Island as a whole.

1.2 - OBJECTIVES

The Carmanville map area (NTS 2E/8) is marked by a complex assemblage of Quaternary deposits and landforms resulting from glaciation and sea-level changes. This assemblage was investigated over a three-year period (1991-1994) with the main objectives being:

1. to describe and map the sediments and landforms of the Carmanville map area;

2. to determine ice-flow patterns in the area by examining all erosional and depositional landforms and associated sediments indicative of ice-flow directions;

3. to determine the glacial history of the area;

4. to determine the sea-level history of the area;

5. to use these data to construct the Quaternary History of the Carmanville area.

1.3 - LOCATION AND ACCESS

The study area is located in northeast Newfoundland between 49°15'N and 49°30'N latitude, and 54°00'W and 54°30'W longitude (Figure 1.1). It lies approximately 50km north of Gander, includes Gander Bay in the western part, and extends eastward to the Musgrave Harbour area. Carmanville, the major settlement in the area, is located on the coast of Hamilton Sound, in the centre of the study area. Smaller settlements include Frederickton, Noggin Cove, Clarke's Head, Mann Point, Aspen Cove, and Ladle Cove.

Access to and within the area is generally good (Figure 1.1). Route 330 (Gander to Wesleyville) crosses the map area from southwest to northeast. Other main routes include Route 331 (Clarke's Head to Twillingate) and Route 332 (Frederickton to Carmanville). These roads are all paved. Gravel logging roads are common throughout the area, and are generally well-maintained.

1.4 - BEDROCK GEOLOGY

A thorough understanding of bedrock geology of an area is important when trying to determine clast provenance. Any clast found in glacially derived material is an indicator of mineral and ore deposits and distinctive rock types (Dreimanis, 1956; Shilts, 1976), and thus aids in determining the approximate ice flow direction, and the source of an ice mass in a particular study area.

The bedrock geology of the Carmanville area is complex. During the Cambro-Ordovician, this area formed part of the southeastern margin of the

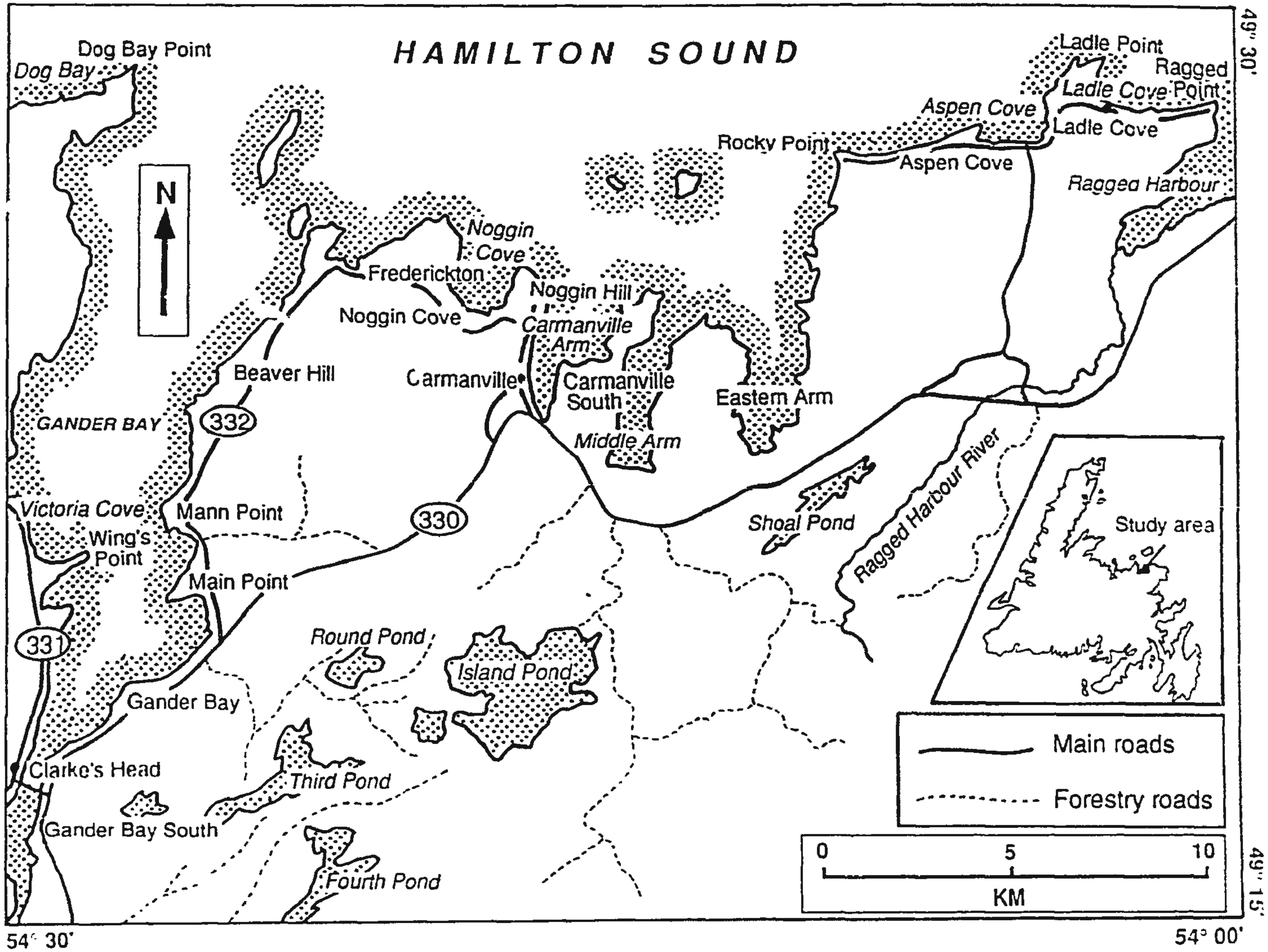


Figure 1.1: Map showing the location of all places mentioned in the text. with a location map of the study area (inset).

Iapetus Ocean (Currie and Pajari, 1981). The area underwent a complex series of sedimentary and deformational processes leading to the collapse of the continental rise prism, obduction of oceanic crust, and eventual cratonization (Currie *et al.*, 1980).

The bedrock geology of the area was mapped by Currie *et al.* (1980), and has since been the subject of more detailed investigation (Currie *et al.*, 1981; Pickerill *et al.*, 1981; Williams *et al.*, 1991; Currie 199[illegible]. It is dominated by five main geological assemblages (Figure 1.2). These are:

1. the Gander River Complex (formerly termed the Gander River Ultrabasic Belt (GRUB)) which was part of the oceanic crust (Currie and Pajari, 1981);

2. the Gander Group, deposited as part of the continental rise prism (Currie and Pajari, 1981);

3. the Davidsville Group, formed in an extensive submarine fan environment by turbidity currents and gravity slides (Currie and Pajari, 1981);

4. the Hamilton Sound Sequence which was formed in a marine environment (Currie, 1992); and

5. granitoid plutons.

The Gander River Complex (Unit 1) (Figure 1.2) represents the oldest rocks in the area (Currie, 1992). Ultramafic rocks in this group include talc-chlorite schist, pyroxenite, and serpentinite probably derived from dunite or peridotite (Currie, 1992). Mafic volcanics include pillow basalts, gabbro, and pervasively fractured and brecciated trondhjemite.

In the Carmanville area the Gander Group (Unit 2) (Figure 1.2) consists of two main formations. The Jonathan's Pond Formation (Unit 2a) is made up of green to grey feldspathic siltstone, with beds of coarser, quartzitic sandstone

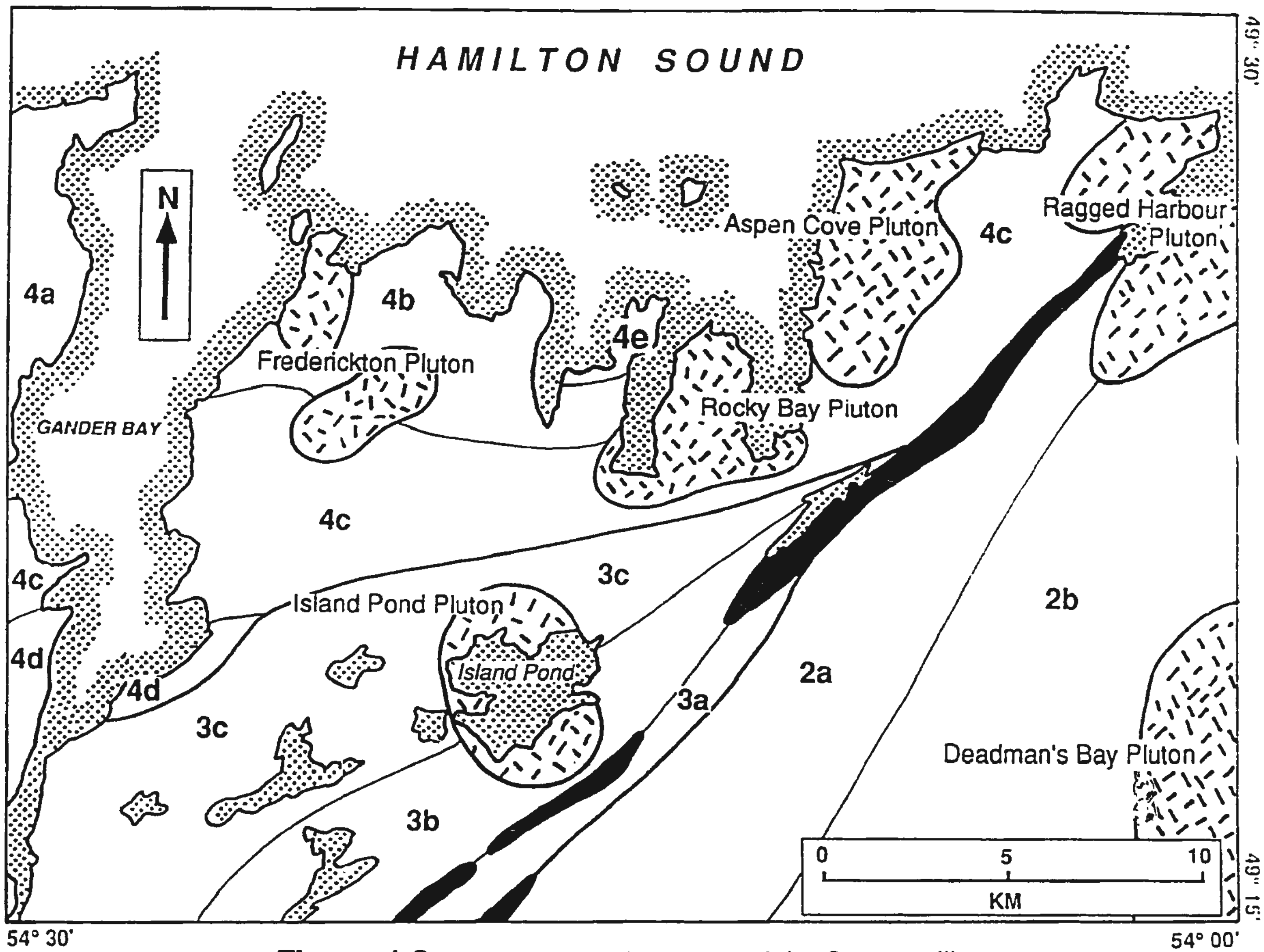


Figure 1.2: Simplified geology map of the Carmanville area (after Currie, 1992)

Unit 1: Gander River Complex

Unit 2: The Gander Group
2a Jonathan Pond Formation
2b Flinns Tickle Complex

Unit 3: The Davidsville Group
3a Weir's Pond Formation
3b Barry's Pond Formation
3c Round Pond Siltstone

Unit 4: The Hamilton Sound Sequence
4a Woody Island Siltstone
4b Noggin Cove Formation
4c Main Point Shale
4d Wing Point Greywacke
4e Carmanville Mélange

Unit 5: Granitoid Plutons

Figure 1.2: Legend

and pebble conglomerate. The Cuff Pond Pelite (Unit 2b) consists of black pelite and thin beds of grey-green psammite. In some areas near the Ragged Harbour pluton, the Cuff Pond pelite is metamorphosed and pervasively deformed. The most common rock types are migmatite, agmatite, amphibolite, pelite and greenschist. This assemblage is also known as the Flinns Tickle Complex (Currie, 1992).

The Davidsville Group (Unit 3) (Figure 1.2) is divided into three main formations (Currie, 1992). The Weir's Pond Formation (Unit 3a) is composed of red shale and minor quantities o[illegible]stone. The Barry's Pond Conglomerate

(Unit 3b) is composed of pebbles from the Gander River Complex in a green sandy to silty matrix. The Round Pond Siltstone (Unit 3c) is composed of thick beds of finely laminated grey to black siltstone and shale.

The Hamilton Sound Sequence (Unit 4) (Figure 1.2) is made up of five main units. The first is the Woody Island Siltstone (Unit 4a) (bands of dark grey to green shales and siltstone). These rocks are easily identified due to their large concentration of manganese-rich beds (Currie, 1992). The second is the Noggin Cove Formation (Unit 4b) composed of basaltic pillow lavas, tuffs, tuff breccia, and resedimented volcaniclastics. The Main Point Shale (Unit 4c) is characterized by black graphitic and pyritic shale which contains graptolites. Fourthly, the Wing Point greywacke (Unit 4d) is made up of dark grey greywacke and diamictites related to late Ordovician (Caradocian) glaciation (Pickerill *et al.*, 1979). The fifth main unit in the Hamilton Sound Sequence is the Carmanville Mélange (Unit 4e) which crops out between Gander Bay and Aspen Cove. The type of rock present is unusual as it displays olistoliths of sandstone, pillowed basalts, mafic volcanics or volcaniclastics, gabbro, trondhjemite, and limestone ranging in size from granules to kilometres across, in a matrix of pyritiferous black shale and silt (Pajari *et al.*, 1979; Williams *et al.*, 1991).

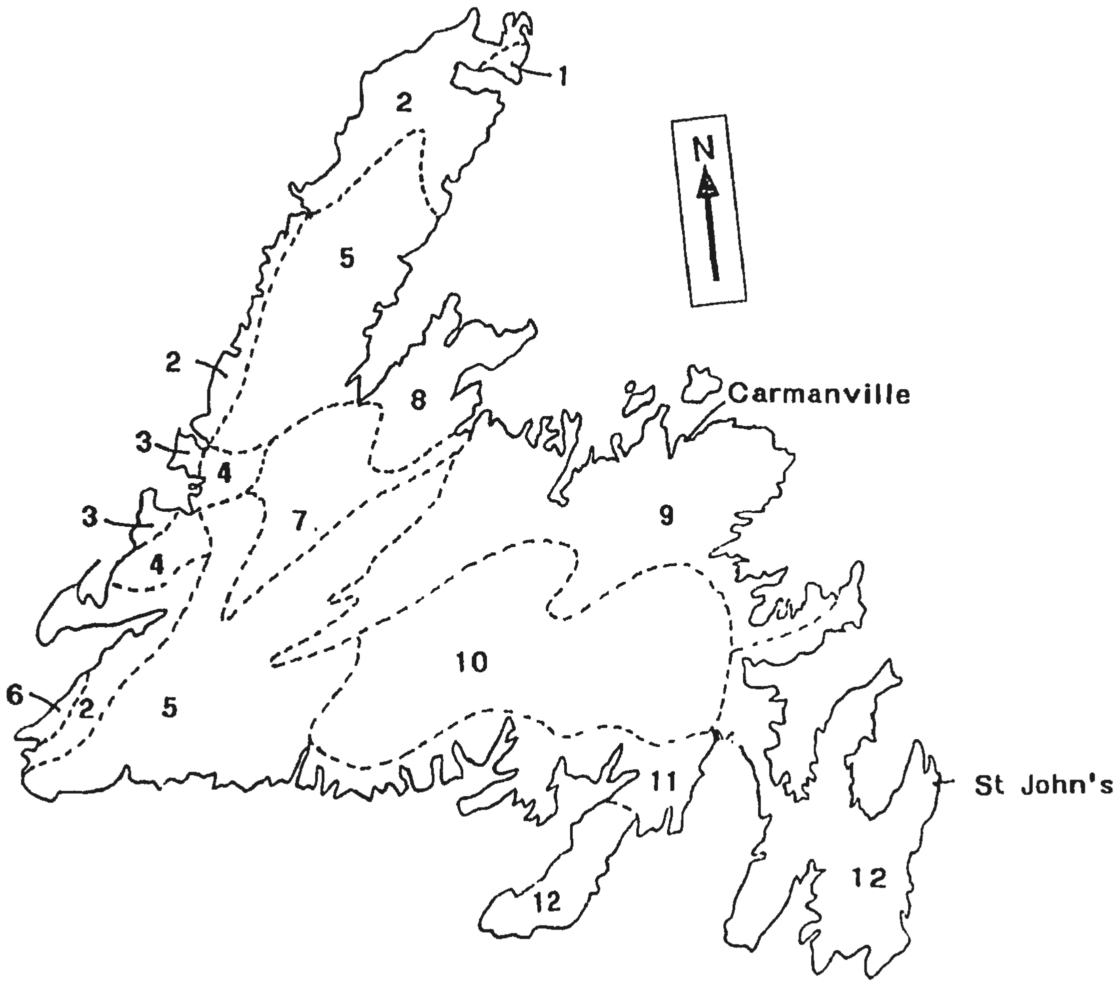
Granitoid rocks in the area are generally petrographically and chemically similar (Currie, 1992). Five major plutons crop out in the area (Figure 1.2). These are the Rocky Bay, Frederickton, Aspen Cove, Ragged Harbour, and Island Pond plutons. Small parts of the Tim's Harbour and Deadman's Bay plutons also crop out. The Rocky Bay and Frederickton plutons are composed of massive biotite granodiorite. The Ragged Harbour and Aspen Cove plutons are formed of biotite granite, muscovite pegmatite, and quartz-rich muscovite-garnet pegmatite. The Island Pond Pluton is a pink massive leucogranite. The Deadman's Bay Pluton is a coarse-grained biotite granodiorite with potassium-feldspar megacrysts.

Clasts from the Gander River Complex are especially useful in clast provenance. Mafics and ultramafics from this group are easily recognisable and minor occurrences of these at some distance either side of the Complex, would indicate glacial transport. Also of use are clasts from the Barry's Pond Conglomerate which have a distinct dark green colour and contain pebbles from the Gander River Complex.

1.5 - PHYSIOGRAPHY

The Carmanville map area lies in the "Northeast Trough" (Zone 9) of Twenhofel and MacClintock's (1940) physiographic classification for the Island of Newfoundland (Figure 1.3). It is characterized by low relief, with maximum elevations reaching 120m asl. This gently undulating zone extends offshore, where Gander Bay rarely exceeds 5m in depth. Hamilton Sound reaches maximum depths of 40m, but depths rarely exceed 20m (Shaw *et al.*, 1990). Long, low-lying elongate ridges cross the map area with southwest-northeast trends and reflect the underlying bedrock geology (Plate 1.1).

Approximately 35% of the map area is covered by marine waters, and coastal processes therefore play a major role in the region. Most of the coastline is bedrock-controlled with low-lying, rocky cliffs. In areas that face the prevailing northeast waves, the coastline is sediment dominated and beach formation is



1. Serpentinized Hills of Hare Bay
2. West Coast Lowland
3. Bay of Islands Serpentinized Range
4. West Coast Calcareous Uplands
5. Long Range Mountains
6. Anguille Mountains
7. Grand Lake-White Bay Basin
8. Burlington Peninsula
9. Northeast Trough
10. Central Plateau
11. South Coast Highlands
12. Eastern Upland

Figure 1.3: Physiographic classification for the Island of Newfoundland (after Twenhofel and MacClintock, 1940).

Plate 1.1: General landscape in the Carmanville region looking north to Ladle Cove. The lowest scarcely vegetated areas are boggy terraces which are remnants of a higher sea-level. The area is generally low-lying and flat. The vegetated areas at the immediate coast are relict beach ridges composed of gravel. The vegetated area in the foreground is a bedrock ridge with a veneer of till. This has a southwest to northeast trend, which is reflective of the structural trends of the underlying bedrock. This photograph was taken in late June when icebergs are dominant.

common. Such areas include parts of Ladle Cove and Aspen Cove, the eastern side of Noggin Hill, and the western side of Gander Bay. Areas which face northwest, away from the local wave direction, are generally devoid of vegetation, with low-lying cliffs. Cliffs are especially prominent on the western side of Ladle Cove and the west flank of Noggin Hill.

The Gander River flows from the southwest into Gander Bay, which is sheltered from wave action. Ragged Harbour River flows from the southwest to the northeast across the map area into Raggéd Harbour. It is a small meandering river, with a flood plain up to 600m wide.

Approximately 15% of the Carmanville map sheet is covered by ponds and 15% by wetland. Wetlands are mainly dominated by sloping fens. There are also some string bogs and aapa.

1.6 - CLIMATE

The Newfoundland climate is classified as cool boreal with humid and sub-humid moisture regimes (Agriculture Canada, 1976). The Carmanville region lies within the "East Coast and Hinterlands Zone" (Banfield, 1981). This extends from the Sops Arm area on the southern part of the Northern Peninsula and includes the northern part of the Baie Verte Peninsula, the Twillingate area, Lewisporte, Gander Bay, Gander, Gambo, and Clarenville. It is characterized by a moderate level of precipitation (1100-1500mm per year) although it has also been described as one of the driest areas on the island (Sudom and Van de Hulst, 1985). Winters tend to be cold with 50-70% of the annual precipitation falling as snow. Spring is late, with sea ice persisting until mid-May. Summers are relatively warm and sunny.

The ocean is the strongest climatic influence on the Island of Newfoundland. The Labrador Current has a major effect on the Carmanville area. Coastal areas are cool and experience a highly variable daily microclimate (Sudom and Van de Hulst, 1985). Prevailing winds are generally southwesterly and have come from overland. There is, however, significant wind activity from the northwest, north, and northeast as recorded at Fogo and Gander (Sudom and Van de Hulst, 1985). Although there are no weather stations in the study area, the three stations nearby (Fogo, Gander, and Twillingate) provide climatic data for this area. Table 1.1 shows some of the

mean annual precipitation	1181.6mm
mean annual rainfall	737.9mm
mean annual snowfall	443.8cm
mean January temperature	-6.8°C
mean maximum January temperature	-2.4°C
mean minimum January temperature	-9.8°C
mean July temperature	16.3°C
mean maximum July temperature	21.9°C
mean minimum July temperature	11.1°C
prevailing winds (summer)	southwest
prevailing wind (winter)	west
average date of last spring frost	June 1 -June15
average date of first fall frost	~October 15

Table 1.1: Climatic data from Gander International Airport (1965-1990) (Banfield, 1981; Atmospheric Environment Service, 1993).

basic climatic data collected at Gander based on observations between 1965 and 1990.

Any region subjected to lengthy periods of frost activity will be affected by frost heave. The Carmanville region is only frost free for approximately four months of the year (Table 1.1). Over time, frost heave can result in the re-orientation of clasts in the frost-heave zone to a near-vertical position (Mackay, 1984; Anderson, 1988). Care therefore must be taken in this region to avoid taking clast fabrics near the surface of the ground, as these may have been altered substantially.

1.7 - SOILS AND VEGETATION

Soils on the island of Newfoundland have been divided into twelve main assemblages (Sudom and Van de Hulst, 1985), related to the physiographic classification of Twenhofel and MacClintock (1940). The North East Trough is generally characterized by Orthic Regosols and Gleyed Eutric and Dystric Brunisols developed along the rivers. Humo-Ferric and Ferro-Humic Podzols and Orthic Gleysols are developed on morainal deposits, and Typic Fibrisols and Typic Mesisols occur on domed bogs (Roberts, 1983). The mapping completed by Sudom and Van de Hulst (1985), indicates that regosols, gleysols, and mesisols are insignificant in the Carmanville area.

Podzols are common, covering approximately 65% of the Botwood and Wesleyville area. Humo-Ferric and Ferro-Humic Podzols occur most commonly. These are frequently gleyed if developed on deposits with a moderate clay content, but are usually associated with coarse-textured till, especially moraine crests where Orthic and Ortstein Humo-Ferric Podzols dominate. The vegetation associated with these soils is balsam fir/black spruce/alder. Grasses occur near Aspen Cove (Sudom and Van de Hulst, 1985).

Typic Fibrisols and Mesic Fibrisols cover approximately 25% of the map area. They constitute thick, undecomposed peat units derived from *Sphagnum* mosses. These are almost entirely developed on flat, sloping, or blanket bogs. Fibrisols are most common at Ladle Cove, Aspen Cove, Carmanville, and south of Island Pond (Sudom and Van de Hulst, 1985).

Brunisols occur to the west of Island Pond and cover less than 5% of the total study area. They are generally Gleyed Eluviated Dystric Brunisols and occur commonly adjacent to Gleyed Humo-Ferric Podzols. The dominant vegetation associated with this soil type is balsam fir (Sudom and Van de Hulst, 1985).

Near Ragged Harbour, Typic and Terric Humisols cover less than 5 percent of the map area. These generally overlie marine deposits or bedrock. The vegetation associated with these soils consists of dense ericaceous shrubs, grasses, sedges, and alders (Sudom and Van de Hulst, 1985).

An understanding of the soils and vegetation needs to be gained prior to undertaking aerial photograph interpretation. Vegetation can often obscure the underlying sediment and thus it is imperative that fieldwork be undertaken in any area, using aerial photograph interpretation only as a preliminary tool.

CHAPTER 2
PREVIOUS WORK

2.1 - INITIAL RESEARCH

The number of glacial events which have affected the Island of Newfoundland is unknown. MacClintock and Twenhofel (1940) proposed that in the absence of evidence to the contrary, any glaciation on the Island can be assumed to be Wisconsinan in age. This assumption has been adopted by many subsequent glacial workers. Some of the early workers believed the Island to be partially ice-free during the Wisconsinan (e.g. Fernald, 1911; Coleman, 1926), whereas others suggested that it was completely overrun by Laurentide ice (e.g. Flint, 1940; Tanner, 1940). Daly (1921) was the first to propose that the Island supported its own ice cap during the Late Wisconsinan, a concept agreed to by many subsequent workers such as Murray (1955) and Jenness (1960). Coleman (1926) proposed that those parts of the Island that were glaciated were only covered by small ice sheets and valley glaciers, rather than by one large ice cap. Murray (1955) expanded on this hypothesis and proposed several ice centres for the central part of the Island with radial spreading of the ice towards the coast.

Jenness (1960) set the trend of much of the work to follow on the Island of Newfoundland. He divided the Island into an "outer drift zone" and an "inner drift zone", based on geomorphology and physiography. These zones are separated by a large discontinuous end moraine between 30 and 50 km inland from the coast in eastern Newfoundland. The "outer drift zone" is characterized by cirques, fiords, kames, and thin till cover, whereas the "inner drift zone" is characterized by thicker till cover, morainic lakes, fluted terrain, kames, and eskers. The Carmanville area is situated in Jenness' "outer drift zone". A small moraine that Jenness believed to represent an ice retreat position is present at Shoal Pond, east of Carmanville.

Jenness also mapped many striations in northeast Newfoundland that indicated the existence of a major ice divide across the Island. These striations indicated north to north-northeastward ice-flow in the Springdale and Bay of Exploits area, north to north-northwestward ice flow in the Birchy Bay area and on Fogo Island, northeastward ice flow in the Carmanville area, and eastward ice flow in the Eastport, Gambo, and Bonavista Bay areas. At the tip of Bonavista Peninsula, ice movement was towards the coast.

A further contribution of Jenness to the study of the glaciation of the Island of Newfoundland involved assessment of isostatic rebound. He combined much of his data on raised deltas around the coast with the data of Flint (1940) and produced a map (Figure 2.1) indicating that isostatic rebound increased to the northwest. His zero isobase passes through Trinity and Placentia Bays, and generally follows the south coast, indicating that some parts of the Island are submerging while others are emerging.

2.2 - SOURCES OF GLACIAL ICE AND ICE FLOW DIRECTIONS

Although some of the earlier workers hypothesized that the Island of Newfoundland was completely overrun by the Laurentide Ice Sheet (e.g. Flint, 1940; Tanner, 1940), most workers have since indicated that during the Late Wisconsinan the Island supported its own ice cap(s) (e.g. Daly 1921; Murray, 1955; Jenness, 1960; Grant, 1974, 1977, 1989; Brookes, 1982; Rogerson,

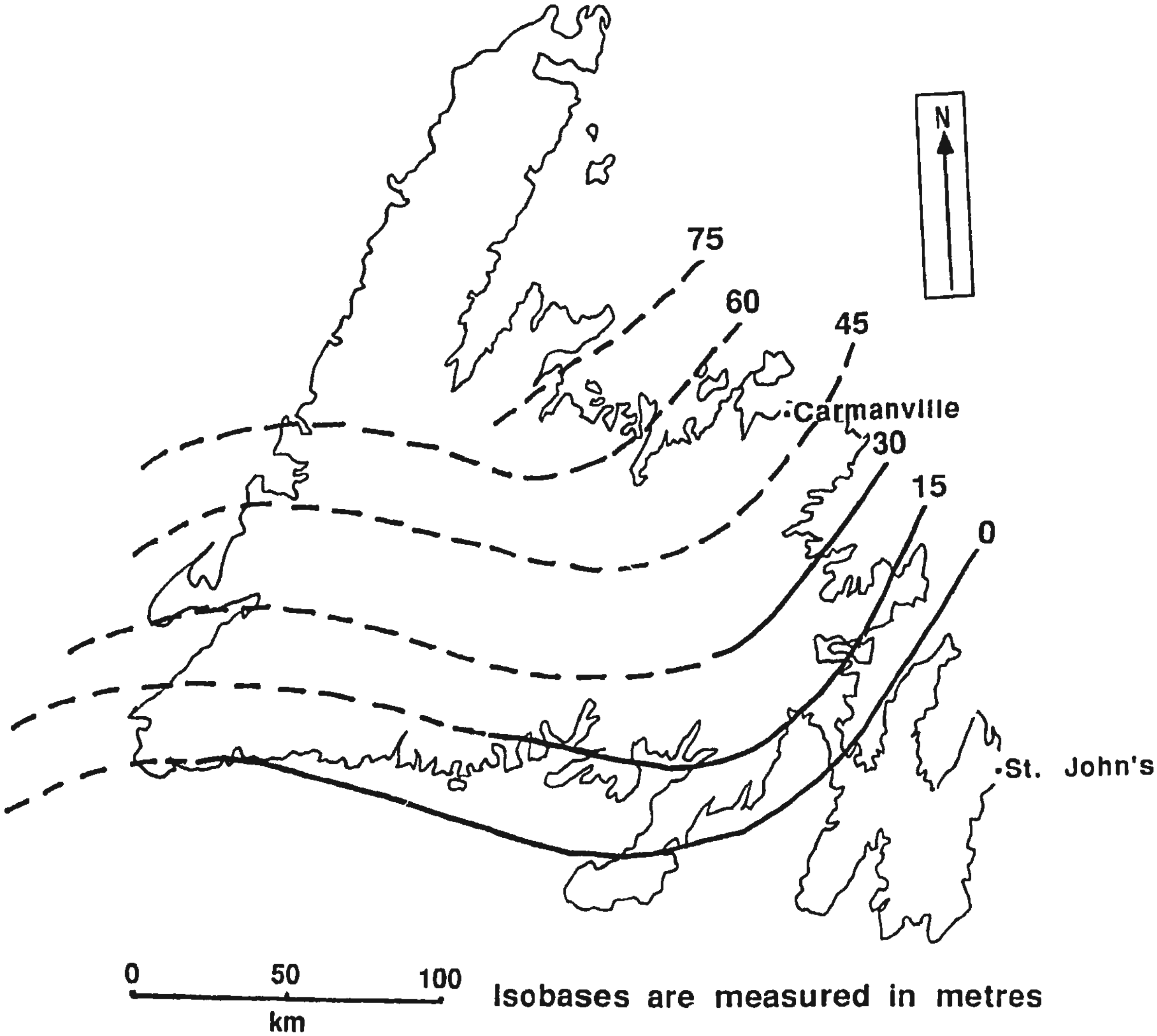


Figure 2.1: Isobases across Newfoundland based on the upper levels of deltas (after Jenness, 1960).

1989). Only on the very tip of the Northern Peninsula are found erratics derived from the Canadian Shield, suggesting that this area was covered by Laurentide ice (Grant, 1989). The sources of glacial ice were determined by most workers by mapping striations and determining the provenance of clasts and minerals in glacial deposits (e.g. Brookes, 1982, 1989; Vanderveer and Taylor, 1987; Liverman and St. Croix, 1989; Liverman and Scott, 1990; St. Croix and Taylor, 1990, 1991; Batterson and Vatcher, 1991; Liverman *et al.*, 1991; Klassen and Henderson, 1992). Many of the most recent projects have been undertaken by the Newfoundland and Labrador Department of Mines and Energy in the course of mapping at a 1:50,000 scale.

Grant (1977, 1989) recognized three distinct glacial accumulation and dispersal zones on the Island of Newfoundland: the Avalon Peninsula, the Northern Peninsula, and the Central Newfoundland Uplands. These areas are characterized by either shallow immature drift or regolith, or by thick till blankets. Many of the thicker deposits are fluted or ribbed, reflecting radial dispersal from these centres. As Jenness (1960) did, Grant (1989) noted that thinner till is peripheral to the thicker deposits.

A recent comprehensive study of ice movement in northeast Newfoundland was undertaken by St. Croix and Taylor (1991). They found evidence of at least three ice-flow events (Figure 2.2). The oldest of these was an eastward flow, which was also detected by Batterson and Vatcher (1991) in the Gander region, and by Liverman *et al.* (1991) and Liverman and Scott (1990) in the Baie Verte region. St. Croix and Taylor hypothesized that ice from the north central Uplands coalesced with ice from the Northern Peninsula resulting in a deflection of ice flow north-central Newfoundland to the east. This

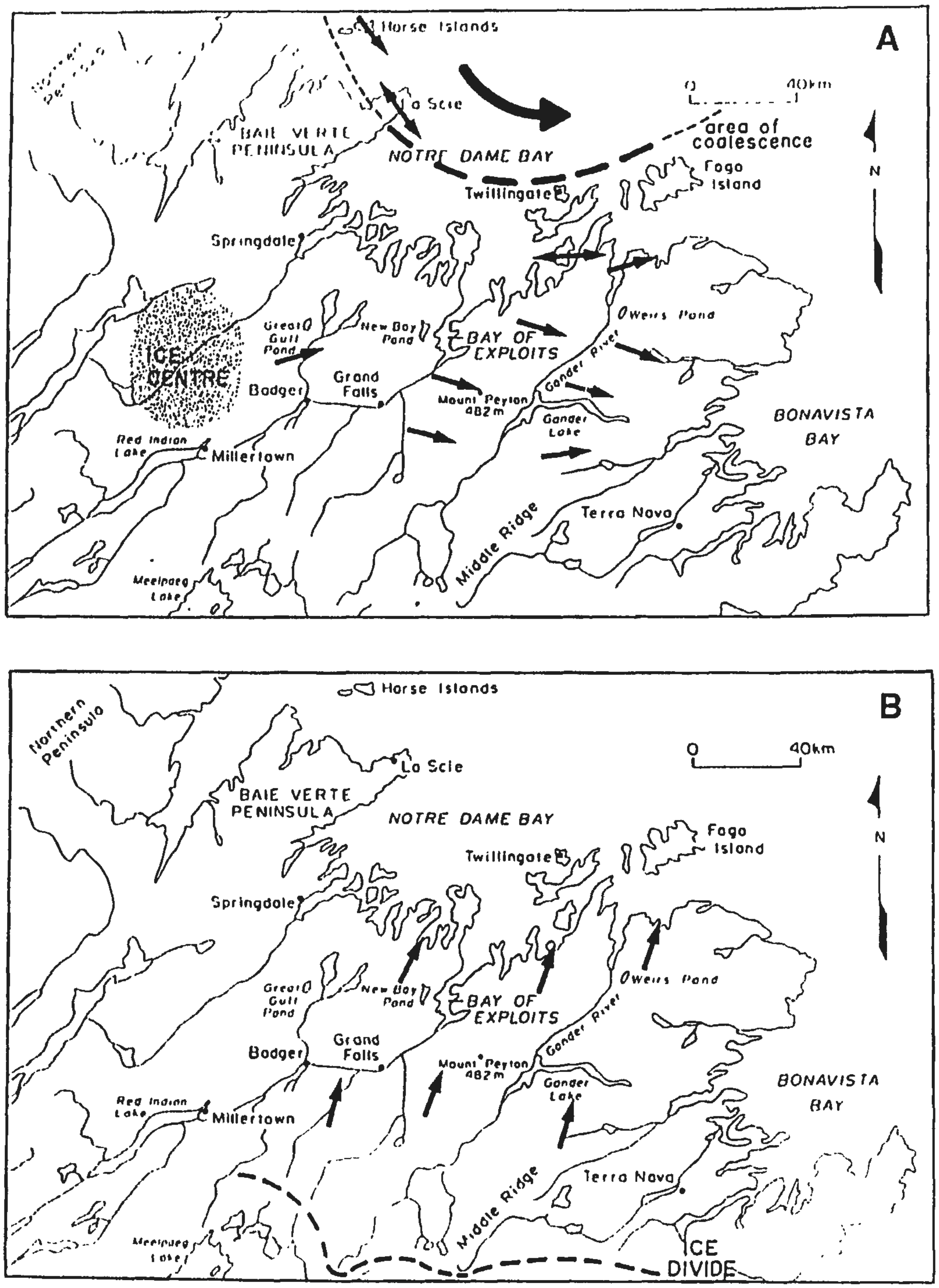


Figure 2.2: The three main Late Wisconsinan ice-flow events that affected Northeast Newfoundland. A is the first ice-flow event, B is the second, and C is the third. (after St. Croix and Taylor, 1990)

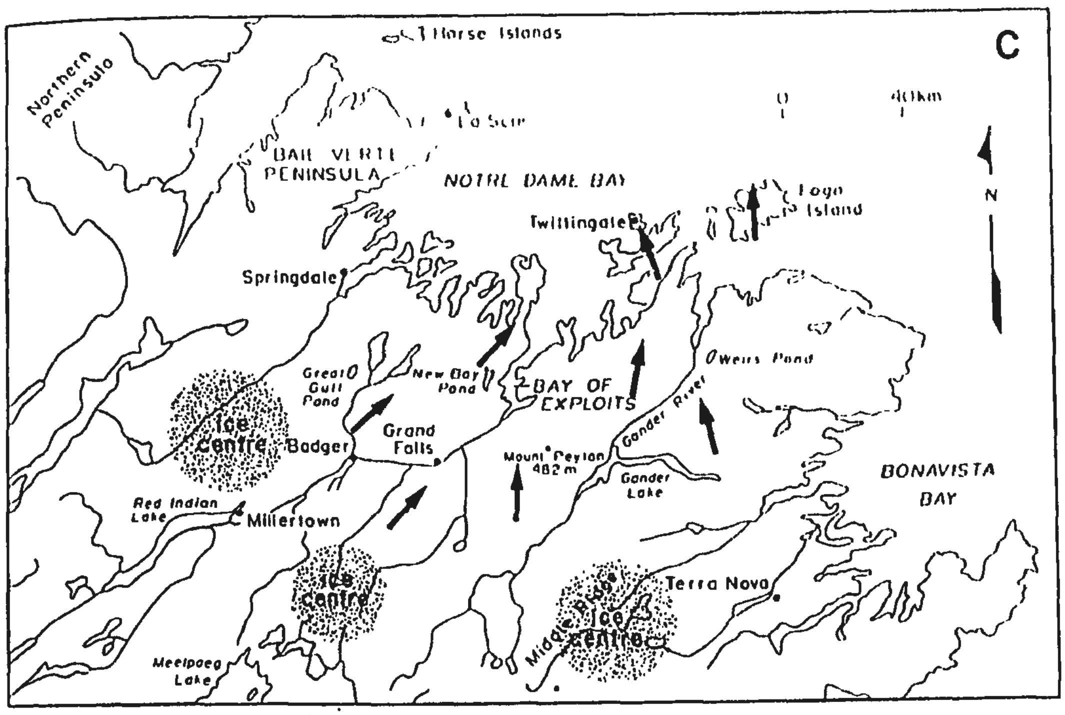


Figure 2.2: (continued).

hypothesis was supported by the observations of Liverman (1992), who noted that southeastward trending striations at the north of the Baie Verte Peninsula, and others on the islands off the Baie Verte coast, could only have formed by ice originating on the Northern Peninsula.

The second flow noted by St. Croix and Taylor (1991) was northeastward. This flow is by far the most commonly recognized in northeast Newfoundland and was also detected by Baird (1950), Hayes (1951), Jenness (1960), Lundqvist (1965), Hornbrook *et al.* (1975), Vanderveer and Sparkes (1980), Brookes (1982), Vanderveer (1985; 1987), Vanderveer and Taylor (1987), Taylor and St. Croix (1989), and Batterson *et al.* (1991).

The third event recognized by St. Croix and Taylor involved at least three ice centres: in the Red Indian Lake area, southeast of Millertown, and in the

Terra Nova Uplands. Flows from these centres were dominantly northwestward. St. Croix and Taylor (1991) also mapped striations younger than those of this northwestward flow and interpreted these to represent subsequent ice-flow from many small remnant ice centres. They suggest that these may represent Younger Dryas glacier movement.

During the Late Wisconsinan, the easternmost section of the Bonavista Peninsula was glaciated by its own ice cap, indicated by the presence of coastward-trending striations (Brookes, 1989). Thus the eastward and northeastward ice-flows recorded in central Newfoundland did not affect the Bonavista Peninsula.

Previous mapping in the Carmanville (NTS 2E/8) map area by Kirby *et al.* (1988) indicated that the area has many striations with general northeastward trends. These probably relate to St. Croix and Taylor's (1991) second ice flow event.

2.3 - EXTENT OF GLACIAL ICE

During the past one hundred years, many authors believed the Island of Newfoundland to be partially ice-free during the Wisconsinan glacial period (e.g. Fernald, 1911; Coleman, 1926; Grant, 1977; Brookes, 1977; Rogerson, 1982). The evidence is in the form of hypothesized nunataks on the west coast, the Burin Peninsula, the easternmost extent of the Bonavista Peninsula, the Gander Peninsula, and the tips of the headlands around the Avalon Peninsula (Grant, 1977; Brookes, 1982). Grant (1989) further modified this idea and termed it the "minimum concept". Only the very tips of the Bonavista Peninsula were unglaciated, more of the Avalon and the west coast was covered by ice, and the Gander Peninsula[1] was completely inundated with ice up to 75km offshore.

Grant (1989) also summarized the evidence for the "maximum concept", which hypothesizes that the entire Island was covered by glacial ice. This is primarily based on the extent of deposits offshore and that there were no unglaciated areas on the Island. Those unglaciated areas previously designated under the minimum model are considered to represent cold-based glaciation where no erosion or till deposition took place (cf. Denton and Hughes, 1981).

Investigation of the Baie Verte Peninsula by Liverman (1992) indicated that this entire region was covered by Late Wisconsinan ice, and hence here, the minimum hypothesis is incorrect. Shaw (1991) also indicated that glacial diamict is present in the region offshore of the Baie Verte Peninsula.

Bonavista Bay was occupied by a grounded ice sheet which extended some distance offshore (Piper *et al.*, 1990; Cumming *et al.*, 1992; Fader and Miller, 1994). Denton and Hughes (1981) and Mayewski *et al.* (1981) suggested that almost all of the continental shelf was inundated by glacial ice, and King and Fader suggest that most of the offshore area was occupied with ice between 29,500 and 15,000 BP. Cumming *et al.* (1992) produced a deglacial sequence of events (Figure 2.3), which had all of the Bonavista Bay area (apart from the nunatak noted by Brookes (1989)), inundated by glacial ice at the Late

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[1]In the literature there is no consistency in the naming of the area of land between Gander Bay and Bonavista Bay. Some authors name it the "Gander Peninsula" (e.g. Cumming *et al.*, 1993), while other authors name it the "Wesleyville Peninsula" (e.g. Brookes, 1989). To avoid confusion this area of land will be referred to as the Gander Peninsula in this thesis.

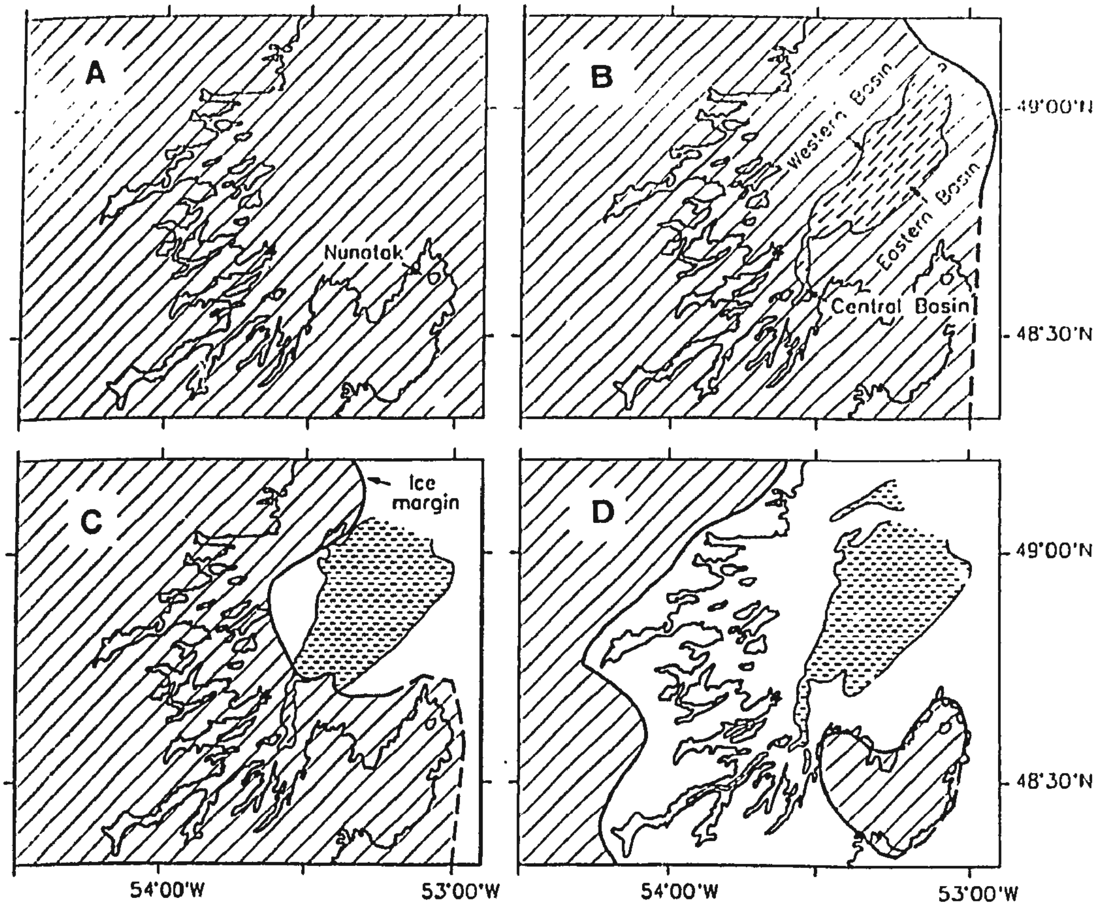


Figure 2.3: a) the extent of the Late Wisconsinan ice in Bonavista Bay: b) ice-shelf development and the location of a buoyant ice-cover (oblique broken lines); c) Recessional ice margin, proglacial deposition (horizontal broken lines); d) open marine conditions (after Cumming *et al.*, 1992).

Wisconsinan maximum. Deglaciation was rapid and occurred prior to 13,500 BP. By 12,790 BP all of Bonavista Bay was ice-free.

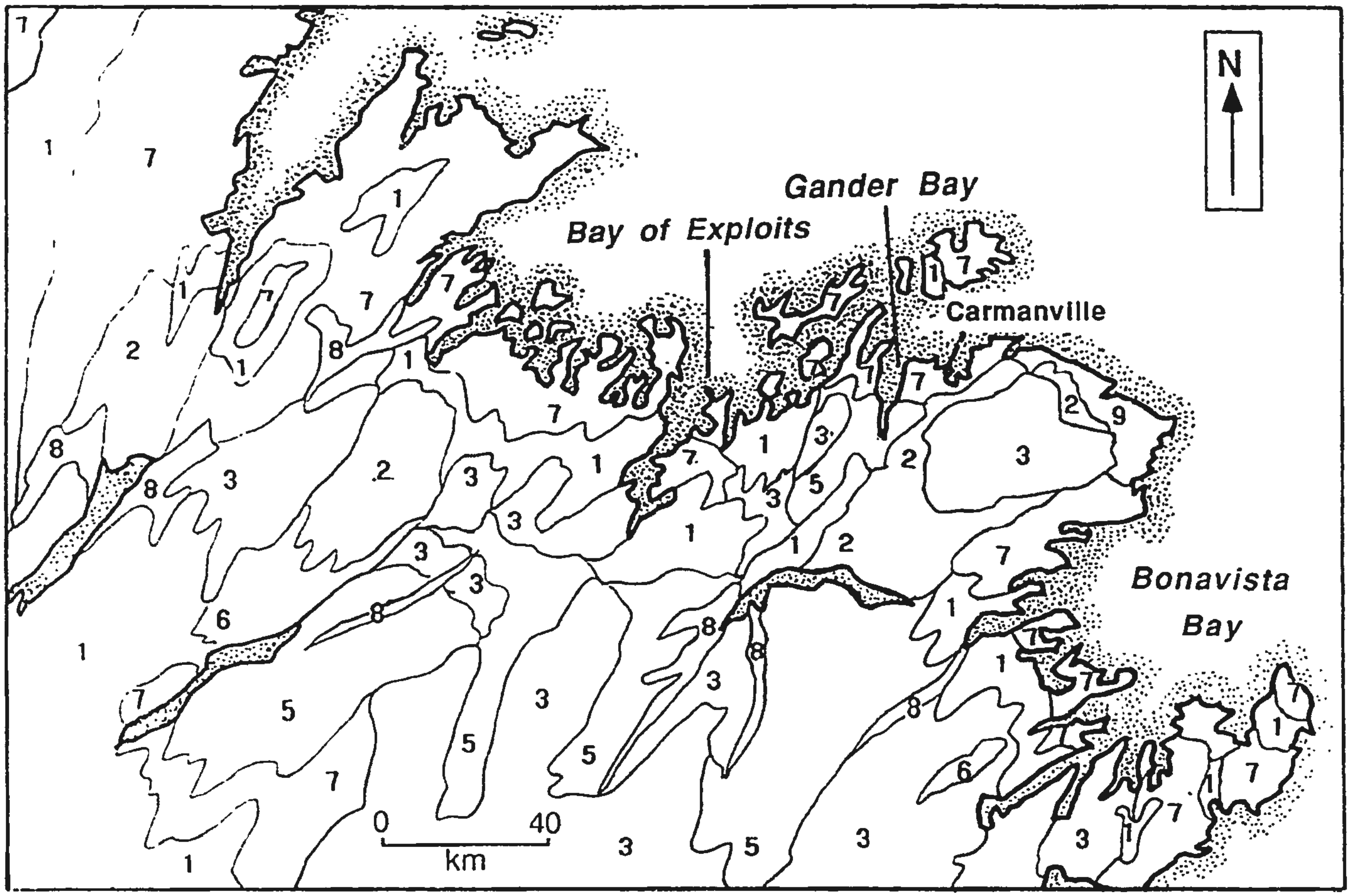
On land, all of the Bonavista Peninsula was glaciated apart from a nunatak mantled by felsenmeer (Brookes, 1989). The inner western area of the peninsula was dominated by eastward flowing ice (as indicated by striations and oriented bedforms) which probably originated from the North Central Newfoundland ice cap. According to Brookes, the outer eastern part of the Bonavista Peninsula was glaciated by a dynamically independent but contiguous ice dome with a major ice divide (the Bonavista Ice Divide) which constituted mainly coastward flowing ice (Brookes, 1989).

Under either scenario proposed by Grant (1989), the entire Carmanville region, the area of the present study, was covered by glacial ice at the Late Wisconsinan maximum. The striations mapped by Kirby *et al.* (1983) suggest this. Striations mapped by Jenness (1960), and St. Croix and Taylor (1991) indicate that ice extended at least as far as Fogo Island. The offshore work of Shaw and Forbes (1990a) indicates that the coastal areas of Hamilton Sound were occupied by glacial ice during the Late Wisconsinan.

2.4 - SURFICIAL GEOLOGY AND GEOMORPHOLOGY

The surficial geology of Newfoundland has been summarized in a map published by Liverman and Taylor (1990). A simplified version of this map for northeast Newfoundland is shown in Figure 2.4. Emerged deltas and marine terraces are very common around the coast in northeastern Newfoundland (e.g. MacClintock and Twenhofel, 1940; Lundqvist, 1965; Tucker, 1976; Brookes, 1989; Liverman and St.Croix, 1989; Liverman and Scott, 1990; Shaw and Forbes, 1990a; Scott *et al.*, 1991; Scott and Liverman, 1991; MacKenzie, *in preparation*).

The surficial geology of the NTS 2E/8 map area was mapped at a reconnaissance level by Kirby *et al.* (1988). Hummocky terrain dominates the area, along with thin eroded till veneers overlying bedrock. Drumlinoid forms



1. Till veneer
2. Hummocky till
3. Ridged till
5. Till Blanket
6. Organic units
7. Bedrock
8. Glaciofluvial sands and gravels
9. Marine units

Figure 2.4: Simplified surficial geology map of Northeast Newfoundland. (after Liverman and Taylor, 1990).

and moulded bedrock features, such as roches moutonnées and crag-and-tails, are also shown and indicate northeastward ice-flow. Striations were mapped and interpreted as being formed by a northeastward ice-flow.

Seismic surveys have been undertaken offshore (Shaw *et al.*, 1990). The substrate is dominantly gravel and bouldery gravel. This is underlain by an acoustic unit with incoherent reflections, interpreted as a glacial diamicton. Drumlin-like forms were detected at the head of Gander Bay oriented south-southwest - north-northeast, and are interpreted as showing north-northeastward ice flow. Sudom and Van de Hulst (1983) noted a large sand and gravel body in the northeast section of the 2E/8 map area and classified this as an esker.

2.5 - POST GLACIAL EVENTS

2.5.1 - The Younger Dryas Cooling Event

Evidence of a severe periglacial climate postdating glacial activity exists at many coastal locations across the Island indicating that these areas were ice free during climatic deterioration. Directly to the west of the Carmanville area, in the Birchy Bay region (NTS 2E/7), Eyles (1977) and Scott (1993) found features that they interpreted as ice wedge casts. Eyles believed that these formed between 12,000 and 10,000 BP. In north central Newfoundland, Liverman and St. Croix (1989), MacKenzie (*in preparation*), and Batterson (1994, *in preparation*) detected features which they interpreted as ice wedge casts. All authors tentatively attribute these to the Younger Dryas cooling event. On the west coast Brookes (1971) also found ice wedge casts, and determined that these formed sometime before 10,600 BP.

The formation of the Ten Mile Lake Moraine was determined by Grant (1969, 1989, 1991) to have formed at approximately 11,000 BP as ice from the Long Range Mountains moved northward into a sea that stood approximately 80m higher than present. Grant (1989) states that this correlates well with the Younger Dryas climatic deterioration. Liverman *et al.* (1991) found evidence of a lateglacial readvance in northcentral Newfoundland and, although there is no numerical dating control, suggest that it may be linked to the Younger Dryas. The probability of these advances being related to the Younger Dryas is enhanced by the well-documented Younger Dryas glacial activity in Nova Scotia (Stea and Mott, 1989).

Palynological successions provide the most convincing evidence of a cooling event after the retreat of the main Late Wisconsinan ice on the Island. Anderson and Macpherson (1994) recognized a cooling episode, and dated it to between approximately 11,000 BP and 10,000 BP on the basis of ^{14}C dates from sites across Newfoundland. Pollen assemblages from this time are generally dominated by herbs, and mineral content is usually high. Anderson and Macpherson attribute this to reduced vegetation cover, and longer lasting snow banks which led to increased slope wash. They indicated that during this time, ice-caps probably persisted or expanded. Wolfe and Butler (1994) provided both pollen and diatom evidence for a climatic deterioration between approximately 11,000 and 10,000 BP at Pine Hill Pond, Terra Nova National Park, approximately 100km southeast of the present study area.

2.5.2 - Relative sea-level change

The pattern of sea level change on the Island of Newfoundland is directly related to the position of the Laurentide Ice Sheet on mainland Canada (Jenness, 1960; Brookes, 1982; Grant, 1989). All studies related to sea level maximum indicate that the marine limit progressively decreases southwestward and southeastward from the Northern Peninsula in accordance with the model first proposed by Jenness (1960). At the tip of the Northern Peninsula, the marine limit is 145m asl (Proudfoot and St. Croix, 1987), at Springdale the marine limit is 75m asl (Tucker, 1976; Scott *et al.*, 1991), in the Bay of Exploits it is at least 65m asl (MacKenzie, *in preparation*), at Eastport it is at least 38.5m asl (Dyke, 1972), and on the Bonavista Peninsula it is at least 26m asl (Brookes, 1989).

Quinlan and Beaumont (1981) developed two sea level models for the Atlantic Provinces based on the minimum and maximum ice hypotheses. Both models relate to the isostatic depression of land under glacial ice and the position of the peripheral forebulge which is forced upwards to compensate for the depression. During and after deglaciation, the depressed areas isostatically rise while the forebulge isostatically depresses as it migrates slowly towards the centre of an ice mass. Quinlan and Beaumont determined that all of Newfoundland lay on the peripheral forebulge regardless of the minimum or maximum ice hypotheses, and the exact position on that forebulge determined the style of sea level change. Four hypothetical curves were developed depending on this position (Figure 2.5).

Using the minimum ice hypothesis, Quinlan and Beaumont placed the Carmanville region in zone D of their model. The region would have originally

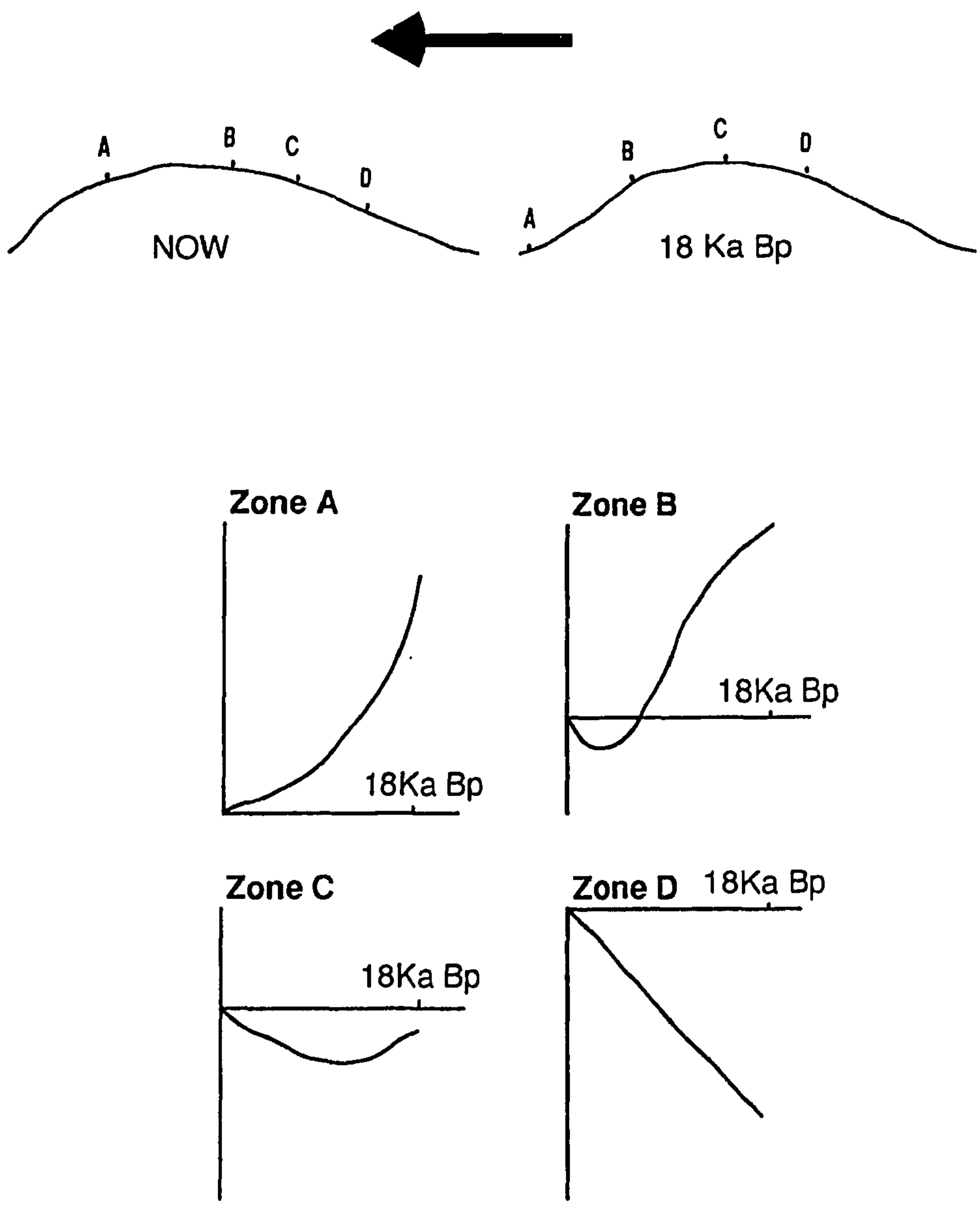


Figure 2.5: Schematic representation of a peripheral bulge at two points in time. The bulge migrates in the direction of the arrow affecting the relative sea level at A, B, C, and D. The sea level curves show the typical pattern of relative sea-level change at each site (after Quinlan and Beaumont, 1981).

been on the outer edge of the forebulge. Such areas are characterized by sea-levels which have constantly risen to the present level (Figure 2.5). Using the maximum ice hypothesis, Quinlan and Beaumont placed the Carmanville region in zone C of their model. In this scenario, Carmanville would have initially been placed on the top of the forebulge. This zone is characterized by relative sea-levels which fluctuated below 0m asl and subsequently rose to present levels (Figure 2.5).

Grant (1989) suggested a high stand of sea level at 40m asl for the Gander Bay region and dated this at approximately 12,500 years BP. Shaw and Forbes (1990a) incorporated this value into a sea-level curve developed for the Cape Freels area (Figure 2.6). They suggested that sea level dropped rapidly between 12,500 and 8,000 years BP, from an elevation of at least 40m to perhaps as low as -25m asl. Shaw *et al.* (*in review*) proposed a minimum level of -17m asl in Gander Bay some time prior to 8600 BP. Sea level has been rising slowly since. This sea-level fluctuation represents a Zone B curve using the terminology of Quinlan and Beaumont (1981). Liverman (1993) indicates that most of the Island has a type B , or modified type B, sea-level curve, with the age of the transition of sea levels below present increasing to the south. The theories of Quinlan and Beaumont (1981) are not discredited, but it is likely that the forebulge affected a much larger area than originally anticipated by these authors.

2.6 - CHRONOLOGY OF EVENTS

All glacial events detected in northeast Newfoundland are assumed to be Late Wisconsinan. These are listed in Table 2.1. It is clear that there is a

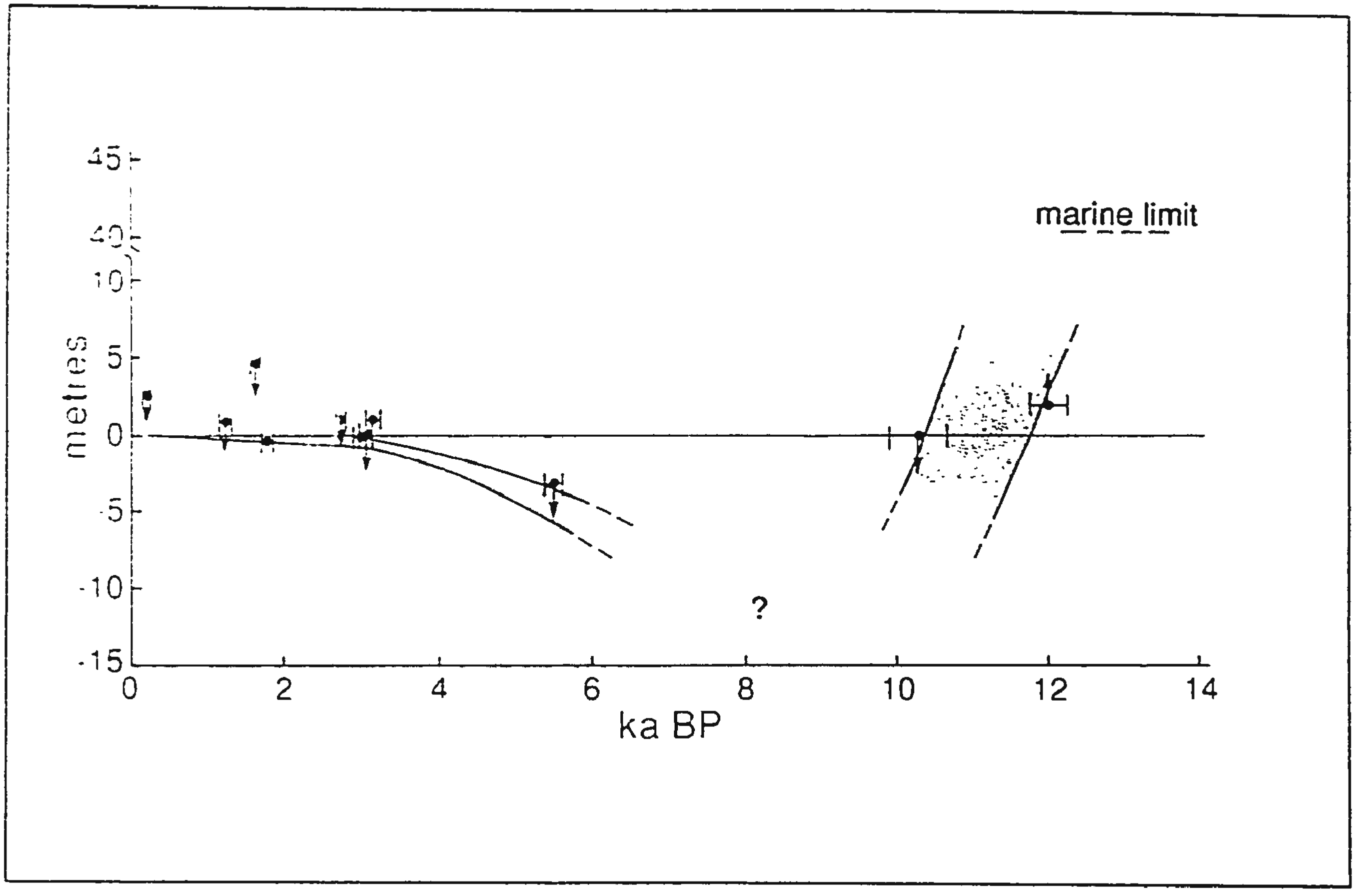
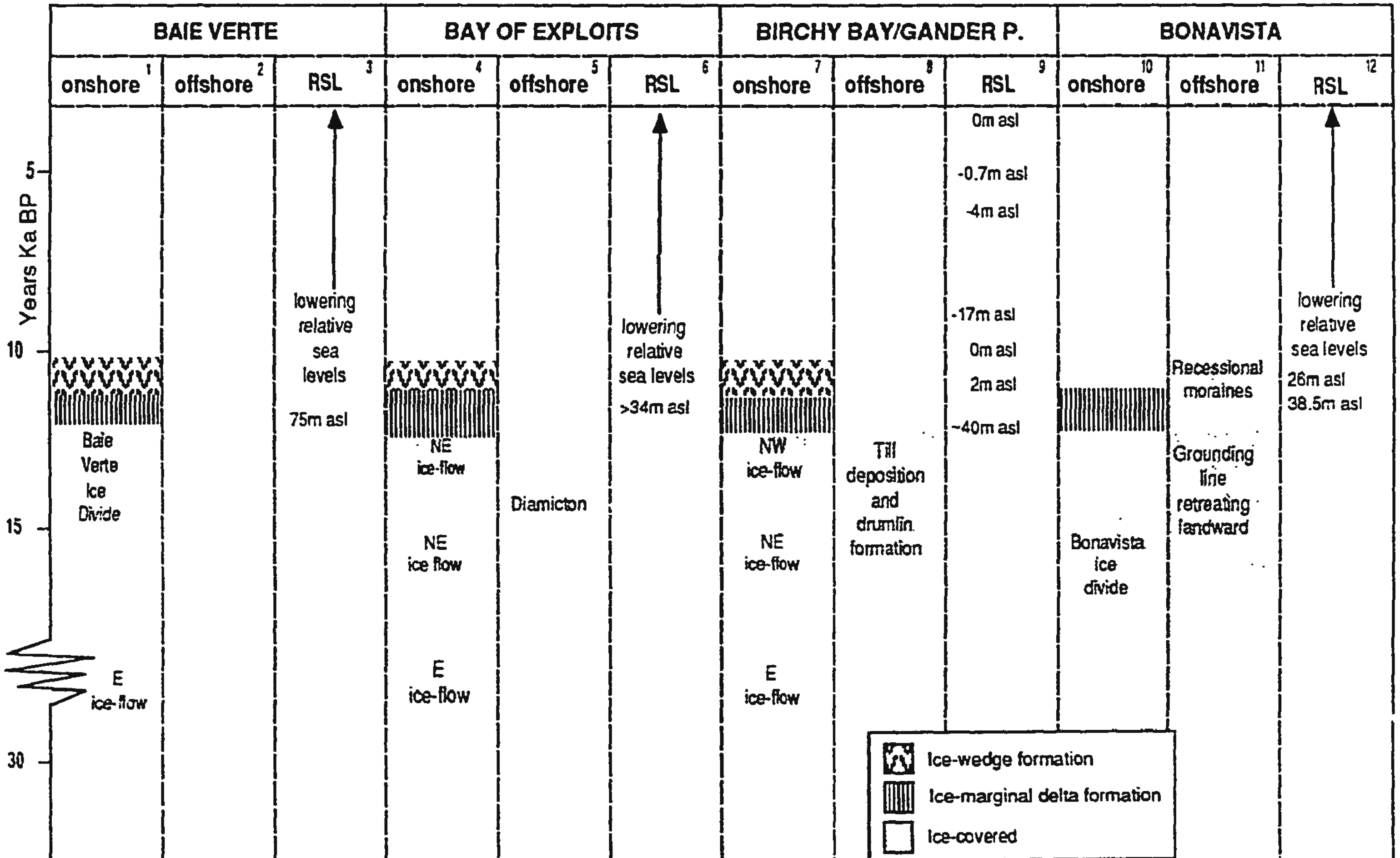


Figure 2.6: Sea level curve developed for northeast Newfoundland (after Shaw and Forbes, 1990a)



1. Liverman and St.Croix (1989)
   Liverman and Scott (1990)
   Scott *et al.* (1991)
   Liverman (1992)
2. Shaw (1991)
   Shaw *et al.* (1991)
3. Scott *et al.* (1991)
4. MacKenzie (*in preparation*)
   St. Croix and Taylor (1991)
5. Shaw *et al.* (1991)
6. MacKenzie (*in preparation*)
7. Scott (1993)
   St. Croix and Taylor (1991)
8. Shaw and Forbes (1990a)
9. Shaw and Forbes (1990a)
10. Dyke (1972)
    Brookes (1989)
11. Denton and Hughes (1981)
    Mayewski and Hughes (1981)
    Piper *et al.* (1990)
    Cumming *et al.* (1993)
    Fader and Miller (1994)
12. Dyke (1972)
    Brookes (1989)

Table 2.1: Chronology of events in northeast Newfoundland based on previous work in the area

correlation between specific events across the area. The Late Wisconsinan maximum occurred between 28,000 and 20,000 BP. During this time almost all of northeast Newfoundland was covered by ice apart from a rocky plateau mantled by felsenmeer at the tip of the Bonavista Peninsula. This ice was centred on the north-central Uplands and flow was predominantly eastward. Only on the easternmost extent of the Bonavista Peninsula were no eastward striations detected.

The onset of deglaciation resulted in the large ice caps shrinking and the formation of an independent ice-divide across central Newfoundland. This resulted in northward to northeastward ice flow to beyond the modern coastline in most areas in northeastern Newfoundland. Further deglaciation occurred which resulted in the formation of several smaller ice caps. This ice-flow event is associated with higher variability in ice flow directions (N-NNW in the Carmanville region, NE in the Bay of Exploits). It is likely that the ice-marginal deltas detected at the coast and up to several kilometres inland are related to this event. This probably occurred prior to 12,470 ±380 BP at Springdale (^{14}C age TO-2305) (Scott *et al.*, 1991). ^{14}C ages of 11,500 ±110 BP (GSC-5527) (MacKenzie, *in preparation*), and 11,600 ±210 BP (Blake, 1983) were obtained from glaciofluvial deposits in the Bay of Exploits region. These also provide minimum ages for the marine limit in the area.

Relative sea levels were at their maximum at approximately 12,500 BP. These fell rapidly from the marine limit to 0m asl at approximately 10,000 BP. In Hamilton Sound sea levels are known to have dropped as low as -17m asl at 8,600 BP and have since slowly risen to subsequent levels (Shaw *et al.*, *in review*).

CHAPTER 3
METHODOLOGY

3.1 - AERIAL PHOTOGRAPH INTERPRETATION

Aerial photograph interpretation was undertaken on vertical black and white photographs (1966) between September 1991 and April 1992, prior to any field work. This is an important preliminary step as it allows one to identify many landforms such as roches moutonnées, crag-and-tail hills, drumlins, emerged beach terraces, etc. Such areas become prime targets for further investigation when undertaking field work.

The tone of the photographs is especially important when determining the sediment type. For instance fine sediment has poor reflectibility and thus appears very dark on photographs. Marine or fluvial silts and clays, in contrast, appear dark grey and uniform on aerial photographs. Well-drained coarse sediments are highly reflective, and are therefore represented by light grey/white tones. The darker the grey tones, the poorer the drainage, and thus the higher the percentage of fine-grained sediment. Of course, vegetation can often act to conceal the underlying sediment, but can also aid in the interpretation of that sediment. For instance, the presence of many pine trees would indicate a well-drained sandy soil.

3.2 - FIELD METHODS

Reconnaissance work was undertaken In June 1992, prior to any detailed fieldwork. All roads were traversed by vehicle, or on foot.

Detailed fieldwork was undertaken between June and August 1992, with additional visits to the field in June 1993 and August 1993. Natural sections such as river banks and coastal exposures, and artificial exposures such as road cuts and gravel pits were located and described. Where sections were small and composed of few units, the exposure was examined as a whole. Where sections were large and complex (such as in gravel pits), several vertical transects were designated and described in detail. Each section was divided into units where appropriate. For each sedimentary unit the thickness, colour, matrix texture, sedimentary structures, clast lithology, clast mineralogy, clast shape, and (where appropriate) clast fabric were recorded. Striated and bullet-shaped clasts was also noted. Fabric analysis involved measuring the orientation and dip of the A-axis of 25 clasts in each diamicton unit using a compass with a clinometer. Clasts chosen for measurement had an A-axis to B-axis ratio of 3:2 (Catto *et al.*, 1987).

All linear landforms were located and described, and their orientations were measured using a Brunton or Silva compass. Orientations were often measured from aerial photographs. The features include: striations, roches moutonnées, crag-and-tails, drumlins, flutings, rogen moraines, and lineated till features.

The heights of all marine terraces, erosional platforms, and marine sedimentary exposures were measured using an altimeter (Wallace Tiernan: Model FA-181220). Readings were taken on calm days to avoid inaccuracies due to pressure changes. Reference points of known elevation, in the form of ordnance datum levels or bench marks are usually used for calibration, but no such points have been established in the area. All elevation references were therefore taken from mean sea-level. An altimeter reading was taken at sea-level before and after each reading on land to avoid inaccuracies. The values are considered accurate to within 5m.

3.3 - LABORATORY METHODS

Laboratory analyses was undertaken between September 1992 and January 1993. Analyses included grain size, clast lithology, roundness, and sphericity.

3.3.1 - Grain size analysis

Grain size distributions were described using the Wentworth-Udden grain size classification (Udden, 1898; Wentworth, 1922) and the phi (ø) scale of Krumbein (1934). The Wentworth and Udden scales are both logarithmic (to the base 2) so that each grade limit is twice the diameter of the limit directly below it. The grades are directly linked to the grain diameter measured in millimetres. Krumbein's phi scale shows the gradation to be the negative log to the base 2 of the diameter in millimetres.

Grain size analysis was performed using both hydrometers and sieves. This was undertaken on the matrix component of a sample only, where granules (-1ø), are the coarsest grain size. Thus all samples were sieved through a number 10 (-1ø) sieve prior to analyses. A visual estimate of the coarse component (> -1ø) was estimated while in the field, using the charts of Shvetsov (1954).

A visual estimate of the particle size was determined before grain size analysis. Those samples which had large quantities of silt and clay were analyzed by hydrometer and then sieved (Figure 3.1). Those samples dominated by sand and granules were sieved only, unless more than 10 grams of sediment were left in the sieve pan (<4ø). In these cases the whole sample would be first subjected to hydrometer analysis and sieving. The methods used are as outlined in Catto *et al.* (1987), and originally proposed in ASTM (1964).

Suspension analysis is based on the theory that clasts suspended in any fluid will settle out. The rate of the particle settling depends on the shape, size, and specific gravity, with larger, denser clasts settling more rapidly than lighter, smaller clasts. The hydrometer method adapts Stokes' Law, which states that a spherical particle settling at a uniform velocity will encounter a resisting force due to the liquid, and thus all particles are assumed to be spheres. A hydrometer measures density changes in suspension as settling occurs (Lindholm, 1987).

For each unit dominated by granules, sands, and silts, at least 100 grams were sieved through a 2mm (-1Ø) sieve (Figures 3.1a and 3.1b show the methods adopted while undertaking this analysis). Fifty grams of the sieved sediment were soaked overnight in 125ml of 4% sodium hexametaphosphate (Calgon) solution, a dispersing agent. The sample and solution were homogenized in a blender. The dispersed solution and sediment were then transferred to a 1000ml graduated cylinder which is topped to the 1000ml mark with distilled water. The cylinder was agitated by hand until all the sample is in suspension, and was then placed on a flat surface. A hydrometer was placed in the cylinder and readings were taken at 15 and 30 seconds; 1, 2, 5, 10, 15, 30, 60, 90, and 120 minutes; and at 4, 8, and 24 hours, after the cylinder was set down.

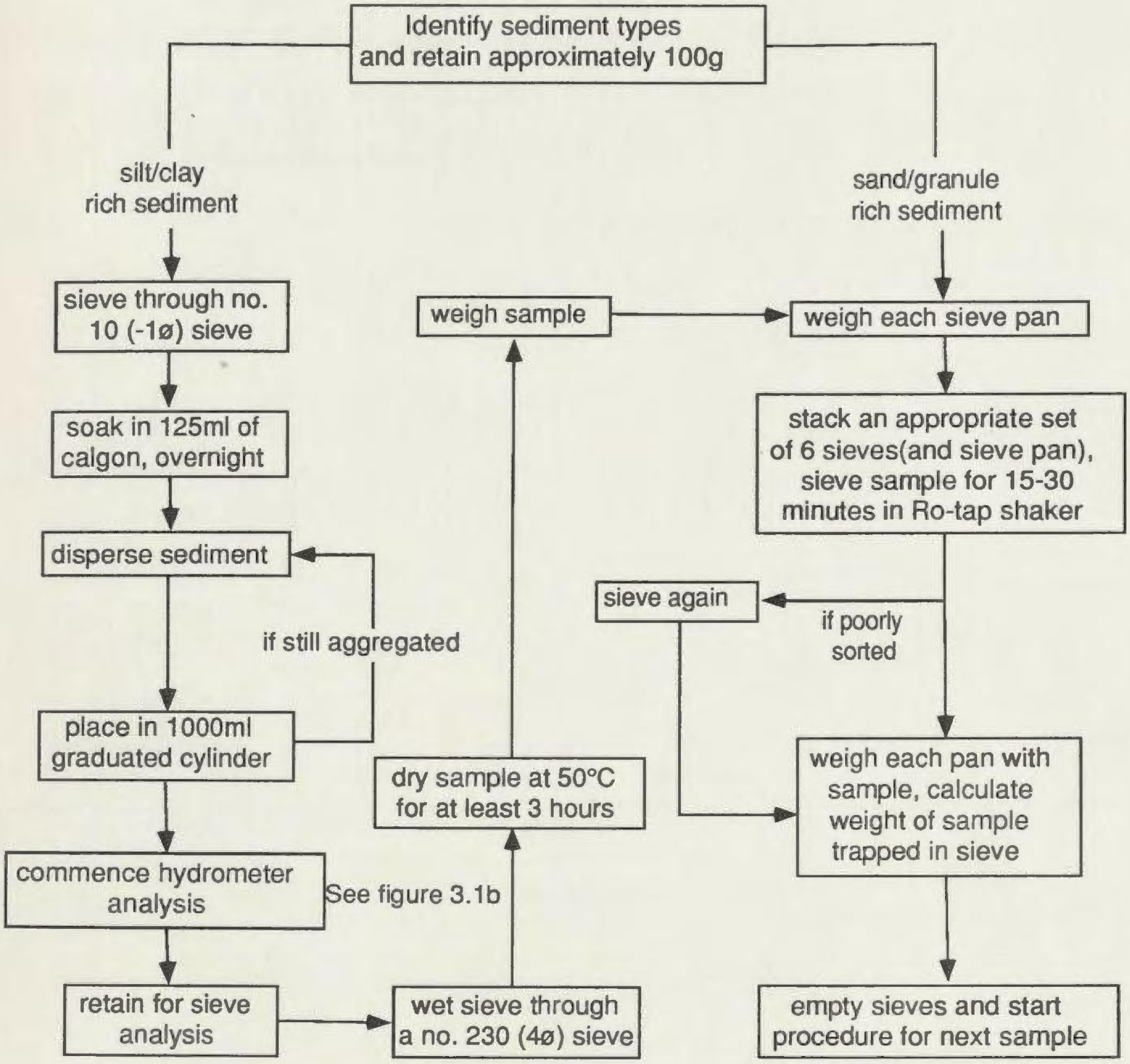


Figure 3.1a: Flow diagram showing procedure for undertaking grain size analysis

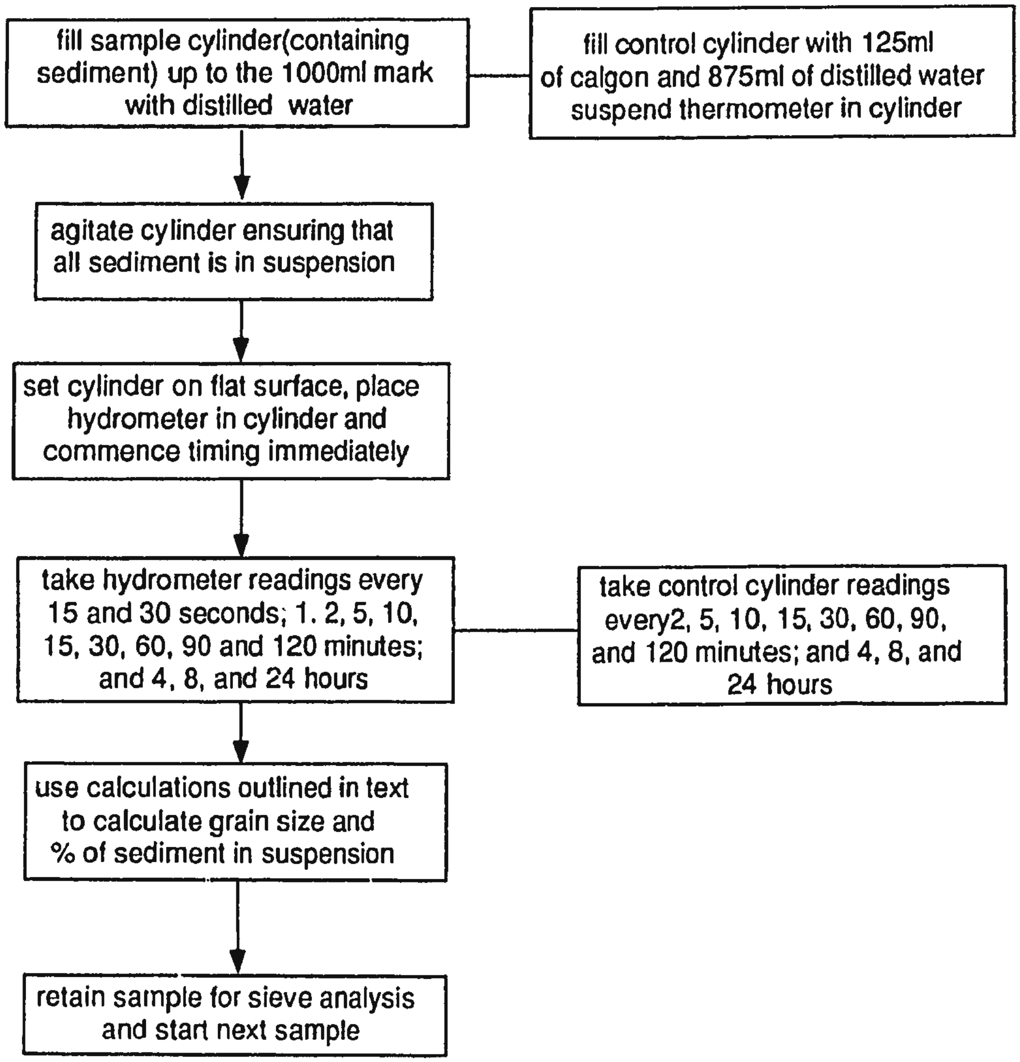


Figure 3.1b: Flow diagram showing the procedure for undertaking hydrometer analysis.

At each time interval the following are involved: the temperature in the control cylinder, the elapsed time since agitation, the hydrometer reading of the sample, the hydrometer reading in the control cylinder, and the corrected hydrometer reading (R)(the difference between the control and sample cylinder). The effective depth (L), the composite correction (K), and the specific gravity (a), can be established by using predetermined tables (American Society for Testing Materials, 1964). Once all the parameters have been established, the clast or grain diameter can be determined using the formula:-

$$d = K(L/T)$$

and the percentage of the sample in suspension can be determined using the formula:-

$$\% = (aR/\text{mass of sample})\text{x}100$$

The fraction coarser than 4ø (sand) is retained and dried in an oven for at least three hours. It is then weighed and sieved through six sieves stacked from the finest size up (Figure 3.1a). All granular samples are sieved through appropriate sieve sizes. For example, a sample with fine granules and coarse sand is sieved through -1ø, 0ø, 1ø, 2ø, 3ø,and 4ø sized sieves.

For each sample a cumulative curve is plotted on log-probability paper (Folk, 1966). This includes both the sieve and the hydrometer results. The particle diameter is plotted along the x-axis and the percentage finer than a given diameter is plotted along the y-axis. Because two types of data are plotted on each graph, there is usually an area of overlap and disjunction known as the "analytical break". This occurs as the result of the differing behavior of fine sand particles under sieving and settling analysis, as well as mechanical difficulties in accurate measurement during the initial stages of hydrometer analysis. This does not have a major effect on the accuracy of the analysis (Catto *et al.*, 1987). The results of grain size analysis are shown in appendices 1 and 2.

3.3.2 - Clast analysis

Fifty or more rock clasts were taken from each site for lithological analysis, the minimum required for statistically significant results (Hornbrooke *et al.*, 1975). All clasts were washed, split and identified. The shapes of the clasts were determined using the classification of Zingg (1935), that groups clasts into spheres, discs, rollers, and blades. Roundness was estimated using the classification of Folk (1955). This groups clasts into categories between 0 (very angular) to 6 (well rounded). In addition to clast identification, the 2ø fraction kept from sieving was also analyzed and the minerals and rock fragments identified.

3.4 - CLAST FABRIC ANALYSIS

Clast fabric analysis involves measuring the dip and direction of the long axis (a-b plane) of 25 clasts within a sedimentary unit. The field measurements taken were analyzed using the Stereo™ program designed for the Macintosh computer (MacEachern, 1989) which plots the 25 observations on an equal area projection net. This gives an immediate impression of the attitude and the shape of the fabric distribution.

The mean orientation of the clasts and the strength of that orientation was calculated using the methods of Woodcock (1977). This involves determining the strength and orientation of a normalized eigenvalue (S_1). Each observation is regarded as a unit vector. Three vectors oriented 90° to each other in three dimensions (v_1, v_2, and v_3) are termed eigenvectors. Eigenvector v_1 can be considered the direction about which the "moment of inertia" of the distribution is minimized (Watson, 1986). That is, it is the estimate of the distribution mean. Eigenvector v_3 is associated with the largest moment of inertia. That is, it is an estimate of the pole of the best-fit girdle to the distribution. Eigenvector v_2 is perpendicular to eigenvectors v_1 and v_3. Each eigenvector has a corresponding eigenvalue which divided by the total number of measurements gives the normalized eigenvalue (S_1). The sum of S_1, S_2, and S_3 is 1.

The normalized principal eigenvalue measures the degree to which the clasts are aligned. A sample that has many clasts lying in the same orientation, will have an S_1 value close to 1. This S_1 value can range from 0.33 (randomly oriented) to 1.0 (perfectly unimodal).

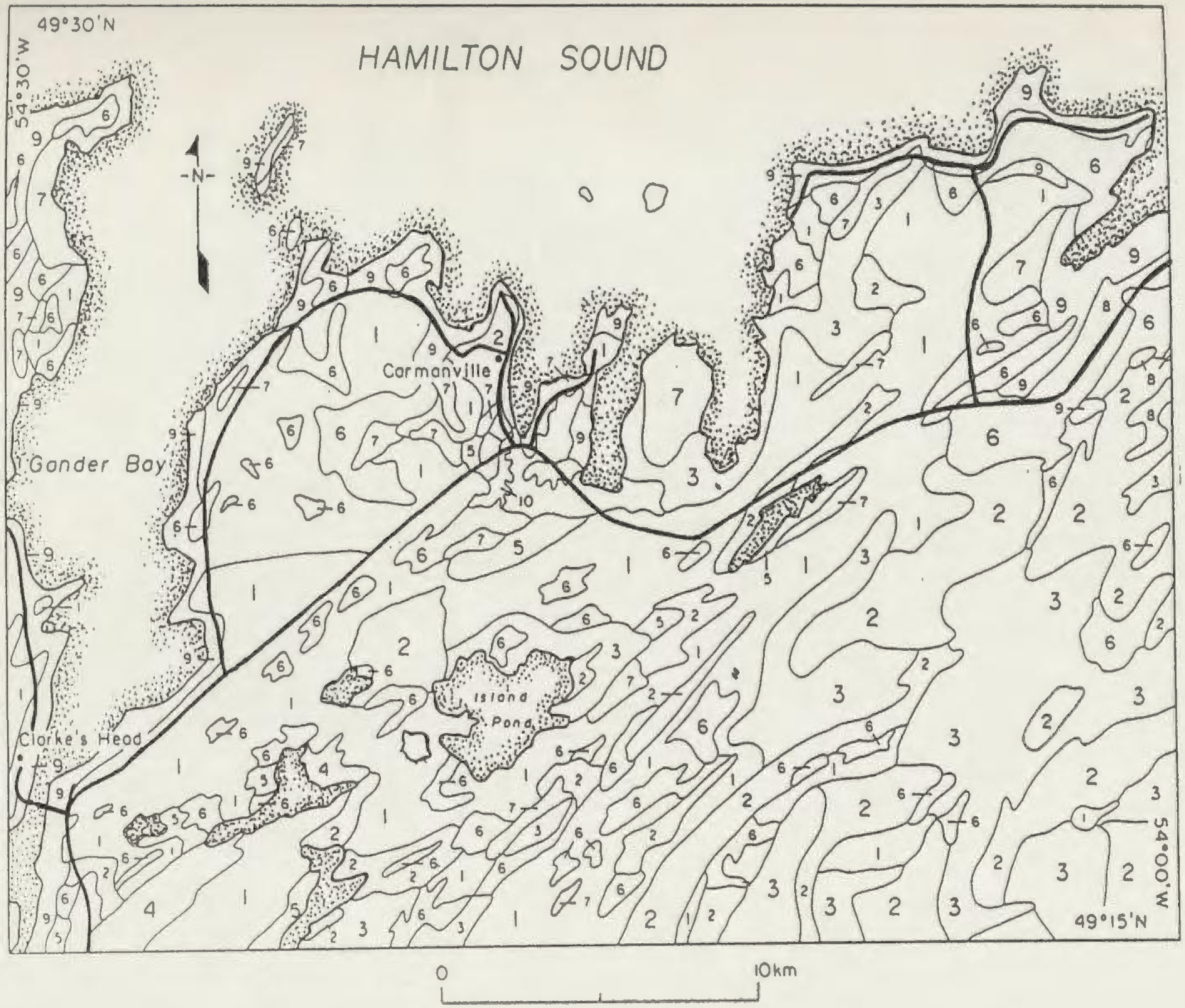
The natural logarithm of (S_1/S_2) divided by the natural logarithm of (S_2/S_3) gives the K value. Distributions that have equal cluster and girdle tendencies have a K value of 1. Where $1 > K \geq 0$, the distribution is described as a girdle. Where $K > 1$ the distribution is described as a cluster. Cluster distributions indicate unimodal fabric alignments.

CHAPTER 4
SURFICIAL GEOLOGY AND GEOMORPHOLOGY

4.1 - INTRODUCTION

Before detailed analysis was undertaken in the Carmanville area, the sediments were mapped using the classification developed by the Newfoundland and Labrador Department of Mines and Energy (e.g. Kirby *et al.*, 1988). Seven major surficial units are recognized in the study area using this classification. These are, in order of diminishing areal importance: glacigenic, organic, bedrock, glaciofluvial, marine, colluvial, and fluvial units. Using this classification, all glacigenic units are classified as "tills". This classification does not distinguish between primary and secondary glacial deposits. Consequently, the deposits are often described as "diamicton", a non-genetic term which simply indicates that the deposit is unsorted and is composed of a mixture of grain sizes from clay to pebbles, cobbles, or boulders.

In any area, as usually more than one sediment type is present, the dominant sediment type is the one that is discussed and shown in Figure 4.1. For example, although glacigenic units are the dominant sediment type in over 70 percent of the area, small patches of organic sediment, fluvial sediment, colluvial sediment, and exposed bedrock occur throughout the areas and are mapped as glacigenic. Figure 4.2 (Munro, 1993)(in the pocket at the back of this thesis) shows the results of the detailed mapping that was undertaken in the Carmanville area. All the sites discussed in this thesis are also shown on Figure 4.2.



1. Till veneer
2. Hummocky till
3. Ridged till
4. Lineated till
5. Eroded till
6. Organic
7. Bedrock
8. Glaciofluvial
9. Marine
10. Colluvial

Figure 4.1: Simplified surficial geology map of the Carmanville area.

4.2 - GLACIGENIC UNITS

Units interpreted to be of glacial origin (units 1-5 on figure 4.1; 'T' on figure 4.2) are the most common surficial materials, covering approximately 70% of the land area. Sediments mapped and interpreted as glacigenic are diamictons and therefore contain mixtures of silt, sand, and larger clasts. These can have been formed due to different genetic processes. They may have been deposited directly from glacial ice and are therefore 'tills' (cf. Dreimanis and Lundqvist, 1984). Other diamictons may, however, be the result of sediment gravity flow beneath or on the surface of glacial ice or into marine waters. In a marine environment diamictons may be substantially reworked.

Sections occur commonly in road cuts, ditches, and on abandoned sawmill or construction sites. Seventeen of the exposures investigated in detail were interpreted to be of primary glacial origin (see Chapter 5). These exposures are between 1m and 4m high. They are cuts in the sides of hummocks, or are found within bedrock controlled topographic lows. The diamictons in these exposures are clast-dominated and are generally structureless, although sorted lenses of material are seen at some locations. Grain size analysis of the matrix component of these sediments (Figure 4.3) shows silty sand (using the classification of Shepard, 1954) to be the dominant sediment type (43%), followed by sand-silt-clay mixtures (32%). Sandy till (14%) and sandy silty till (7%) occur rarely. There is no significant difference in the grain size distribution of different glacigenic units in this area as it is primarily controlled by the bedrock type of the source material (see Chapter 5). Figure 4.4 shows the typical grain size distribution of tills in the region as plots

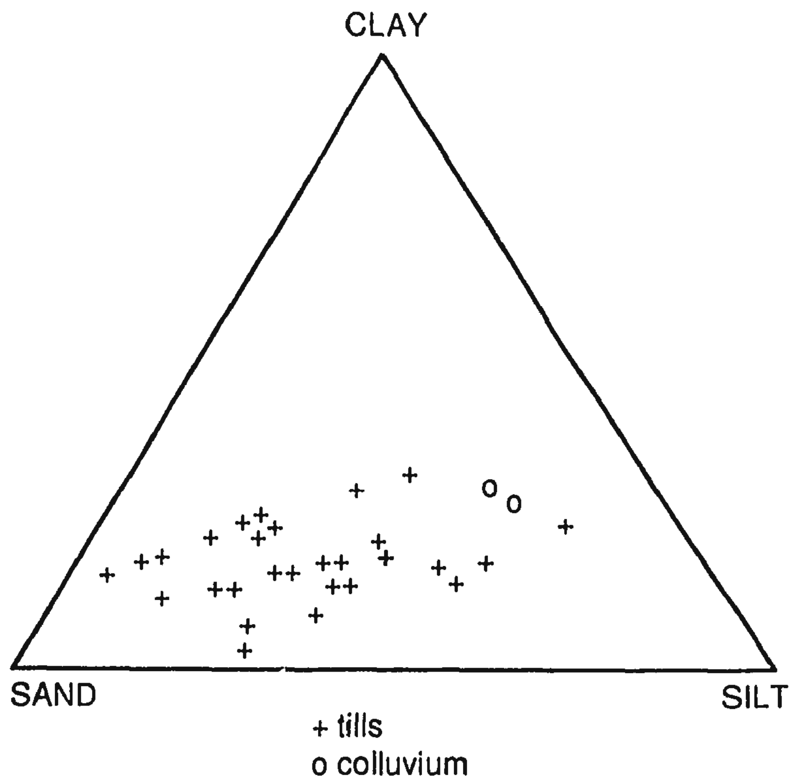


Figure 4.3: Ternary diagram of tills and colluvium in the Carmanville area

on a cumulative curve. Appendix 1 shows the results of grain size analysis on glacigenic units.

Examination of clasts indicates that they are mostly of local provenance. Clasts are often striated, and bullet and barrel shapes are common (see Chapter 5).

4.2.1 - Till Veneers

Till veneers (unit 1 on figure 4.1; 'Tv' on figure 4.2) and eroded till veneers ('Tve' on figure 4.2) are the most extensive surficial glacigenic material

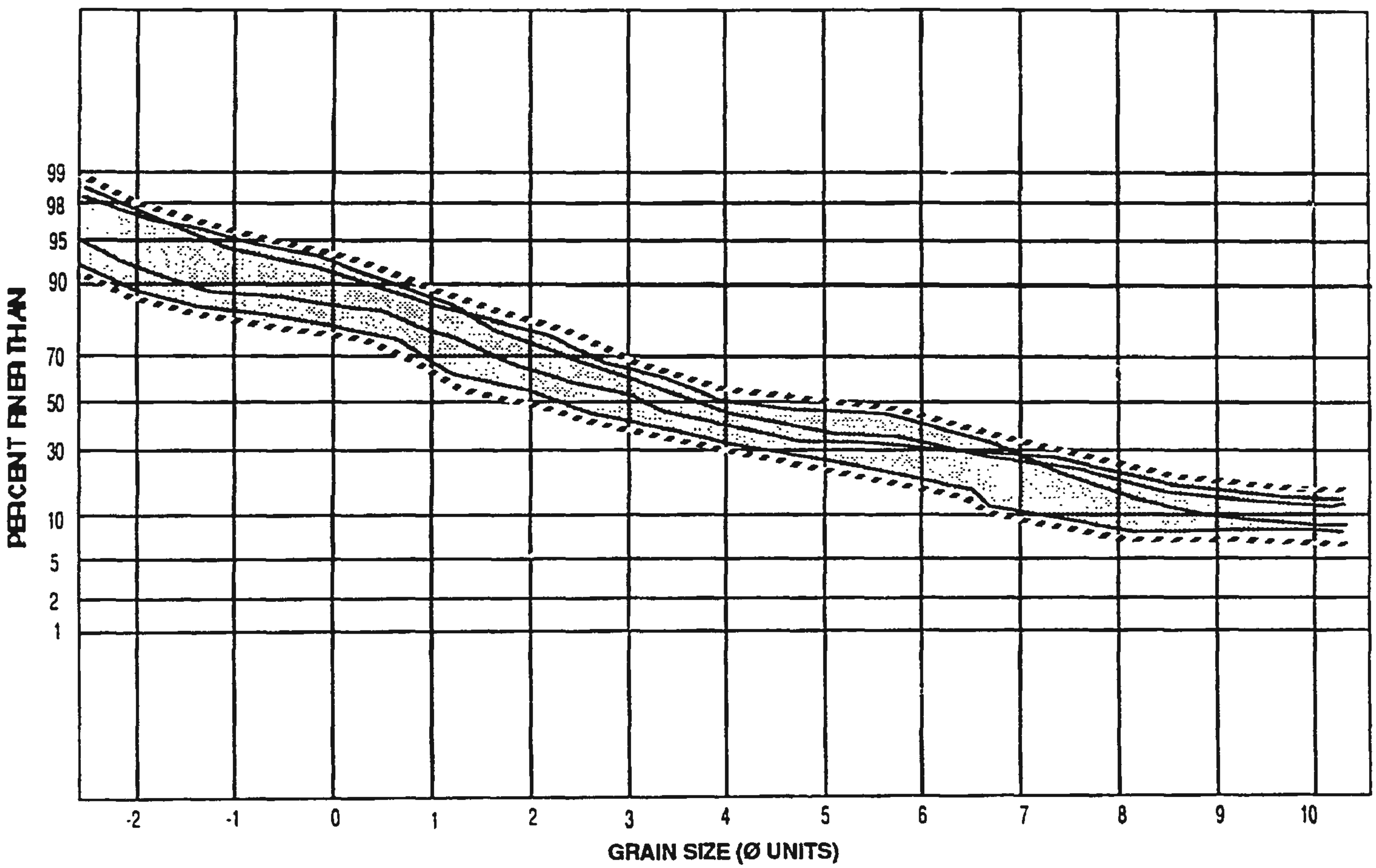


Figure 4.4: Cumulative grain-size plot of all the tills in the Carmanville region. The dashed lines show the upper and lower limits of the plots and the unbroken lines indicate some representative plots of the tills. These are generalized curves and do not show the analytical break.

in the study area, constituting approximately 35% of the total land area. These veneers are defined as any till deposit with little independent surface expression, usually less than 2m thick. They occur most extensively near Frederickton and Carmanville. A typical example of a till veneer is Unit 4 at site MM-30 (see Chapter 5). Vertical exposures are rare and where encountered are 1m or less thick. Post-depositional changes due to ground water movement, pedogenic processes, and frost action have resulted in concentrations of iron and manganese oxides in the upper portions of these eroded till units. Consequently, these tills are characteristically darker coloured than other till deposits in the study area. They vary from dark greyish brown to dark yellowish brown (Munsell 10YR 3/2 m/f - 10 YR 4/2 m/f). Till veneers generally have sandy matrices, and mostly pebble-sized clasts. The matrix accounts for approximately 50 percent of the deposits.

4.2.2 - Hummocky Till

Hummocky till (unit 2 on figure 4.1; 'Th' on figure 4.2) is the dominant sediment type over approximately 20% of the study area. It occurs most commonly in the interior, but extends to coastal areas near Carmanville and Musgrave Harbour. Fifteen of the diamicton exposures studied occur in hummocky till. Hummocks are 10m or less high and are up to 100m in diameter. They are generally restricted to broad topographic lows. These deposits are generally shades of olive grey and olive brown (Munsell 5Y 3/2 - 5Y 5/2 m/f). They are moderately compact, although along the eastern edge of Gander Bay overconsolidated deposits occur. Matrices are generally silty sand but may contain up to 33% clay. The matrix accounts for 50 percent of the sediment and clasts range mainly from pebbles to cobbles.

4.2.3 - Ridged Till

Ridged till (unit 3 on figure 4.1; 'Tr' on figure 4.2) covers approximately 15% of the study area and is represented by rogen and cross-valley moraines. It is generally restricted to the interior of the area, and occurs commonly in association with hummocky till deposits. In Eastern Arm, till ridges are seen as linear islands, composed of relatively coarse cobbles and boulders which are now exposed due to the winnowing of fine sediments by marine erosion. Similar moraines are also recognized at Aspen Cove. An extensive complex (120 km^2) of these moraines is present in the southeast part of the map area, but poor access precluded field investigation.

Aerial photograph analysis has shown these features to be between 100 and 1000m long, attaining maximum widths of 150m and rarely exceeding 20m in height. They are crescent-shaped with the arms of the moraines pointing northeast. They appear to be well-drained, as shown by the extensive spruce growth on their surfaces. Their genesis is probably similar to the moraines observed at Eastern Arm. Similar features observed on the Avalon Peninsula are made up of sandy or muddy gravel, and are often crudely stratified (Fisher and Shaw, 1992). These features are interpreted as rogen or cross-valley moraines. The genesis of rogen moraines is discussed in Chapter 6.

4.2.4 - Lineated Till

Lineated till (unit 4 on figure 4.1; 'Tl' on figure 4.2) occurs in one area, directly to the west and southwest of Island Pond. It covers less than 1 percent of the study area and takes the form of drumlinoid features and flutings. These are between 150m and 800m long and range between 5m and 20m in height. The drumlins are generally highest and steepest at their southwestern ends and taper in a northeastward direction. All the drumlins in the Carmanville area are made up of unconsolidated material. No sections through these features could be studied. However, because drainage is very poor, it is assumed that the matrix has a high proportion of fine grained sediments. These features are mainly oriented southwest-northeast.

4.2.5 - Eroded Till

Eroded till (unit 5 on figure 4.1; 'Te' on figure 4.2) is rarely recognized as a distinct unit in the study area. Eroded till units are mostly classified as eroded till veneer ('Tve' on figure 4.2), eroded hummocky till ('The' on figure 4.2), eroded ridged till ('Tre' on figure 4.2), and eroded lineated till ('Tle' on figure 4.2). Erosion was due to meltwater activity.

4.3 - ORGANIC UNITS

Organic units (unit 6 on figure 4.1; 'O' on figure 4.2) cover approximately 15% of the study area. They occur throughout and are recognisable as flat marshy areas, sometimes in the form of string bogs or aapa. Fibric peat is the most common surficial material (Sudom and Van de Hulst, 1985). Organic units commonly overlie marine deposits and consequently occur most frequently at elevations below the marine limit (at least 57m asl). Exposures at Ladle Cove and south of Musgrave Harbour indicate that these deposits are typically between 0.5m and 1.5m thick. Most of these boggy areas are domed bogs or sloping fen (cf. Wells and Pollett, 1983).

4.4 - BEDROCK

Exposed bedrock (unit 7 on figure 4.1; 'R' on figure 4.2), or bedrock covered by vegetation ('Rc' on figure 4.2) covers approximately 10% of the land area. It is most prominent around the coastline where vegetation cover is thin. The physiography of many parts of the map area is controlled by the underlying bedrock structure (see Chapter 1). The bedrock is locally shaped into roches moutonnées. These are discussed in detail in Chapter 6.

4.5 - GLACIOFLUVIAL SAND AND GRAVEL

Sand and gravel deposits (unit 8 on figure 4.1; 'G' on figure 4.2) are restricted to the north-east part of the study area, directly west of Musgrave Harbour. They occupy an area of 12km^2 (<1%). These have a hummocky surface expression with marine sediments occupying some of the depressions overlying the gravel. The deposits are texturally and structurally variable. For example, horizontally bedded, finely laminated clays, silts and sands are found adjacent to sharply contrasting steeply dipping thick beds of cobbles.

Grain size analysis was undertaken on the finer size fractions (Figure 4.5; 4.6). The results of these are shown in appendix 2. Two ternary diagrams were used to plot the results because of the textural variability of the sediments.

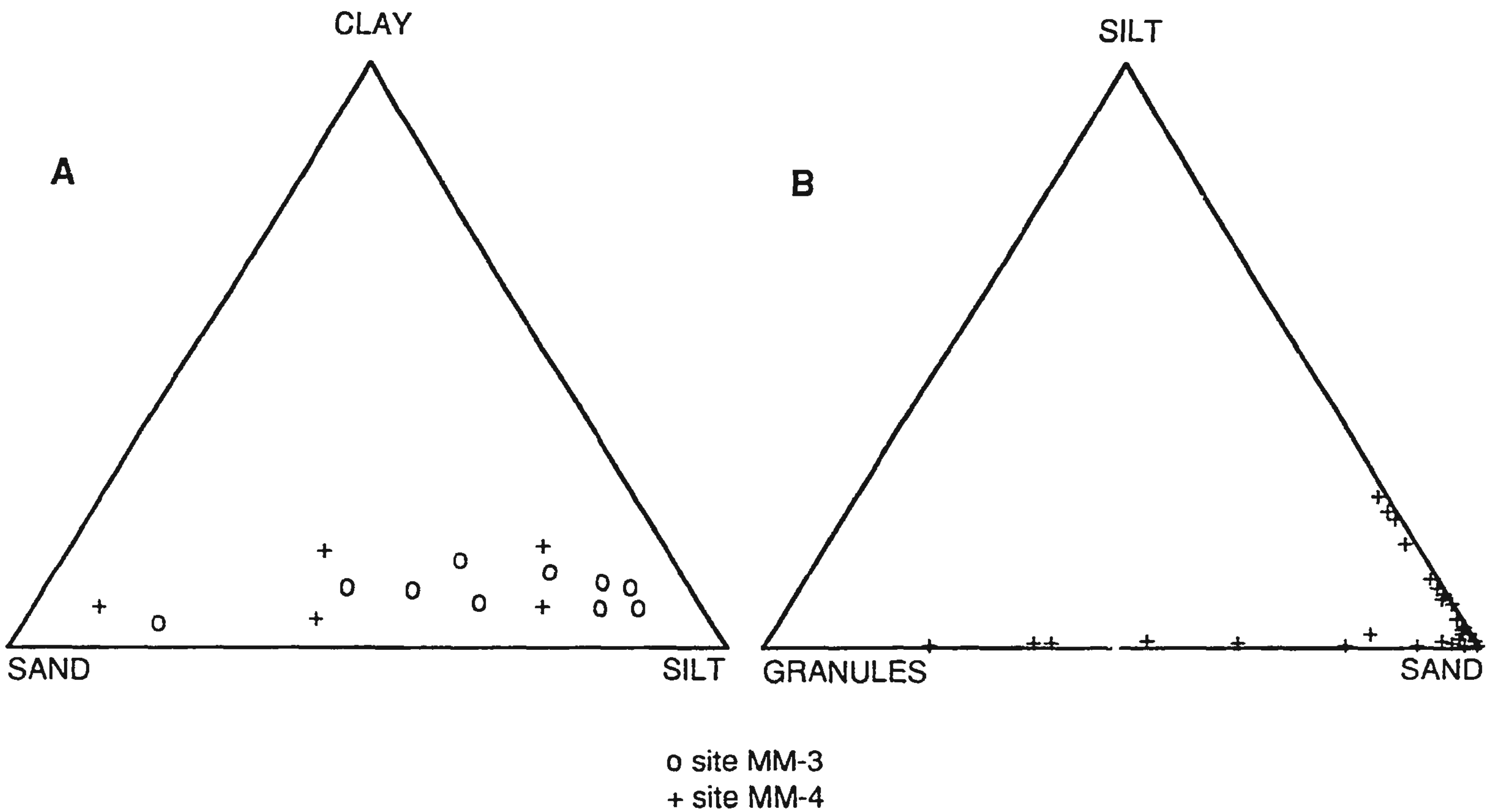


Figure 4.5: Ternary diagram of sands and gravels in the Carmanville area. (A) all samples plotted on this diagram have no particles coarser than sand. (B) all samples plotted on this diagram have no particles finer than silt. Note how the sediments at site MM-3 are generally finer than those at site MM-4

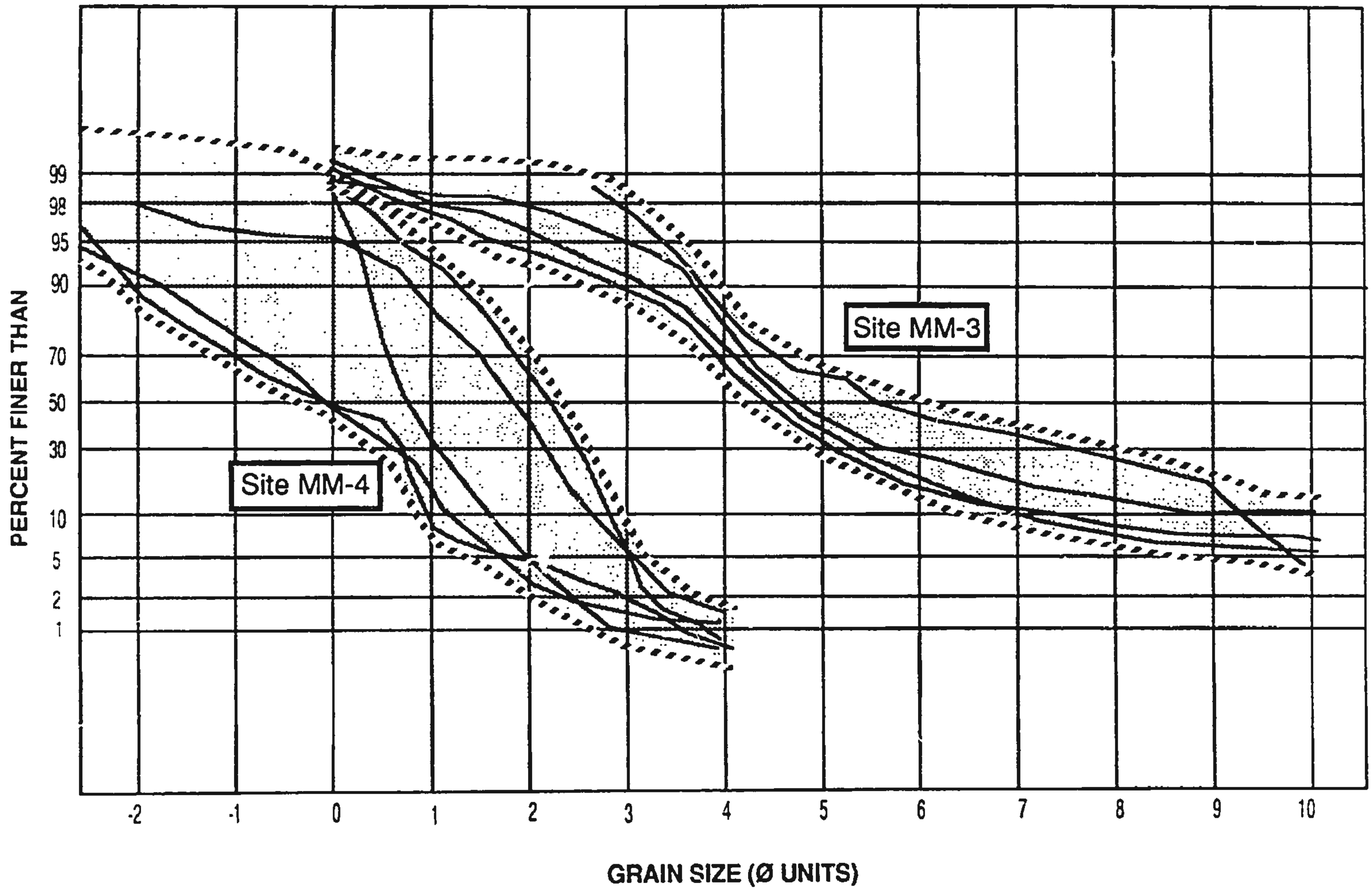


Figure 4.6: Cumulative grain-size plot of glaciofluvial sand and gravel in the Carmanville region. Note how the deposits at MM-3 are finer than those at MM-4.

Deposits with no coarse sand and granules are plotted on Figure 4.5a and deposits with coarse sand and granules are plotted on Figure 4.5b. Deposits containing granules have no or minor silt components (maximum 3%). Deposits which contain more than 3% silt contain no granule-sized material. Silty gravels represent more than 20% of the sands and gravels observed in the study area.

The finer grained deposits (Figure 4.5a) are classified as sandy silts. Deformation in these finer silts and sands is common, particularly where overlain by thick gravels. Common sedimentary structures include high-angle normal and reverse faults, diapirs, flame structures, ball and pillow, box folds, spiral folds, and recumbent folds. There is also extensive evidence for sediment gravity flows. The coarser grained deposits (pebbles to cobbles) are trough or planar cross-bedded. There are many slump structures associated with these. The internal structures and stratigraphy of the glaciofluvial successions are discussed in detail in Chapter 7.

At elevations up to 40m, the surfaces of these deposits have commonly been reworked by marine action, resulting in the deposition of marine beach gravels. Beach sediments form veneers up to 1m thick. Marine reworking subsequent to deposition has resulted in the formation of flat platforms on the surface of the glaciofluvial deposits.

4.5.1 - Moraine Deposits

A large moraine approximately 15km long is located south of Ragged Harbour. Route 330 follows the trend of this moraine from Ragged Harbour River to Musgrave Harbour, approximately 1km east of the study area. The width of the feature varies from a maximum of 2km in the southwestern and northeastern parts, to a minimum of approximately 0.8km in the central part. Most of the surface of the moraine is flat with a slope of 1.5° towards the shoreline, and it has a maximum elevation of 43m asl. The surface of the central area (the narrowest part) is higher with maximum elevations reaching 46m asl. It is also distinctly asymmetrical with a steep slope (approximately 20°) on the southern edge and a long, gentle slope (1.5°) marking the northern boundary of the feature. The moraine is composed of glaciofluvial silts, sands, gravel, and cobbles (see Chapter 7).

4.5.2 - Meltwater Channels

Meltwater channels occur throughout the map area (see figure 4.2), and generally range from a few metres wide to 0.5km across, and are up to 100m deep. They are aligned roughly southwest to northeast, following the structural trend of the underlying bedrock. Although no rivers or large streams currently occupy the channels, they commonly connect many small lakes. One channel exits Island Pond, trends northeastward to Shoal Pond, extends northeastward where it ends abruptly before reaching the modern shoreline. This channel is approximately 15km long. The size and alignment of these channels suggest that they are most likely subglacial forms (cf. Sugden and John, 1976). Although the topography of the area is generally flat, these channels have undulating profiles along their lengths with maximum reliefs of approximately 10m. Undulating profiles are characteristic of subglacial meltwater channels (Sissons, 1961). The bases of the channels are generally flat and contain veneers of sand.

Several ice-marginal and submarginal channels occur on the eastern flanks of Gander Bay. These are mainly submarginal forms as their gradients are relatively steep (between 1:15 and 1:20). Channels which have gradients of 1:50 or greater can be considered marginal meltwater channels (Sissons, 1961). Two distinctive channels are formed at 45m asl, and 23m asl. These channels are generally less flat-bottomed than the subglacial forms. Steeper valley slopes have aided in increased erosion and subsequent infilling of the channels.

4.6 - MARINE UNITS

Marine deposits (unit 9 on figure 4.1; 'M' on figure 4.2) occur mostly as emerged beach gravels although a marine kettle fill at the Ragged Harbour 2 site (see Chapter 7) containing silts and sands is also present.

Emerged; terraced beach gravels ('Mt' on figure 4.2) occur up to elevations of 57m asl, although erosional platforms occur up to elevations of 67m asl. Exposed marine gravels are found at three principal locations. In the Aspen-Ladle Cove area, these deposits extend along the coastline from Rocky Point to Ragged Point, and are found up to 300 metres inland. Beach sediments occur up to 2km inland, between Victoria Cove and Clarke's Head. These areas are currently being exploited for aggregate. At Noggin Hill, the deposits extend inland from the coastline for about 500m, and extend north to south for 4km. These sediments consist of well sorted gravel and vary from granules to cobbles.

A series of marine terraces in the Wing's Point area can be seen on aerial photographs. These reach maximum elevations of 51m asl. A good exposure occurs at 32m asl at Wing's Point. Over 3m of gravels overlie bedrock in terraces which extend laterally for 2-3km (see Chapter 8). Over 4m of gravel is exposed at Victoria Cove at an elevation of 11m asl, representing part of a lower elevation terrace (see Chapter 8). It is possible that similar thicknesses of gravels may be present at Noggin Hill and Ladle Cove at similar elevations.

All marine gravels observed in the Carmanville area are open-work and consequently well-sorted. They are planar bedded and locally dip with low angles towards the modern shoreline. Beach ridges are evident in some exposures such as those at Wing's Point. Clast sizes range from granule gravel to cobbles, and are likely controlled by differing energy levels. For example, at Ladle Point almost 2m of cobbles are present in section at 5m asl. This part of the coastline was directly exposed to oncoming waves when sea-levels were higher. Less than 3km to the west at Ladle Cove, however, gravel at the same elevation is much finer. This part of the coastline lay oblique to oncoming waves when sea-levels were higher. At most locations, gravel consists of medium pebbles. Clasts are generally disc-shaped with some spheres.

Marine silts and clays were observed overlying sands and gravels at Ragged Harbour. They are also likely to exist at Aspen Cove and Ladle Cove. This is reflected in the present poor drainage of the area and bog formation. These boggy areas represent ancient lagoonal or back-barrier systems which are generally characterized by silt and clay sedimentation.

4.7 - COLLUVIAL UNITS

Colluvium (unit 10 on figure 4.1; 'C' on figure 4.2) is rare in the study area (<1% by area), due to the lack of steep slopes on which mass movement can take place. Colluvium is found near Carmanville and Carmanville South, and along Route 330 near the Aspen Cove exit, as fan shaped lobes at the base of slopes. The texture and structures within the sediment vary considerably from the local till. In general, it is much more compact, the clay content is higher, and slump structures are evident. The most common matrix type is clayey silt with approximately 35% clay and 50% silt (Figure 4.2). Lithologically and mineralogically, these are similar to the tills. This is primarily because the colluvial deposits in the Carmanville region are reworked till. Fabrics from these deposits commonly have girdle distributions. Eigenvalues range from 0.43 to 0.56, and K-values are uniformly less than 1.

4.8 - FLUVIAL UNITS

Fluvial deposits ('F' on figure 4.2[1]) cover less than 1 percent of the surface area. They are present along the banks of the Ragged Harbour River on the modern flood plains. These are frequently covered by thin organic veneers. Characteristically, the fluvial deposits are made up of fine clays and silts, as reflected in the poor drainage of the area.

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[1] Fluvial units are not indicated on figure 4.1 as their extent is minimal and the scale on figure 4.1 is too small to indicate their presence.

CHAPTER 5
SEDIMENTOLOGY 1: GLACIGENIC SECTIONS

5.1 INTRODUCTION

Seventeen exposures in the study area (Figure 5.1) containing diamicton were described and analyzed in detail. The results of this analysis are shown in Appendix 1. These were determined to have been directly deposited by glacial ice and are thus interpreted as tills. It is important that careful analysis of these deposits is undertaken, as tills may give an indication of the direction of ice-flow, the distance of transport of the entrained sediments, and the mode of genesis of the till. These are important factors in discussing the glacial history of any area. Five of the exposures are discussed in detail here, and all exposures studied are considered in the summary.

5.2 CARMANVILLE SOUTH 1 (MM-32)

Description

MM-32 is a diamicton composed of four units (Figure 5.2). The section is approximately 2.5m high and is situated in a small hummock (UTM 699650, 540450). The thickness of the diamicton is unknown as the basal contact was not observed.

The lowermost unit has a minimum thickness of 32cm, and is clast-dominated (40% clasts.). Approximately 35% (by volume) of the diamicton matrix is sand, 11% is silt, and 4% is clay. The remainder is composed of granules. The matrix colour is dark olive grey (Munsell: 5Y 5/2 mf). Clasts are generally subangular to subrounded and clast types are dominantly shale

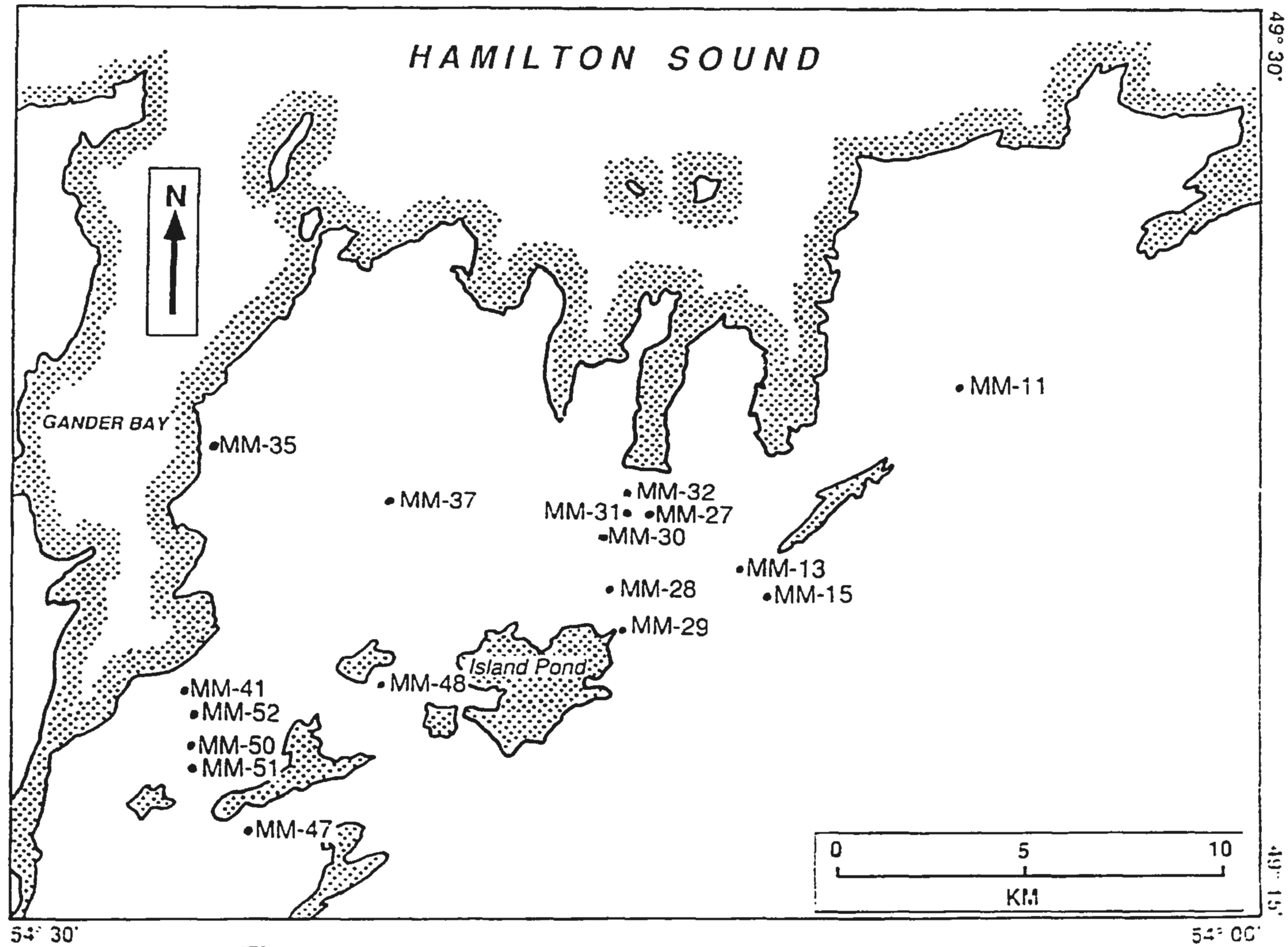


Figure 5.1: Location map of all studied glacigenic sections

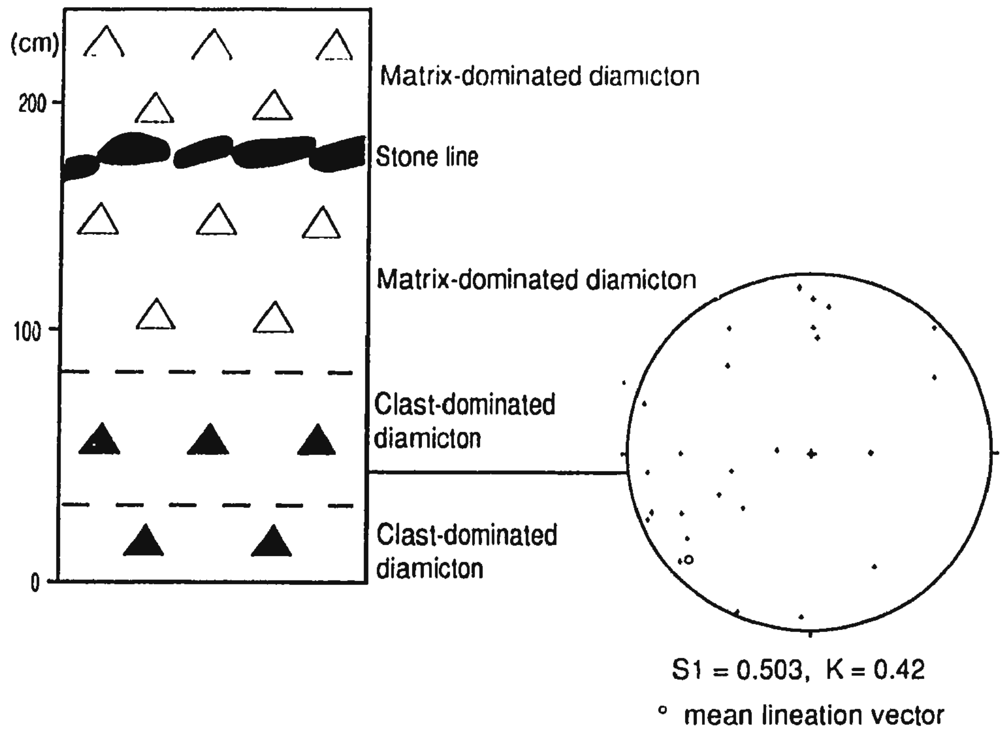


Figure 5.2: Vertical transect through sediments at Carmanville South 1 (MM-32), with fabric information.

(36%), and mafics and ultramafics from the Gander River Complex (32%). Clasts derived from the Rocky Bay Pluton (granites), the local bedrock, comprise 20% of the assemblage. Clasts are commonly striated.

The lower diamicton grades vertically into unit 2, a 50cm thick clast-dominated diamicton (50% clasts), marked by striated cobbles. The clast assemblage is similar to that of the underlying sediments, with mostly shales from the Hamilton Sound Sequence, and ultramafics from the Gander River Complex. The matrix consists of 58% sand, 28% silt, and 11% clay. It is dark olive grey (Munsell: 5Y 5/2, mf). The 2ø fraction of the diamicton matrix is composed of 75% crushed shale and schist, and quartz and feldspar grains. The remaining 25% is composed of crushed ultramafics, biotite, muscovite, hornblende, and pyroxene. A clast fabric from this unit has a weak orientation with an S_1 value of 0.503 and a K value of 0.42. The mean trend orientation is 227.3° and the mean plunge is 11.2°.

The cobbly diamicton grades vertically into unit 3, a 1m thick matrix-dominated diamicton (<20% clasts) of similar colour to the underlying unit (dark olive grey; Munsell: 5Y 3/2, mf). The matrix is composed of 51% sand, 33% silt and 16% clay. The clast assemblage is dominated by rocks from the Hamilton Sound Sequence (shale and slate, 40%), and the Gander River Complex (diabase and basalt, 36%). The remainder of the clasts are derived from local granites. Most of the clasts are striated. A single distinct stone pavement is evident towards the top of this unit, marked by a monolayer of pebbles and cobbles in direct contact with each other (Plate 5.1 shows a similar stone pavement at the Beaver Hill site, p.72). The line of clasts generally dips towards the south at an angle of approximately 10° with many of the individual clasts also dipping towards the south with a low angle of plunge (<10°). They are faceted and their upper surfaces are striated. The mineral assemblage in the matrix is similar to that in the underlying unit, and is dominated by crushed ultramafics, pyroxene, hornblende, and mica.

Directly overlying the stone line is a 65cm thick matrix-dominated diamicton, which is olive in colour (Munsell: 5Y 4/4, mf). The basal part of this unit is very similar to the underlying diamicton. Clast content is low (5%), and most clasts are faceted and striated with barrel or bullet shapes predominating. Clast lithology is similar to the underlying units with clasts from the Hamilton

Sound Sequence and the Gander River Complex predominating. The matrix is slightly finer than the underlying material with 45% sand, 37% silt, and 18% clay. It is relatively compact and fissile towards the top of the unit. The 2ø fraction of the matrix is composed of at least 60% crushed shale and slate, 25% is quartz and feldspar, and 15% is mica, pyroxene, and crushed ultramafics.

Interpretation

The sediments at Carmanville South 1 are interpreted as having been deposited directly by glacial ice, i.e. they are till. Throughout the section, faceted and striated clasts are present which indicate glacial transport (Dreimanis, 1982). These indicate that abrasion played a major role at some point in time during the transport and depositional history of the clasts. All the clasts are of local origin, mainly from the Rocky Bay pluton, the Hamilton Sound Sequence which extends 1km south of the site, and the Gander River Complex which crops out 4km south of the site. This suggests that the sediment is of basal origin as the clasts have not been transported laterally for great distances and have not been transported within the body of the glacier. The clast assemblage also indicates that ice flowed in a northerly direction.

Abrasion is the most common mode of comminution in lodgement tills, and produces a siltier matrix than the crushing action more commonly associated with melt-out tills (Haldorsen, 1983). In general, basal tills of different origins are characterized by differing grain sizes and concentrations of particular minerals in those size fractions (Dreimanis and Vagners, 1972; Shilts, 1982; Haldorsen, 1983; Shaw, 1989; Hicock and Dreimanis, 1992). Minerals have specific terminal grades: the final grain sizes which result from glacial comminution (Dreimanis and Vagners, 1972). When a mineral has been abraded to its terminal grade, further glacial transport will not result in additional abrasion. Igneous and metamorphic minerals such as feldspar, quartz, pyroxene, and garnet, have the coarsest terminal grades, and generally dominate the sand and coarse silt fraction. Most minerals derived from sedimentary rocks (apart from quartz in sandstone) have their terminal grades concentrated in the silt size fraction and (less commonly) in the clay size fraction (Dreimanis and Vagners, 1972).

The genesis of till deposits cannot be determined on the basis of grain size alone. For example, deformation tills generally have finer matrices than lodgement tills, and lodgement tills have finer matrices than melt-out tills (Ashley *et al.*, 1985), but their grain size distribution may be entirely dependent on the bedrock source and the minerals in that bedrock rather than on the mode of deposition.

The lowermost unit in the Carmanville South 1 exposure is relatively coarse which may suggest that it is a melt-out till (cf. Haldorsen, 1983). The only evidence, however, of glacial transport is the presence of striated clasts which does not necessary indicate that this is a primary till.

Unit 2 in the exposure is relatively coarse (85% sand and granules) which may indicate that crushing was the major comminution process (cf. Haldorsen, 1983). This unit contains high percentages of shale, both as clasts and matrix, and minerals from this rock type would normally have their terminal grades concentrated in the silt size fraction (Dreimanis and Vagners,1972). The low proportion of silt, therefore, suggests that either there was not enough time or enough distance traveled for the minerals to reach their terminal grades, or that abrasion was not the major mode of comminution in this lower till unit, or that the fines may have been winnowed out of the deposit by meltwater.

The fabric from this unit is weak, in contrast to the relatively high S_1 values (strong orientations) characteristic of both lodgement tills and melt-out tills (Dowdeswell and Sharp, 1986; Haldorsen and Shaw, 1982). Clast plunge angles are consistently low however, suggesting that this unit was probably deposited basally (cf. Ashley *et al.*, 1985). Melt-out tills may have weak and inconsistent clast fabric orientations, caused by clasts falling through flowing water and thus have their preferred orientations altered (Boulton *et al.*, 1984). Melting of clast-rich basal debris may lead to frequent clast interactions, and thus may increase the chance of a weakened fabric orientation (May *et al.*, 1980; Rappol, 1985). Consequently, weak fabrics may be indicative of melt-out tills. Although there is no on piece of evidence to suggest that this is a melt-out till, the unit was deposited basally, as indicated by the low angle of plunge of the clasts, and probably represents melt-out as the grain size is much coarser than would normally be expected for other basal tills.

Unit 3 is texturally similar to the two underlying units. The presence of striated clasts indicates that basal glacial transport has occurred, and because of the coarse texture of the deposit it may be a till of melt-out origin.

Units 1, 2, and 3 grade vertically into each other and are lithologically and mineralogically similar. They probably, therefore, represent one depositional event. The vertical changes in the exposure may represent vertical changes in the ice profile. Although no stratification or laminae were observed in the lower three units of the section, it is dominated by sands and granules suggesting that there may have been some winnowing of fines. Sorting is considered to be indicative of basal melt-out till (Krüger, 1979; Shaw, 1979; Dreimanis, 1982). The unsorted nature of the sediments, the presence of striated clasts, consistently low angles of clast plunge, and the presence of primarily locally derived material, suggest that this is a till of basal origin. It may be a melt-out till, because of the large grain size of the sediment particles (using Haldorsen's (1983) criteria), but this cannot be absolutely determined.

Unit 4 is finer grained than the underlying till units, and it contains many striated and faceted clasts suggesting that abrasion was a major component of the comminution process. This unit, therefore, may be a lodgement till using the criteria of Haldorsen (1983). Fissile sediments of glacial origin are generally considered to be representative of lodgement till (e.g. Dreimanis, 1982; Ashley *et al.*, 1985), and reflect minor failure planes within the sediment that originate during the lodgement process. This unit is most probably, therefore, a lodgement till. Lithological and mineralogical examinations have demonstrated that the mineral content and clast type do not differ substantially between this unit and the lower three units and thus all were deposited by ice moving in a similar direction.

The clast pavement occurs between units 3 and 4. At other locations, clast pavements are generally 1 clast thick but their areal extent can range from square metres to square kilometres. They often occur at the base of structureless diamictons (e.g. Dreimanis, 1976) but can also be found at various levels within a diamicton (e.g. Dreimanis *et al.*,1987; Clark, 1991). Clasts are always faceted, surfaces are usually striated, and bullet shapes are common. The diamictons associated with clast pavements usually have a substantial fine-grained matrix, the silt and clay components representing more than 50% of the matrix (Wright *et al.*, 1973; Skinner, 1973; Sauer, 1974; Thorliefson, 1989; Clark, 1991).

As the stratigraphic position of the clast pavements within diamictons is highly variable, various theories of formation have been proposed. These include subaerial erosion of pre-existing sediment (Wright *et al.*, 1973), clast ploughing (Clark and Hansel, 1989), and selective lodgement or accumulation of coarse clasts accompanying removal of fine particles during lulls in till deposition (Johansson, 1972; Stea, 1984; Stea and Brown, 1989).

Hicock (1991) postulated that a wet-based glacier moving over viscously deforming till gradually releases stones that will accumulate at the glacier sole. Assuming that the stones are of similar lithology, they will have similar densities and consequently will sink to a common level in the newly deposited till depending on the velocity, density, and the viscosity gradients of the deforming till flow. Hicock named this zone of accumulation the 'viscous suspension zone'. Depending on the parameters listed above, the clast pavement could ultimately be located at any level within a till unit. Because the till is deforming, stones should be subject to rotation. Therefore, not all stones should be lying flat, and all surfaces could potentially be striated. These stones would probably form an undulating or irregular surface rather than a flat surface.

Clark (1991) equated the formation of clast pavements in till with the formation of traction carpet sediments in debris flows. Debris flows commonly have larger clasts at their bases which never become incorporated into the body of debris flow. Clasts as traction load will thus move by sliding and rolling along the base of the flow (Lowe, 1979; Lawson, 1981). Clark states that the low sediment strength of subglacially deforming sediment may permit the settling of larger clasts to the base of the sediment. If the overriding deforming sediment acts in a similar manner to a debris flow, then abrasion of the upper surfaces of the clasts will result.

At Carmanville South 1, the clast pavement is situated between units 3 and 4. This may suggest that it is a subaerial lag deposit between two periods of till deposition. This interpretation is considered unlikely. The clast lithology and mineralogy of sediments above and below the pavement are similar. This indicates that the tills were probably deposited by the same glacier which overrode the Gander River Complex and the Hamilton Sound Sequence directly to the south and flowed northward.

At MM-32, the sediments overlying the clast pavement are generally finer than those underlying it. Consequently, density gradients may be different in these two areas, allowing Hicock's (1991) theory of differential settling of clasts to be tested here. If the clasts had settled to a common level in the till, it is unlikely that this level would be a flat surface. The clast pavement at MM-32 is a planar monolayer. In addition, although most of the clasts have similar lithologic origins, they are not identical. Granites, shales, slate, and basalt dominate, and these all have differing densities. Consequently, all the clasts would not necessarily sink to the same level in the till. There is no evidence of rotating clasts in the upper zone of the till. If the clasts had been subjected to rotation, then all the surfaces of the clasts could potentially be striated. Clasts associated with deforming till would be expected to lie at various angles, rather than in a flat smooth pavement. It is unlikely, therefore, that these deposits represent deformation till. Hence, Hicock's theory of differential settling of clasts through the till is not applicable at this locality. Clark's (1991) theory also depends on the till being subjected to deformation and therefore cannot be applied here.

The most likely origin for the clast pavement is that clasts were preferentially melted out from glacial ice, perhaps from a foliation plane, while ice was actively ablating. Ice then started to readvance, overriding the clast pavement causing the upper surfaces of the clasts to become more striated than the lower surfaces. Clasts were not removed and reincorporated into glacial ice because clasts ploughing into each other would allow them to become firmly lodged next to each other (cf. Clark and Hansel, 1989). Under these circumstances, the clast pavement would not be areally extensive. Several adjacent sections do not display clast pavements.

In summary, the two tills at Carmanville South 1, were deposited either by the same ice mass or by two ice-flows that moved in similar directions. Lithologies and mineralogies from all the units are similar, and all four units were deposited basally. The lower three units probably represent ice that was ablating. A series of larger clasts were then melted out from the ice, probably from a foliation plane. There may then have been a lull in deposition while the entire area was still covered in glacial ice. Ice then started to readvance, lodging the larger clasts against each other and striating their upper surfaces to produce the clast pavement. This advancing ice may have deposited the upper till unit by lodgement, which would explain its finer-grained matrix . There is not sufficient evidence, however, to indicate the genesis of this deposit as anything more specific than a basal till.

5.3 BEAVER HILL SITE (MM-35)

Description

The Beaver Hill site (MM-35) is situated in a roadside cut in a low lying hummock, approximately 2km south of Beaver Hill along the eastern flanks of Gander Bay (UTM 687750, 5472450). The section is approximately 2m high, although the uppermost 0.5m are slumped (Figure 5.3). The basal contact was not observed. Two units were present.

The basal unit has a minimum thickness of 45cm. It is a clast-dominated diamicton (~50% clasts), and clasts are generally large (>10cm). They are very weathered, breaking easily when struck with a trowel. No matrix samples were taken or fabric analysis attempted. A visual estimate, however, determined the matrix to be relatively coarse, with sand and silt dominating.

The lower diamicton grades into a 90cm thick matrix-dominated diamicton. It becomes more compact and fissile towards the top. It has a low concentration of clasts (<40%). The matrix is olive (Munsell: 5Y 4/4, mf) with high concentrations of silt (62%) and clay (26%). Clasts are generally fine pebbles (2-5cm A-axis), are commonly striated, and are bullet shaped. Most of these are shale from the Hamilton Sound Sequence (60%), with a further 20% from the Gander River Complex (mafics and ultramafics).

Rare, isolated clay stringers were observed within the diamicton unit. These are usually less than 1cm thick and extend laterally for 20-30cm. Upper and lower contacts are relatively abrupt and planar. These laminae do not drape over clasts but are often bisected by them.

One clast fabric analysis was undertaken on this unit. An S_1 value of 0.843 and a K value of 10.69 were determined. The mean trend of the clasts is

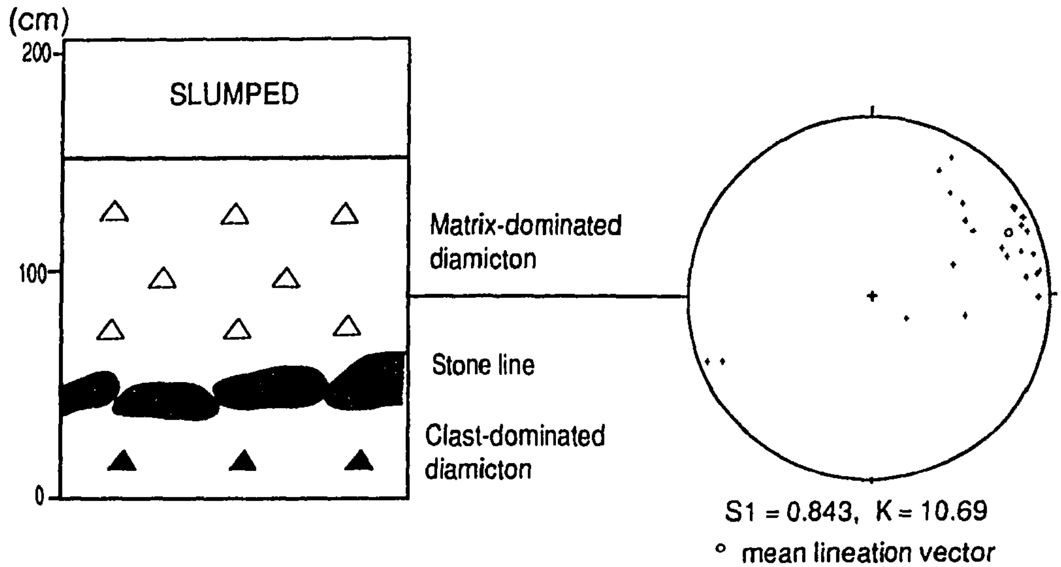


Figure 5.3: Vertical transect through sediments at Beaver Hill (MM-35), with fabric information

65.7° and the mean plunge is 17.1° (Figure 5.3). These values represent a strong orientation and thus would be placed on the cluster field of Woodcock's (1977) graph.

The clast pavement is generally one clast thick, although in some cases the clasts overlap (Plate 5.1). The surface of the pavement is flat and it slopes to the north with a dip of approximately 10°. The clasts are dominantly cobbles (A-axis 15-20cm) and are thus much larger than the pebbles in the upper and lower diamictons. Clasts are striated on their upper surfaces only.

Plate 5.1: Clast pavement at Beaver Hill (MM-35). The pavement is composed of a monolayer of cobbles with some pebbles. The surface of the pavement is relatively flat but dips to the north at an angle of approximately 10°. The upper surfaces of the clasts are striated. The trowel is approximately 25cm long.

Interpretation

The sediments at the Beaver Hill site are interpreted as primary till deposits. The site is located within the Hamilton Sound Sequence outcrop area, bordered by the Carmanville Mélange. The prominent bedrock types are shale, slate, and siltstone, as is reflected in the till lithologies. Most of the minerals derived from these sedimentary and altered sedimentary rocks have their terminal grades concentrated in the silt and clay size fractions (Dreimanis and

Vagners, 1972), and thus both abrasion and crushing could potentially produce a similar grain size distribution.

At Beaver Hill, similarly to site MM-32, the lower unit is coarser grained than the upper unit. As no detailed analysis was undertaken on the lower unit, its genesis cannot be absolutely determined.

As the upper unit is finer than the lower unit it is likely that its genesis is different to the lower unit . As the sediment is fissile it is likely that this is a lodgement till (cf. Dreimanis, 1982; Ashley *et al.*, 1985). The clast fabric from this unit is strongly unimodally oriented. Unimodal clast fabrics are generally considered indicative of basal melt-out tills and lodgement tills (Lawson, 1979; Dowdeswell and Sharp, 1986), and are good ice-flow indicators. The fabric pattern would initially suggest that ice was flowing towards the southwest, as the mean trend and plunge of the clasts is towards 65.7°. Up-ice plunges are common in basal tills (e.g. Dowdeswell and Sharp, 1986; Clark and Hansel, 1989) suggesting that the up-ice end of the glaciers in this region was to the northeast. Striation evidence, however, shows that there was no ice-flow from the northeast to the southwest, but there was ice-flow towards the northeast (See chapter 6: Section 6.1.1). Lithological data also indicates that the source areas for glaciation were located to the southwest and thus this fabric cannot represent southwestward flowing ice.

Catto (1990) encountered a similar problem when interpreting clast fabrics associated with roches moutonnées. He suggested that where there is diamicton overlying a resistant substrate (e.g. igneous/metamorphic bedrock) continued deformation and stress on that diamicton will result in re-orientation of the clasts. Ultimately, they could preferentially plunge up- or down-ice. The depth of the surface of the bedrock at this site is unknown, but it is likely that a similar pattern of deformation could have occurred.

The clast pavement at this site is similar to that noted at the Carmanville South 1 site with a relatively fine-grained diamicton overlying a coarser diamicton. Like Carmanville South 1 the clast pavement probably represents coarser deposits that were preferentially melted out from glacial ice and then were overridden by readvancing ice.

The presence of clay laminae in the upper unit suggests that water may have been present, suggesting that this sediment was deposited by melt-out rather than lodgement. However, a thin film of water due to pressure melting (e.g. Hicock and Dreimanis, 1992), can be present beneath ice during the lodgement process which would produce thin sorted laminae. Alternatively, the laminae could represent shear planes in freshly deposited till or in the overlying ice mass (Dreimanis, 1976). This is more common in lodgement till or deformation till.

In summary, the unimodally oriented clast fabric, the fine nature of this upper till, the presence of striated and bullet-shaped clasts, and the undeformed nature of the laminae indicate a lodgement origin. The laminae could either have been formed by sorting at the glacier sole or they could represent shear planes. There is no sedimentary evidence to distinguish these.

5.4 SHOAL POND SITE (MM-13)

Description

The Shoal Pond site (MM-13) is situated on a forestry access road approximately 1km south of the southwestern shore of Shoal Pond (UTM

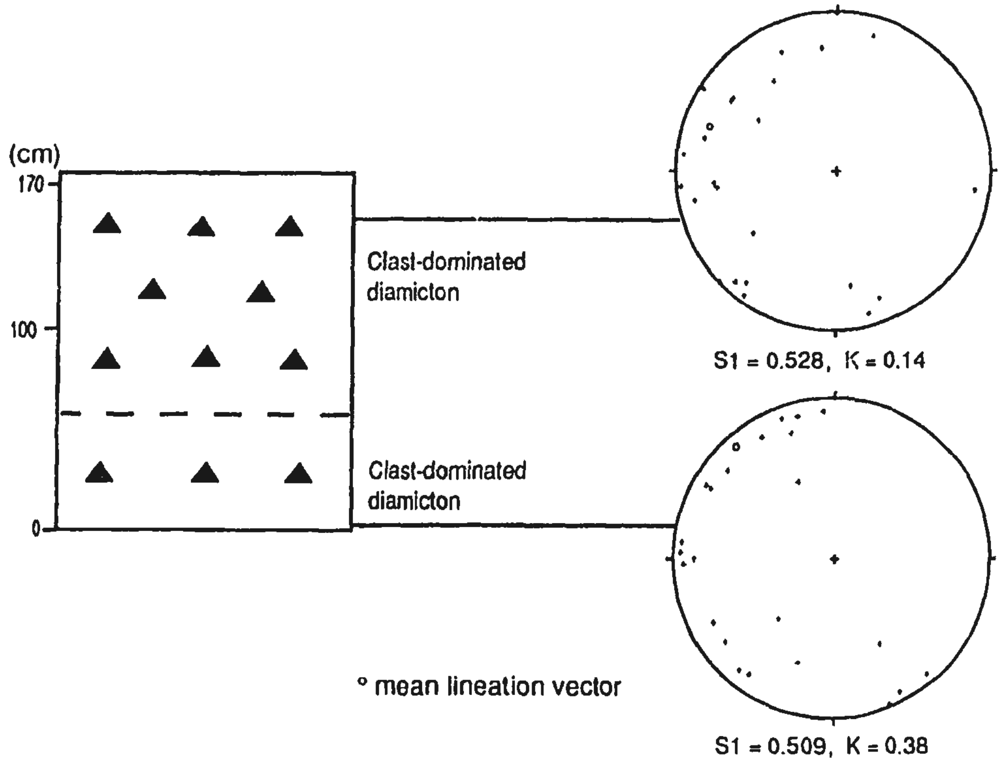


Figure 5.4: Vertical transect through sediments at Shoal Pond (MM-13), with fabric information

704250, 5469200). The exposure is approximately 170cm high and is composed of two units (Figure 5.4).

The lower unit has a minimum thickness of 60cm. The basal contact was not observed. It is a clast-dominated diamicton (approximately 40% clasts), with cobbles dominating. Pebbles and granules are also present. Approximately 68% of the clasts are mafics and ultramafics from the Gander River Complex and approximately 24% are sandstones derived from the Davidsville Group. The matrix of the diamicton is olive grey (Munsell: 5Y 3/2, mf) and generally loose. It is composed of 68% sand, 20% silt, and 12% clay.

Sand lenses are present in the unit, although these are obscured by iron and manganese staining. These lenses are up to 40cm long and are usually 2-5cm thick but their lateral extent is unknown. In cross-section, the lenses have a slight sinusoidal shape. Their upper and lower contacts tend to be abrupt and undulating. The absence of folding and faulting indicates that these lenses were not subjected to glaciotectonic deformation. A clast fabric from this unit has a poor orientation with an S_1 value of 0.509 and a K value of 0.38. It has a mean trend of 319.3° and a mean plunge of 7.7° (Figure 5.4).

This lower unit grades into an overlying clast-dominated diamicton approximately 110cm thick which is slightly more compact. Clasts range from pebbles to cobbles and are predominantly from the Gander River Complex (42%) and the Davidsville Group (36%). Many of these clasts are friable and break up easily in the hand. Some have weathered to clay particles. The matrix is composed of 72% sand, 26% silt, and 2% clay. Lenses of sorted material are much more common in this uppermost part of the section. These consist mainly of silt and have a lateral extent of 30-40cm, and are generally 2-3cm thick. The lenses are draped over some of the clasts. A clast fabric from this unit has a poor orientation with an S_1 value of 0.528, and a K value of 0.14. The mean orientation is 289.1° and the mean plunge is 17° (Figure 5.4).

Interpretation

The sediments at MM-13 are interpreted as till. Striated clasts and lenses of sorted material are generally indicative of basal till (Krüger, 1979; Shaw, 1979). Sandy till with weakly developed fabrics and broken and friable clasts can be indicative of flowing water (Drake, 1971; Flint, 1971; Krüger, 1979), and friable clasts can only be preserved under transport and depositional conditions where they are not subjected to extensive reworking. Such clasts would not survive extensive abrasion during lodgement, but would survive the melt-out process (e.g. Lawson, 1979; Shaw, 1982; Haldorsen and Shaw, 1982). Consequently this deposit may be a basal melt-out till.

The lenses at Shoal Pond are undeformed and therefore may be contemporaneous with till deposition; temporary concentrations of water beneath the ice or within the ice allowed enough size sorting to form the lenses (Hicock *et al.*, 1981; Haldorsen and Shaw, 1982; Shaw, 1982; Domack, 1984; Ashley *et al.*, 1985). Characteristically, lenses in melt-out till have convex upwards upper surfaces, showing that the unit was deposited in a closed conduit. Some authors have suggested that lenses of sorted material may originally have been proglacial and subsequently ploughed up by ice and incorporated into till (Lundqvist, 1959). This would result in deformed lenses.

Clasts falling through flowing water may have their orientations substantially altered (Boulton *et al.*, 1974), thus resulting in a weakly oriented clast fabric. This, however, seems unlikely at the Shoal Pond site as the cavities in which the sorted sediment was deposited were small, perhaps reaching a maximum thickness of 1cm. Thus the preferred orientations of the clasts in the till should mimic that of their original englacial positions. Weak fabrics do not preclude a melt-out origin (cf. Lawson, 1979; Dowdeswell and Sharp, 1986).

Both clast fabrics have girdle distributions and low S_1 values and are therefore not good indicators of ice-flow. Ice flowed from the south in this area as indicated by the presence of large concentrations of clasts from the Gander

River Complex which crops out directly to the south of the exposure. A precise direction, however, cannot be determined.

In summary, this site represents the deposition of melt-out till by a glacier moving northwards.

5.5 CARMANVILLE SOUTH 2 (MM-30)

Description

A well exposed succession of glacigenic sediment is located on Route 330, 4.5km south of Carmanville South (MM-30). It is part of a low-lying hummock (UTM 699500, 5470350). A backhoe excavation revealed 4.1m of sediment with four distinct units (Figure 5.5) (Plate 5.2). The top of the section lies between 43m and 46m asl, below the marine limit which is at least 57m asl.

The basal unit has a minimum thickness of 1.5m and is composed of open-work moderately sorted granule gravels. No current-produced structures are evident but the unit generally fines upwards. The basal contact was not observed.

Abruptly overlying this is a 25cm-thick unit of relatively compact, dark olive-grey (Munsell-5Y 3/2, mf), clast-dominated (50% clasts) diamicton, containing faceted clasts. It has approximately 30% greenschist, 25% basalt, and 20% ultramafic clasts. Other clast types include slate, quartzite, granite, and diabase. The diamicton matrix is dominated by sand and granules (80%), with small amounts of silt (4%), and clay (16%). Two fabric analyses from this unit have moderate orientations with S_1 values of 0.683 and 0.653, and K values of less than 1 (0.78 and 0.66) (Figure 5.5). These fabrics were taken within 3m of each other laterally .

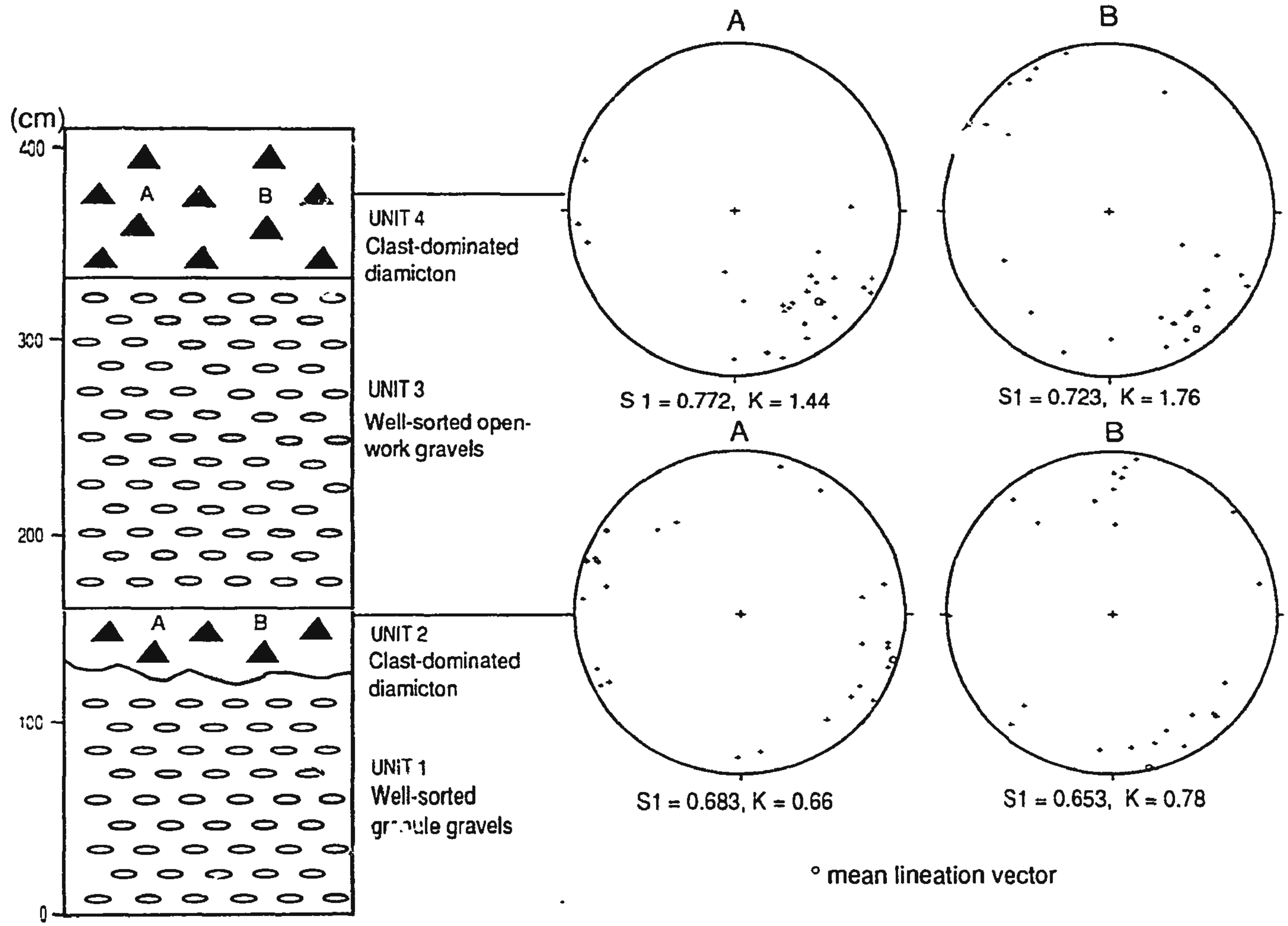


Figure 5.5: Vertical transect through sediments at Carmanville South 2 (MM-30). Fabric information is provided for the upper and lower diamictons.

Plate 5.2: Section at Carmanville South 2 (MM-30). Till (Unit 4) overlies beach gravel (Unit 3) along a horizontal contact. Units 1 and 2 are concealed below the slumped material.

Abruptly overlying the diamicton is a 165-cm thick unit of granule gravel, which generally fines upwards. It is capped by approximately 20cm of much larger pebbles and cobbles (5-10cm A-axis), which are disc-shaped with some blades (Plate 5.3). The gravel is open-work and shows some evidence of crude horizontal planar bedding. There is no clay in the unit and only minor quantities of silt and sand are present (less than 5%).

Abruptly overlying the gravel is a 75-cm thick, moderately compact, light olive green (Munsell-5Y 4/4, mf), matrix-dominated (15% clasts) diamicton. The clast assemblage in this upper unit consists of approximately 40% ultramafic clasts, 20% greenschist, 10% basalt, and minor percentages of greywacke, quartz, and siltstone. The matrix is finer grained than the lower diamicton unit,

Plate 5.3: Unit 3 at Carmanville South 2 (MM-30) shows characteristics of beach gravels. The gravel is moderately to well sorted, shows evidence of crude horizontal bedding, and contains sub-rounded to well rounded discs and blades. These may have been deposited in the upper beach face area just below the berm crest. (cf. Komar, 1976. Massari and Parea, 1988).

with 32% sand, 48% silt, and 20% clay. Mineralogical analysis of the 2ø fraction of the diamicton showed it to be composed of at least 90% crushed shale. Approximately 8% is quartz and feldspar grains, and the remainder is made up of minor quantities of talc, feldspar and chlorite. This unit contains many striated and faceted clasts with bullet shapes predominating. Fabric analysis (Figure 5.5) shows the clasts to have strong orientations with S_1 values of 0.723 and 0.772, and K values of 1.76 and 1.44. They have mean orientations of 143.2° and 137.2°, and mean plunges of 12° and 24.9°. They are therefore classified as clusters on Woodcock's (1977) graph. These fabrics were taken approximately 4m apart laterally.

Interpretation

The Carmanville South site represents alternation between marine and glacial sedimentation. Unit 1 is composed of open-work gravels which are indicative of deposits reworked by fluvial or marine waters (Komar, 1976). Because the clasts are mainly disc- and blade-shaped, they are interpreted as forming in a coastal rather than a fluvial environment. As the clast size is relatively small, the beach environment may have been a relatively low energy system. Comparison of the sediment to modern medium and high energy beaches at Aspen Cove, however, indicates that various clast sizes can be present in any one beach environment. At Aspen Cove, the seaward zone of the modern beach is dominated by cobbles (20-30cm A-axis) which are well imbricated, dipping seawards. There is a very abrupt landward change to small granule gravels and then a general increase in the clast size towards the back beach environment (Komar, 1976; Massari and Parea, 1988). The deposits in this lower unit probably represent the depositional environment at the seaward zone of the beach. This finer deposition in the seaward zone is common in many pebble beach environments (Komar, 1976; Massari and Parea, 1988). Also, as the unit is moderately sorted, it cannot have formed in the backbeach area which is dominated by well sorted gravels, in distinct beds (Massari and Parea, 1988).

The diamicton of unit 2 probably does not represent a primary till. Clast fabrics are moderately oriented with medium S_1 and K values. Mean orientations differ substantially between the two fabrics: 107° and 166°. This would suggest that this represents a slump, a collapse deposit, or a debris flow rather than a till. The abrupt and irregular contact between the two units also suggests this. The area directly to the south is of slightly higher elevation (~2-3m) and may have been subjected to erosion when sea-levels were higher. This would have involved waves breaking against a small unconsolidated bluff, undermining it to cause slumping of the overlying sediments.

Bank-collapse deposits usually have erosional curved or planar contact surfaces (Gibling and Rust, 1984; Postma, 1984; Rust and Jones, 1987; Williams and Flint, 1990). At this location, however, the contact is undulatory and it is not erosive. Failure planes should also be observed (cf. Gibling and Rust, 1984), but at this site they are not. It is possible that the lateral extent of the exposure is too limited to observe the failure planes. Alternatively, the deposit may simply represent blocks toppling from an unconsolidated bluff, and thus the contact would be irregular and failure planes would not be observed.

The sediment may also represent a debris flow deposit which would produce the irregular but abrupt basal contact. However, there is no substantially elevated topography surrounding this area. Debris flow could only have occurred, therefore, if ice was nearby, creating the necessary elevation needed to induce debris flow activity. The clast fabric orientation is moderate which is characteristic of many debris flow deposits (Lawson, 1979; Rappol, 1985; Dowdeswell and Sharp, 1986), but clast plunge angles are consistently low, which is uncharacteristic (Lawson, 1979; Rappol, 1985). There is also no evidence to suggest that ice was nearby when this unit formed. Thus the most likely origin for this unit is thus as blocks toppling from an unconsolidated bluff.

The sediment in unit 3 is similar to open-work beach gravels exposed in the modern beach at Victoria Cove. The finer gravels at the base of the unit were probably deposited at the seaward zone of a beach (cf. Komar, 1976;

Massari and Parea, 1988), and are likely part of the same sequence as unit 1. Beach sedimentation was interrupted by the slope failure represented by unit 2. The larger disc-shaped clasts in the upper part of the unit were probably part of the upper beach face just below the berm crest, indicated by the low angle of dip of the clasts (cf. Komar, 1976; Massari and Parea, 1988). This unit may, therefore, represent continuous marine deposition as sea-levels were falling.

The two fabrics from unit 4 have strong S_1 values, and K values greater than 1 which, using the classification of Woodcock (1977), shows that these are unimodal and are therefore good indicators of ice-flow directions (cf. Lawson, 1979), and are of a basal origin (cf. Lawson, 1979; Dowdeswell and Sharp, 1986). The till matrix is fine (68% silt and clay) which suggests that this may be a lodgement till using Haldorsen's (1983) criteria and thus abrasion was the more common comminution process. However, the terminal grades of the minerals associated with the clasts in the till (mainly ultramafics) are concentrated in the silt size fraction (cf. Dreimanis and Vagners, 1972) and could have thus been subjected to both crushing and abrasion. The matrix texture at this site cannot, therefore, be considered diagnostic of till genesis because of the bedrock inheritance.

This till has many of the characteristics of a lodgement till (bullet- and barrel-shaped clasts, a fine matrix, and strong unimodal fabrics) but it cannot be clearly distinguished from a melt-out till. It, therefore, can only be determined to be of basal origin. At this site ice was flowing from the southeast to the northwest. A northwestward ice-flow was detected in the striation record (see Chapter 6).

The stratigraphic succession at Carmanville South 2, with basal till overlying beach deposits is of regional significance. This indicates that there was a readvance of ice over the beach sediments following initial deglaciation. This stratigraphy is confined to this site and other directly adjacent sections. Although there is no direct chronological control on this unit, it must have been deposited after sea-level reached its maximum elevation (at approximately 12,500 BP), and before complete glacier retreat at approximately 10,000 BP. Sea-levels are known to have been above present between 12,500 and 10,000 BP (Shaw and Forbes, 1990a), and ice had retreated onland by 12,000 BP (Shaw *et al., in review.*) The beach deposits therefore formed between approximately 12,000 BP and 10,000 BP.

There are two alternatives for the age of this upper till unit. It may have formed during a late oscillation of the main Late Wisconsinan ice prior to approximately 11,500 BP. Alternatively, this event may be correlative with the Younger Dryas cooling event, dated to between 11,000 and 10,000BP in Newfoundland (Anderson and Macpherson, 1994), at the end of the Late Wisconsinan. The till is areally restricted, yet emerged beach deposits are areally extensive in the coastal environment of northeast Newfoundland generally overlying glacially deposited sediments. It can be determined that the sea-level responsible for the formation of the beach deposits is older than 11,400 BP by comparison with the Bay of Exploits region (MacKenzie, *in preparation*)(see Chapter 10). The till overlying these beach deposits therefore formed after this date, and is probably co-incident with the timing of the Younger Dryas cooling event recorded elsewhere in Atlantic Canada (e.g. Stea and Mott, 1989). Therefore, regardless of the precise mechanism responsible for glacial advance, it occurred between 11,400 and 10,000 BP.

The till at the Carmanville South site was deposited subsequently to the beach formation (12,000-10,000 BP) which occurred after till formation in other parts of the region. The restraints placed upon this unit suggest that it can be perhaps placed within the Younger Dryas which occurred on the Island of Newfoundland between 11,000 BP and 10,000 BP (Anderson and Macpherson, 1994). Deposition probably occurred from a local ice cap emanating from the higher land approximately 25-35 km south-southeast of Ragged Harbour indicated by the presence of greenschist clasts in the till originating in the Gander Group. All other known glacial events had their origins to the west or southwest of the site.

5.6 ISLAND POND SITE (MM-29)

Description

MM-29 is situated in the centre of the study area approximately 1 kilometre east of Island Pond (UTM 699500, 5467200). The exposure is over 2m thick and displays three main units (Figure 5.6), (Plate 5.4). The basal unit consists of at least 1 metre of deformed slate. The basal contact was not observed. The slate is folded with the most common forms being recumbent and s-folds. The folds have commonly been displaced by thrust faults on the scale of a few centimetres. Directional measurements of slate clasts along these faults were undertaken, and a thrust direction of 038° was determined. The local bedrock surrounding the area is not deformed.

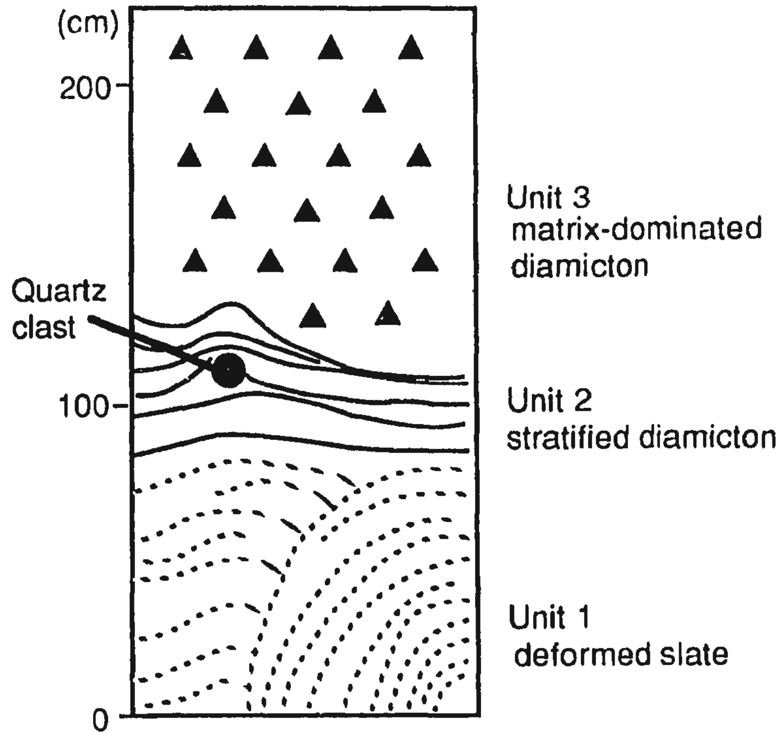


Figure 5.6: Vertical transect through sediments at Island Pond

Abruptly overlying the deformed slate is a relatively thin (average 25 cm thick) stratified diamicton unit that is moderately compact. This unit is extensively brecciated. Laminae of very dark greyish brown diamicton (Munsell 2.5Y 3/2, mf) are interstratified with dark greyish brown diamicton (Munsell 2.5Y 4/2, mf). Each lamina has a maximum thickness of 1 centimetre. The unit contains substantial clay (approximately 20%) and silt (approximately 26%). Clasts are dominated by shattered slate, representing approximately 98% of the

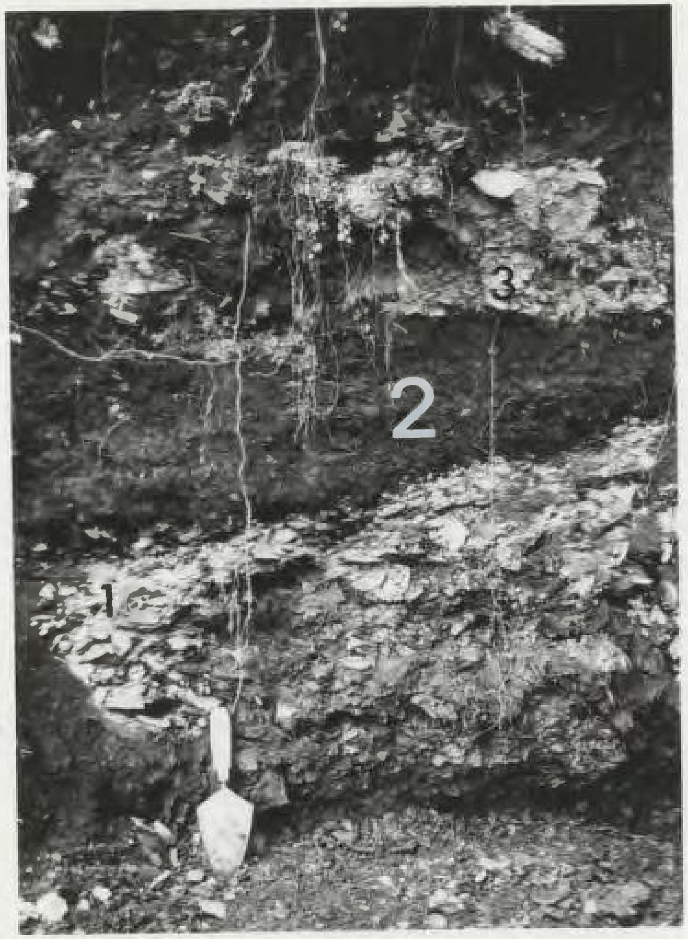


Plate 5.4: Section at Island Pond (MM-29). Unit 1 is composed of folded and faulted slate, unit 2 is composed of stratified brecciated diamicton, and unit 3 is composed of till containing small rafts of slate. Exposure of unit 3 is poor.

total. The 2ø grain-size fraction of the matrix is dominated by at least 95% crushed shale and slate fragments.

Larger, more resistant clasts have been incorporated into this unit and show evidence of directional shearing. For example, a quartz clast (3.5cm A-axis) has been pushed into the darker brecciated material, deforming the surrounding beds (Plate 5.5). The clast is covered with many small striae.

Plate 5.5: A quartz clast at Island Pond (MM-29), has been pushed down into the underlying diamicton. It has been subjected to rotation indicated by the striations on the surface of the clast. The strata have been deformed on either side of the clast (Figure 5.7b). The trowel is approximately 25cm long.

Directly NNE of the clast is a zone where the laminae are wider than SSW of the clast. There is also evidence of folding of the laminae directly to the SSW of the clast.

This second unit grades vertically into an 80-cm-thick unit of matrix-dominated diamicton. The lower part of this diamicton is relatively compact and olive brown (Munsell: 2.5Y 4/4, m/f). The upper part of the unit has been affected by pedogenesis, and is consequently less compact. It is yellowish brown (Munsell: 10 YR 5/8, m/f). Approximately 90 percent of the diamicton is made up of locally derived shattered slate. The matrix contains 52% sand, 24% silt, and

24% clay. Also present in this unit are very small rafts of shattered slate. They have a maximum thickness of 10cm and maximum lengths of 40cm. Apart from the rafts of shattered bedrock, clast content in the unit is low (<5%) and consequently clast fabric analysis was not undertaken.

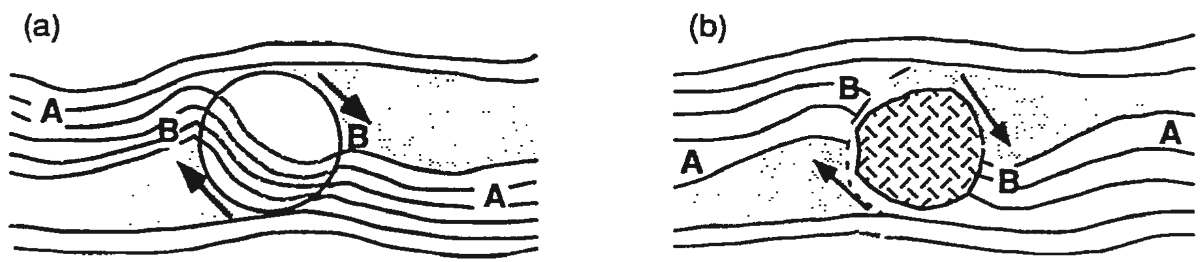
Interpretation

An almost identical sedimentary assemblage to that found at MM-29 was described by Hart and Boulton (1991) and interpreted as being the result of glaciotectonic deformation. A lower zone consisted of material which had been overturned and folded in a down-ice direction. An intermediate zone oflaminated diamicton contained mainly locally derived material but also had a far traveled component. An upper zone contained more far traveled material and was generally homogeneous in appearance. This sequence was interpreted by Hart and Boulton as having been formed by continuous deposition and deformation. Firstly, when ice advanced across the area, sediment was subjected to simple shearing (zone of overturning). Then, as it deformed the sediment, the ice also began to deposit more far-traveled debris (zone of shearing). Finally, the sediment became so deformed that individual beds could not be distinguished (homogeneous zone).

A similar zone of overturning, a zone of shearing, and a homogeneous zone can be identified at MM-29. Unit 1 can be regarded as the zone of overturning. It is made of local slate, a relatively weak rock that will shear easily. Unit 1 would have formed as ice was advancing into the area and represents proglacial thrusting as described by Moran (1971), Aber (1979), Moran *et al.*, (1980), Evans (1989), and Hart (1990). As ice advanced over this, material began to be deposited and extensive shearing and brecciation occurred due to alternating compressional and extensional flow. Finally, the upper sediment was subjected to sufficient deformation to produce the laminae which are so thin that the unit appears homogeneous. Although this unit appears homogeneous, shearing occurred during deposition, as indicated by the thin rafts of bedrock. The upper two units can be termed `deformation till' using the definition of Hicock and Dreimanis (1992).

The style of deformation (brecciation) present in unit 2 and the many small striae on the surface of the quartz clast suggests that it has been subjected to rotation. Rotation is indicative of shearing (Simpson and Schmidt, 1983; Hart and Boulton, 1991; Hicock, 1991; Hicock and Dreimanis, 1992). Consequently, the approximate direction of shear and therefore also the approximate direction of ice movement can be determined. The direction of shearing and folding in the basal slate is along azimuth 038°. Hence, ice movement was towards the northeast.

Simpson and Schmidt (1983) studied the more resistant grains in foliated gneisses which have also been subjected to ductile shearing, and noted that the foliation planes on the upper left-hand side and lower right-hand side of all clockwise rotating grains are usually more closely spaced (Figure 5.7a). Often there are small associated microfolds. Usually, the most recently deposited material on the upper right-hand side of the rotating grain shows more widely spaced foliation planes. Hart and Boulton (1991) noted similar phenomena in glacigenic material which has been subjected to shearing. Similar forms were observed in the laminae surrounding the quartz clast at site MM-29 (Figure



(after Simpson and Schmidt, 1983)

A: closer spacing of pre-existing foliation planes on the upper left-hand and lower right-hand side of the rotating grains.
B: microfolds in laminae next to rotating grain.

Figure 5.7: (a) Diagrammatic representation of a grain subjected to ductile shear in gneiss. The grain rotates and the adjacent laminae become deformed; **(b)** A quartz clast in till at Island Pond shows evidence of rotation and is evidence of ductile shear in the till.

5.7b). The style of deformation is analogous to that producing mylonite in lithified material (Hart and Boulton, 1991).

The lack of clasts in the upper two units is significant as coarser, more permeable, clast-rich tills are more resistant to deformation than finer, more impermeable, clast-poor tills (Boulton and Hindmarsh, 1987; Hicock and Dreimanis, 1992). Deposition of fine-grained till may occur in depressions which contain deformable sediment and/or weak bedrock, and may be dominated by subglacial meltwater sheet flow (Hicock and Dreimanis, 1992). Site MM-29 lies in a depression, the present basin of Island Pond. The basin contains relatively weak slate, and may have contained fine-grained glacio-lacustrine or lacustrine deposits prior to ice override. This would explain the fine clast-poor till present at MM-29. All of the above conditions are conducive to shearing between the glacier sole and rigid (probably frozen) sediment or bedrock (Hicock *et al.*, 1989; Hicock and Dreimanis, 1992).

The Island Pond exposure was the only glacigenic exposure determined to have been glaciotectonically deformed. Although this appears to be a very local phenomenon, it is important in determining the glacial history of this area. If these deposits do represent proglacial thrusting, then ice had to have retreated at least as far south as Island Pond prior to the northeasterly readvance.

5.7 CLASSIFICATION OF TILLS IN THE CARMANVILLE REGION (Summary)

The seventeen diamicton exposures in the Carmanville area have been classified into various genetic types based on sedimentary features and clast fabric analysis. These are listed in Table 5.1. Most of the tills in the Carmanville region are homogeneous. The most commonly observed sedimentary features are clay laminae and nodules, sand laminae, and stone lines. Stratified laminae are usually characteristic of melt-out tills (Shaw, 1982), and stone lines have usually been considered to be indicative of lodgement till, deformed lodgement till, and deformation till (Clark, 1991; Hicock, 1991).

Grain size can sometimes be used as a determinant in classifying tills. Abrasion (more common in lodgement tills) produces a significant silt component in a till, whereas crushing (more common in melt-out tills) produces a sandier matrix (Haldorsen, 1983). Therefore, in theory, the lodgement tills in the Carmanville region should be finer than the melt-out tills. However, this concept is difficult to apply to the tills of the Carmanville region, because where they occur, the bedrock along the direction of glacial flow is dominated by shale, slate, and siltstone from the Davidsville Group and the Hamilton Sound Sequence. Most of the minerals derived from these sedimentary and altered sedimentary rocks have their terminal grades concentrated in the silt and clay size fractions (Dreimanis and Vagners, 1972). Consequently, all tills, no matter how they were formed, could potentially have high concentrations of silt in their matrices. Consequently, grain size is not useful in determining till genesis in the Carmanville region.

SITE	S_1	S_3	K	STRIATED CLASTS	BULLETS/ BARRELS	PRIMARY STRUCTURES	PROBABLE GENESIS
11	0.712 0.717	0.017 0.019	1.22 1.78	no	no	clay stringers, sand laminae, extremely fine grained matrix.	basal melt-out
13	0.509 0.528	0.136 0.065	0.38 0.14	yes	yes	clay nodules, sand laminae, sand nodules.	basal melt-out
15	0.718	0.038	0.58	yes	yes	stone line, faceted clasts, clay nodules, fissile towards top of unit.	basal melt-out overlain by lodgement
27	0.814	0.022	0.79	yes	no	stone line, faceted clasts.	basal till
28	0.543	0.64	0.64	yes	no	none	basal till
29	--------	--------	--------	yes	no	laminated diamicton, thin rafts of bedrock, rotating clasts, fine grained matrix.	deformation till
30	0.723 0.772	0.088 0.059	1.76 1.44	yes	yes	sand and silt laminae, faceted clasts.	basal till
31	0.572	0.089	0.39	yes	no	stone line, faceted clasts, sand laminae	basal melt-out overlain by lodgement

Table 5.1: Classification of tills in the Carmanville region.

SITE	S_1	S_3	K	STRIATED CLASTS	BULLETS/ BARRELS	PRIMARY STRUCTURES	PROBABLE GENESIS
32	0.503	0.147	0.42	yes	yes	stone line, faceted clasts, sand laminae.	basal melt-out overlain by lodgement
35	0.843	0.070	10.69	yes	yes	stone line, faceted clasts, sand laminae.	basal melt-out overlain by lodgement
37	0.558	0.157	0.98	yes	no	stone line, faceted clasts.	basal till
41	--------	--------	--------	yes	no	fissile towards top of unit.	basal till
47	0.543	0.132	0.53	yes	no	stone line, faceted clasts.	basal till
48	0.556	0.079	0.28	yes	no	stone line, faceted clasts, high percentages of sand.	basal melt-out
50	0.748	0.097	3.39	yes	yes	stone line, faceted clasts, clay nodules.	basal melt-out
51	0.595 0.434	0.114 0.165	0.77 0.09	yes	no	sand laminae.	basal melt-out
52	0.709	0.022	0.39	yes	no	clay nodules, extremely fine matrix, sand and silt clasts.	basal melt-out

Table 5.1 (continued)

Twenty-nine grain size analyses were undertaken on the tills of the Carmanville region. Of these, thirteen are characterized by more than 50% silt and clay, and 21 contain more than 35% silt and clay (Figure 5.8). These textures are much finer than the generalized till textures for the Appalachian region which are generally dominated by more than 60% sand, with till finer than this being typically composed of reworked glaciomarine or glaciolacustrine deposits (Scott, 1976). In the Carmanville region, grain size is very much a function of bedrock source.

Comparisons of differing diamicton units within a single exposure can aid in genetic analysis. Each of the upper and lower diamictons at MM-15, MM-31, MM-32, and MM-35, had similar mineralogies and lithologies, but differing texture. These units probably therefore represent differing genetic processes and distinctions can be made. Many of the diamicton units observed do not display such changes and thus their genesis is more difficult to establish.

On the basis of clast fabric analysis, all of the tills in the exposures investigated in detail in the Carmanville region have been interpreted as basal tills. Whether clast trend orientations are strong or weak, the average clast plunge is low (Figure 5.9). Debris flow deposits may have well-oriented trend patterns, but clast plunges should vary considerably (Rappol, 1985; Dowdeswell and Sharp, 1986; Gravenor, 1986). If the deposits were of supraglacial origin, then many of the clasts should have high-angles of plunge. Also, the mineralogy should suggest distal origins, reflecting the long process of incorporation at the glacier sole and subsequent transport through the ice mass. At most of the exposures studied the finest grain fractions were composed of at least 90% locally derived material.

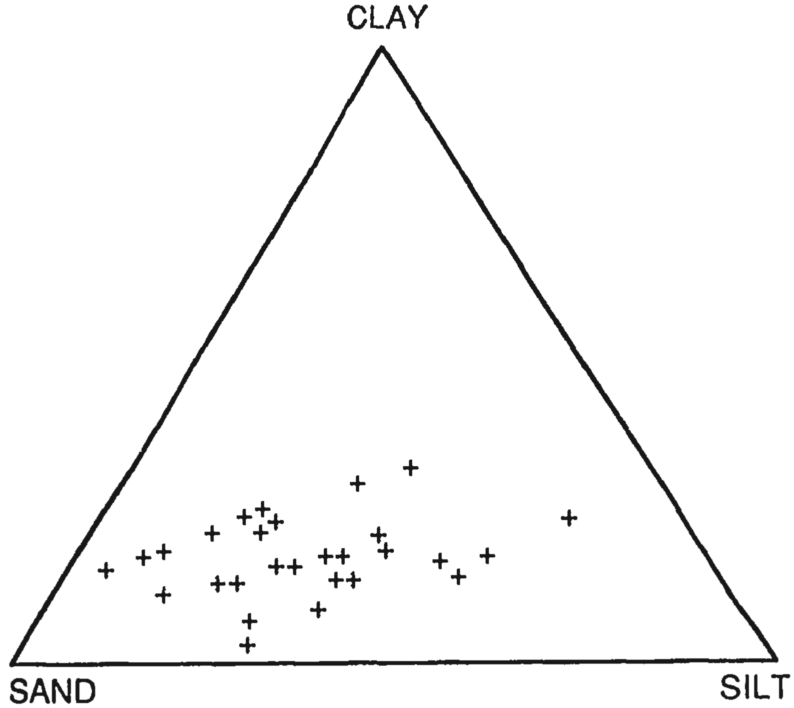


Figure 5.8: Ternary diagram of tills in the Carmanville area

Ten deposits were interpreted as basal melt-out tills. These are located at sites MM-11, MM-13, MM-15, MM-31, MM-32, MM-35, MM-48, MM-50, MM-51, and MM-52. All of these tills contain sorted stratified laminae or nodules which are characteristic of basal melt-out till (Shaw, 1979, 1982, 1987; 1989; Krüger, 1979; Haldorsen and Shaw, 1982; Dreimanis, 1982). Grain size varies substantially between the deposits, with silt and clay accounting for between 24% and 74% of the till matrices. This strengthens the argument that grain size is not a reliable indicator of till genesis in the Carmanville region.

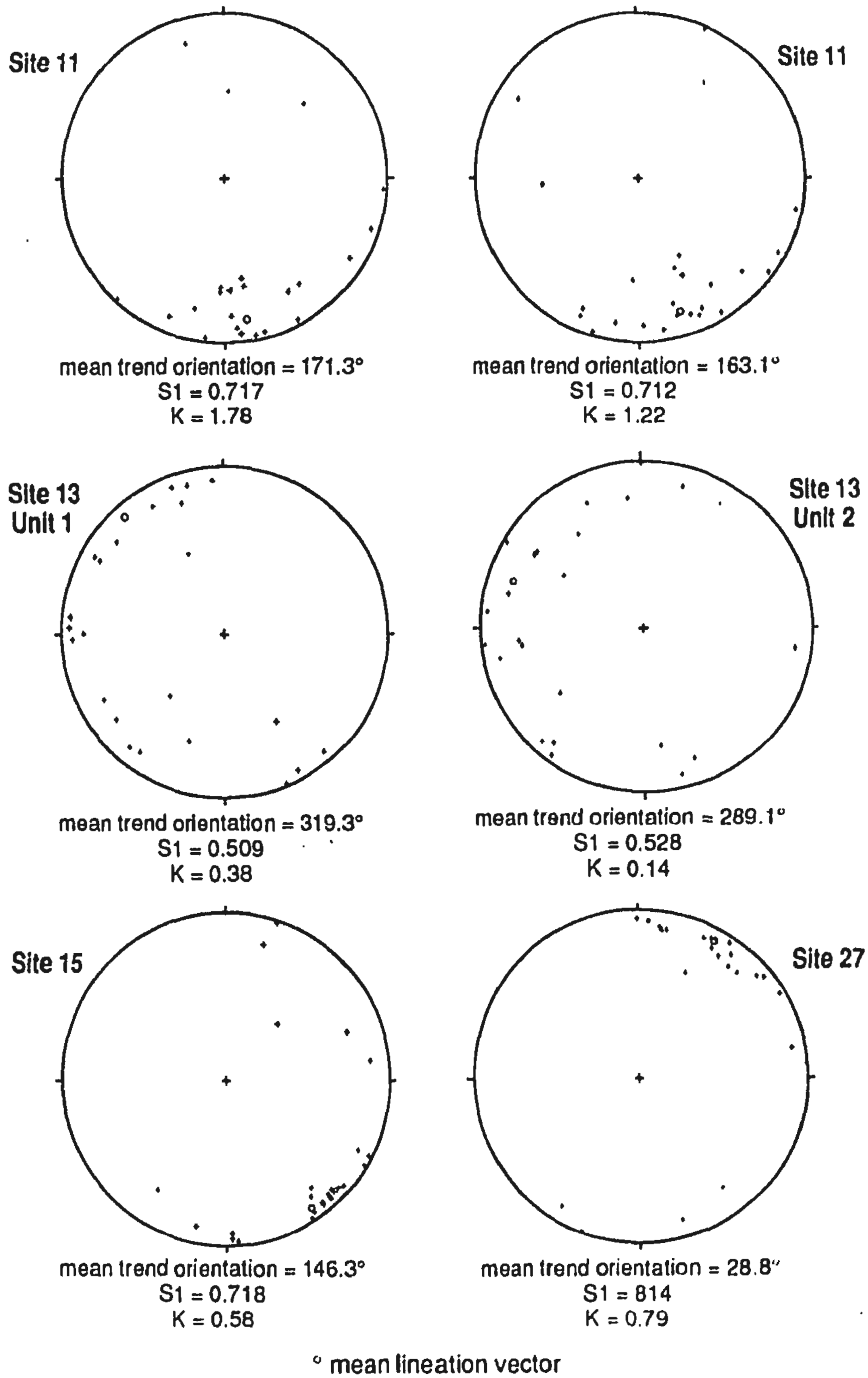


Figure 5.9: Fabric plots of all the tills examined in the Carmanville region.

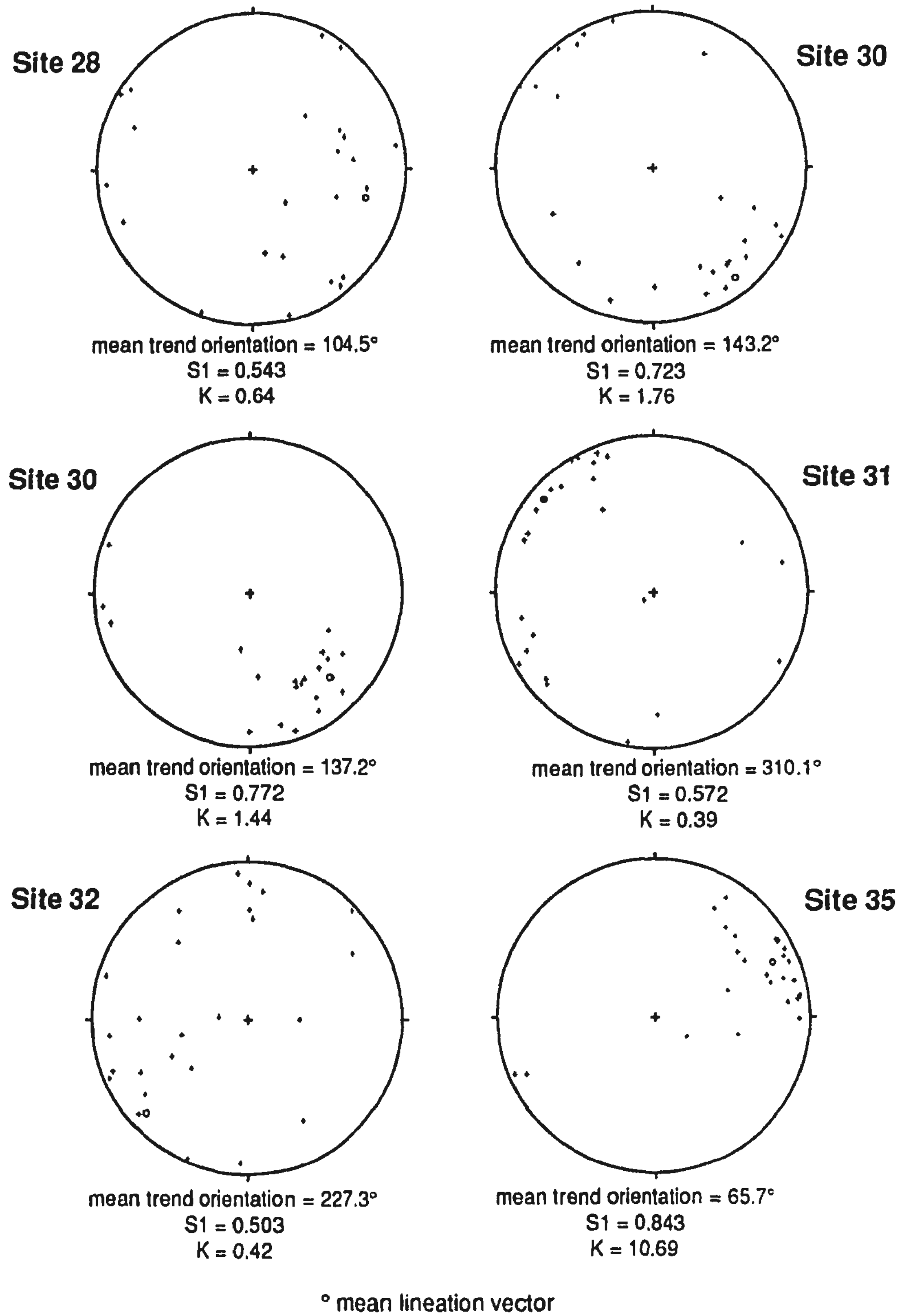
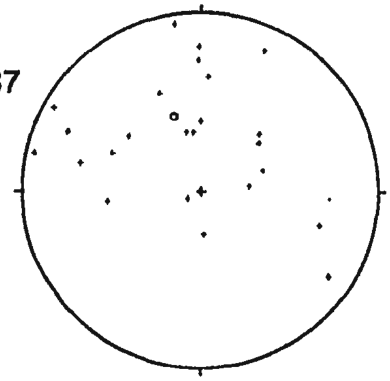


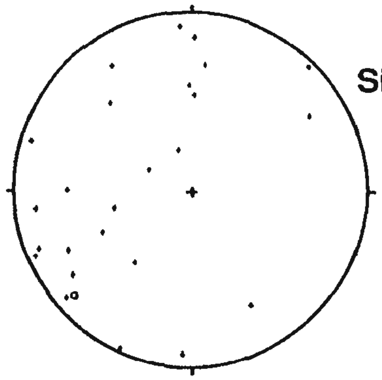
Figure 5.9 (continued)

Site 37



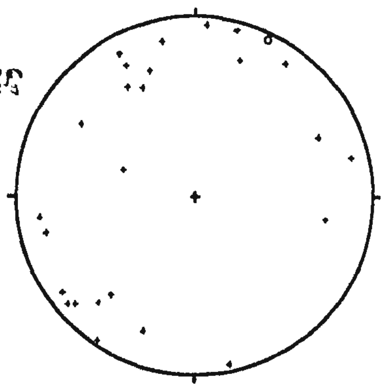
mean trend orientation = 340.3°
S1 = 0.558
K = 0.98

Site 47



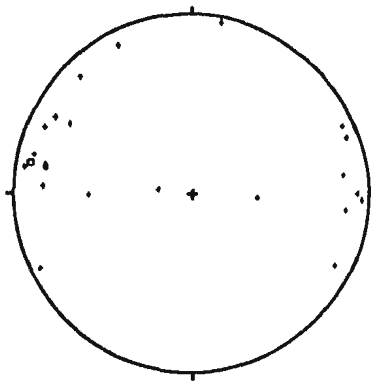
mean trend orientation = 229.1°
S1 = 0.543
K = 0.53

Site 4[illegible]



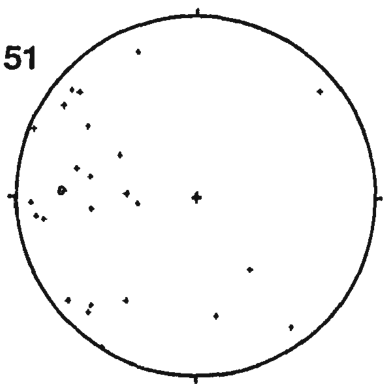
mean trend orientation = 25.4°
S 1 = 0.556
K = 0.28

Site 50



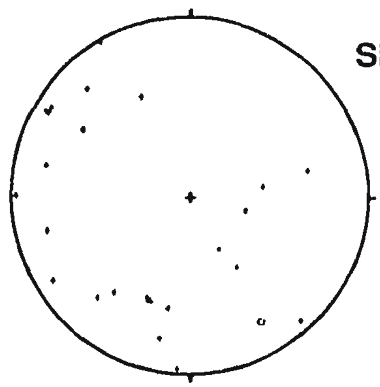
mean trend orientation = 280.6°
S1 = 0.748
K = 3.39

Site 51



mean trend orientation = 272.3°
S1 = 0.595
K = 0.77

Site 51

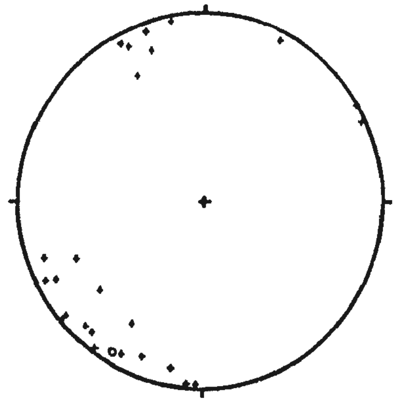


mean trend orientation = 149 7
S1 = 0.434
K = 0.09

° mean lineation vector

Figure 5.9 (continued)

Site 52



mean trend orientation = 211.3
S1 = 0.709
K = 0.39

° mean lineation vector

Figure 5.9 (continued)

Of the ten diamictons interpreted as melt-out tills, fabric analysis showed that only three had strongly unimodal fabrics, with high S_1 values and K values greater than 1 (MM-11, MM-35, and MM-50)(Figure 5.9). Two deposits were relatively well oriented with S_1 values greater than 0.7 (MM-15 and MM-52)(Figure 5.9), but are still categorized in the girdle field of Woodcock's (1977) graph. The other five till deposits in this category have both low S_1 values and K values less than 1. Thus, although basal melt-out tills may have well oriented fabrics, in practice they often do not. Clasts falling through flowing water (laminar and turbulent), and perhaps interacting with other clasts can result in these weakened fabric orientations (May *et al.*, 1980; Boulton *et al.*, 1984; Rappol, 1985). Clast fabric orientations considered in isolation are thus unreliable indicators of basal melt-out till in the Carmanville region.

Four diamicton exposures were determined to contain lodgement till (MM-15, MM-31, MM-32, and MM-35). At each site there is a transition between a lower basal melt-out till and the lodgement till, marked by a clast pavement. These units were determined to be of lodgement origin because of the high content of bullet- and barrel-shaped clasts which are associated with abrasion (more common in lodgement tills) (Haldorsen, 1983); and because of the extremely fine-grained nature of the tills (these deposits were much finer than any other till deposit in the Carmanville region). Silt and clay accounted for between 42% and 88% of the matrices. Direct comparison between these lodgement units and the underlying basal melt-out tills aided in the recognition of genetic differences. At each of the four sites, the grain size of the overlying till was always finer than the underlying till, yet there were no lithological or mineralogical differences. Consequently, both tills were probably deposited by the same ice mass but the mode of clast alteration changed from a process of crushing in the underlying till, to a process of abrasion in the overlying till.

Five deposits were determined to be basal till (MM-27, MM28, MM-30, MM-41, and MM-47). There is not enough evidence to determine whether these are basal melt-out tills, lodgement tills, or deformation tills. Fabrics range from well oriented to poorly oriented. One fabric is strongly unimodal (MM-30 - Carmanville South 2), and one fabric has an S_1 value greater than 0.7 but a K value less than 1. Three other fabric analyses have low S_1 values and K values less than 1. Fabric analysis was not undertaken on MM-41. Grain size also varies, with silt and clay accounting for 33% to 70% of the matrices. Deposits marked by moderately aligned clast fabric patterns could represent melt-out or lodgement tills, whereas those with poorly aligned clast fabrics could represent disturbed melt-out, disturbed lodgement, deformed lodgement, or deformation till.

Clast fabric analysis was not undertaken on MM-29 because of the absence of clasts. It has, however, been interpreted as a deformation till on the basis of lamination, brecciation, rotation of clasts, and rafts of local bedrock in the till. These features are all associated with ice thrusting.

Of the 17 diamicton exposures studied, stone lines (clast pavements) were observed at 9 sections. They occurred at MM-15, MM-27, MM-31, MM-37, MM-47, MM-48, and MM-50, as well as at MM-32 and MM-35 discussed above. At MM-32, MM-35, MM-15, and MM-31, textural changes were observed above and below the clast pavements, with the sediments above the clast pavement being finer than the sediments below. Hence, at these sites differing density gradients above and below the clast pavement could have resulted in differential settling of the clasts to a common level. As indicated in the discussion of MM-32, if differential settling occurred, it is unlikely that these surfaces would be flat, as clasts of different densities occur within the pavement. It was determined that the pavement occurred between the deposition of basal melt-out till and the deposition of lodgement till. Consequently, the pavements could have formed as a direct result of a change from melting ice to advancing ice. Clasts previously melted out from the ice would become lodged next to each other as ice started to advance over them, allowing the upper surfaces of the clasts to become striated.

The stone lines at the other 5 sites are marked by no textural differences in the tills above and below the pavements. These stone lines lie within a structurally and texturally homogeneous till. Consequently, there is no evidence indicating changes in the density gradients of the tills and clasts are thus unlikely to have sunk to a common level in the till. Subglacial deformation may have occurred in the tills although this cannot be absolutely determined. Subglacial deformation may result in an homogeneous appearance (e.g. site MM-29), but homogeneous tills are often also found at sites not subject to deformation. Using similar criteria to those listed above, the tills can only be determined to be of basal origin. The origins of the clast pavements cannot be determined.

In Chapter 4, the tills of the Carmanville region were classified geomorphically, mainly as veneers and hummocks. In this section, however, it has been shown that these hummocks and veneers can be composed of very different genetic till types, and thus the tills of the Carmanville region are far more complex than would be suggested by their geomorphic expression.

CHAPTER 6
ICE-FLOW HISTORY

6.1 - GLACIAL FLOW FEATURES

6.1.1 - Striations

A glacial striation is an erosional scratch or groove on the surface of an ice-abraded rock, produced by the scoring action of rocks transported along the base of glacial ice. Striation orientations indicate local ice-flow directions. Although individual striations reflect the motion of single clasts, the regional pattern of measurements can be used to infer the pattern of glacial flow. Striations are frequently associated with deglaciation, with many of the features formed by glacial advance being destroyed.

Twenty-two single and multiple striation sites were discovered in the Carmanville area. These, combined with previously known sites (Taylor *et al.*, 1991), are interpreted as showing the influence of three ice-flow events: successively eastward, northeastward, and northwestward. Commonly, striations from the eastward flow are cross-cut by striations from the northeastward flow, and both are cross-cut from striations from the northwestward flow. Occasionally evidence of the eastward flow is preserved in the lee side of rock outcrops, the tops of which show evidence of the northeastward and northwestward flows. Two striation directions are well represented at one site on the Aspen Cove highway (Plate 6.1).

Individual exposures show that the eastward flow is well preserved with striations being on the order of a millimetre or more deep. It is not, however, the most extensive. The northeasterly flow is also well-represented in the striation

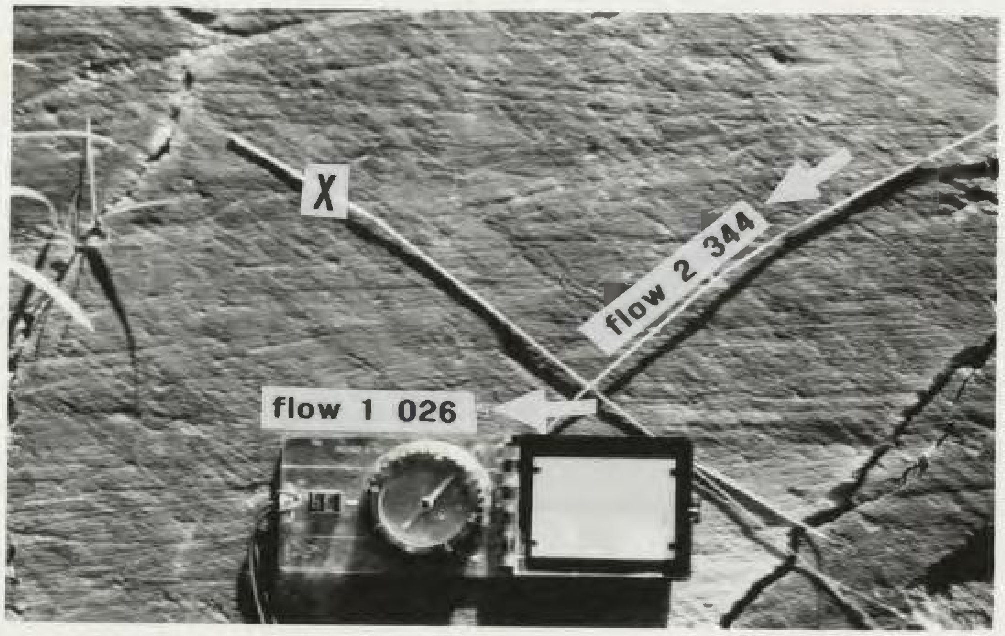


Plate 6.1: Striations (from the Aspen Cove/Ladle Cove highway), preserved in the leeside (sloping surface) of this outcrop represent two ice-flow directions. These are (1) north-northeastward ice-flow (026°), and (2) north-northwestward ice-flow (344°). The northeastward flow is the oldest as striations from it are cross-cut by striations from the northwestward ice-flow. The straw marked with an "x" does not represent a striation orientation. The compass is for scale only and cannot be used to determine ice-flow directions, as the surface it is on is sloping.

record and is found in all parts of the study area. The northwestward flow is less well represented with only minor "scratches" evident. No evidence for this flow was found on the western side of Gander Bay or in the Carmanville, Frederickton, and Mann Point areas. Striation orientations are shown in Appendix 3.

6.1.2 - Roches Moutonnées

Roches moutonnées are asymmetrically eroded rock bodies with a gently sloping smooth abraded surface developed on the stoss side, and a rougher, steeper slope developed on the lee or down-glacier side. The long axes of roches moutonnées are oriented parallel to ice-flow direction. They are produced by the action of advancing ice which moulds and abrades the up-ice side of a rock outcrop. A cavity is formed on the down-ice lee-side which is plucked by frost action facilitated by the low pressure conditions, and thus the lee-side becomes steeper than the stoss side (Sugden and John, 1976). The form of roches moutonnées is closely related to bedrock structure, especially dip and strike of bedding planes, layering, joints, and other lines of structural weakness (Gordon, 1981; Lindström, 1988). Where joints are few and small, large roches moutonnées will be formed. Where there are many large joints, smaller roches moutonnées will be formed (Rastas and Seppälä, 1981).

Although roches moutonnées can be developed by any flowing glacier, they are most prominent in areas where the trends of rock strata or structures are aligned parallel to glacial flow. Sometimes the bedrock weaknesses are aligned in such a way that roches moutonnées form transverse or obliquely to glacial flow (Gordon, 1981). This usually occurs where joint systems exist in the rock oblique to the bedrock strike or where the bedrock strike is at a relatively low angle. Thus, at certain locations roches moutonnées may be a result of glacial-induced fracturing alone rather than of moulding and fracturing (Gordon, 1981). In many areas, the initial direction of glacial flow is modified by the pre-existing bedrock topography. Consequently, roche moutonnée orientations in areas of multiple glaciation commonly reflect only a single flow event, marked by ice movement parallel to the structural weakness.

Roches moutonnées occur extensively throughout the study area, and they range from approximately 20m to 80m in length. They all have strong northeast to north-northeast orientations which parallel the strike of the bedrock. The youngest north-northwest flow, advancing approximately normal to the direction of structural weakness, apparently did not modify the previously formed roches moutonnées significantly. The earlier easterly event had an erosional effect on the landscape as indicated by striations. This ice-flow may have abraded the leeside of some outcrops although the trend of the resultant roches moutonnées may have been north-northeast. No features trending eastward were detected in the Carmanville area.

6.1.3 - Crag-and-Tails

Crag-and-tails are elongate ridges of rock outcrop and glacial debris which form parallel to ice-flow direction. They have a steep end (crag) which faces up-ice. This is usually made up of resistant rock which obstructs glacier movement, thus creating a cavity and inducing deposition of the sediments (tail) directly down-ice of the bedrock crag.

Crag-and-tails occur rarely in the study area. They occur directly to the northeast of Shoal Pond and are oriented in northerly to northeasterly directions. They range from approximately 50m to a kilometre in length.

6.1.4 - Rogen Moraines

Rogen moraines are believed by some authors to form transverse to ice-flow direction (Shaw, 1979; Bouchard, 1989). They normally occur as a series of arcuate ridges concave in the down-ice direction. The origin of rogen moraine is still controversial (Lundqvist, 1969; Minell, 1977; Shaw, 1979; Bouchard, 1989; Fisher and Shaw, 1992) but the fabric patterns, sediment texture, and clast lithology of the materials indicate formation at the glacier sole. Strong fabric orientations parallel to regional ice-flow are usually, but not always, present within rogen moraine sediments.

One possible origin of rogen moraines is deposition transverse to ice-flow as a direct result of localized compressive stresses allowing shearing and stacking of slices of near-base englacial ice (Bouchard, 1989). Moraines formed by basal shearing are clear indicators of ice-flow directions.

Rogen moraines may also form as a result of fluvial processes and debris flow activity as suggested by Fisher and Shaw (1992). This process involves the formation of cavities below the ice by subglacial meltwater erosion and the subsequent filling of these cavities with sediment. The moraines could also have formed due to subglacial erosion of pre-existing sediment in a manner similar to the formation of drumlins (Shaw and Sharpe, 1987; Shaw *et al.*, 1989). In both cases the features form transverse to meltwater flow, which in a deglacial environment may not be the same as the ice-flow direction.

Rogen moraines occur at several locations in the Carmanville region. The most extensive belt in the southeast part of the study area covers an area of approximately 120km^2. Individual moraines are between 100 and 1000m long and attain maximum widths of 150m. They rarely exceed 20m in height. On aerial photographs they have a ribbed appearance with the horns of the moraines pointing in a northerly to northeasterly direction. These features may, therefore, indicate northward to northeastward flowing ice, northeastward flowing meltwater, or both.

Smaller areas of rogen moraines are located at Island Pond, in Middle Arm and Eastern Arm, and near Aspen Cove. All of these areas are marked by topographic lows. In Middle Arm and Eastern Arm, rogen moraines form linear islands. Flow directions in all areas were northerly to northeasterly.

6.1.5 - Drumlins

Theories on the origin of drumlins are controversial. They can be composed of sand and gravel, laminated clay, till, or solid rock (Menzies, 1979). Their most notable traits are their streamlined form with a blunt steeper slope at the up-ice side and a long narrow slope on the down-ice side. There are, however, many deviations from this form (e.g. Doornkamp and King 1971;.Shaw and Kvill, 1984; Shaw *et al.*1989).

Researchers have traditionally linked drumlin form directly to the direction of ice-flow, the deposition of till, and more specifically with the deposition of lodgement till (Boulton, 1979; Menzies, 1979). Shaw (1983), Shaw and Kvill (1984), Shaw and Sharpe (1987), and Shaw *et al.*, (1989) showed that many drumlins are composed of stratified material or even bedrock, and thus these cannot have formed due to lodgement processes. Shaw and his co-workers noted how drumlin form was very similar to inverted melt-water erosional marks as described by Allen (1971). This led Shaw (1983), and Shaw and Kvill (1984), to build a model whereby subglacial meltwater produced similar erosional marks on the underside of an overriding glacier due to flow separation. Drumlins are therefore the casts of these erosional marks and formed when the cavities filled with sediment.

Shaw and Sharpe (1987), and Shaw *et al.* (1989), noted how many drumlins are eroded into bedrock. They compared these with small scale scour remnant ridges as described by Allen (1982). These features are commonly termed rat tails and form on the downstream side of an obstacle. The obstacle can be any resistant inclusion such as a pebble, or a garnet or quartz crystal. Directly upstream of the obstacle is a large concentric scour, and directly downstream is a ridge paralleled by two furrows. Drumlins can be formed in the same way. If large sheet floods were to exist subglacially and were responsible for eroding upwards into the ice, then it is possible that the meltwater could erode down into the underlying sediment or bedrock to form drumlins. Shaw and Kvill (1984) suggest that meltwater flow directions may be different to ice-flow directions but the characteristics of the two flows and their effects on the landscape is at present unknown. Consequently, like rogen moraines, these features may be indicative of meltwater flow directions rather than ice-flow conditions and thus may not be reliable indicators of ice-flow directions.

Drumlins occur in the Carmanville region to the south and east of Island Pond. They are generally oriented southwest to northeast. Other drumlins were noted at the head of Gander Bay below sea-level (Shaw *et al.*, 1990). These are between 150m and 800m long and range between 5m and 20m in height. These drumlins may represent northeastward ice-flow, northeastward meltwater flow, or both.

6.2 - CLAST FABRIC

When a clast is incorporated into glacial ice, the long axis is commonly aligned parallel to the direction of ice-flow, dipping up-ice (Dowdeswell and Sharp, 1986). Well-oriented clast fabrics from glacial sediments are considered to be good palaeo-ice-flow indicators (Dowdeswell and Sharp, 1986). Strong unimodal or bimodal fabrics usually indicate basal melt-out till or lodgement till (Lawson, 1979). These may, however, be subsequently reworked resulting in a more random fabric (Boulton, 1971; Lawson, 1979). Subaerial debris flows can result in less aligned girdle distribution clast alignments (Boulton, 1971), and care must be taken in distinguishing these deposits from primary till deposits. (See Chapter 3: Section 3.4 for clast fabric techniques).

Twenty one clast fabrics were obtained from diamictons throughout the map area. The principal eigenvalues of these range from 0.434 (randomly oriented) to 0.843 (unimodal). The K-values of these fabrics range from 0.09 to 10.69. Using the methods of Woodcock (1977) these values can be plotted two-dimensionally (Figure 6.1).

Six of these have K-values greater than 1, ranging from 1.22 to 10.69. The principal eigenvalues of these fabrics range from 0.712 to 0.843. These fabrics are unimodal and are thus interpreted to represent the local ice-flow directions (cf. Lawson, 1979). All of these are interpreted as basal melt-out till although two sections are capped with lodgement till. Four fabrics have a northwest orientation, one has a northeast orientation, and one has a easterly orientation. These fabric plots are shown in Figure 6.2.

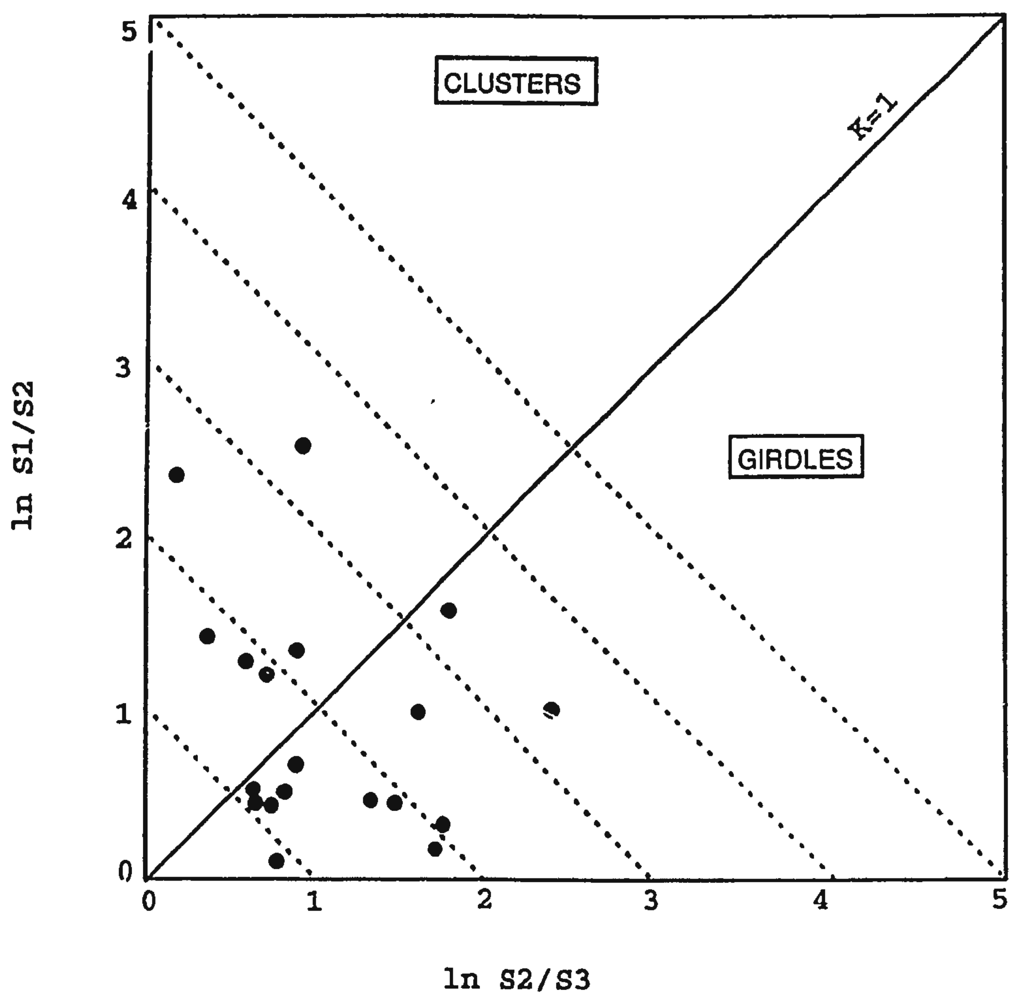


Figure 6.1 : Diagram of clast fabric eigenvalues using the method of Woodcock (1977). The y-axis is the logarithm of the ratio between S1 and S2, and the x-axis is the logarithm of the ratio between S2 and S3.

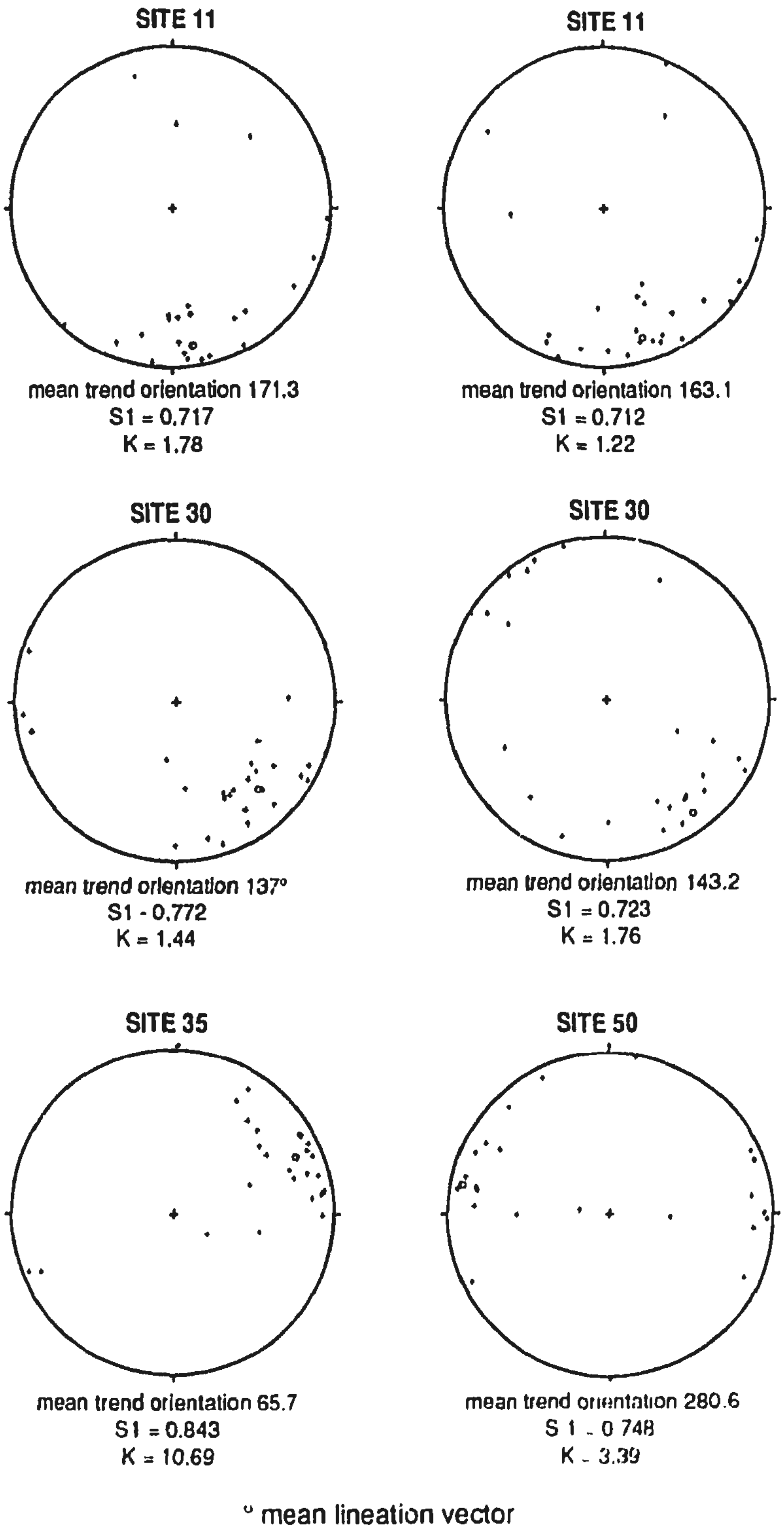


Figure 6.2: Stereoplots of fabrics used to determine ice flow

Fabrics which have girdle distributions have principal eigenvalues ranging from 0.434 (randomly oriented) to 0.814. These represent disturbed basal tills and debris flow deposits (see Chapter 5).

6.3 - CLAST PROVENANCE AND ICE-FLOW

Clasts and grains were categorized into six main lithological groups based on provenance. These are:

1. the Gander River Complex;
2. the Davidsville Group;
3. the Hamilton Sound Sequence;
4. white granites (e.g. the Aspen Cove Pluton);
5. pink granites (the Island Pond Pluton);
6. the Gander Group.

When clast provenance could be assigned to a specific outcrop, the location of that outcrop was noted.

The distribution of the clast assemblages from tills in the study area is shown in figure 6.3. Not all lithological groups are useful for determining transport directions. For example, white granites are found in almost all samples. Most of these granites are petrographically similar (Currie, 1992), and as the granites crop out in several locations, determining the provenance of a particular clast to a particular outcrop is impossible.

The most useful clast type indicating transport directions originates from the Gander River Complex. These clasts are relatively soft and thus will be incorporated into glacial ice easily (cf. Shilts, 1982). The Gander River Complex is a belt which extends across the map area from southwest to northeast, and

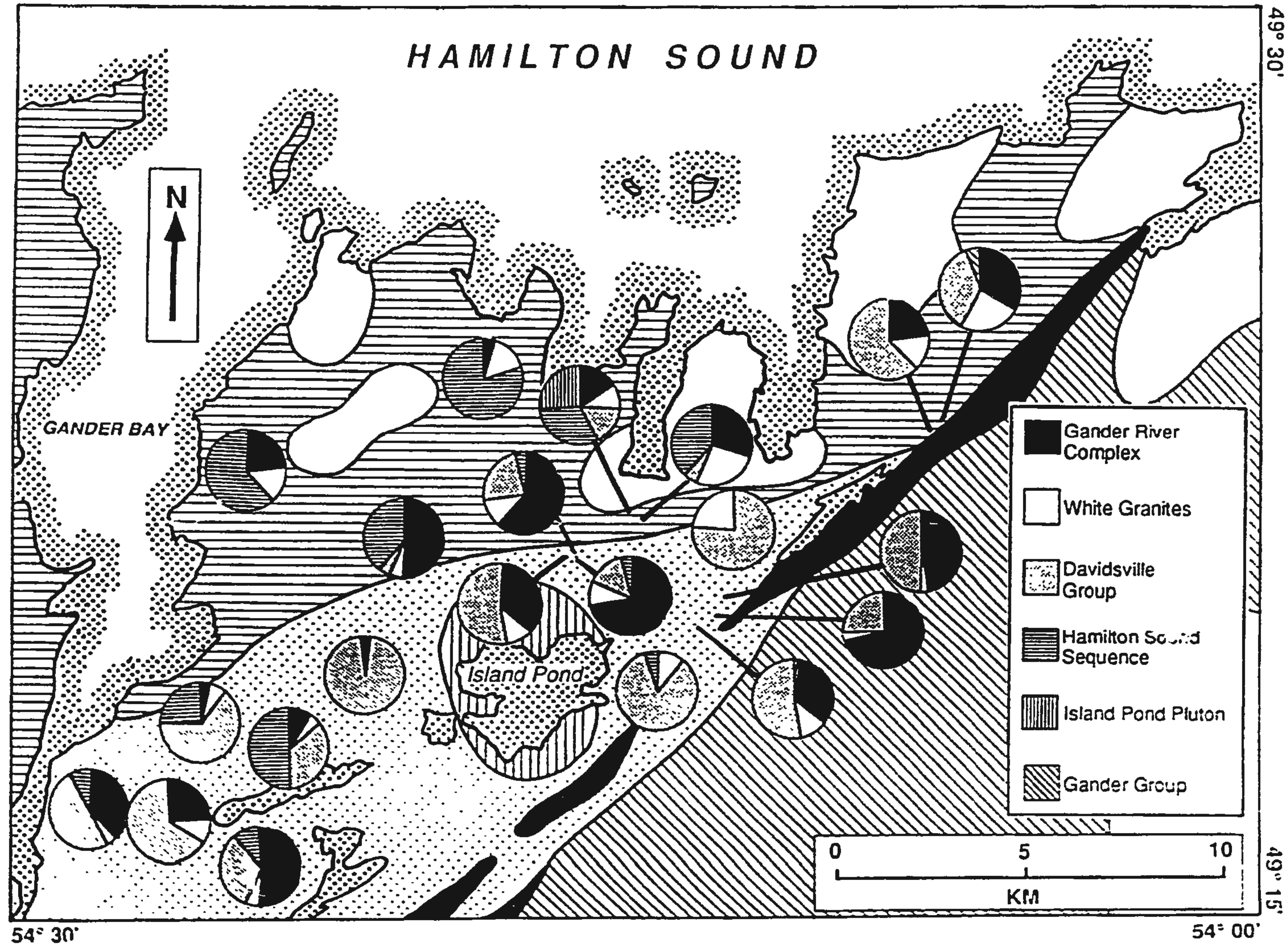


Figure 6.3: Clast Lithology of tills relative to the bedrock geology in the Carmanville

hence minor occurrences of clasts at some distance on either side of this belt would indicate ice movement. Clasts from the Gander Group will not be incorporated as easily into glacial ice, although it may be possible to detect minerals or small grains of abraded rock from this group.

Clasts from the Gander River Complex are found at most locations in the study area. Their highest concentrations are directly to the northwest of the belt. Where the Gander River Complex does not crop out (e.g. between Island Pond and Shoal Pond) concentrations directly to the northwest are very low, generally between 5% and 10%. The highest concentration of Gander River Complex clasts observed in the study area is located directly overlying the Gander River Complex near Shoal Pond (Site MM-13). The sample is from about 3m depth, much deeper than most of the other samples taken in the area. Five kilometres northwest of this site concentrations decrease to between 34% and 64%. Approximately 6km to the northwest, concentrations decrease to about 28%. Concentrations decrease to about 10%, 10km north of the site. Consequently, approximately 10km from the source area 10 percent of all clasts identified originated from the Gander River Complex. This dispersal pattern is clearly related to northwestward ice-flow. A similar pattern of concentrations can be observed related to the outcrop at Barry's Pond.

Although no obvious patterns can be observed in the distribution of clasts derived from other rock types in this area, there are a few interesting occurrences. One clast identified as being from the Island Bay Pluton was detected 4km east of that outcrop. Clasts from the Davidsville Group were detected east of the Gander River Complex. The distribution of these clasts relates to transport by eastward ice-flow. They have probably been subsequently reworked by later ice-flows.

Clasts from the Hamilton Sound Sequence are found in large concentrations in the source area. Concentrations decrease in the samples directly to the east of the Hamilton Sound Sequence where they are found in small concentrations (less than 1%).

Clasts from the Barry's Pond Conglomerate (part of the Davidsville Group) are easily recognizable, as they contain pebbles from the Gander River Complex in a dark green sandy silty matrix (Currie, 1992). These clasts are usually substantially weathered and thus break easily. Clasts from this unit are found up to 5km northwestward of the source area. This distribution thus reflects transport by northwestward moving ice.

Studies of the finer grain sized fractions proved relatively inconclusive. At most sites, 75% or more of the sample was made up of crushed slate, which in most cases reflects the underlying bedrock. At almost all sites there is a minor component of crushed schist or minerals derived from this such as talc, chlorite, biotite, and hornblende. These minerals and rock fragments have their origins in the rocks of the Gander Group and consequently represent dispersal by northwestward moving ice.

6.4 - LAKE SEDIMENT GEOCHEMISTRY

Using the maps published by Davenport and Nolan (1988), trends can be observed in the distribution of certain trace elements in the Carmanville area. The most notable of these are the elements associated with the Gander River complex: Cr, Sb, Sc, Au, As, and Ba. In each case there is a definite concentration of material directly to the northwest of outcrop of the Gander River Complex, at the head of Gander Bay, and in Eastern Arm. The material to the northwest of the Gander Complex shows an abrupt increase up to 2km from the outcrops. Concentrations decrease slowly with distance in a northwest direction.

6.5 - ICE-FLOW HISTORY

The Carmanville area has been subjected to at least three ice-flow events during the Late Wisconsinan. Striations indicate that the first of these was an easterly ice-flow that had little influence on the landscape. It trends approximately 075°. This ice covered all of the study area but it did not flow directly towards the coast (Figure 6.4). In accordance with the model of St. Croix and Taylor (1990), this ice-flow probably was associated with the maximum extent of glacial cover, which persisted until the onset of deglaciation. This ice flowed obliquely to the strike of the underlying bedrock, and hence did not form roches moutonnées or influence the landscape morphology. St. Croix and Taylor (1990) postulated that ice may have flowed down from the Northern Peninsula into the Notre Dame Bay area causing a deflection of the ice moving from the Central Uplands in an easterly direction. Subsequent ice-flows may have erased most of the evidence of the easterly flow with few striations preserved in the lee of bedrock outcrops. Evidence of this flow is found extensively throughout the surrounding region, between Grand Falls and Gander Bay (St. Croix and Taylor, 1991), and in the Gander Region (Taylor and St. Croix, 1989; Batterson *et al.*, 1991). The one till fabric indicative of easterly flow was taken at site MM-50.

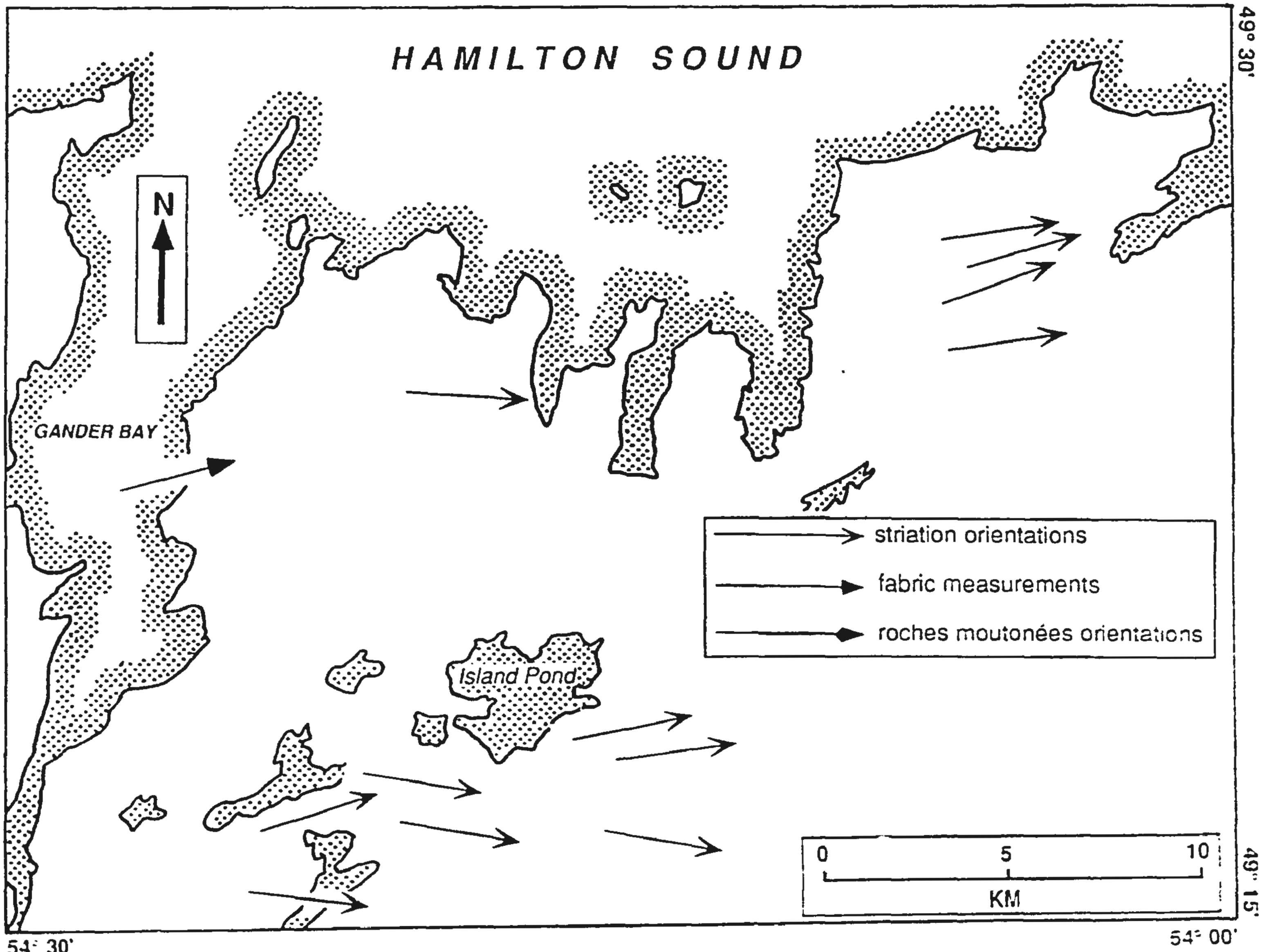


Figure 6.4: The first (eastward) ice-flow to affect the Carmanville region.

The second ice-flow moved northeasterly to north-northeasterly (Figure 6.5). This event had the largest influence on the landscape. Striations, roches moutonnées, flutings, and crag-and-tails throughout the Carmanville area reflect this event. The erosional features have a strong northeast to north-northeast orientation, which reflects the underlying bedrock trends. This flow has been found between the Baie Verte Peninsula and Bonavista Bay (St. Croix and Taylor, 1990). It is consistent across the whole of northeast Newfoundland suggesting that there may have been an ice divide crossing the Island eastward between approximately Red Indian Lake and the Bonavista Peninsula (Jenness, 1960; Grant, 1974; St. Croix and Taylor, 1991). The one fabric indicative of northeasterly ice-flow was taken at site MM-35.

The third ice-flow event was northwesterly (Figure 6.6). This event is not represented in terms of large-scale landforms. Fabric analyses and striations, however, show that a northwesterly ice-flow was influential throughout the entire Carmanville region apart from the area to the west of Gander Bay. The patterns of ice-flow documented here conform to those proposed by St Croix and Taylor (1990) for north central Newfoundland.

Drumlins and rogen moraines have northeastward trends. If these were formed by northeastward ice-flow, they would probably have been destroyed by the subsequent northwesterly ice-flow, or would have fluted surfaces, evidence for which is absent. These deposits are therefore interpreted as representing northwestward ice-flow and were probably moulded by subglacial meltwater flowing in a northeasterly direction, following the main bedrock trends in the area.

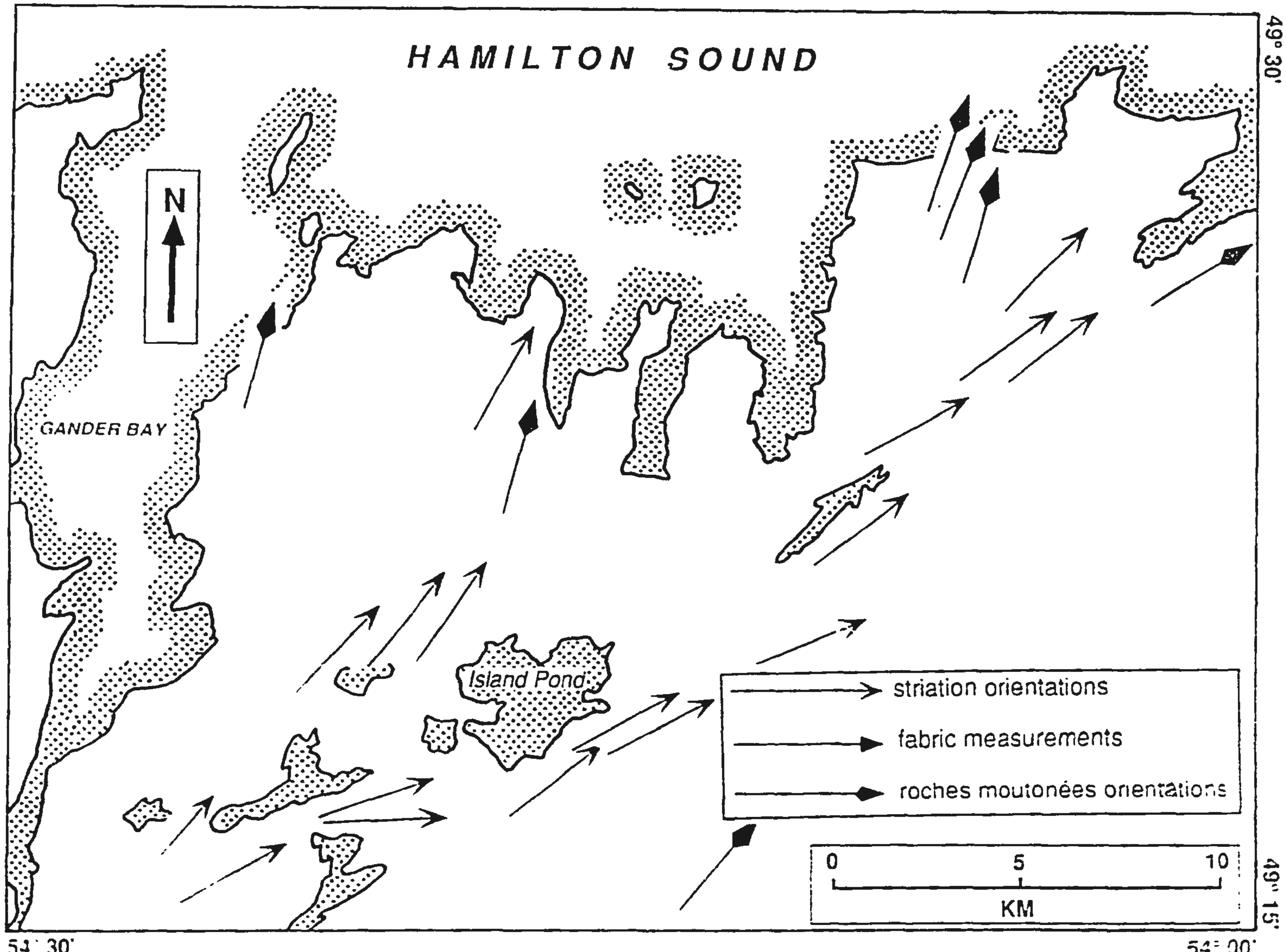


Figure 6.5: The second (northeastward) ice-flow to affect the Carmanville region.

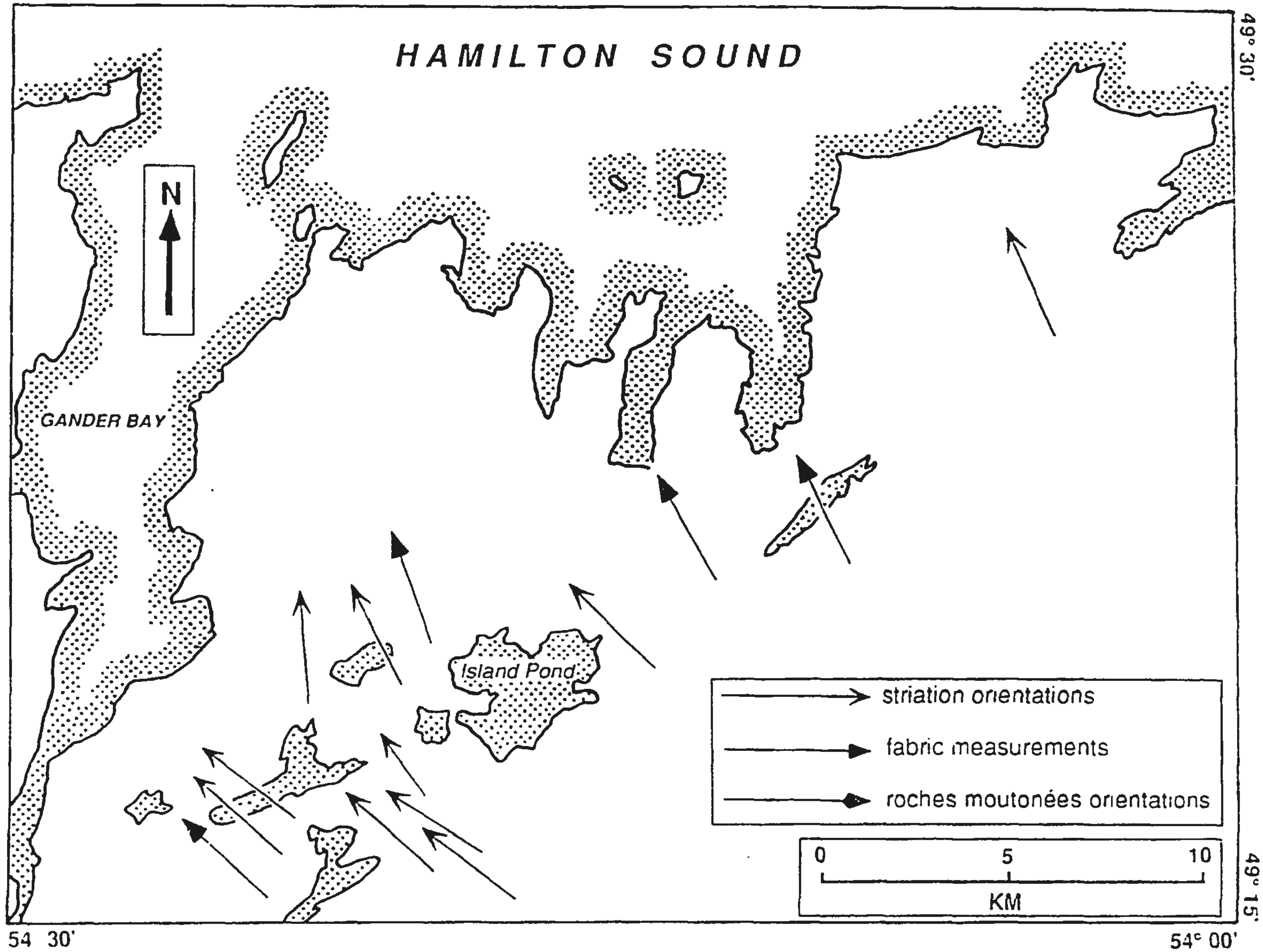


Figure 6.6: The third (northwestward) ice-flow to affect the Carmanville region.

There was a final period of glacial activity in the Carmanville area occurring subsequent to high sea-levels. This is evident in the form of till deposition by ice that was flowing northwestward. The evidence for this is limited and it is therefore difficult to determine whether this period of glacial activity is related to the final phase of Late Wisconsinan ice or is related to the Younger Dryas cooling event. The evidence suggests that it may represent Younger Dryas activity (see Chapter 5: Section 5.5).

CHAPTER 7
SEDIMENTOLOGY 2: GLACIOFLUVIAL SANDS AND GRAVELS

7.1 - INTRODUCTION

The largest, most complex, and best developed areas of glaciofluvial sands and gravel sections occur along Route 330 between Musgrave Harbour and the Aspen Cove access road. These sediments comprise the "Ragged Harbour Moraine" which was described in chapter 4. Sections were studied at six locations (MM-1, MM-2, MM-3, MM-4, MM-5, and MM-6)(Figure 7.1). Sedimentary assemblages are complex and the number of units is large. Consequently, nineteen vertical transects were described and analyzed in detail. Two large sites (MM-3 and MM-4) show typical sequences for the area and thus will be described in detail here. MM-4 (Ragged Harbour 1) is a large active gravel pit, and MM-3 (Ragged Harbour 2)s a large abandoned gravel pit.

7.2 - RAGGED HARBOUR 1 (MM-4)

Ragged Harbour 1 (MM-4) is a large gravel pit situated 3km south of Ragged Harbour. It displays a series of complexly interbedded silts, sands, gravels, and diamicton (Figure 7.2), which were analyzed and described as a series of 10 vertical transects (Figure 7.3). The contacts between these units are characteristically abrupt. Structureless and high-angle planar cross-bedded gravels are overlain by rippled sands, which in turn are overlain by deformed silt and sand. The assemblage comprises a series of high-angled planar cross-bedded gravels, with beds commonly separated by upper and lower bounding

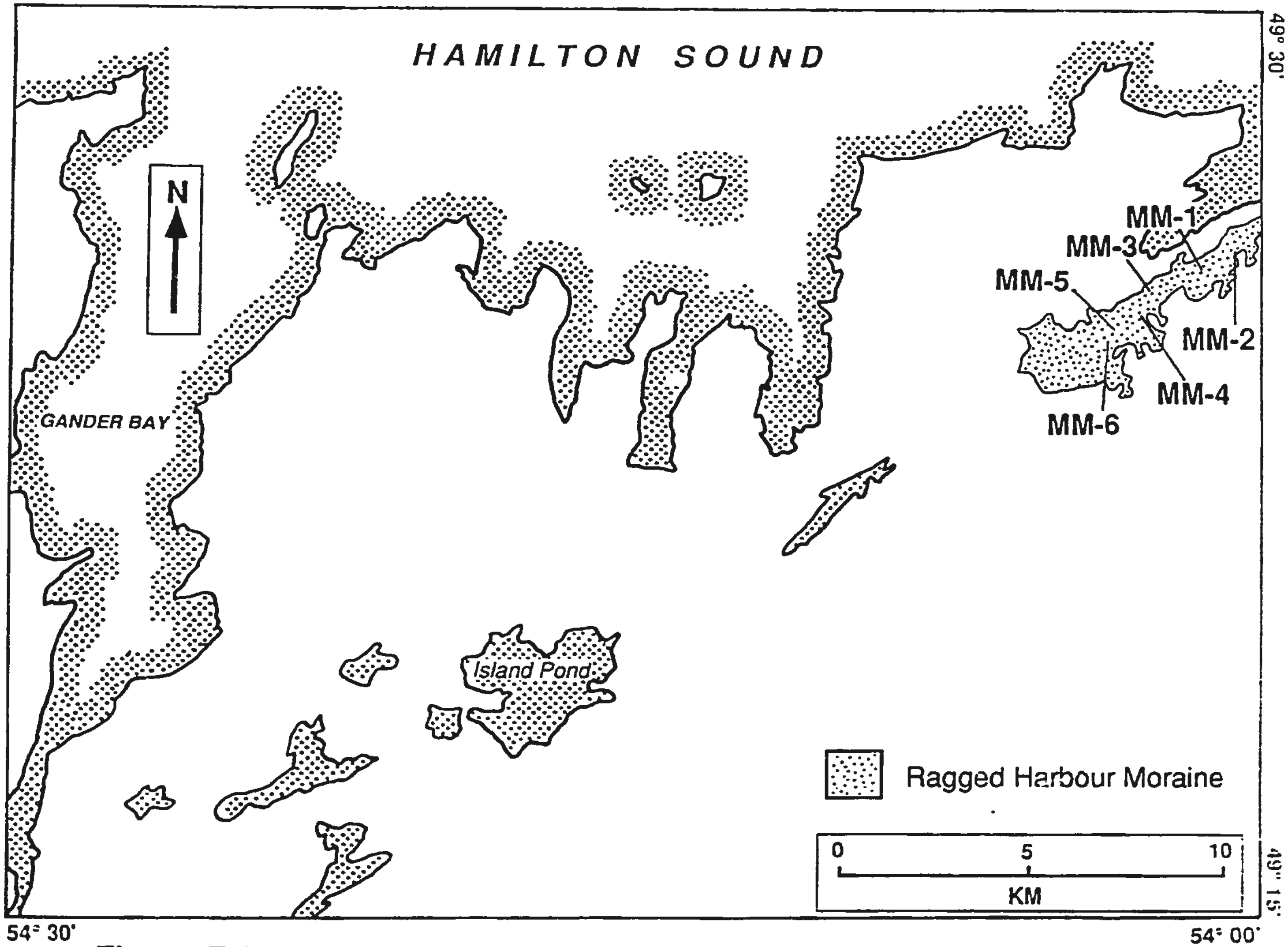


Figure 7.1: Location map of glaciofluvial deposits and all glaciofluvial sites studied in the Carmanville region.

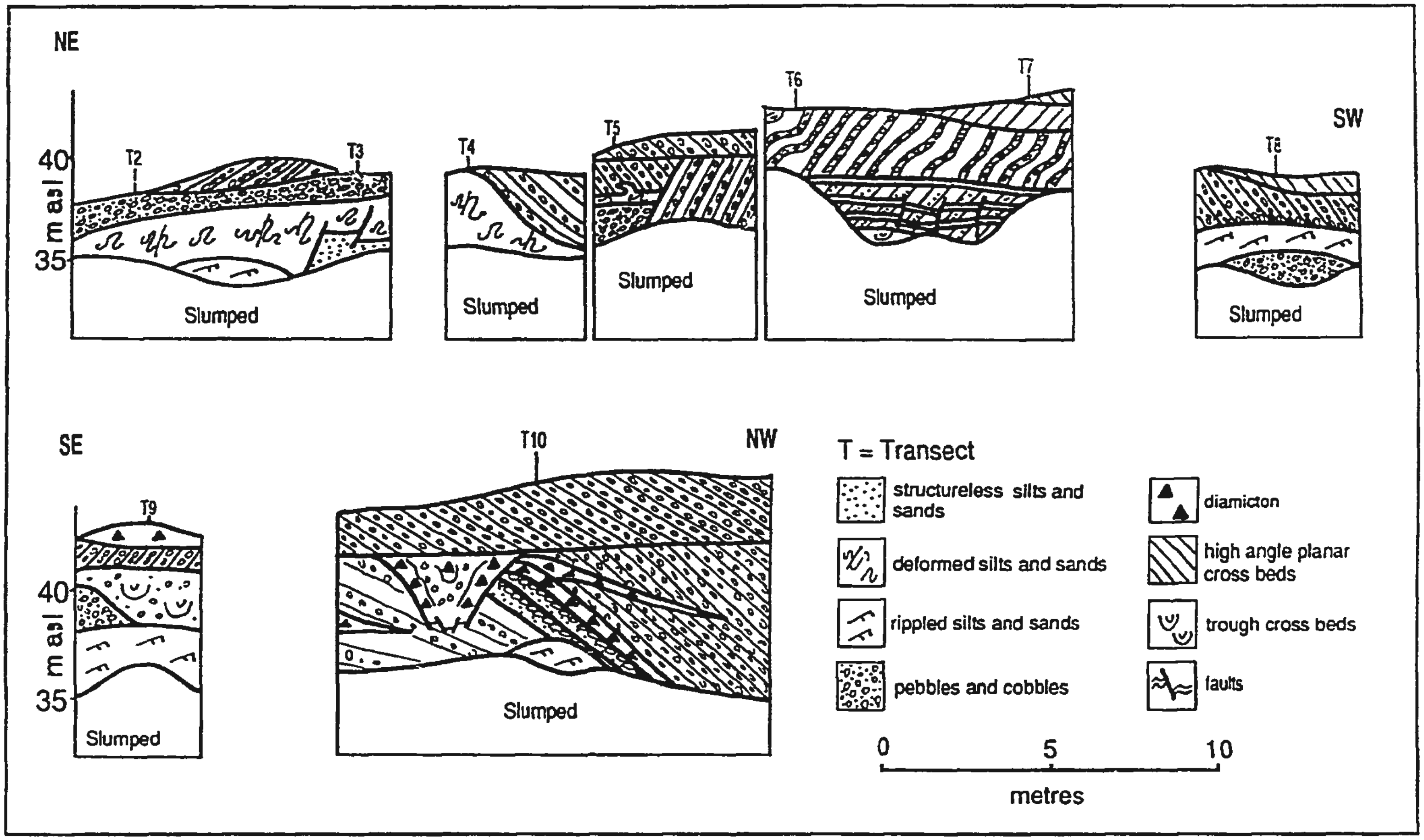


Figure 7.2: Generalized stratigraphy at site MM-4 showing the location of transects

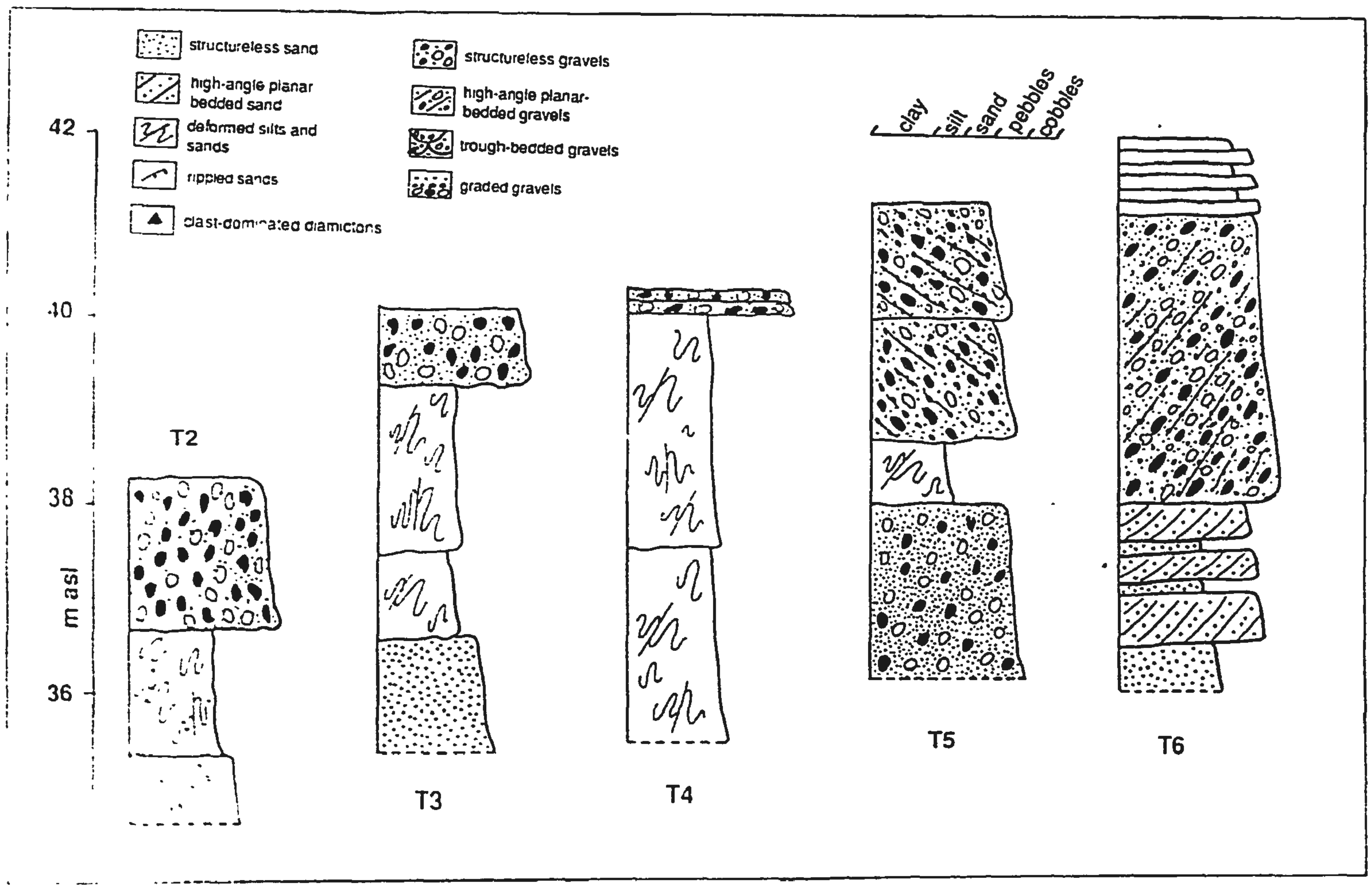


Figure 7.3: Sedimentary transects at Ragged Harbour 1 (MM-4).

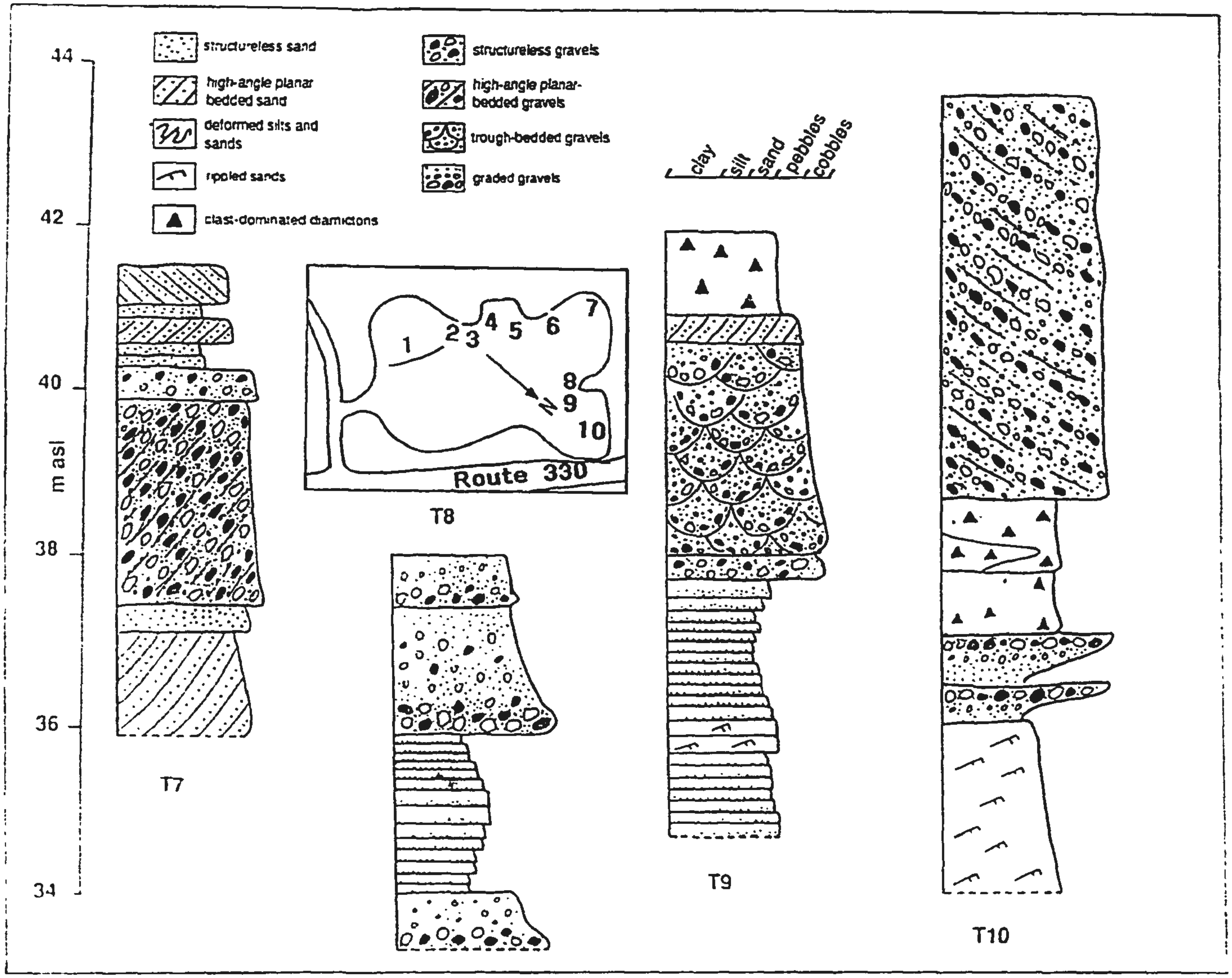


Figure 7.3: (continued)

surfaces. Discontinuous beds of diamicton locally overlie the gravels. The uppermost unit of the sequence is trough cross-bedded and high-angle planar cross-bedded gravels. The upper surface of these gravels has been reworked by marine action to depths of up to 1m to form a terrace at 46m asl.

As the site is complex and the number of units is large, the sedimentary succession will be described as a series of four textural units:

1. a lower gravel unit;
2. a sand and silt unit;
3. a diamicton unit; and
4. an upper gravel unit.

7.2.1 - LOWER GRAVEL UNIT

Description

The stratigraphically lowest sedimentary unit observed at MM-4 is gravel-dominated and varies from structureless cobble gravels to high-angle (18°-32°), planar cross-stratified, interbedded fine to coarse gravels and sands. The structureless gravels were observed at one location only (transect 8), and limited exposure restricted analysis. The unit is best exposed at transect 1. The whole exposure is described here.

The height of transect 1 is 165cm, and the lateral extent of the exposure is 40m (Figure 7.4). It is composed of a sequence of interbedded, high-angle planar cross-bedded cobble and pebble gravel, and sand. The beds at the western end and the eastern end of the exposure dip at angles between 30° to 32°, and in the central part of the section they dip at 18°. The direction of dip

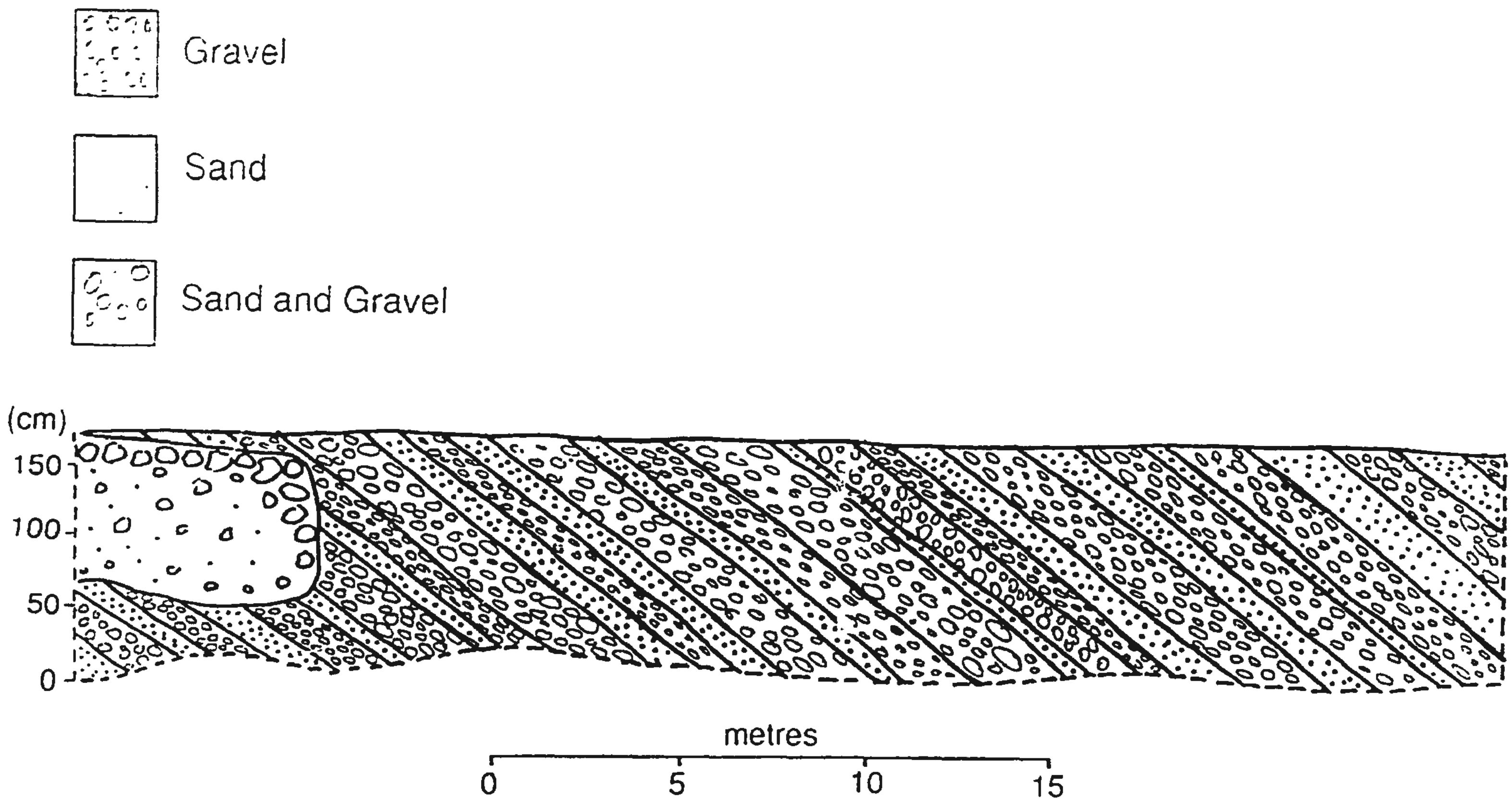


Figure 7.4: Location of transect 1 within exposure at Ragged Harbour 1 (MM-4)

varies throughout the exposure between 070° and 104°. The contacts between beds are abrupt.

Pebble and cobble gravel beds range from 8cm to 35cm thick and are not graded. The gravel is clast-supported, with approximately 25% to 30% of each bed being matrix, composed of well sorted coarse sand. Clasts are generally spheres, using the Zingg (1935) classification, and they are mostly subrounded to rounded with some rounded clasts. They are not imbricated. Schists account for 48% of all the clasts observed, 26% are granite, and 9% are gneiss. Minor quantities of sandstone, diorite, gabbro, and shale are also present.

The sand units range from 4cm to 21cm thick and are made up almost entirely of coarse sand with minor quantities of fine sand and silt (<5%). No grading or structures were observed.

One bed is anomalous within this sequence. It is a clast-dominated diamicton that contains granules and pebbles and grades upwards to pebbles and cobbles at the top of the bed. Its maximum thickness is 1 metre and it extends laterally for at least 3 metres. The unit tapers east to west from a thickness of 1m to approximately 50cm. The westernmost extent is concealed. It overlies high-angle planar cross-beds above an erosional contact, and is overlain by other cross-stratified gravels, above a horizontal and planar contact. Clasts are mostly subangular to subrounded. The matrix is composed of 40% sand, 44% silt, and 6% clay, and the diamicton is therefore finer and more compact than the surrounding beds. The clast assemblage is dominated by 37% white granite, 33% schist, 10% granite from the Deadman's Bay Pluton, 6% gneiss, and small quantities of chlorite schist, basalt, sandstone, and diabase.

Interpretation

High-angle planar dipping beds are often indicative of deltaic deposition or linguoid bar formation (Allen, 1968, 1970, 1983; Miall, 1978; Harms *et al.*, 1982). Deltas are deposited as foreset beds or avalanche faces of prograding sloping surfaces and this is the likely origin of this unit. The dip angle and the size of the material transported would suggest an ice-proximal location for most of these pebbles and cobbles (cf. Thomas, 1984). As the rate of deposition and mean grain size decreases with distance from the sediment source (Allen, 1968), the angle of dip of the beds will decrease with transport distance due to the reduced angle of repose. The foreset beds have the greatest rate of deposition (Weimer, 1970; Harms *et al.*, 1982) and are closest to the sediment supply. The alternating bands of coarse clasts and sand can be explained either by fluctuating flow regimes, or avalanching of oversteepened foresets, which is more common in ice-proximal environments where slope angles are higher. Prior to deposition of the overlying beds, observed at other sections at MM-4, the topsets and an undetermined thickness of the foresets were eroded.

The diamicton bed is interpreted as a debris flow deposit. In a deltaic environment debris flow would have taken place due either to oversteepening of the delta slope or as a result of some other mass instability. Debris flow deposits are typically structureless and are composed of diamicton, but inverse grading can occur due to dispersive pressure within the flow (Middleton and Hampton, 1973). They form due to the downslope movement of mixtures of clasts, clay and water due to gravity (Middleton and Hampton, 1973), and can flow down slopes as low as 1° (Curry, 1966). The terminus of a debris flow deposit usually has a blunt steep-sided profile, and an uneven contact such as that observed at MM-4.

7.2.2 - SAND UNIT

The sand unit accounts for approximately 25% of all the deposits observed at the Ragged Harbour 1 site. It can be divided into two types. The first of these is a finely laminated (with structureless beds) coarse silt and fine sand with extensive soft sediment deformation structures. The second style of sand deposition is rippled and rhythmically bedded.

7.2.2.1 - Deformed fine sand and coarse silt unit

Description

The fine sand/coarse silt unit is faintly laminated throughout its extent. It has, however, been deformed, which has severely disrupted and obscured the primary lamination, often causing the deposit to appear homogeneous. The deformed sediments are up to 4 metres thick, and extend laterally up to 25 metres. The deposits commonly have matrix compositions between 26% and 52% sand, and 38% and 74% silt (Appendix 2). The clay component is approximately 10%. High-angle (70°-85°) normal and reverse faults are common throughout with displacements reaching a maximum of up to 2 metres.

Soft-sediment deformation structures observed at MM-4 include diapirs and flame structures, ball and pillow, dish and pillar, recumbent folds, and convolute lamination in the form of box folds and spiral folds. These are water escape structures and structures related to reverse density gradients (cf. Allen, 1983) and occur as sediment is rapidly deposited over underlying saturated sediments (Lowe, 1975). They occur most commonly in subaqueous sediments which were deposited rapidly, in environments such as deep water basins, storm influenced shallow marine settings, deltas, or river floodplains. Development is in response to stresses originating in four ways: 1) gravitational instability caused by vertical contrasts in bulk density; 2) non-uniform confining loads; 3) the presence of sediment on a slope; and 4) fluid flow, either at the sediment-fluid interface or internally (Allen, 1983). Deformation structures are discussed here in four groups. These are: 1) dish and pillar structures; 2) diapirs and flame structures; 3) ball and pillow structures; and 4) convolute lamination.

Dish and Pillar Structures

The dish structures at MM-4 range from a few centimetres to approximately 50cm across. Heights of these features are consistently between 5 and 8 cm. The large dish structures generally have flat bases whereas the small features are strongly concave upward. There is a concentration of finer sediments in the base of each feature. The dishes tend to be concentrated towards the top of the deformed silt and sand beds but are best developed along a single plane in the central part of the unit (Plate 7.1). It cannot be determined whether the dishes at MM-4 formed due to the trapping of mobilized fines (cf. Lowe and LoPiccolo, 1974), or formed due to cavity collapses (cf. Pederson and Surlyk, 1977). The mode of formation is not necessarily important as both theories involve the rapid upward escape of porewater.

Plate 7.1: Large scale and small scale dish structures at Ragged Harbour 1 (MM-4). These are best developed towards the top of the deformed silt and sand unit. The photograph shows approximately 2 metres of sediment from top to bottom.

In most environments pillar structures are strongly associated with dish structures. They occur as circular columns or sheet-like clay-free curtains of structureless or swirled sand and/or silt. They can intrude through metres of laminated or cross-laminated sands. The formation of pillars involves the localized breakdown of grain-supported sand frameworks by one of two processes: 1) fluidization where the weight of individual sediment grains becomes partially or entirely supported by the drag of the upwards moving water; and 2) liquefaction where loosely packed, metastable grains are shaken loose from each other and briefly settle unsupported through the pore fluid (Allen, 1982). Consequently, pillar formation is also related to the upward movement of escaping pore water (liquefaction) in rapidly deposited thick silt and sand beds. Unlike dish structures, the mode of deposition is rapid.

At MM-4, pillar structures are found throughout the extent of the deformed silts and sands. The central area of each pillar structure is composed of medium grained sand and is homogeneous. Laminae in the adjacent sediments are sharply deformed upwards adjacent to the pillar structure (Plate 7.2). Unlike dish structures, pillars are not concentrated in the upper parts of the unit, but occur throughout the deformed silts and sands. They are generally large, ranging from 0.5m to 1.5m high and up to 0.5m wide. They often dissect the deformed and laminated silts and sands. Like dish structures, pillars form due to the rapid upward escape of porewater (liquefaction) resulting in the intrusion of structureless sand through metres of laminated or cross-laminated sand or silt. Some pillars at MM-4 are topped by a plug of fine grained silt, marking the point where pore water escape ceased. The effect of rapidly upward moving pore water locally caused failure and faulting of overlying beds.

Diapirs and Flame Structures

Diapirs and flame structures form as a consequence of inverted density gradients in stratified successions (Mills, 1983). Instability develops between liquefied sand and finer sediment layers, which establishes a reverse density gradient due to the differential porosities of sand (approximately 45%) and clay (70-90%) (Mills, 1983). If the underlying clays are more viscous, broad forms develop with sharp crested diapiric intrusions between them. These occur most

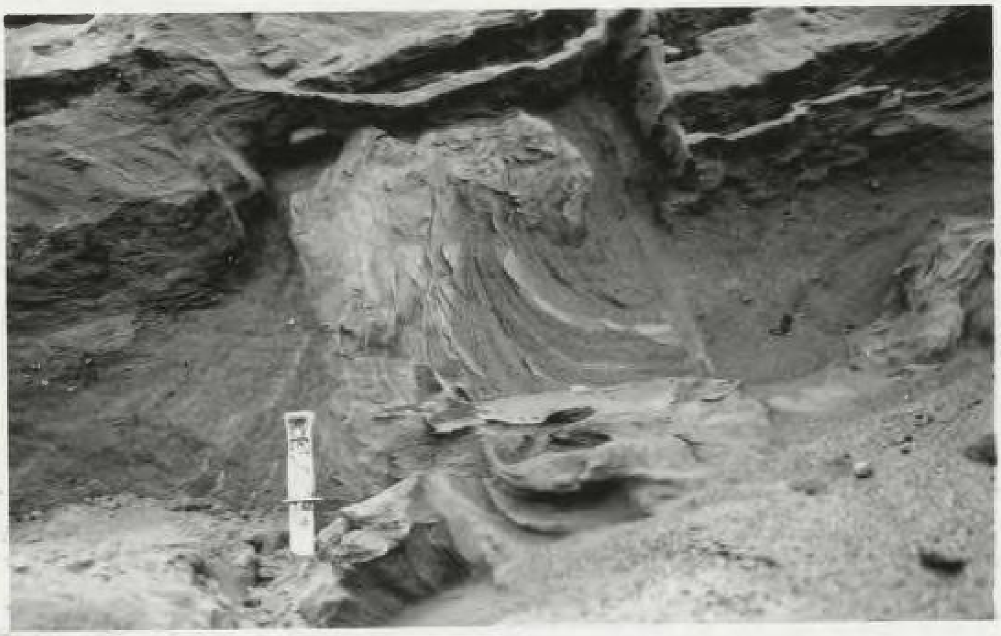


Plate 7.2: Pillar structure at Ragged Harbour 1 (MM-4). Porewater tried to escape vertically removing the finer grained clays and silts and leaving behind relatively coarse structureless sand. The overlying silts restricted the complete escape of porewater at this location although the force of the escaping porewater was strong enough to cause upthrusting of the overlying sediments. The knife is approximately 25cm long.

often when there is differential loading of sand beds (Dzulynski and Walton, 1965).

Diapirs and flame structures were observed at transects 2, 3, 4, and 5 at MM-4. They are also common at sites MM-1, MM-2, MM-5, and MM-6. They range in scale from 1cm to 0.5m wide, and from 5cm to 1 metre high. The tops of the features are sharp and they often penetrate one or two of the overlying strata without deforming the strata substantially (Plate 7.3). Strata within the diapir are perfectly preserved and follow the outline of the diapir shape. These

Plate 7.3: Large diapir at Ragged Harbour 1 (MM-4). Coarse sand and fine granules overlie a fine silt bed which produced a reverse density gradient when the sediments were saturated. The more viscous silts hence intruded into the coarser sands and granules. The hammer is approximately 40cm long.

features are commonly overturned or recumbent at the top. Beds containing diapirs are always overlain by coarse sand beds or units of small pebbles confirming that they formed due to reverse density gradients. Palaeoflow measurements taken from these deformed loading structures show that almost all are deformed to the NE/NNE (Figure 7.5). They are deformed due to directional loading of the sediments from above. These features are generally composed of 20-25% clay, approximately 55% silt, and 20% sand.

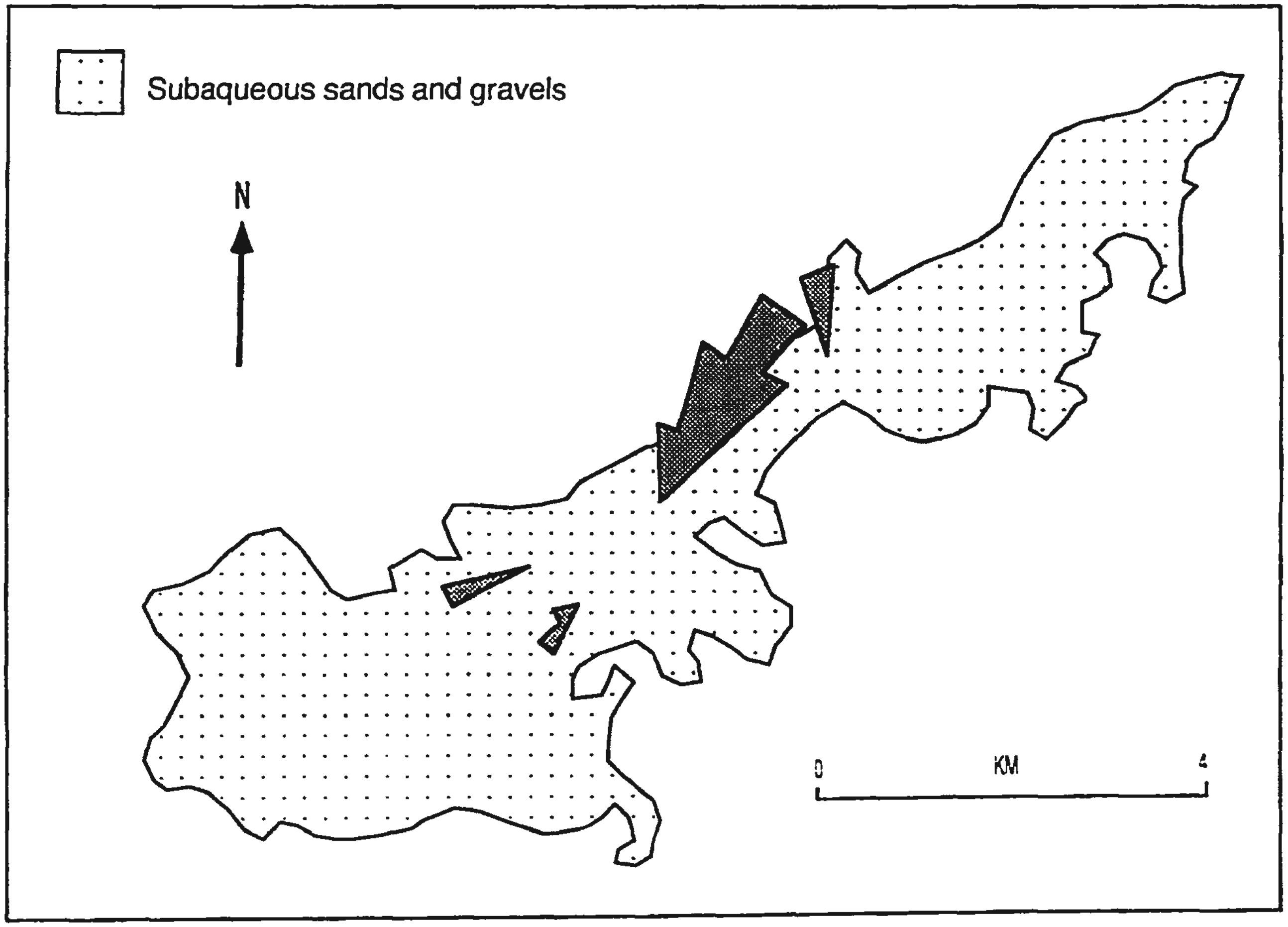


Figure 7.5: Directional measurements of structures subjected to deformation due to loading at all glaciofluvial sites in the Carmanville region.

Ball and Pillow Structures

Ball and pillow structures are genetically related to diapirs and flame structures but occur rarely at MM-4. They are most common in transects 2, 3, and 4. They are generally isolated examples, but are easily recognizable as medium to coarse sand beds are completely detached from their related overlying sand beds. They are usually about 20cm in diameter. The sand laminae within the structures are well preserved and follow the shape of the "ball", losing their original expression. The finer silt laminae which envelop the sand are also well preserved and hug the outside of the "ball" structure.

Ball and pillow structures form due to differential loading of soft fine sediments by rapidly deposited sands (McBride *et al.*, 1975). The instability between the liquefied sand and the finer sediment layers establishes a reverse density gradient and allows "balls" of the sand to sink to into the finer sediment.

Convolute Lamination

Convolute laminations occur at transects 2 and 3 at MM-4 (Plate 7.4). They are best developed in the upper part of the deformed silts and sands. Although they may occur throughout a sedimentary unit, they are best developed towards the top of a unit (Allen, 1983) as observed at MM-4. Like ball and pillow formation, convolute laminations form due to the instability between liquefied sand and finer sediment layers. The folds are generally open and sometimes distinctly box-shaped. Each fold is approximately 50cm in amplitude and has a wavelength of 50cm. The upper contact between the deformed beds and the overlying gravels is erosive, truncating the upper part of the folds. The convolutions at MM-4 are therefore either syndepositional or metadepositional.

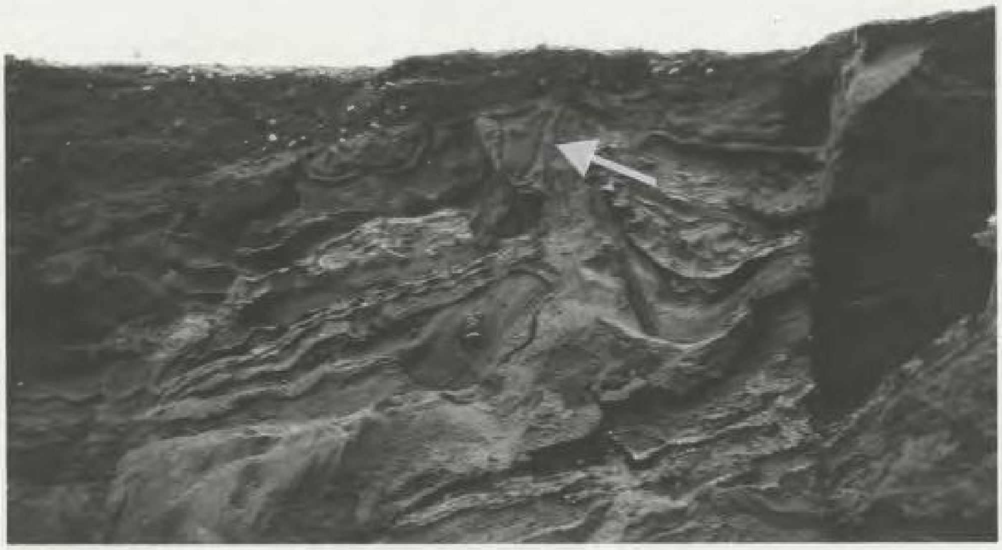


Plate 7.4: Convolute lamination at Ragged Harbour 1 (MM-4). These occur at the top of the deformed silt and sand unit. These are either syndepositional or meta-depositional as they are truncated. Each fold is approximately 50cm wide.

Summary and Interpretation

In summary, almost all the deformation structures present at MM-4 were formed shortly after the deposition of the individual sedimentary units involved, indicated by the preservation of many of the laminae within the structures. All the dominant mechanisms for soft-sediment deformation such as liquefaction, reverse density gradation, slump and slope failure, and shear stress, are likely to have occurred at MM-4. Gravitational processes occurred mainly in the form of coherent slumping where laminae are preserved rather than destroyed (Mills, 1983). Many of the beds are slumped to angles greater than their angle of repose, to maxima of 45° (Plate 7.5), and deformation of the beds has

Plate 7.5: Slump structures at Ragged Harbour 1 (MM-4). This may have resulted due to the presence of ice which was buried rapidly by sediment. The ice subsequently melted causing the overlying sediment to slump. The core of the structure is approximately 1m wide.

resulted. In subaqueous environments, gravitational slumping can take place on slopes with angles as low as 1° (Shepard, 1955; Mills, 1983). Liquefaction was responsible for dish and pillar structure formation. Reverse density gradation was responsible for the formation of diapirs and flame structures, and in part for convolute lamination.

The silt and sand deposits containing deformation structures are interpreted as fluidized sediment flows. Dish structures are especially common within such flows (Middleton and Hampton, 1973). Deposition takes place because of the loss of pore fluids and consequently the flows will gradually "freeze" from the base up (Middleton and Hampton, 1973). Where grains are supported by pore fluid (excess pore pressure) sediments have little strength and consequently behave like fluids with a viscosity approximately 1000 times that of water. These liquefied sediments can flow very rapidly down gentle slopes (Middleton, 1969). Loss of pore fluids cause excess pore pressures to be rapidly dissipated (Middleton and Hampton, 1973). Van der Knapp and Eijpe (1969) and Middleton (1969) calculated that for a layer of sand 10 metres thick, excess pore pressures were dissipated within a few hours. When grain size increases and the thickness of the units decreases, these times became shorter. The deformed sediments at MM-4 attain maximum thicknesses of 4 metres, and therefore the time required for loss of pore fluid would be less than that suggested by Middleton (1969), and Van der Knapp and Eijpe (1969). This would explain the presence of many large pillar structures at MM-4, which formed rapidly during liquefaction.

7.2.2.2 - Rhythmically bedded sands

Description

Rhythmically bedded sands at MM-4 have maximum thicknesses of 2m (Plate 7.6). All exposures showed a vertical upward thickening of individual beds. At the base of the succession each bed is approximately 1cm thick, and thicknesses progressively increase to approximately 4cm in the upper part. Some strata are up to 15cm thick. The entire sequence coarsens upwards. Each stratum consists of fine to medium sand at the base and grades normally to silt.

Plate 7.6: Rhythmically bedded sands at Ragged Harbour 1 (MM-4). These grade laterally into rippled sands. Two small trough cross-beds are present towards the bottom of the unit indicating that these probably represent upper-flow regime deposits.

Rhythmically bedded sands grade laterally into rippled sands. The vertical extent of the rippled sands is 70cm, and they only occur over a few metres laterally (Plate 7.7). The ripples are asymmetrical with flat crests, have stoss angles between 6° and 17°, and have lee angles between 20° and 26°. Characteristic wavelengths range from 6cm to 18cm with amplitudes ranging between 1.5cm and 4cm. Ripple indices (wavelength/amplitude) range from 4 to 12. They are low-angle (14°-20°) climbing ripples with partial or total stoss-side

Plate 7.7: Rippled sands at Ragged Harbour 1 (MM-4). These are Type B (cf. Jopling and Walker, 1968), or "depositional stoss" (cf. Ashley *et al.*, 1982) ripples. The trowel is approximately 25cm long.

preservation. Ripples are smaller towards the base of the unit and increase in size vertically. There is no accompanying change in the angle of climb. These ripples are climbing ripples of Type B in the Jopling and Walker (1968) scheme, and "depositional stoss" of the Ashley *et al.*, (1982) scheme. Where the stoss sides are partially eroded, the ripples contain characteristics of both type A and type B using Jopling and Walker's scheme, and "erosional" and "depositional stoss" using Ashley *et al.'s* classification.

Palaeoflow directions measured from ripples at site MM-4 and at other sites in the glaciofluvial complex are highly variable. At MM-4, the ripples at the western extent of the pit have orientations that vary from WNW to WSW. Elsewhere, ripple directions are concentrated between ENE and ESE with some ripples oriented towards the north (Figure 7.6). At no location are the ripples oriented towards the south.

Ripples are composed mainly of coarse to fine sand with some silt. Grain size analyses show the sand component to range between 44% and 100%. The coarsest deposits are concentrated in the lee side of the ripples (medium to coarse sand). On the stoss sides, medium grained sand dominates. Upper surfaces of the ripples commonly have draped laminae composed of silt or fine sand. Rippled sands contain high-angle normal and reverse faults (65°-80°), but displacements are generally only a few centimetres.

Interpretation

Climbing ripples often occur in turbidite sequences (Bouma, 1962; Walker, 1965), in deltaic environments (e.g. Williams, 1971), and in ice-contact and proglacial fluvial deposits (Jopling and Walker, 1968; Gustavson *et al.*, 1975; Rust and Romanelli, 1975).

The ripples observed at Ragged Harbour 1 represent fluvial ripples, commonly with a ripple index of between 7 and 20, and with flat crests (Allen, 1982).

Climbing ripples commonly occur when the bedload transport rate over migrating ripples decreases downflow or when ripples receive sediment from suspension (Ashley *et al.*, 1982). Velocities are relatively low and sediment supply is relatively high. Ripples of this nature usually form at velocities of 15 cm sec^{-1} to 25 cm sec^{-1} (Ashley *et al.*, 1982). At MM-4, velocities were at the upper

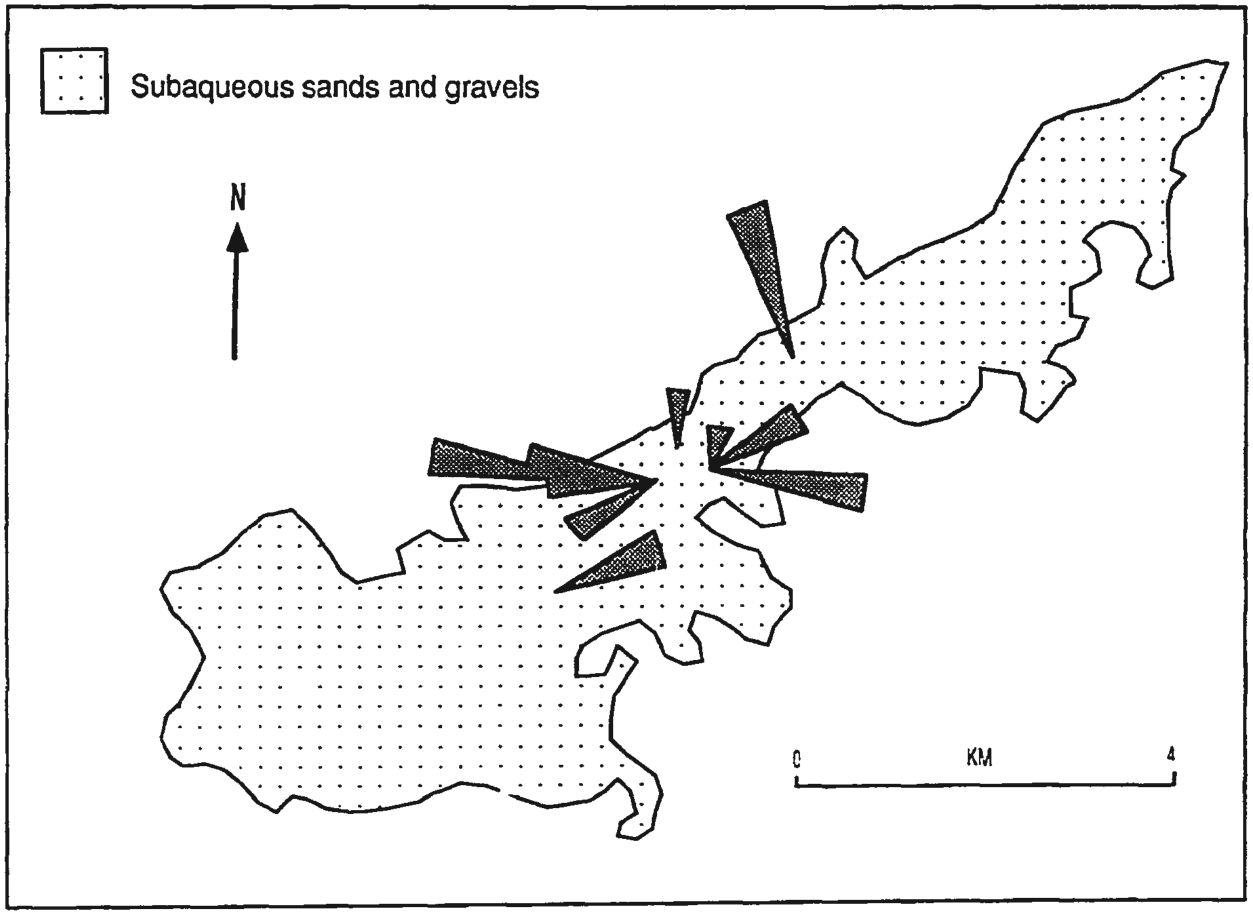


Figure 7.6: Orientations of ripples at all observed glaciofluvial sites in the Carmanville area.

end of this spectrum, as indicated by the low angle of climb of the ripples and the presence of erosional stoss characteristics. These usually occur at velocities greater than 25cm sec^{-1}. The steeper the angle of climb, the greater the role of suspension sedimentation and the smaller the role of traction sedimentation (Allen, 1982; Ashley *et al.*, 1982). High-angles of climb in supercritical sets were shown by Kuenen (1965, 1967) and Kuenen and Humbert (1969) to form under rapid flow deceleration, and hence the last structures formed are draped lamination. Traction is associated with higher velocities than suspension sedimentation. At MM-4 the flow velocity had to be higher than the sediment entrainment threshold, giving the ripples a low angle of climb. Thus it was above 15cm sec^{-1}, but below 25cm sec^{-1}.

7.2.3 DIAMICTON UNIT

Description

Diamicton beds were observed at transects 9 and 10 (Plate 7.8), and extend laterally for approximately 25 metres. Their maximum thickness is 2.5m, but at most locations they are 75cm to 120 cm thick. Three distinct diamicton units were observed overlying each other at transect 10. Matrix textures are similar (approximately 60% sand, 20% silt and 20% clay). Colours are also very similar, ranging from dark grey (Munsell: 5Y 4/1 mf) to olive grey (Munsell: 5Y 4/3 mf).

The lowermost bed is a 25 cm thick matrix-dominated diamicton with approximately 20% clasts. The bed contains some striated cobbles which are subangular to subrounded. Although there are no structures within the bed, it

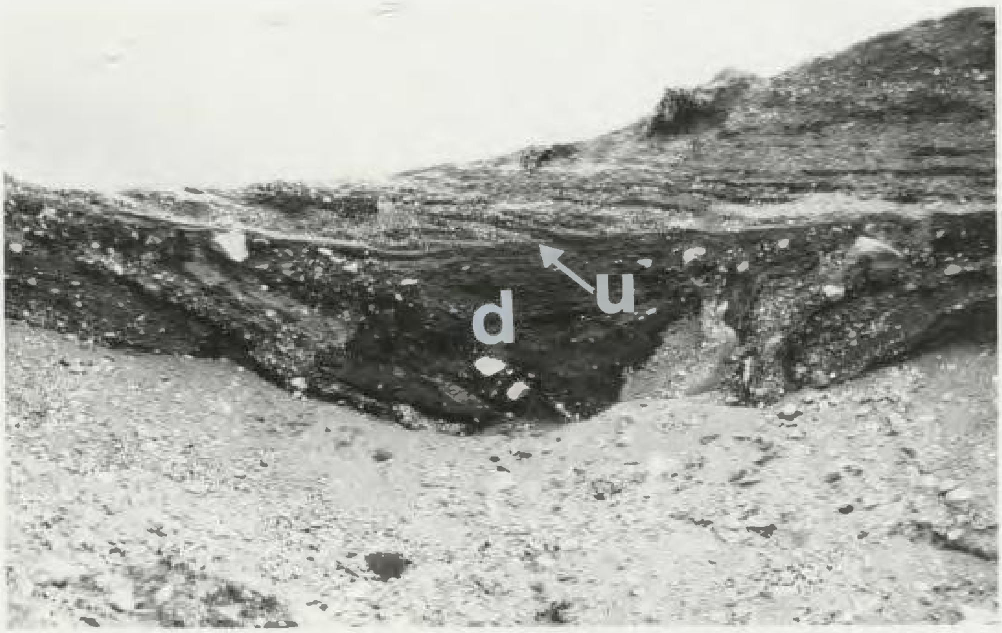


Plate 7.8: Diamicton at Ragged Harbour 1 (MM-4). Diamictons at this location are faintly stratified, and underlie an unconformity. They may represent accumulations of sediment at an ice front which subsequently failed and moved downslope as debris flows. "d" is diamicton, and "u" is the unconformity.

generally fines upward to silty clay. The mean trend orientation of clasts is 47.5°, and the mean plunge is 5.1° (Figure 7.7). Plunge angles up to 75° were recorded. An S_1 value of 0.530 and a K value of 0.3 indicate a girdle distribution.

The basal diamicton is abruptly overlain by a 20 cm thick unit of disturbed and deformed matrix-dominated (<5% clasts) diamicton. The unit has relatively undisturbed upper and lower contacts, but within the unit there are many flame structures. These flame structures are composed entirely of medium to coarse

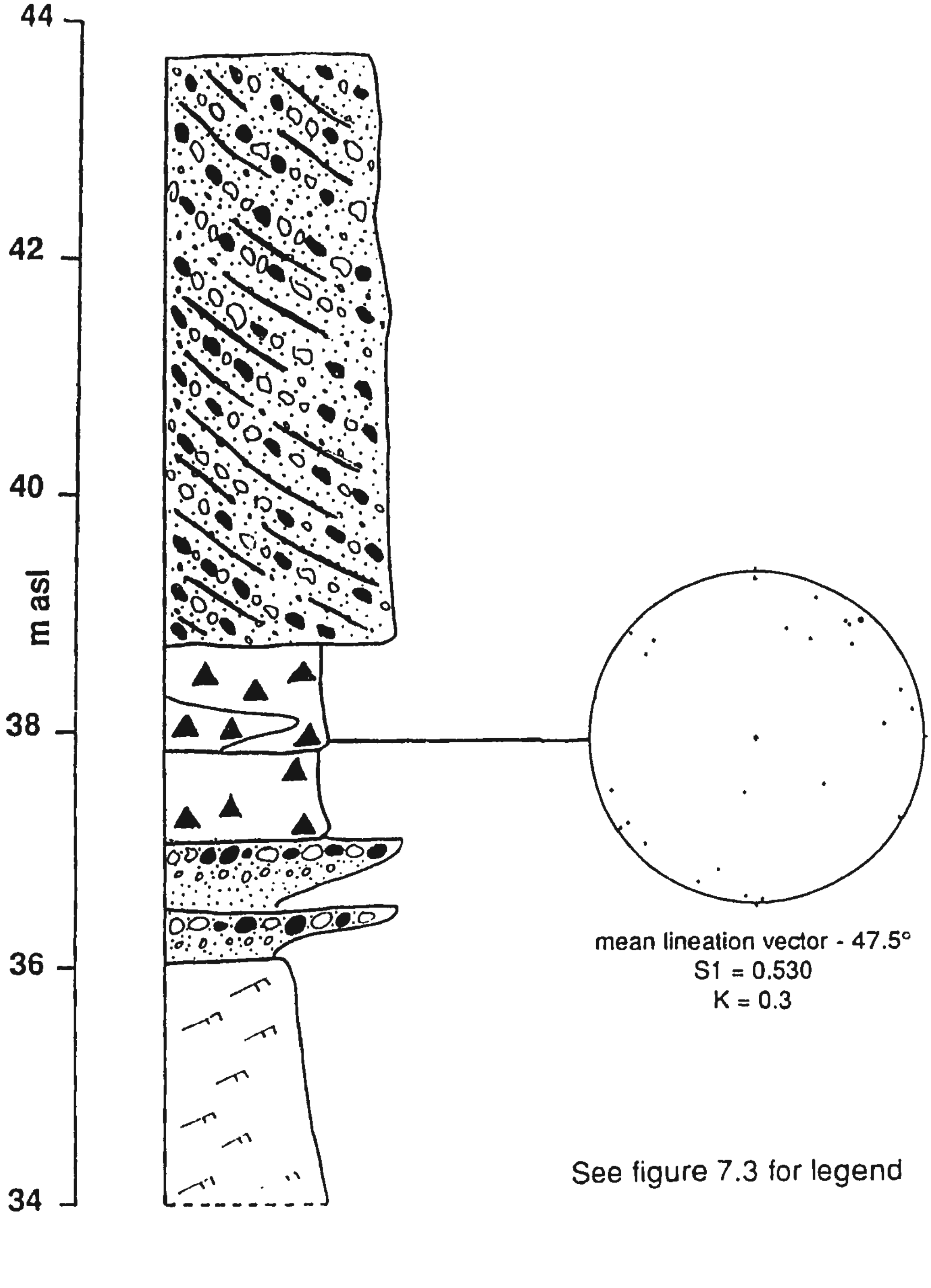


Figure 7.7: Vertical transect through sediments at Ragged Harbour 1 (MM-4)

sand. They are commonly 1-2cm wide and 15cm in height. They do not pierce the upper and lower contacts of the bed. The surrounding matrix contains deformed laminae.

The uppermost diamicton unit is a 75cm thick unit of contorted interbedded diamicton, clay and sand. Commonly, there are beds within the unit that fine upwards from granular mixtures to clay. The beds vary from approximately 5cm to 20cm thick. The silty clay beds have been subjected to far more contortion than have any of the other beds in this diamicton.

Interpretation

These diamictons are not primary till deposits. Although striated clasts are present, the clast fabric analysis shows the strength of the orientations to be poor. Although many tills can have their fabrics substantially altered post-depositionally so that the clast fabric becomes less well oriented, this was probably not the case at MM-4. The diamictons were deposited in water as indicated by the presence of extensive clay laminae that could only have formed by suspension settling. The presence of contorted laminae and flame structures indicates that these diamictons were subjected to the same reverse density gradients which produced instability between liquefied sand and finer sediments.

As the diamictons display many of the characteristics of the underlying deformed silts and sands, they were probably also laid down in an subaqueous environment. They probably originated as accumulations of sediment (primary till) at the ice front and subsequently failed, moving downslope as a series of small debris flows. Although many of these features are characteristic of melt-out till (see chapter 5), a melt-out origin for these deposits is unlikely. If the diamictons represented till they would be laterally continuous over hundreds of metres or kilometres rather than over only 25m.

7.2.4 UPPER GRAVEL UNIT

Description

The upper gravel unit accounts for approximately 75% of all the sediments observed at the site. These sediments were observed in all transects at the site apart from transect 1. They range from fine granules/pebbles to very coarse cobbles. The coarsest gravel units tend to be structureless. Most of the sediments are either crudely trough cross-bedded, or high-angle (25°-40°) planar cross-bedded. Horizontally bedded deposits are also present. They commonly have slightly deformed silt and sand interbeds.

High-angle planar cross-beds are the most common bed type in this unit. They are generally enclosed by upper and lower horizontal bounding surfaces, marking abrupt and erosional contacts. Sets of cross-beds vary from 30cm thick to 3m thick, and locally contain thick (30-40cm) beds of subrounded and rounded cobbles with abrupt upper and lower contacts. Larger clasts are commonly concentrated at the base of each bed. Interbedded with these are thin, faulted and deformed (<10cm) beds of silt and sand. The faults are high-angled (75°-85°) and are both reverse and normal. The direction of dip of these beds is vertically and laterally inconsistent. Most dip towards the north, but dip directions vary from WNW to east (Figure 7.8). No beds dip southwards.

The coarser gravel units are clast-supported, containing small amounts of sand (<5%). The finer gravel units are also clast-supported but can contain

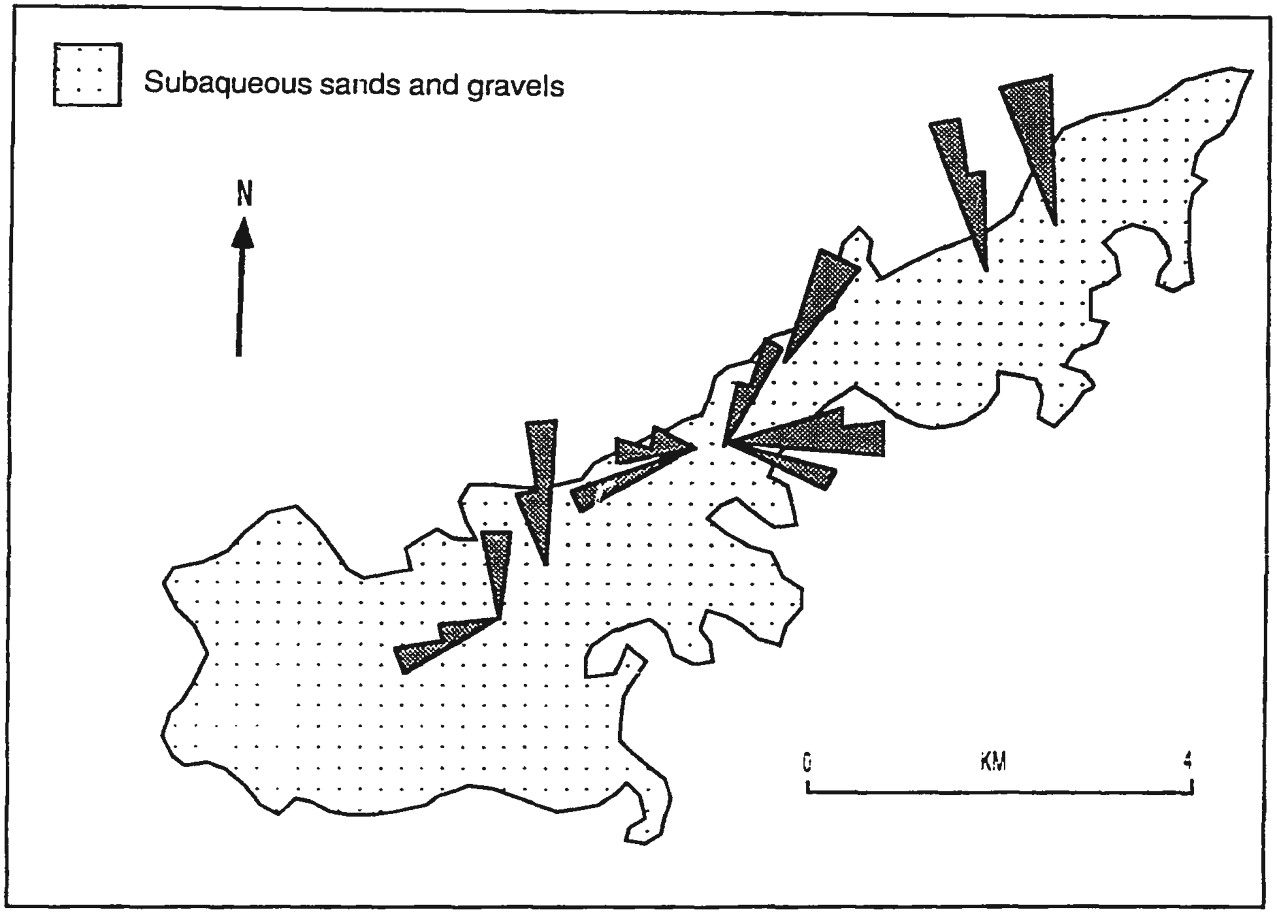


Figure 7.8: Orientations of high angle planar cross beds at all observed glaciofluvial sections in the Carmanville area.

up to 40% sand with minor quantities of silt (<5%). Faulting occurs extensively throughout these beds. Displacements can be up to 2 metres but are commonly between 15cm and 50cm. Some isolated low-angle thrust faults are present, ranging between 8° and 12°. Displacements along these are generally on the scale of only a few centimetres. All faults also affect the underlying deformed silts and sands (Plate 7.9).

Two distinct reverse-graded beds were observed at transect 8 at MM-4. They range from fine and medium sand at the base of the sediments, to granules, pebbles, and small cobbles at the top. The finer sediment is faintly laminated. Within the sand beds are monolayers of gravels a few millimetres thick. Lenses of coarser granules within fine granule beds, and fine granules within coarse sand beds are common. There are also some small slump structures (10cm x 6cm) towards the base of these units. Laminae infill the structures and follow the shape of them (Plate 7.10). The beds dip between approximately 18° and 21°. Some of the finer deposits are faintly deformed.

Unconformably overlying all other deposits at section 8 is a 3 metre thick unit of trough cross-bedded gravels. The contact with the lower beds is horizontal and planar. This deposit was not observed at any other exposure at the site, nor was a similar deposit observed at any other section in the Ragged Harbour Complex. It is composed of small pebble gravels interbedded with coarse sands. The pebble units are clast-supported but contain up to 20% sand. Clasts are predominantly spheres with some rollers using the Zingg (1935) classification. They are generally rounded to very rounded. The clast assemblage is dominated by 22% basalt, 19% schist, 19% white granite, 10%

Plate 7.9: Faults in the upper gravel unit at Ragged Harbour 1(MM-4). These are normal and reverse high-angle faults which affect the underlying silts and sands as well as the upper gravel unit. The section is approximately 4m high.

Plate 7.10: Fluidized grain flow deposits at Ragged Harbour 1 (MM-4). These are represented by reversely graded beds of fine sand to small cobbles. Small slump structures are evident along the base of the lower flow deposit.

gneiss, 5% diabase, 5% shale, 5% diorite, and minor quantities of sandstone, ultramafics, and pegmatite.

Interpretation

High-angle planar cross-beds are usually found in deltas or fluvial bars (Allen, 1968; 1970; 1983; Harms *et al.*, 1982). Similar planar bedded cosets have been described from the Ottawa region (Rust and Romanelli, 1975; Rust, 1977; Burbridge and Rust, 1988; Rust, 1988), in Western Ontario (Sharpe and Cowan, 1990), and in Scotland (Thomas, 1984). These authors interpreted their deposits as having been formed subaqueously.

Thomas (1984) interpreted similar features to those described here as small fans where sedimentation was not restricted by tunnel or channel walls but was free to expand laterally as a result of decelerating flow. The authors working in Ontario interpreted such features as large bedforms with avalanche faces (i.e., bars). These were either cross-channel bars or prograding bars on a submerged fan surface. At MM-4 it is difficult to determine whether these sediments represent bars or fans. The palaeoflow information (Figure 7.8) indicates that they were generally unconstrained by ice and were able to form in directions ranging from NW to NE. This could be true for either fan or bar formation.

Whichever interpretation is accepted for the formation of these features, they can generally be described as representing large bedforms with avalanche faces, as indicated by the concentration of coarser material to the bottom of many beds. This indicates high energy conditions (Sharpe and Cowan, 1990). Such features usually form within a channel where flow is suppressed (Rust, 1988), or directly in front of the ice where velocities are initially extremely high and sediments are subjected to rapid deceleration and thus deposition (McDonald and Vincent, 1972; Thomas, 1984; Sharpe and Cowan, 1990; Shaw and Gorrel, 1990). Shaw (1975) and Shaw and Gorrel (1990) associated large bedforms (cross-bedded gravels) with deposition from large circulation gyres, or large separation eddies. Consequently, deposition is related to strong currents and return currents, and is thus rapid.

The inversely graded beds are interpreted as fluidized grain flow deposits. Grain flows occur under slow flow rates and thus require a relatively steep slope (between 18° and 37°) in a subaqueous environment to sustain movement (Middleton and Hampton, 1973; Lowe, 1982). They form due to dispersive pressure derived from the interaction of grains and interstitial fluid in close proximity. The larger the dispersive pressures, the smaller the slope angle required for initiating the grain flow (Middleton and Hampton, 1973). The dispersive pressure acts to sort grains by size, by pushing them to the top of the flow, i.e. to the area of least shear-strain rate (Bagnold, 1968). The smaller grains thus move to the bottom. Middleton (1970) suggested that reverse grading in grain flows is the product of a kinematic sieve mechanism where the smaller grains fall downward thus displacing the larger grains upward. The lamination at MM-4 possibly would not be preserved under these circumstances, so Middleton's (1970) theory may not be applicable here.

Deposition of the grain flow sediment takes place by mass emplacement, involving sudden stagnation of the flow. This occurs when gravity is no longer the driving force in sediment movement as a consequence of either a reduction in the slope angle or consolidation of the flow (Middleton and Hampton, 1973).

The small slump structures are probably a result of small pieces of ice being incorporated into the grain flow at the glacier front. They became rapidly buried and subsequently melted causing the overlying beds to fall into the space previously occupied by the ice. Alternatively, the structures may be erosional scour marks, which were subsequently infilled with sediment. This is unlikely as the beds within the scour would be horizontal and planar rather than following the shape of the cavity.

Superficially the trough cross-beds at the top of transect 8 bear some resemblance to braided stream deposits. There is, however, a lack of small cut-and-fill structures considered to be representative of braided stream deposits (cf. Miall, 1977), and braided channels are not usually developed in subaqueous environments (e.g. Thornburg and Kulm, 1987).

The formation of the unconformity may represent wave action during storms, but unconformities at this elevation (approximately 43m asl) were not detected elsewhere. It could also be interpreted as an erosional surface that formed as relative sea-level was dropping, but the absence of similar features at the same elevation makes this unlikely. Also, beach gravels are not present along the unconformity. Deposits above 43m asl at other transects at site MM-4, and at other sites in the Ragged Harbour Complex are interpreted as subaqueous outwash, and thus these deposits may also be of subaqueous origin. They may represent an ice-proximal subaqueous channel. If these gravels were carried down over the fan surface as a large surge, they would have eroded the upper parts of the underlying beds. This explains the presence of the unconformity, but does not explain why it is horizontal and planar.

The channels may alternatively represent a subaerial channel fill. This, however, could only have occurred while ice was still in direct contact with all the deposits at MM-4, and when sea-levels had dropped below 43m asl. The deformed sediments, however, indicate that ice had retreated prior to the deposition of this upper unit. If ice was still present, then all the sediment at the site, including this upper unit, would have been subjected to slumping when the ice retreated from the area.

Faulting can result due to loading from, for example, grounded ice (Thomas, 1984; Shaw, 1987) or because of collapse due to the melting of buried ice (McDonald and Shilts, 1975; Rust and Romanelli, 1975). At MM-4 most of the faulting can be attributed to the melting of buried ice as faults are commonly associated with slump structures.

7.2.5 - DISCUSSION

The palaeoflow information (Figures 7.5, 7.6, 7.8) indicates that the sediments at MM-4 were deposited generally by northwesterly flow transverse to the geomorphic feature, rather than along its length as would be expected in a subaerial river system. On the basis of this and other sedimentary evidence at MM-4, the successions have been interpreted as pro-glacial subaqueous outwash.

Other proglacial subaqueous deposits are characterized by clast-supported gravels which have little or no horizontal stratification and abundant cross strata (Rust, 1977). Extensive sand units with ripple lamination, high-angle planar beds, and deformation structures including ball and pillow, convolute lamination, and dish structures, are also characteristic (Rust, 1977).

All of these features are present at MM-4. Similar sand and gravel deposits which contain extensive sand and silt bodies and show abrupt changes have been described from several localities in North America (Rust and Romanelli, 1975; Rust, 1977; Cheel and Rust, 1986; Burbridge and Rust, 1988; Rust, 1988; Sharpe and Cowan, 1990; Lundqvist *et al.*, 1993), in Europe (Ruegg, 1983; McCabe *et al.*, 1984; Thomas, 1984; Eronen and Vesajoki, 1988; Fyfe, 1990; Brandal and Heder, 1991), in Australia (Fitzsimons, 1992), and in South Africa (Visser *et al.*, 1984). These deposits contain similar diagnostic sedimentary structures and are all interpreted as forming in proglacial subaqueous environments.

The sediments at the Ragged Harbour 1 site were deposited in an ocean that was at least 46m above present levels, as determined from the highest elevations of the deposits. Sedimentological evidence suggests that they were laid down in an ice-proximal and ice-contact environment. Evidence includes the presence of many faults, especially high-angle normal and reverse faults which are often associated with the melting of buried ice, either as isolated blocks, or at the sediment laden foot of the glacier. The water into which these sediments were being deposited was shallow. The marine limit was at most 67m asl (see Chapter 8) and the sediments at MM-4 occur at elevations between 35m and 46m asl. These were therefore deposited into waters which were 32 metres deep or less. Ice calving off the front of the glacier could easily have become grounded in such shallow water. Rapid sediment deposition would allow the ice to be buried rapidly, and its subsequent melting would cause slumping.

South of transects 2, 3, and 4, there is a rapid drop in elevation marking the edge of the geomorphic feature (see Chapter 4 for discussion). This position marks the ice front while the proglacial sediments at MM-4 were forming. When the ice melted the sediments were left standing well beyond their angle of repose and consequently failed and slumped. Both normal and reverse faults are common under these circumstances.

The presence of diamictons also indicates an ice-proximal location. Although these diamictons are not primary till, they do contain striated clasts. Sediments slumped off the front of the ice or from the base of the ice. There were lulls in sedimentation, as indicated by the presence of clay laminae due to suspension settling. There is no other possible source for these deposits as there are no topographic highs surrounding the deposits.

High-angle planar cross-beds require high velocities, and in this glacial environment are indicative of ice-proximal locations. They generally occur within meltwater conduits in the ice near the glacial margin, or develop at the point where the conduit enters standing water.

7.3 - RAGGED HARBOUR 2 (MM-3)

The Ragged Harbour 2 site is situated approximately 1.5km south of Ragged Harbour (1.5km north of Ragged Harbour 1), on Route 330. It is a large abandoned gravel pit. Five transects were studied at this site, four within one very distinct sedimentary unit dominated by silts, very fine sand and some clay. Ragged Harbour 2 as a whole, however, is dominated almost entirely by medium pebbles to cobbles. The upper parts of the transects contain many large boulders up to 2m in diameter which have large percussion marks.

7.3.1 - GRAVEL ASSEMBLAGE

Exposure of gravels at MM-3 is poor, making sediment description very difficult. Exposures vary from 2 metres to 10 metres high but in almost all places the sections are badly slumped. One large exposure (transect 5) is approximately 10 metres high and has 3 metres of sediment exposed at its upper extent. Four main units were observed in this transect (Figure 7.9).

The lowest unit has a minimum thickness of 140cm. Beds are mainly low-angled (<15°) trough cross-beds with undeformed interbeds of massive coarse sand. These beds are predominantly composed of rounded to subrounded pebbles. They are generally spheres using the Zingg (1935) classification. The main lithologies observed are white granites, gneiss, pegmatite, and granite from the Deadman's Bay Pluton.

Erosionally overlying this unit is a 70cm thick unit of high-angle planar cross-beds composed of sands and silts dipping at up to 70°, well beyond their natural angle of repose. Each bed is structureless, and loading structures are not present.

Embedded within these lowest two units is a large white granite boulder, approximately 2 metres in diameter. No other large boulders were observed in this section although approximately twenty granite boulders of similar dimensions were observed at the top of other sections within the pit or on the pit floor. The boulders are subrounded to rounded, are generally spheres, and are consistently covered with large percussion marks.

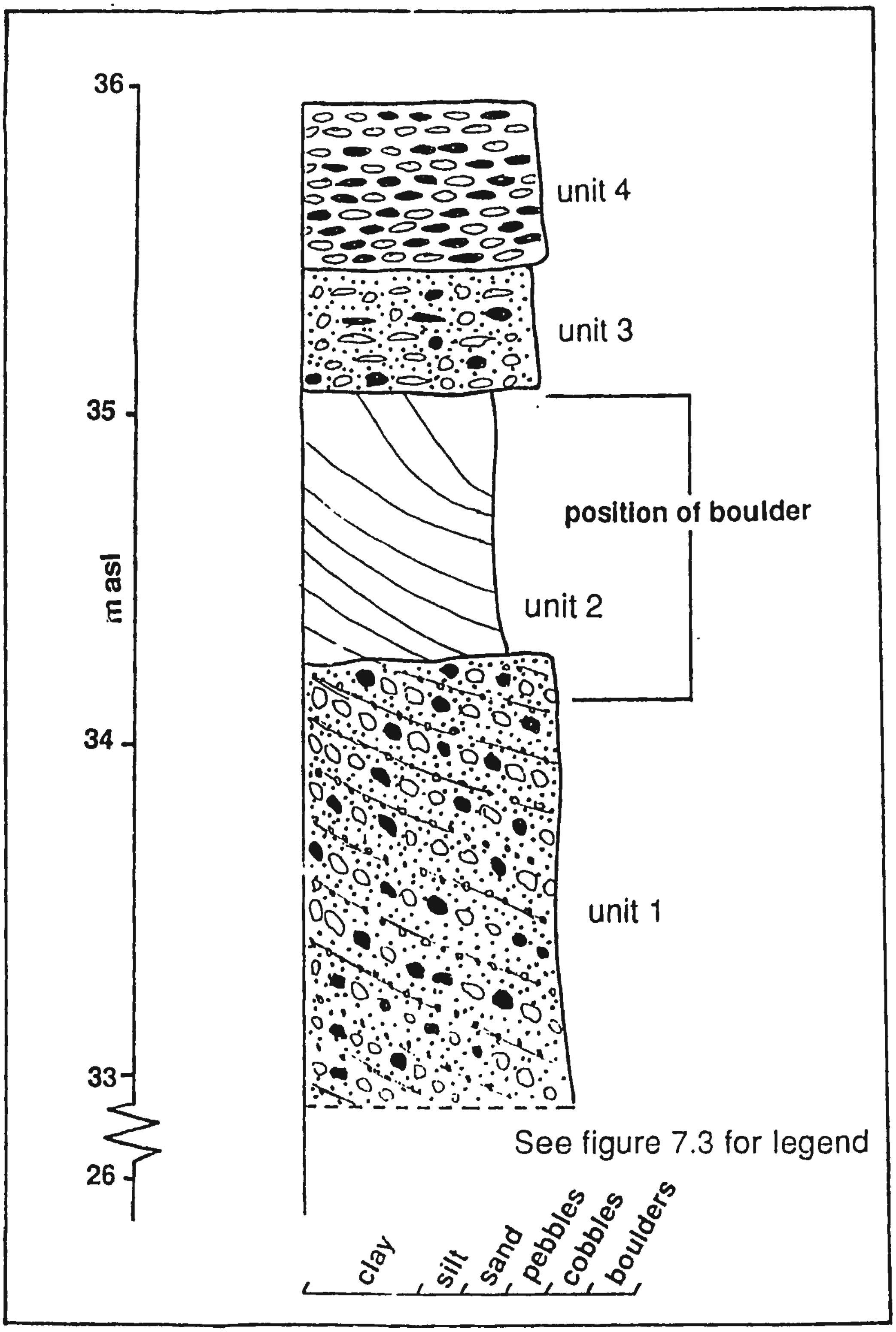


Figure 7.9: Stratigraphic column at section 5 at Ragged Harbour 2 (MM-3).

Abruptly overlying unit 2 is is unit 3 which is composed of well sorted small to medium pebbles with approximately 20% sand. Beds dip to the east at an angle of 5°. Imbrication is poor. Clasts are rounded to well rounded, and are generally spheres with some rollers. They are predominantly local white granite, granite from the Deadman's Bay Pluton, gneiss, pegmatite and schist.

Unit 3 grades vertically into unit 4. This is 40cm thick and is composed of horizontally planar bedded pebble and small cobble gravel. The gravel is clast-supported. Isolated large cobbles (A-axis 20cm) are also present. Clasts are rounded and well rounded and are generally spheres and rollers. Small beds of sand approximately 2cm thick are interbedded with the pebbles. Approximately 20% of the entire deposit is composed of sand. Clasts are well imbricated and generally dip towards the north with low angles (<5°). The unit is very compact due to high concentrations of iron and manganese oxides in these layers. The upper surface of the unit is flat and planar, and marks the surface of a terrace at 36m asl.

7.3.2 - SILT ASSEMBLAGE

Rhythmically bedded silt deposits are restricted to the eastern section of MM-3. Four vertical transects through the silts were examined (Figure 7.10). Transect 1 is in the northeast area of the pit. It contains up to 65cm of rhythmically bedded silt deposits. Approximately 15m to the west, transect 2 contains 60cm of silt and clay deposits. Transect 3, approximately 10m to the south, contains 165cm of silts and clays. The southernmost transect 4, 6m south of section 3 contains a minimum thickness of 220cm of silt and clay.

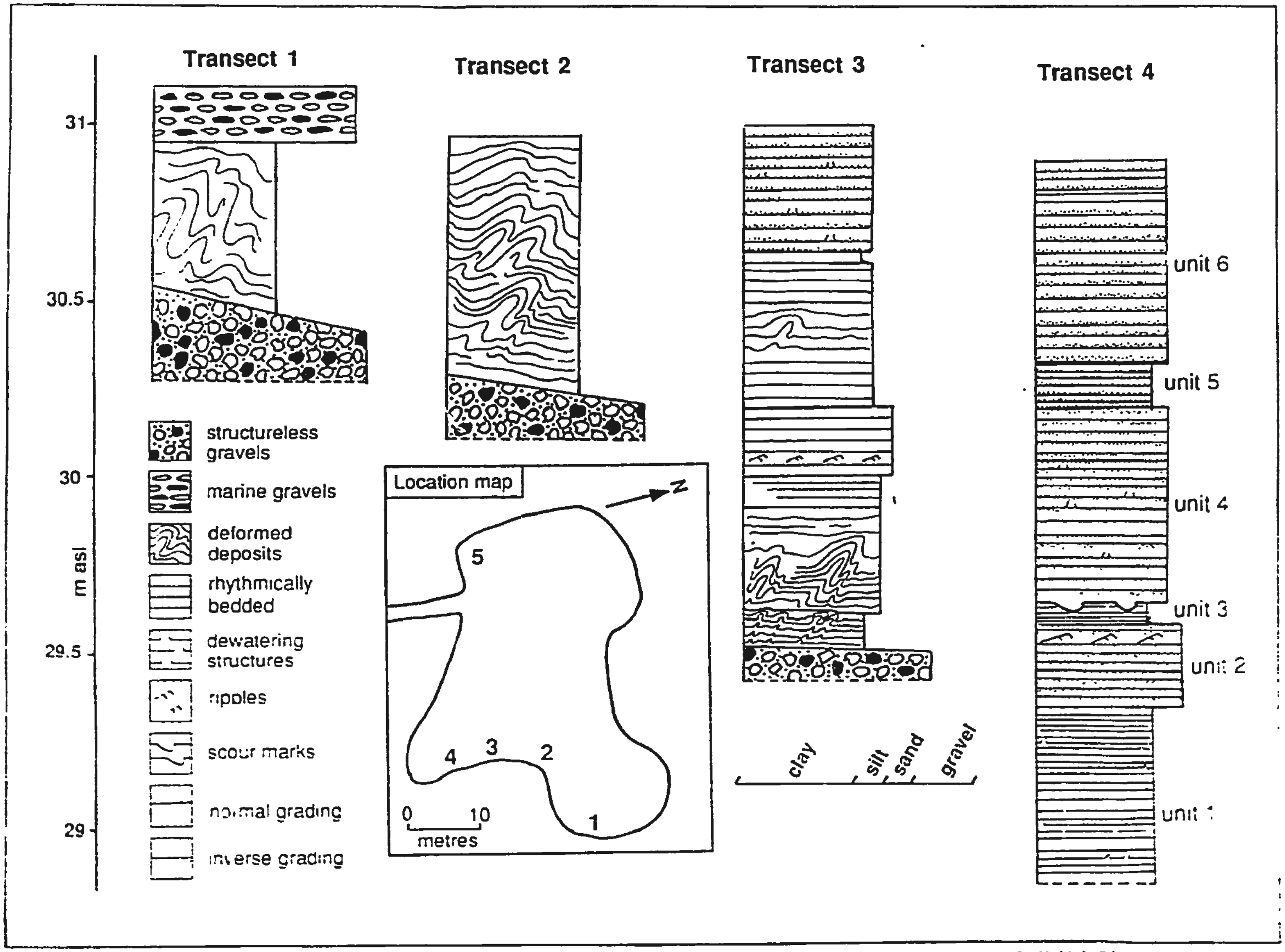


Figure 7.10: Sedimentary transects 1, 2, 3, and 4 at Ragged Harbour 2 (MM-3).

The deposits can be observed for another 15 metres to the south, where they taper out.

At transects 1 and 2, these deposits are abruptly overlain by a bed of well aligned discs and blades, interpreted as beach gravels. The upper surface of the beach deposits forms a terrace at 32m asl, evident above all the glaciofluvial deposits in the area. Directly underlying the silts are well sorted, bedded gravels. Transect 4 is the thickest sequence at this site and consequently will be described in detail.

Throughout its extent transect 4 is composed of rhythmically bedded silt, clay, and sand. Large clasts were not observed. Six major units are described (Figure 7.10). Unit 1 is at least 60cm thick. The lower contact was covered. This unit is composed of relatively compact silt with sand and clay (42% sand, 47% silt, 11% clay). The beds in this unit are thin, generally about 1cm thick, although some strata are as thin as 1mm. All beds are normally graded from sand to clay. Minor flame and other dewatering structures are present along the upper surfaces of most of the beds.

Unit 1 grades into unit 2. This is 25cm thick and is texturally different from all the other units in the transect as it is dominated by sand (84%). Silt accounts for 14% of the unit, and only 2% is clay. Strata are generally thicker than in the underlying unit with most being 2-3cm thick. Most of these beds coarsen upwards, but are often capped by up to 0.25cm of clay, forming a distinct couplet. One bed in the upper part of the unit is 6cm thick and has faint ripple cross-lamination. Laminae maintain approximately constant thickness over the ripple forms. The ripples have an amplitude of 1.5cm and a wavelength of 12cm, with a ripple index of 8. Shapes are asymmetrical, with stoss angles of 8°

and lee angles of 18°. They are flat-crested. Palaeoflow measurements indicate a mean flow direction of 336° (Figure 7.6).

Unit 3 abruptly overlies unit 2 and consists of 6cm of rhythmically bedded silt (28% sand, 63% silt, and 9% clay). Beds are generally thin (3-10mm), with little evidence of grading. Thin laminae (<2mm) of consolidated manganese oxide lie between most of the strata.

Unit 4 overlies unit 3 along an erosional contact which has small scour marks present. This unit is approximately 60cm thick and consists of strata between 0.5 and 1.5cm thick. The unit is generally coarser than the underlying unit with 39% sand, 49% silt, and 12% clay. Most of the beds show faint reverse grading but have no other sedimentary structures. Each bed is capped with clay.

Unit 5 is 12cm thick and lies conformably over unit 4. It is finer than the underlying unit with 14% sand, 77% silt, and 9% clay. Beds are generally thin (<1cm) but are much less well defined than in the underlying units. Contacts are defined where the strata coarsen upwards slightly from finer silt to coarser silt.

Abruptly overlying unit 5 is unit 6. The boundary is well defined by a consolidated manganese oxide layer approximately 0.5cm thick. The exposed thickness of this unit is 25cm, but the upper layers have been quarried away so that the complete thickness of the unit is unknown. Beds are normally graded and each is capped by clay, therefore forming a couplet. Each couplet is approximately 1cm thick and is well defined, with fine sand at the base and a clay cap. This is reflected in the textural measurements of the unit which has more clay and more sand than the underlying unit (15% clay, 65% silt, and 20% sand).

Transect 4 shows no evidence of slumping or deformation. In contrast, section 3, 6m to the north shows minor normal faulting, small recumbent folds, and diapirism in every major unit. The lower contact of the silt unit in this section slopes at 5° southward and overlies structureless gravels. All the beds within the unit dip at approximately 5° to the south.

Transect 1, approximately 25m north of transect 4, is the most deformed of all the silt sections observed. The silts and sands are 65cm thick, are underlain by glaciofluvial gravels, and are overlain by marine gravels. The lower contact slopes at about 15° south. Individual rhythmite beds cannot be identified due to the extensive deformation by dewatering and faulting. These sediments taper out approximately 3 metres to the west.

7.3.3 - INTERPRETATION

The gravel sequence observed at MM-3 is interpreted as subaqueous outwash capped by marine gravels. The lower 2 units represent subaqueous outwash, and resemble the gravels at MM-4. Beds are slumped in excess of their natural angle of repose, due to the melting of buried ice in a proglacial environment. High-angle planar cross-beds are associated with fan or delta formation, or prograding bars (Allen, 1968, 1970, 1983; Harms *et al.*, 1982), both of which occur commonly in ice-proximal environments (McDonald and Vincent, 1972; Thomas, 1984; Sharpe and Cowan, 1990; Shaw and Gorrel, 1990). The dip angle of the planar cross-beds at this site is generally lower than at MM-4, and clasts are generally smaller, suggesting that these deposits may represent a more distal part of the same assemblage (cf. Allen, 1968). The sand beds within this transect have no dewatering or loading features comparable to those observed at MM-4. This suggests that these deposits were laid down at a slower rate than the sands observed at MM-4.

Large boulders with percussion marks were noted by Sharpe and Cowan (1990) in similar deposits in Ontario. They suggested that these may have been swept from a scoured bedrock surface during high-velocity subglacial meltwater flow in keeping with the theories of Shaw *et al.* (1989). Sharpe and Cowan argued that the boulders were not dropstones, as no large clasts were observed in the associated finer sediments. Percussion marks preclude the possibility of deposition by passive ablation. These arguments can be applied to the deposits at MM-3. They also cannot be part of an ablation moraine, as no till deposits were observed in the area.

Extremely high velocity flows that could have swept away the loose gravels, silts, and sand can explain the presence of percussion marks on the boulders. They would have been transported as bedload and collisions with bedrock and other boulders would have resulted in the percussion marks. This is termed overpassing where larger particles move easily over beds of small particles (eg Foley, 1977; Allen, 1983). When clasts are subrounded to well-rounded and are spherical in shape (as they are at MM-3), they can be rolled for relatively long distances (Foley, 1977). The underlying beds act as a smooth surface over which these larger clasts can travel.

Alternatively, the boulders may have simply been melted out of the ice and then rolled down over the fan surface. This helps to explain the presence of the boulders in a distal location rather than a proximal location, but it does not explain why the boulders are isolated in the tops of the sections only.

The upper two units have been reworked, with clasts showing a well aligned fabric (better developed in unit 4) and imbrication. These deposits are interpreted as beach deposits. As sea-levels started dropping towards the end of the Late Wisconsinan, the upper level of the subaqueous deposits would have been reworked by waves, forming beach deposits. The upper surface marks a terrace at 36m asl which extends parallel to the modern shore for over 10km. Energy levels were moderate to low as indicated by the presence of small and medium pebbles.

The rhythmically bedded silt deposits at MM-3 are interpreted as a subaqueous kettle fill. Stratigraphically, they overlie the gravels and therefore must represent the final stage of subaqueous deposition prior to beach formation. The kettle has a lateral extent of approximately 34 metres from north to south. Prior excavation precludes determination of the east-west extent. The depth of the kettle is at least 220cm as determined from the thickness of the sequence at MM-4. Depths, however, do not greatly exceed this, as the lower contact of the silt deposits was observed only metres to the north and south.

Catto (1989), interpreted a similar deposit, in Sweden, of smaller dimensions (0.52m thick x 8.9m wide), as a subaerial pond fill which formed on the top of stagnating ice, as opposed to a subaqueous kettle fill. Laminae were of similar thicknesses and the contacts between laminae were sharp and planar. The sequence is similar to that at Ragged Harbour 2 and emphasizes the need for examining all sedimentary units together before interpretation of the environment of deposition is determined.

Transect 4, measured through the central area of the sedimentary feature, shows no evidence of deformation. The degree of deformation increases systematically toward the northern margin of the kettle fill (sections 1, 2, and 3). Slope angles also increase towards the edge of the feature. Failure can occur on slopes as low as 1° in a subaqueous environment (Shepard, 1955; Mills, 1983). This slope failure can result in faulting and can enhance diapirism. As the slopes at the edge of the kettle fill are relatively steep (5°-15°), gravitational processes would be initiated easily, preferentially inducing slope failure in these areas. Slumping would thus produce the greatest amount of deformation at the edges, where the slope angles are highest. Where beds are horizontal (in transect 4), there is no slumping and thus no deformation. Rust (1988) noted similar kettle fills in the Ottawa region. He suggested that such depressions, in sheltered positions below wave base, except during major storms, would be marked by low energy conditions.

The lowermost unit at transect 4 shows no evidence of traction sedimentation. Rhythmites are small and normally graded. It was most likely that this unit was formed by several episodes of direct suspension settling.

Unit 2, dominated by sand, is texturally different to all the other units observed at transect 4. The asymmetric ripples in this sequence show that traction current sedimentation prevailed throughout the deposition of this unit. These ripples are type B ripples of Jopling and Walker's (1968) classification, and depositional stoss of Ashley *et al.*'s (1982) classification. These represent relatively low velocities of between 15 cm sec^{-1} and 25cm sec^{-1} (Ashley *et al.*, 1982). Velocities were probably at the lower end of this spectrum, as the laminae on the stoss and lee sides of the ripples are approximately the same thickness. This indicates that suspension sedimentation was also important in the formation of these features (cf. Jopling and Walker, 1968). Coarsening-upwards sequences are usually indicative of grain flows (Middleton and Hampton, 1973). Ripple sedimentation usually occurs at rates of centimetres to decimetres per hour (Allen, 1971), and thus this thin unit was deposited relatively quickly.

The lower thin rhythmites in this unit each represent a period of traction sedimentation (inversely graded sands) followed by a period of suspension settling (clay cap). These may represent the Bouma T_d and T_e divisions within a turbidite sequence which are representative of lower flow regime traction sedimentation (cf. Middleton and Hampton, 1973). These beds, therefore, may represent several periods of low density turbidity current sedimentation. Most turbidity currents occur as surges (Middleton and Hampton, 1973), and these beds in unit 2, therefore, probably represent several such surges The rippled unit is representative of the Bouma T_c division which usually contains ripples and/or wavy or convolute laminae, and represents lower flow regime traction sedimentation.

Unit 3 shows no evidence of traction current sedimentation. Because most strata are thin and fine grained, they probably resulted from direct suspension sedimentation. The manganese oxide layers lying between most strata were formed because clay layers at the surface of each stratum would have restricted water percolation, allowing manganese oxide to become concentrated on the surface of each stratum. The surface of each clay stratum thus also represents a time lag before the deposition of the overlying stratum.

Unit 4 represents several periods of either turbidity current sedimentation or grain flow. This is shown by the reverse grading in most of the beds. Scouring of the upper surfaces of the underlying beds is also associated with these modes of deposition (Middleton and Hampton, 1973). Like the lowermost beds in unit 2, these deposits probably represent the Bouma T_d and T_e divisions. The clay caps, however, are less prominent in these beds suggesting that surges occurred at closely spaced intervals.

There was a time lag between the deposition of unit 5 and unit 6. No structures were observed in the units and therefore it is difficult to establish their origin. They may have formed entirely by suspension settling, or by a combination of traction and suspension settling.

Because of the fine nature of these sediments and the thickness of the beds, they represent very low energy conditions. As the ice mass was ablating, pulses of meltwater were emitted either diurnally, seasonally, or annually. These pulses could have disturbed proximal sandy sediments causing slumps and slides which developed into turbidity currents as they moved down the surface of the subaqueous fan(s). Also, deposits slumping at the sediment-ice interface could trigger similar gravity flows. Some of the finer deposits within the kettle fill may represent small scale turbidity currents or grain flows initiated at the unstable edges of the feature.

Site MM-3 is interpreted as an ice-proximal subaqueous assemblage that is more distal than the deposits at MM-4. Evidence for this includes high-angle planar bedding (lower angle than at MM-4), slumped sands, and large boulders with percussion marks. The high-angle planar beds and slumped sands represent ice-proximal and ice-contact deposition, and high velocity meltwater flow. The kettle originally would have formed due to the melting of buried ice and subsequently would have filled with sediment.

7.4 - DEPOSITIONAL ENVIRONMENT

In summary, all the sediments in the Ragged Harbour exposures represent rapidly deposited proglacial, ice-contact and ice-proximal subaqueous outwash. The surface morphology and the sedimentology indicate a subaqueous origin. These sediments form a ridge 15km long, running ENE to WSW, which is interpreted as several outwash fans which join together to form a moraine (refer to Chapter 4; Section 4.4). Similar features have been described in Ontario (Sharpe and Cowan, 1990), in Wisconsin (Lundqvist *et al.*, 1993), and in Finland (Fyfe, 1990). Generally these features are associated with eskers on the proximal sides. In the Carmanville region, no eskers were observed. Subglacial meltwater likely moved through interconnected subglacial and englacial cavities. Meltwater escaping from cavities at the front of the ice formed a series of fans and subaqueous channel deposits, which joined to form the moraine (Fig 7.11).

Palaeoflow information from site MM-4 proved to be variable, whereas at sites MM-1, MM-2, and MM-3, flow directions are distinctly concentrated between NNW and NW. The stratigraphy at MM-4 suggests that the directions of flow changed very rapidly and thus flow was unrestricted. This pattern of deposition is most likely related to migrating outlets along the ice front in this area. Only one or two outlets were open at any one time.

As shown at MM-3, crude trough cross-bedding is present in the northeastern section of the moraine. The presence of cross-bedding suggests that individual point sources for the entry of meltwater into the ocean existed at the ice front. Point sources for meltwater transfer, however, should result in sediment being deposited in a fan shape. This should be reflected in the

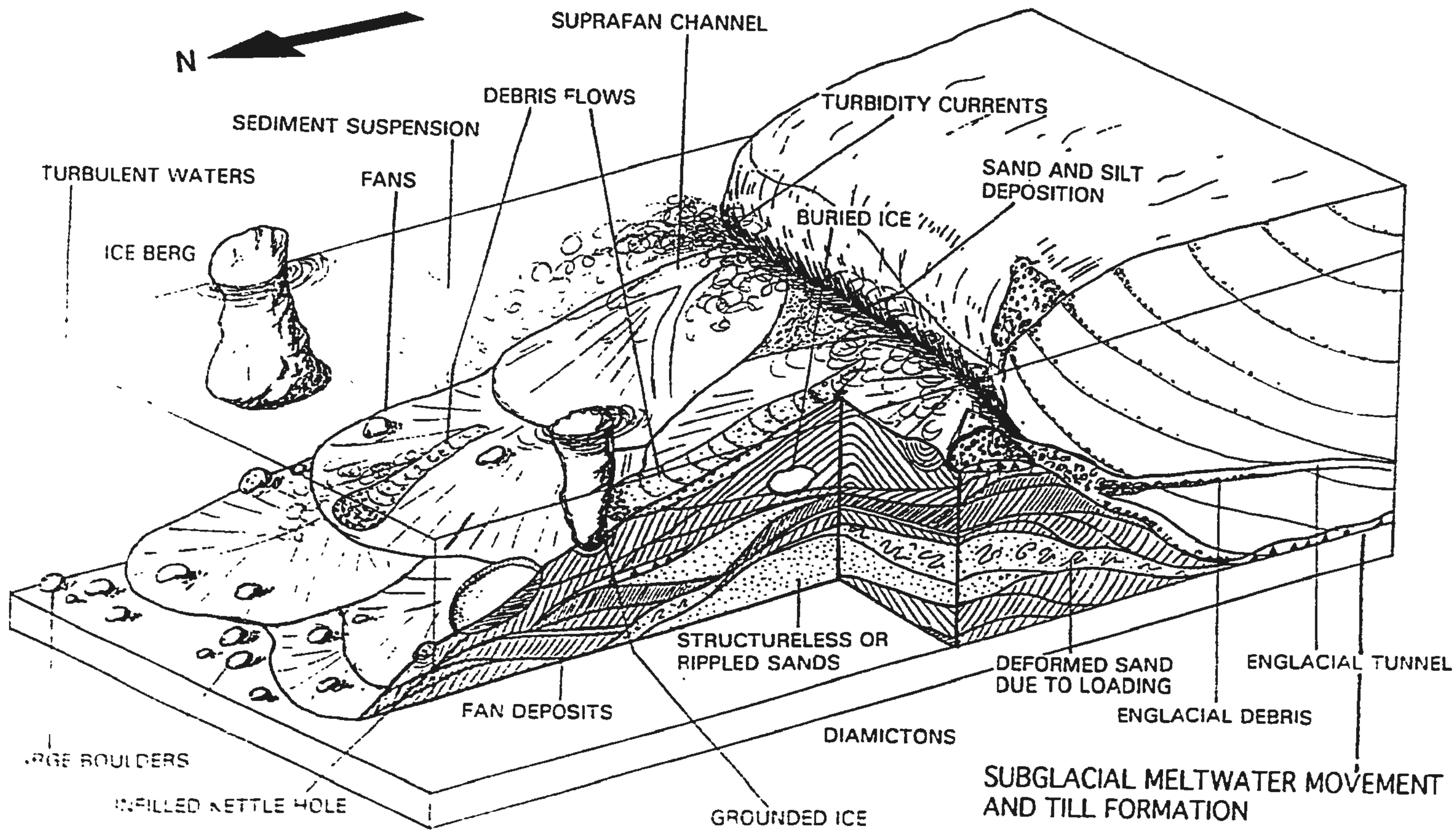


Figure 7.11: Schematic diagram showing the style of subaqueous deposition as ice was calving into high sea levels of at least 46m asl. Deposition was rapid, indicated by the fan formation and the deformed sands and silts.

palaeoflow data. The palaeoflow information at these sites suggests relatively unidirectional flow. This can be explained by postulating the existence of many outlets at the ice front. If each outlet formed a fan, the deposits from each outflow would amalgamate and perhaps overlap in places. All flow would be forced out from the ice front, longitudinally directed by other meltwater flows.

As the ice ablated, these fans formed rapidly in front of the ice. Rapid sand deposition took place between the fans forming the proximal rhythmites (Figure 7.11). Each rhythmite could represent diurnal, seasonal, or annual accretions, related to periods of increased meltwater activity. Crowley (1984) noted that a decrease in sedimentation rate filters out low-magnitude events. Rhythmically bedded deposits which are related to decreased sedimentation rates may not, therefore, be representative of all events during the time of their deposition. It is therefore difficult to ascertain the timing of these events.

No shells or other marine skeletal fauna were found in the subaqueous deposits at the Ragged Harbour Complex. It is unlikely that life would be attracted to areas of extremely rapid deposition marked by turbulent waters with large quantities of sand, silt, and clay in suspension, and calving ice (Figure 7.11). This would make waters cold and brackish, and unsuitable for marine life (cf. Powell, 1984).

Further downwasting would have allowed the newly exposed sediment faces along the length of the moraine to fail (Figure 7.12). Many of the deformation structures observed in the sediments are post depositional and would have formed at this time. As sea-levels dropped rapidly in the area (Shaw and Forbes, 1990a), those deposits facing northwest were reworked

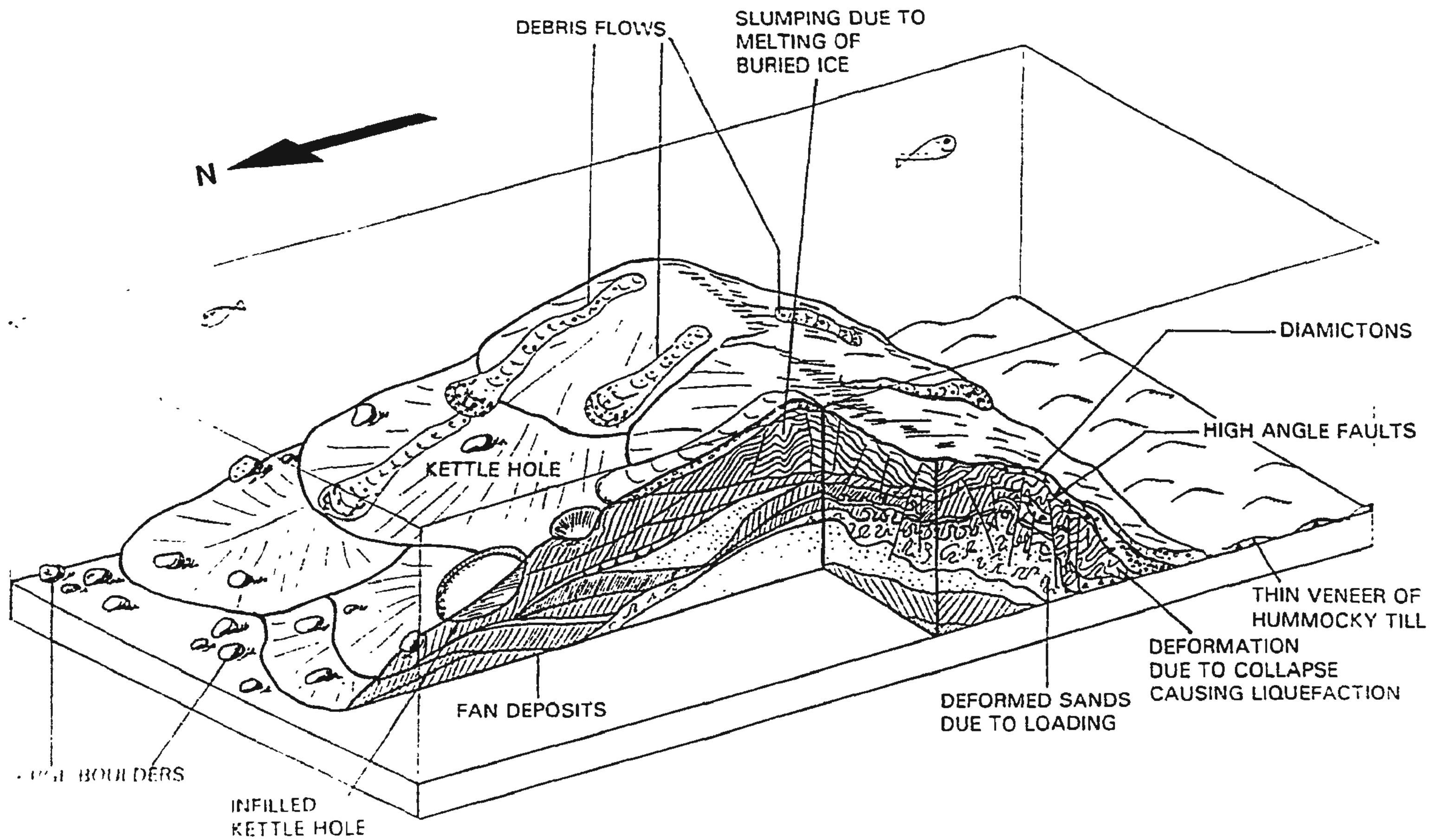


Figure 7.12: Schematic diagram showing the modification of the subaqueous deposits once ice had retreated from the area, buried ice had melted, but when sea levels were still high. The waters were less turbid and less cold and may have been able to sustain marine life.

periodically by marine action. This resulted in the formation of beach terraces composed of approximately 1m of beach sediment, overlying the subaqueous deposits (Figure 7.13).

It is unlikely that the Ragged Harbour Moraine represents a significant stillstand of ice. Sedimentary analysis suggests extremely rapid sedimentation, and also, as sea-levels were dropping rapidly in this area, there was a limited time during which these deposits formed. Ice had retreated from Hamilton Sound by 12,000 BP (Shaw *et al., in review*) thus the deposits post-date this time. MacKenzie (*in preparation*), noted, that on the basis of ^{14}C dates, and areal and stratigraphic relationships, all sea-level stands higher than or equal to 34m asl are older than 11,400 BP. These sand and gravel deposits, forming in waters of at least 46m asl, therefore predate 11,400 BP. As sea-levels were dropping rapidly during this time, this deposit probably represents a maximum of approximately 100 years of deposition some time during the period 12,000 BP to 11,400 BP. Sharpe and Cowan (1990) noted that a similar feature in Ontario did not represent a significant stillstand of ice.

Subaqueous outwash has often been associated with glacial surging and flow expansion (Thomas, 1984; Burbridge and Rust, 1988; Sharpe and Cowan, 1990; Lundqvist *et al.*, 1993). There is no conclusive evidence at Ragged Harbour that these deposits were associated with advancing ice. No melt-out tills are present and there is no evidence of glaciotectonism. These deposits were not deformed by overriding ice.

Sharpe and Cowan (1990) suggested that deposition of subaqueous deposits in Ontario may have been initiated by lowering lake levels. Lower lake levels would result in a reduced potentiometric surface within the ice, and this

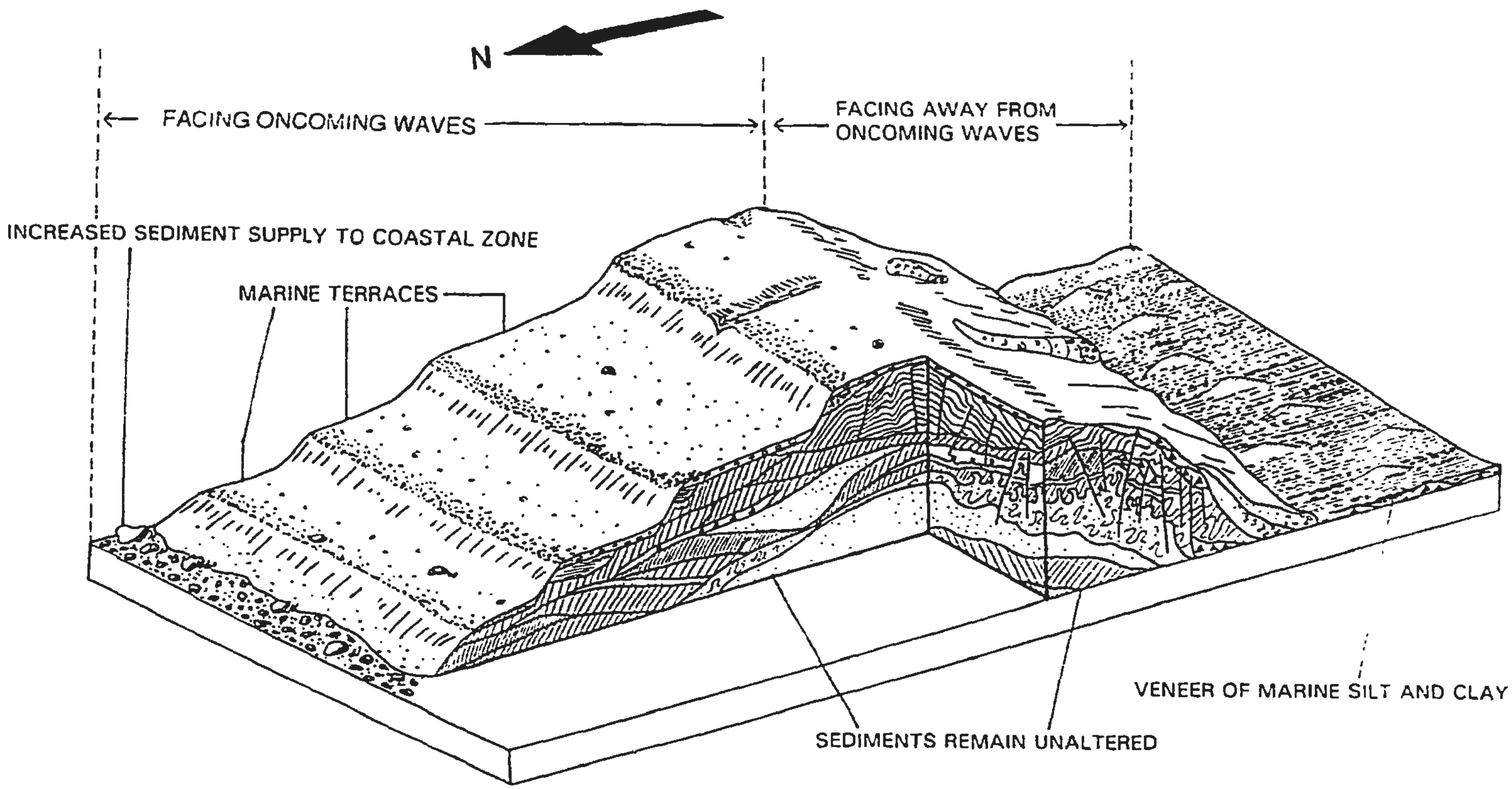


Figure 7.13: Schematic diagram showing the surface morphology and internal stratigraphy of the subaqueous deposits as they are present times. Lowering sea levels modified the upper 1m or less of all the deposits.

would result in increased meltwater discharge and thus increased proglacial sediment deposition. It is possible that the Ragged Harbour moraine formed in response to a rapid drop in sea-level. Similar deposits should have formed around the coast of Newfoundland close to the marine limit if this was a common process.

In summary, the glaciofluvial deposits observed in the Carmanville region represent a 15km long linear feature interpreted as several outwash fans which coalesced to form a moraine. An array of sediments were deposited proglacially which relate to extremely high flow velocities. Liquefaction took place in the silts and sands, deforming them. Calving ice grounded immediately and subsequently melted to induce slumping. Further melting of the glacial ice caused major slumping at the former ice margins.

CHAPTER 8
SEDIMENTOLOGY 3: EMERGED BEACH SEDIMENT

8.1 - INTRODUCTION

Almost all of the modern shoreline in the Carmanville area is wave-dominated micro-tidal (cf. Davis and Hayes, 1984), characterized by moderately- steep faced gravel beaches (cf. Forbes and Taylor, 1987). As the entire region has been affected by glacial deposition (mainly till veneers), there is a ready sediment supply to the coast. The widespread occurrence of gravel beaches in Atlantic Canada can be attributed, in part, to the high proportion of coarse clastic material in glacigenic sediments (Forbes and Taylor, 1987). As well as till and diamicton deposits, the subaqueous deposits in the northeast of the study area near Musgrave Harbour provide an abundant source of both gravel and sand. Large flat sandy beaches exist east of the study area at Musgrave Harbour. A small dune field also exists behind the beaches.

The coastline in Gander Bay is not wave dominated. The beaches are poorly developed with small (<5m wide) areas of poorly sorted gravel, sand, and mud on the western shore. Mud flats occur on the eastern shore.

Quinlan and Beaumont (1981) classified the region as zone C in their sea-level model, although subsequent workers have shown the area to be classified as zone B (see chapter 2). This implies that the area had relatively high sea-levels which fell rapidly at first, continued to fall below present, and subsequently rose back to 0m asl. Because of the large change in relative sea-level, older beach sediments should be left high above the reach of wave action (Forbes and Taylor, 1987). This is true for the Carmanville area where emerged beaches are observed all around the coast at elevations up to 57m asl. Sections were studied at Victoria Cove, Wing's Point, Aspen Cove, Ladle Point, and Ragged Point (Figure 8.1). Many of these areas are currently being exploited for aggregate, and consequently many well exposed sections were available for study. Each of the sections discussed here has a slightly different geomorphic setting.

8.2 - VICTORIA COVE (MM-45)

Description

Five transects were studied in detail at site MM-45. The upper surface of each of these marks the top of a terrace at 11m asl. Four distinct units were observed, although not all units occurred in all sections. The lowermost and thickest unit observed consists of interbedded sand, granule and pebble gravel. In some areas this unit is horizontally bedded and in others the beds dip east with low angles (<10°). At transects 1, 2, and 3, unit 1 is overlain by a diamicton unit. The diamicton is overlain by a thick manganese oxide layer. Where the diamicton is absent, this layer directly overlies the lowermost unit. Directly overlying the manganese oxide is a unit of horizontal planar or dipping planar beds of cobbles. Transect 3 contains all four of the above units (Figure 8.2), and consequently will be described in detail here.

Transect 3 is located in the centre of the pit at Victoria Cove and faces south. All units are dominated by the same clast types, (slate, shale, basalt, greywacke, and schist). Minor quantities of siltstone, gabbro, greenschist, and sandstone are also present.

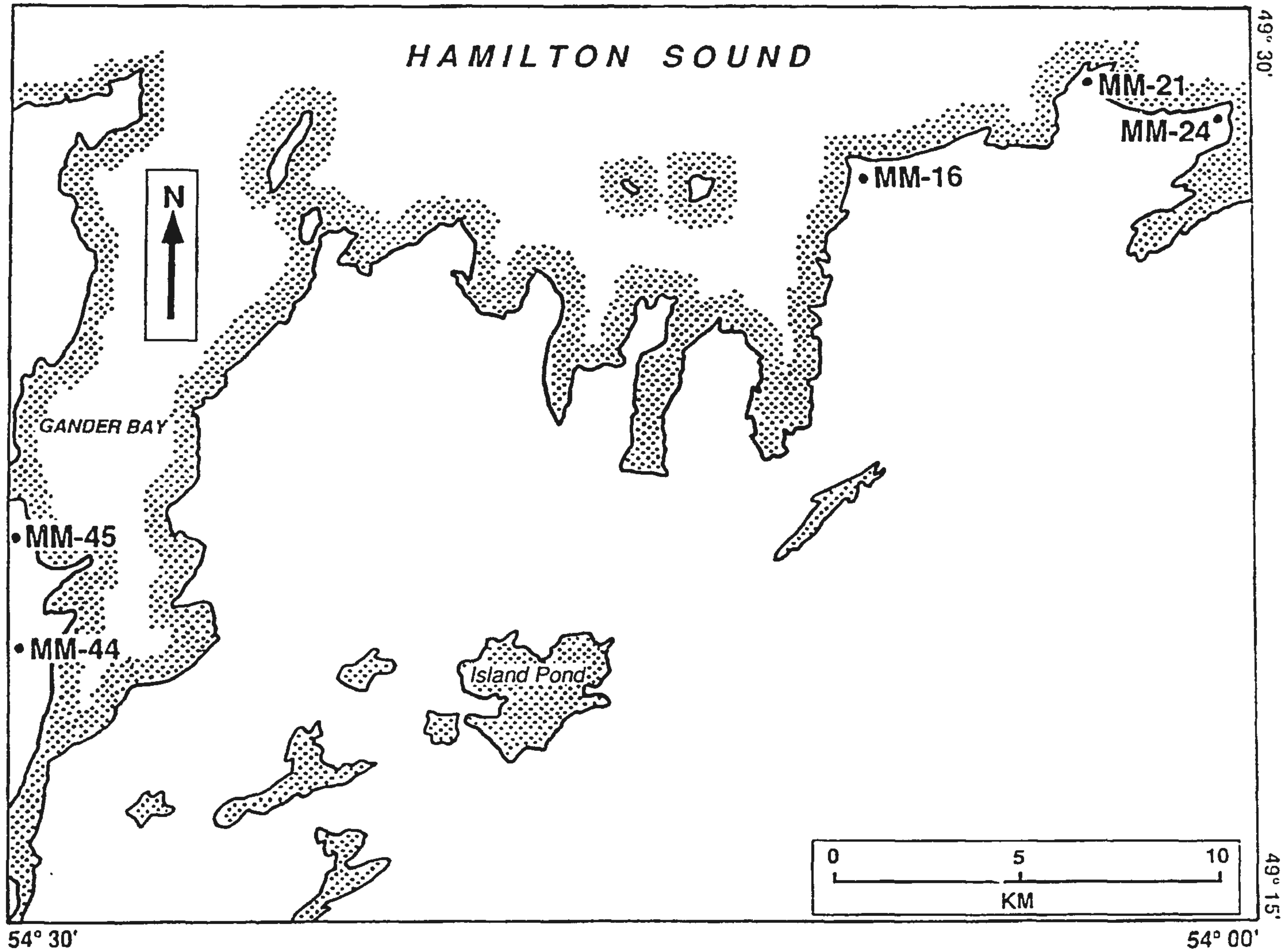


Figure 8.1: Location map of all observed sites containing raised marine sediment.

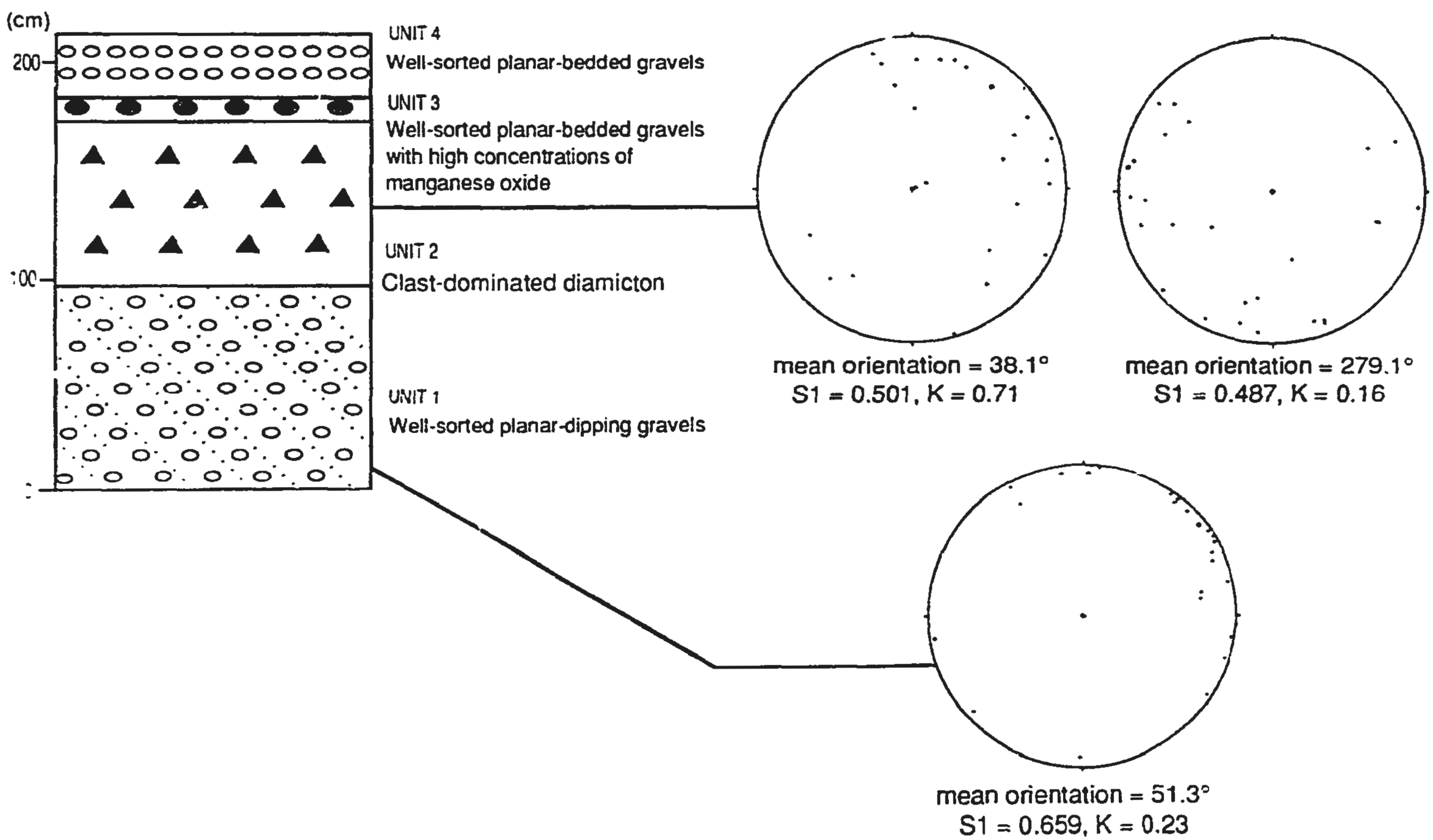


Figure 8.2. Vertical transect through sediments at section 3, Victoria Cove

The lowermost unit, having a minimum thickness of 120cm, is a clast-supported gravel consisting of approximately 80% aligned pebbles and granules with some cobbles. The basal contact was not observed. Sorting is moderate to good. Medium and coarse sand account for approximately 20% of the matrix. Iron staining is common, making the unit very dark brown (Munsell: 7.5 YR 3/4 mf). A series of coarsening upwards and fining upwards sequences are present within this unit, and the beds generally dip east at 10°. Rounded to well rounded blades, discs, and rollers are present in approximately equal numbers. Fabric analysis indicates that most of the clasts are lying approximately flat or have very low plunge angles averaging 4.6°. The sample has a mean trend orientation of 51.3°, with an S_1 value of 0.659, and a K value of 0.23. This fabric therefore has a girdle distribution using Woodcock's (1977) classification.

Unit 1 is abruptly overlain by a matrix-supported diamicton. The basal contact is undulatory. It varies from 45cm- 75cm-thick. The unit is grey towards the top (Munsell: 2.5YR N5/ mf), and dark yellowish brown (Munsell: 10YR 4/6 mf), towards the base. The basal part of the unit is silt and clay dominated with 33% clay, 45% silt, and 22% sand. The unit fines upwards to clay. No structures are present in the basal part of the unit. The upper part contains some clay nodules which have a maximum diameter of 4cm. These contain approximately 65% clay. Faint stratification is present. Approximately 25% of the unit is composed of pebbles and small cobbles, although one large (75cm A-axis) clast of sandstone is present. There are equal numbers of sphere, disc, and roller clast shapes. Clast fabric analysis on unit 2 shows orientations to be variable, with mean trends of 38.1° and 279.1°, and mean plunges of 17.2° and

4.9°. Some clasts have dips up to 70°. This is reflected in the relatively high S_3 values of 0.176 and 0.115. The fabrics are poorly oriented with S_1 values of 0.501 and 0.487, and K values of 0.71 and 0.16. These therefore represent girdle distributions using the classification of Woodcock (1977).

Unit 2 is abruptly overlain by unit 3 which is 15cm thick and composed of horizontally planar-bedded pebbles. These are moderately to well sorted. Clasts generally lie flat although some are oriented vertically. They are predominantly blades and discs and are rounded to well rounded. The unit is lithified due to an extremely high concentration of manganese oxide. As a result the entire unit is stained black.

Unit 3 grades vertically into unit 4. In contrast to unit 3, unit 4 is not lithified or stained black. The composition is similar, marked by well sorted horizontally planar-bedded, rounded to well rounded pebbles which are mainly discs and blades. Vertically aligned clasts are prominent in this unit.

Interpretation

The sediments in transect 3 are interpreted primarily as beach deposits due to the presence of discs and blades (cf. Komar, 1976; Bourgeois and Leithold, 1984). As all the beds, apart from the diamicton, are moderately to well sorted it is likely that they formed in the upper beach face where sorting is generally better than lower in the beach profile (Komar, 1976; Massari and Parea, 1988). An upper beach origin is also reflected in the low eastward dip of the beds towards the modern coastline.

The diamicton may have formed either as a bank collapse or an iceberg dump. It is not a till as it does not contain striated clasts, the basal contact is undulatory, the matrix is very fine, and the clast fabric orientation is poor. This diamicton was also detected in sections 1 and 2, which are 40m apart, indicating that its lateral extent is at least 40m by 15m. The undulatory basal contact suggests that it was deposited suddenly. The weight of the large clast deformed the underlying beds. Clay nodules within the deposit are probably post-depositional.

The deposit may perhaps represent a sediment gravity flow deposit. Ice would have to be present to provide the necessary elevation. There is no evidence, however, to indicate that ice was present in the area as all the marine terraces above 11m asl are perfectly preserved. They would have been destroyed or extensively deformed if glacial ice was present. This deposit is not an iceberg dump as deep water sedimentation is not compatible with beach formation.

This, therefore, may be a bank collapse deposit, perhaps initiated due to wave undercutting during a storm. As discussed in Chapter 5 (Carmanville South 2), bank collapse deposits usually have erosional, curved or planar contact surfaces (Gibling and Rust, 1984; Rust and Jones, 1987; Williams and Flint, 1990), and this deposit does not. Failure planes should also be observed (cf. Gibling and Rust, 1984), but at site MM-45 they are not. As noted for the deposit at Carmanville South 2, this may represent blocks toppling from an unconsolidated bluff as it was undermined by waves, and thus the failure planes are not observed. The undermining of an unconsolidated bluff is therefore the most likely origin for this diamicton.

The upper two units in this sequence are interpreted as beach gravels which, like unit 1, formed in the upper beach face area. These units are generally better sorted than unit 1 and contain higher proportions of discs, suggesting that they probably formed higher in the beach profile where sorting increases (Massari and Parea, 1988), and where disc-shaped clasts become more common (Bourgeois and Leithold, 1984). Clasts are normally larger in this zone than lower in the beach profile. Clasts in units 3 and 4 are not larger than clasts in unit 1, but this is not uncommon in gravel beaches. The upper units in the sequence probably relate to a storm event of differing intensity than the clasts in the lower unit.

Manganese oxide deposits are post-depositional. It is possible that water carrying heavy minerals was unable to infiltrate the clay-dominated diamicton directly below consequently enhancing manganese oxide deposition. However, the manganese oxide layer occurs at the same elevation at all the sections in the Victoria Cove pit, whereas the diamicton is not always present. Consequently, it is probably coincidence that this layer overlies the diamicton at this section. Its presence may be explained by a high water table that remained at that level over a lengthy period of time. This may have enhanced manganese deposition directly above the water table. The water table would thus be related to a lower sea-level than the 11m stand. The next major stand in the area is 8m asl. Manganese oxide deposition cannot be related to minor stands between 11m asl and 8m asl, as a relatively long time would be required for this thickness of manganese oxide to form.

In summary, the deposits at Victoria Cove (MM-45) represent storm deposits, which have a greater chance of survival in the sedimentary record than fair weather sediments (Massari and Parea, 1988). It is difficult to distinguish how many events occurred during the deposition of these beds.

They may represent the effects of one main storm or because of the dynamic nature of beach sedimentation they may represent a lengthy period of storm activity. Sea-levels may have dropped a few metres and then remained level over a long period of time, resulting in a stable water table at site MM-45 at approximately 10.5m asl. Sea-level continued to drop leaving the elevated deposits at Victoria Cove.

8.3 - WING'S POINT (MM-44)

Description

The upper surface of the site at Wing's Point forms a terrace at 34m asl. It is situated approximately 1km inland from the modern coast. Five main units are described at the site and they have a maximum thickness of 260cm (Figure 8.3). These directly overlie shale bedrock that has been eroded into a flat wave-polished platform.

Unit 1 is approximately 110cm thick. It is dark yellowish brown (Munsell: 10YR 3/6, mf) towards its base and is lighter in colour and less compact towards the top. The basal part is stained with manganese oxide. The unit is dominated by fine granule gravels and small pebbles. There are no clasts coarser than pebbles and none finer than medium sand in this unit. Most of the larger clasts are discs and blades, although preferential shaping is restricted to shale and slate clasts. Granules are generally spheres. All clasts are subrounded to rounded. At the base of the unit, beds are horizontal and planar. Towards the top of the unit, the beds dip to the west at an angle of approximately 5°. Sorting improves towards the top of the unit, and imbrication becomes more prominent.

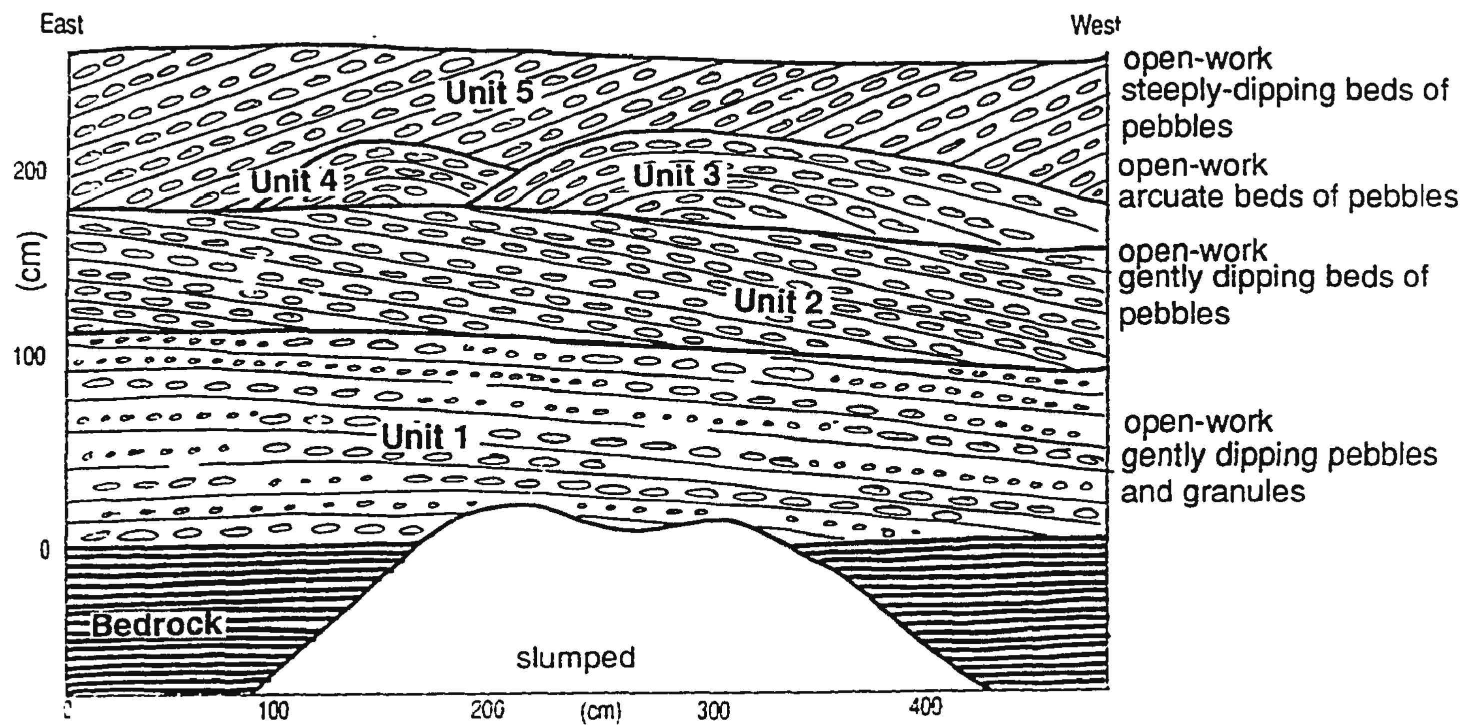


Figure 8.3: Section at Wing's Point (MM-44)

Unit 1 grades into unit 2 which is a 55-cm-thick moderately compact gravel containing well-sorted, mainly disc-shaped medium pebbles and fine pebbles and granules. Subrounded to rounded clasts dominate. The matrix consists of a high proportion of fine and medium sand (30%), and is more compact than the underlying layers. This unit is well stratified and contains both fining- and coarsening-upwards pebble-granule sequences. Manganese staining is also less apparent in this unit. Beds dip towards the west at angles of between 5° and 10°.

Abruptly overlying unit 2 is unit 3. The unit is 60cm thick at its thickest point and extends laterally for 170cm. It consists of several arcuate strata which differ texturally. Clasts range from rounded to well-rounded granules to large pebbles, and the predominant shapes are discs and blades. The beds dip, on average, about 15° towards the east, with the angle decreasing upwards asymmetrically to the top. The upper beds lie horizontally. They then dip to the west, with the lower parts of the beds dipping more steeply (17°), than the upper parts (7°). Clasts are generally well imbricated, although on the west side they tend to lie obliquely to the dip direction.

Abruptly overlying unit 3 is a thinner unit (20cm) with similar sedimentary characteristics. It is 20cm thick at its thickest point and extends laterally for approximately 40cm.

The uppermost unit consists of 40cm of steeply dipping (25°) beds of disc- and blade-shaped pebbles. Clasts are rounded to well rounded. Sorting is good in the basal part of the unit, but the upper parts are moderately to poorly sorted. Some of the clasts are aligned vertically. Beds generally dip towards the east at an angle of approximately 3°.

Interpretation

The deposits at MM-44 are interpreted as a beach succession. Like the sediments at Victoria Cove, these contain high proportions of disc-shaped clasts. This site is unusual in terms of emerged beach deposits in the Carmanville region, as it is much thicker than deposits at similar elevations (34m asl). Coastal processes, therefore, played a more prominent role on this part of the coast in the past than at present. The polished bedrock surface underlying the sediment is interpreted as a wave cut platform. These occur in modern times along the more exposed coastline between Ladle Point and Aspen Cove. The thickness of the deposits may also be, in part, a function of the bedrock type underlying the deposits. Slate is very fissile and thus will break up easily in a coastal environment and be reworked by marine action.

When sea-level was at 34m asl, the coastline configuration was quite different to that of today (see Chapter 9). All of the Dog Bay headland was under water and all of the Mann Point, Main Point, Gander Bay Brook and Frederickton areas were inundated by coastal waters. There were scattered islands at Noggin Hill and the highland area between Carmanville and Frederickton. The whole coastline was thus very exposed when sea-levels were at 34m asl. It would also have been relatively open to the oncoming prevailing waves from the northeast, as it faces easterly. The west coast of Gander Bay was thus a good area for beach formation when sea-levels were at this level. Energy levels were probably not as extreme as those observed on the present, relatively high energy coastline at Ladle Cove but were high enough to erode the bedrock and deposit pebble-sized beach sediment.

In a beach environment, the areas which tend to slope away from the ocean are the shoreward edges of storm ridges, and the backshore area behind the berm crest (Davis, 1978; Bourgeois and Leithold, 1984). Both of these features were observed on the modern beaches at Aspen Cove and Ladle Cove. Units 1 and 2 represent the backshore area of a beach, as they dip away from the coastline. Their overall extent shows the beds to be consistent laterally for about 6m. This is too large to be the shoreward edge of a storm ridge, as such ridges in both the modern coastline here and at other localities are characteristically less than 2 metres wide.

Units 3 and 4 are interpreted as storm ridges. These were observed on almost all of the beaches in the study area. These occur most commonly in the transition zone between the backshore and the foreshore zone in a beach face (Davis, 1978; Bourgeois and Leithold, 1984). Storm ridges are usually relatively small, with the largest features being up to 2m in width. The features at MM-44 are therefore typical in size. Sediment should be well imbricated on the coastward side of the ridge and less well so on the landward side (Komar, 1976). Both ridges at MM-44 show these characteristics (Plate 8.1). It is unusual that these features are preserved in the sedimentary record, as usually they are eroded by waves and replaced by gently dipping beds (Elliot, 1986; Massari and Parea, 1988). This indicates, therefore, that the features must have been buried rapidly after their formation.

Unit 5 represents the beach face in the foreshore zone of a beach. As the beds are well sorted in the basal part of the unit, they probably formed in the upper beach face area just below the berm crest area, where sorting is

Plate 8.1: Relict beach ridges preserved in emerged beach gravels at Wing's Point (MM-44).

generally much better than in the lower parts of the beach profile (Komar, 1976; Massari and Parea, 1988). This is also reflected in the relatively low angle of dip of the beds towards the modern coastline. The upper part of the unit is disturbed and as a result appears poorly sorted. Most of the disturbance can be accounted for by post-depositional changes. For instance, the vertically oriented pebbles may indicate frost heave processes (cf. Mackay, 1984; Anderson, 1988). Pedogenesis could account for the remainder of the disturbance in the upper 40cm of unit 5.

This section is unusual in terms of preservation of the sedimentary structures. Storm ridges on beaches are usually planed flat by wave action

(Massari and Parea, 1988). The features at Wing's Point can only have been preserved if they were buried soon after their formation. It is likely that at least one storm ridge formed above units 3 and 4 and then was planed flat to form the planar beds of unit 5.

8.4 - ASPEN COVE (MM-16)

Description

All the sediments exposed at Aspen Cove and at Ladle Cove occur on one major terrace with its upper surface at 7m asl. Exposures examined were small (<1m thick), reflecting the overall sediment thickness. Although the sediment cover is thin, a diverse assemblage of material was observed along this northeastern coastline.

Section MM-16 is approximately 1m thick and is composed of 2 main units (Figure 8.4). The lowermost unit has a minimum thickness of 90cm. The lower contact was not observed. This unit is composed of interstratified granule and pebble gravel. Each bed fines or coarsens upwards and is moderately to well sorted. The smaller granules are generally subrounded spheres, whereas the larger clasts are rounded blades. The beds dip gently oceanward (towards the north) at an angle of 5°.

The lowermost unit is overlain abruptly by a thin (10cm thick) unit of larger pebbles interstratified with fine open-work granular gravels. The larger clasts are characteristically discs and blades. Sorting is moderate to good. This upper unit is rich in manganese and iron oxides, making it almost black in colour, and it is much more compact than the underlying units. This is

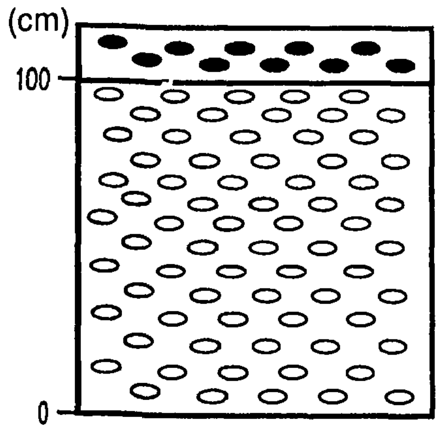


Figure 8.4: Vertical transect through sediments at Aspen Cove

characteristic of the upper layers of many of the beach deposits in the study area. The beds in this unit also dip 5° oceanward.

Clast types are predominantly the same in these 2 units. They consist of 48% white granite, 21% slate, 12% granodiorite, and 12% sandstone. Minor quantities of basalt are also present.

Interpretation

The Aspen Cove site represents a shoreline with moderate exposure to oncoming waves, which developed when sea-levels were higher than present. The low angles of di of the beds, and the moderate to good sorting are characteristic of deposition in the upper beach face area of a beach (cf. Komar, 1976; Massari and Parea, 1988).

Most beach deposits preserved in the sedimentary record are storm deposits (Massari and Parea, 1988). The larger clasts at the top of the section probably represent deposits higher in the beach face where sorting is better and clast sizes are larger (Bourgeois and Leithold, 1984). These, therefore represent either a second less intense storm event which deposited larger clasts further down the beach face, or the waning stages of a storm that deposited unit 1. Energy conditions decreased and thus the discs in this upper unit were not carried as far by the lower energy waves.

8.5 - LADLE POINT (MM-21)

Description

Recent removal of sediment for aggregate at Ladle Point has resulted in the exposure of site MM-21. This site lies within a topographic low between two buried rock headlands. This type of feature is common along the coastline at Ladle Point. The coastline is dominated by slate and shale. When sea-levels were higher, this area would have consisted of many small rocky headlands with small coves between them. The coves filled with sediment, whereas the rocky areas only have a few beach clasts thrown onto the top of them by storm waves. The sediment-filled coves are termed pocket beaches (Davis, 1978) and are commonly higher energy environments than open beaches. These deposits and headlands are now represented in the landscape as one main emerged terrace that occurs at 5.5m asl. A ^{14}C age of 2930±50 BP (GSC-5559) was obtained from basal peat overlying this marine terrace.

The section at Ladle Point is 1.5m high and is composed of one main unit. Beds of large cobbles and pebbles grade both normally and inversely into each other. These clasts are mostly discs and rollers and are well rounded. Granites account for 65% of the clast assemblage, and slate and shale account for 22% of the assemblage. Shale and slate clasts are predominantly discs, whereas granites are mostly rollers. Sand content is low (<5%) making this site different from the Aspen Cove site. Although the unit is well sorted, distinct beds cannot be seen.

Interpretation

This sediment at Ladle Point was deposited as a pocket beach in a small cove which faces northeast into the dominant modern wave directions (Figure 8.5). As there is no evidence suggesting a change in the predominant wind direction , the energy levels in this area were much higher than in the Aspen Cove area, allowing much larger well-sorted material to be deposited. Also, wave energy was directed into the small cove in which these deposits formed. All the sediments in this section were probably deposited during one storm event indicated by the lack of stratification in the unit. The ^{14}C age holds little significance in dating the formation of the beach deposits at this location, as regional relationships indicate that the beach formed considerably earlier than 2930±50 BP. Sea-level fell below 0m asl at approximately 10,000 BP (Shaw and Forbes, 1990a), and thus the beach must have formed prior to this, as there is no evidence in the area for subsequent rises in sea-level above the present marine limit. The date, however, does conform to the general time of initiation of the most recent phase of peat formation, between 4000 BP and 2500 BP (Davis, 1984).

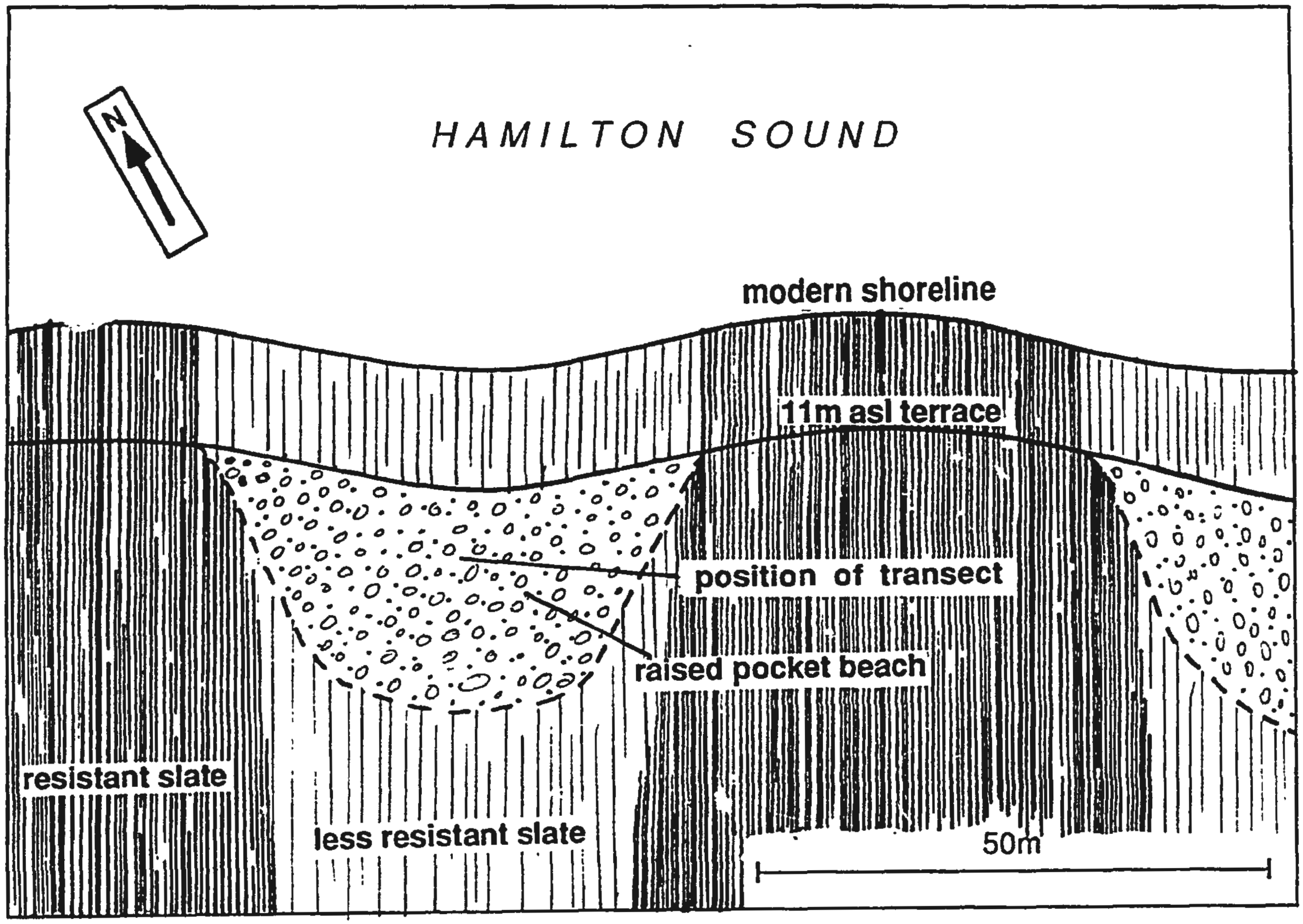


Figure 8.5: Sketch map showing the location of the transect at Ladle Point (MM-21)in relation to raised pocket beaches (raised marine gravels).

8.6 - RAGGED POINT (MM-24)

Description

The section at Ragged Point has been exposed through aggregate extraction. It consists of two main units (Figure 8.6). The lower unit is 30cm thick and is composed of well sorted open-work beds of pebbles and granules. These beds grade normally and inversely into each other. The smaller granules are predominantly spherical and pebbles are discs and blades. Granites account for 52% of the clast assemblage, 13% are granodiorite, and 21% are shales and slates. Minor quantities of diabase are also present. Granules are generally subrounded and pebbles are mostly subrounded to rounded. The beds dip relatively steeply towards the west with the upper strata dipping at 14°, and the lower strata dipping at 24°. The beds are stained with iron and manganese oxide making them dark reddish brown (Munsell: 5YR 3/3 mf).

Overlying the basal unit is a thick (110cm) unit of open-work pebbles and granules. The unit contains multiple fining- and coarsening-upwards sequences. Clasts finer than coarse sand are not present. As in the underlying unit, the smaller clasts are predominantly spherical and the larger clasts are discs and blades. The granules are generally subrounded and the pebbles are predominantly subrounded to rounded. These beds dip gently to the north at approximately 5°. The upper 30cm of this section is consolidated by iron and manganese oxide deposition. The entire unit is stained dark with manganese (Munsell: 7.5 YR 4/6 mf).

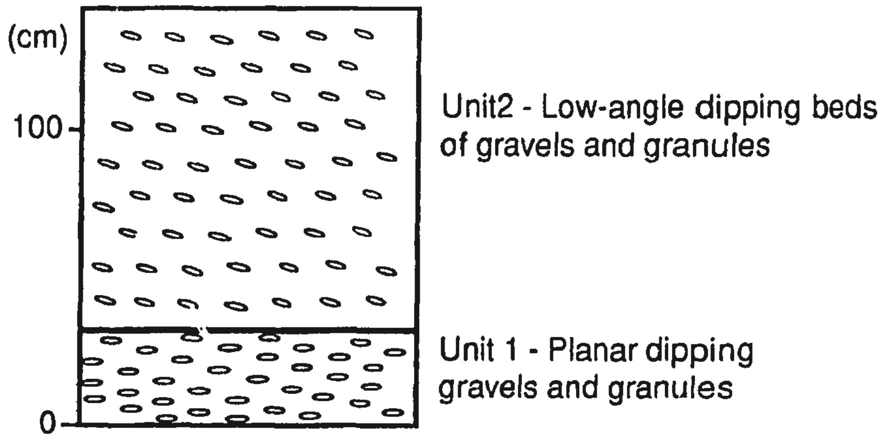


Figure 8.6: Transect at Ragged Point (MM-24)

Interpretation

The units at Ragged Point Are interpreted as beach deposits due to the presence of rounded to well rounded discs and blades (cf. Komar, 1976; Bourgeois and Leithold, 1984). They represent two separate geomorphic features. A relict beach ridge aligned SW-NE overlies another relict ridge aligned NW-SE (Figure 8.7). Both ridges can be observed clearly on aerial photographs as they rise higher than the marine terrace to the north and the fen area lying directly to the south. They are the only parts of the immediate coast that are vegetated with trees. The surrounding area is colonized by low-lying shrub and heath. Both units formed relatively high in the beach profile where sorting is generally better than lower in the beach profile (Massari and Parea, 1988).

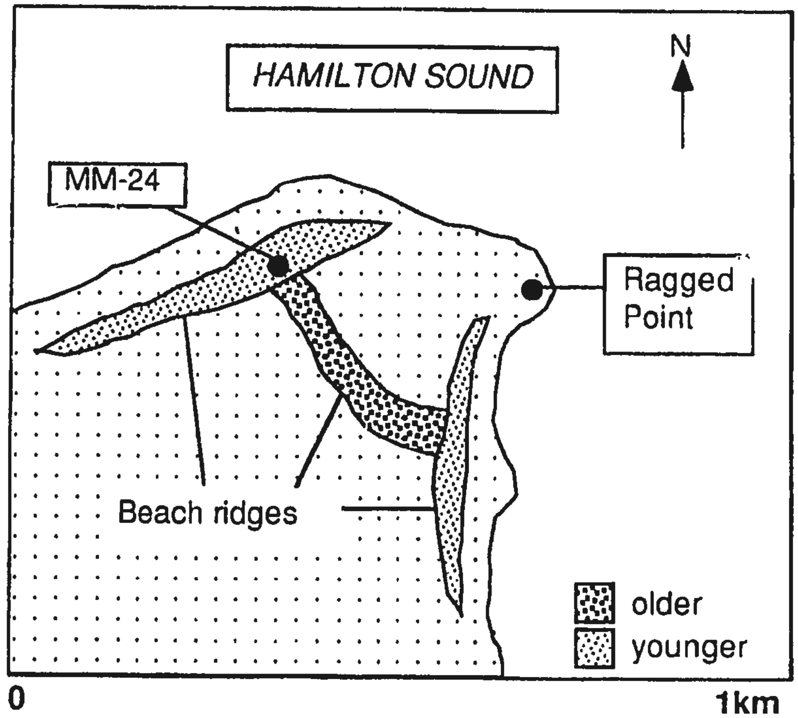


Figure 8.7: Sketch map showing the location of site MM-24 in relation to the relic beach ridges.

Unit 1 represents deposition on the backbeach area of the older relict beach ridge oriented NW-SE. This is indicated by beds dipping steeply to the west. As determined for site MM-44, sediments that dip away from the ocean tend to form in the backbeach area behind the berm crest (Davis, 1978; Bourgeois and Leithold, 1984).

Unit 2 represents deposition in the centre of the younger beach ridge, oriented SW-NE. This is indicated by the low angle of dip of the beds which would have formed on the crest of the ridge rather than the backshore or foreshore area.

8.7 - SUMMARY

Emerged beach deposits around the coast of the Carmanville region contain similar sedimentary characteristics. Despite this, they are developed in many different geomorphic settings. Most of the deposits represent falling sea-levels, or represent a short period of beach activity with previous sedimentary features being destroyed and the deposits reworked. The thick manganese oxide layer at Victoria Cove, at 10.5m asl, probably formed as a result of a emerged water table that remained at 10.5m asl over a period of time. This is related to the 8m sea-level stand. Relative sea-levels therefore probably did not fall at an even rate, as suggested by the models of Quinlan and Beaumont (1981) and Shaw and Forbes (1990a). Rather, there were short periods of stability.

CHAPTER 9
RELATIVE SEA-LEVEL CHANGES

9.1 - INTRODUCTION

Postglacial sea-level change has occurred as a direct result of isostatic recovery related to glacial unloading. There is extensive geomorphic and stratigraphic evidence for emerged sea-levels throughout the study area.

9.2 - LANDFORMS RELATING TO EMERGED SEA-LEVELS

The most common geomorphic features related to emerged sea-levels are emerged marine terraces. Most of the sediments described in Chapter 8 are terrace sediments. Forty-one elevated marine terraces and platforms were measured throughout the area (Figure 9.1). These occur commonly at 52m, 38m, 34m, 17m, 11m, 5m, and 2m asl and are most extensive at Wing's Point, Aspen Cove, and Noggin Hill. There are many less well developed terraces which occur between these elevations. At Aspen Cove and Wing's Point terraces occur between 17m and 34m asl, but at Noggin Hill there are no terraces at these elevations.

Lower elevation marine terraces (≤17m asl) are well developed, and are dominated by beach gravels or capped by blanket bog. At Aspen Cove and Ladle Cove, bog-covered terraces lie behind relict beach ridges which currently appear as vegetated ridges. The beach ridges are easy to detect both from the ground and on aerial photographs. They have been colonized by spruce trees in contrast to the surrounding area which is dominated by low-lying shrub and heath. Some ridges have overwash fans lying behind them that are now

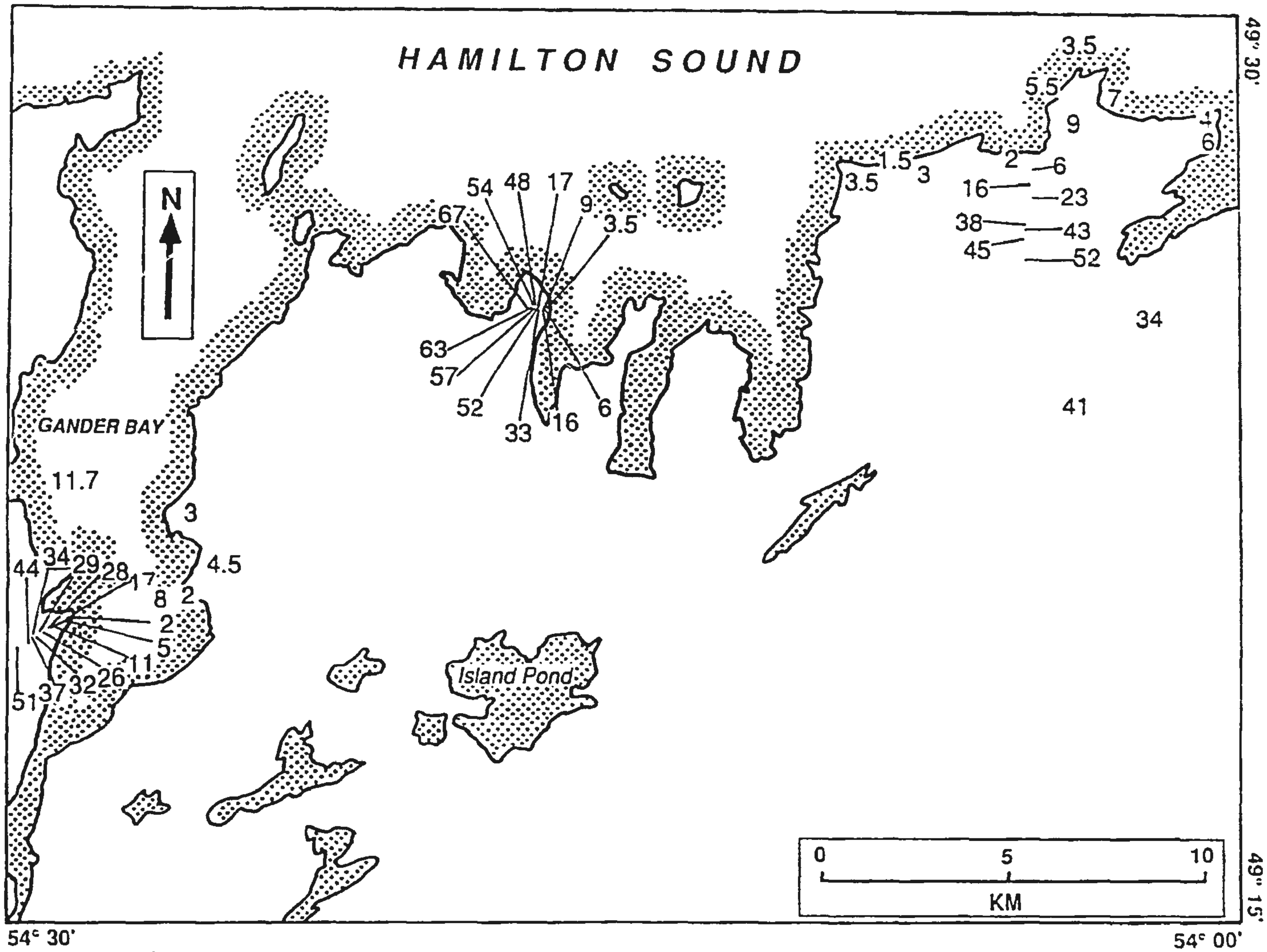


Figure 9.1: Altimeter measurements of all observed raised terraces (m asl).

vegetated with grasses. The terraces at Aspen Cove are up to 2km wide, and slope landward from elevations of approximately 6m asl to approximately 12m asl. The geomorphic relationships between the terraces and the relict beach ridges indicate that the ridges may relate to old lagoonal systems which subsequently drained when sea-level dropped. Silts and clays overlying bedrock in the lagoonal area would restrict drainage, resulting in bog formation in these areas.

Higher elevation marine terraces (above 34m asl) are less well developed. They tend to be veneered by beach gravels, or are associated with wave cut platforms. South of Ragged Harbour, the uppermost parts of glaciofluvial sand and gravel deposits (<1m below the surface) have been reworked resulting in the formation of gravel terraces.

9.3 - SEA-LEVEL HISTORY

A possible marine limit of 67m asl has been established, based on the recorded elevation of an eroded platform on the east side of Noggin Hill. Beach gravels occur up to elevations of 57m asl at Noggin Hill and Wing's Point and therefore the marine limit was at least 57m asl. This value is 17m higher than Grant's (1989) estimated limit for the area. The marine limit is (tentatively) dated to prior to 12,500 BP. Although there are no ^{14}C ages from the Carmanville region, Scott *et al.* (1991) obtained an age of 12,470±380 BP (TO-2305) from silt and clay deposits in the Springdale region. These deposits may have been the distal component of deltas at 75m asl, the marine limit in the Springdale area. In the Bay of Exploits region, high sea-levels can generally be dated to prior to 11,600 BP based on ^{14}C ages of 11,600±210 (GSC-2134)(Blake, 1983), and 11,500±110 BP (GSC-5527)(MacKenzie, *in preparation*). Shaw and Forbes (1990a) date the marine limit in Northeast Newfoundland to approximately 12,500 BP.

Correlation of elevations across the study area shows that there were at least five major stands of sea-level between the onset of deglaciation and present. These were at 52m, 38m, 34m, 17m, and 11m asl. Lower beaches at 5m, and 2m asl may represent either relative sea level changes or may indicate extreme storm events. Figure 9.2 shows the relationship between sea-level and ice limits when sea-level stood at 46m asl. Ice stood some distance from the coast, with the exception of the area surrounding Gander Bay.

Terraces between 17m and 34m asl occur at Aspen Cove and Wing's Point, but no terraces at this elevation occur at Noggin Hill. Marine terrace formation would be restricted if the area was occupied by ice while relative sea-level was between approximately 17m and 34m asl. It is possible that Younger Dryas ice may have affected this central part of the study area. Evidence for the Younger Dryas event is provided by till overlying marine gravels at 46m asl at site MM-30 (see Chapter 5). The terraces at 23m, 26m, 28m, and 29m asl at Aspen Cove and Wing's Point may therefore have formed between approximately 11,000 and 10,000 BP , the timing of the Younger Dryas cooling event (Anderson and Macpherson, 1994). Figure 9.3 shows probable ice limits when relative sea-level stood between 23m and 29m asl. Carmanville Arm was occupied by ice, probably of Younger Dryas origin.

Sea-level dropped less rapidly after the initiation of the Younger Dryas, as indicated by the well developed lower elevation terraces. The terrace at 11m asl is up to 1.5 kilometres wide. Figure 9.4 shows the relative sea-level stand at

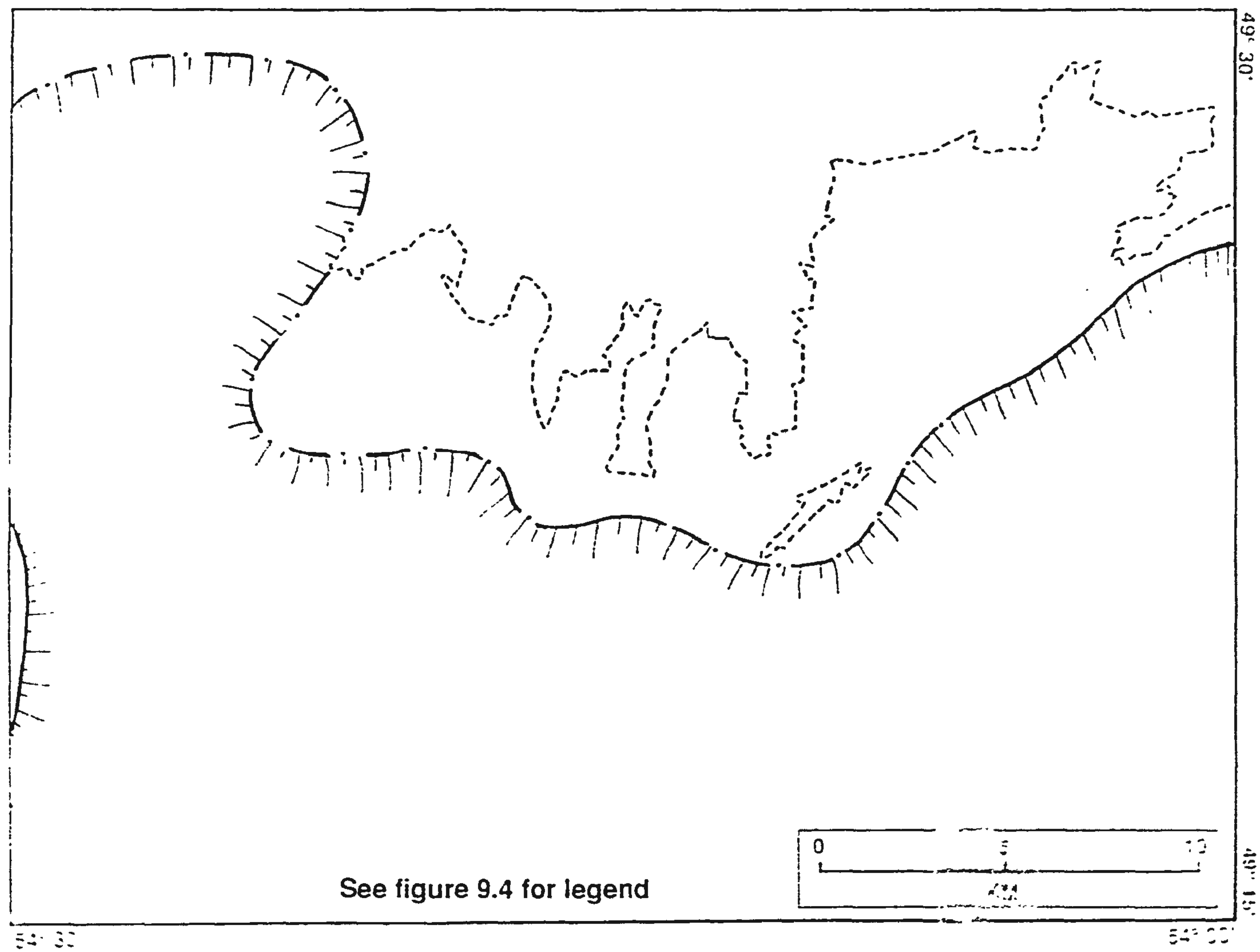


Figure 9.2: Probable ice limits when sea level stood at 46m asl. The ocean stood high against the retreating ice margins.

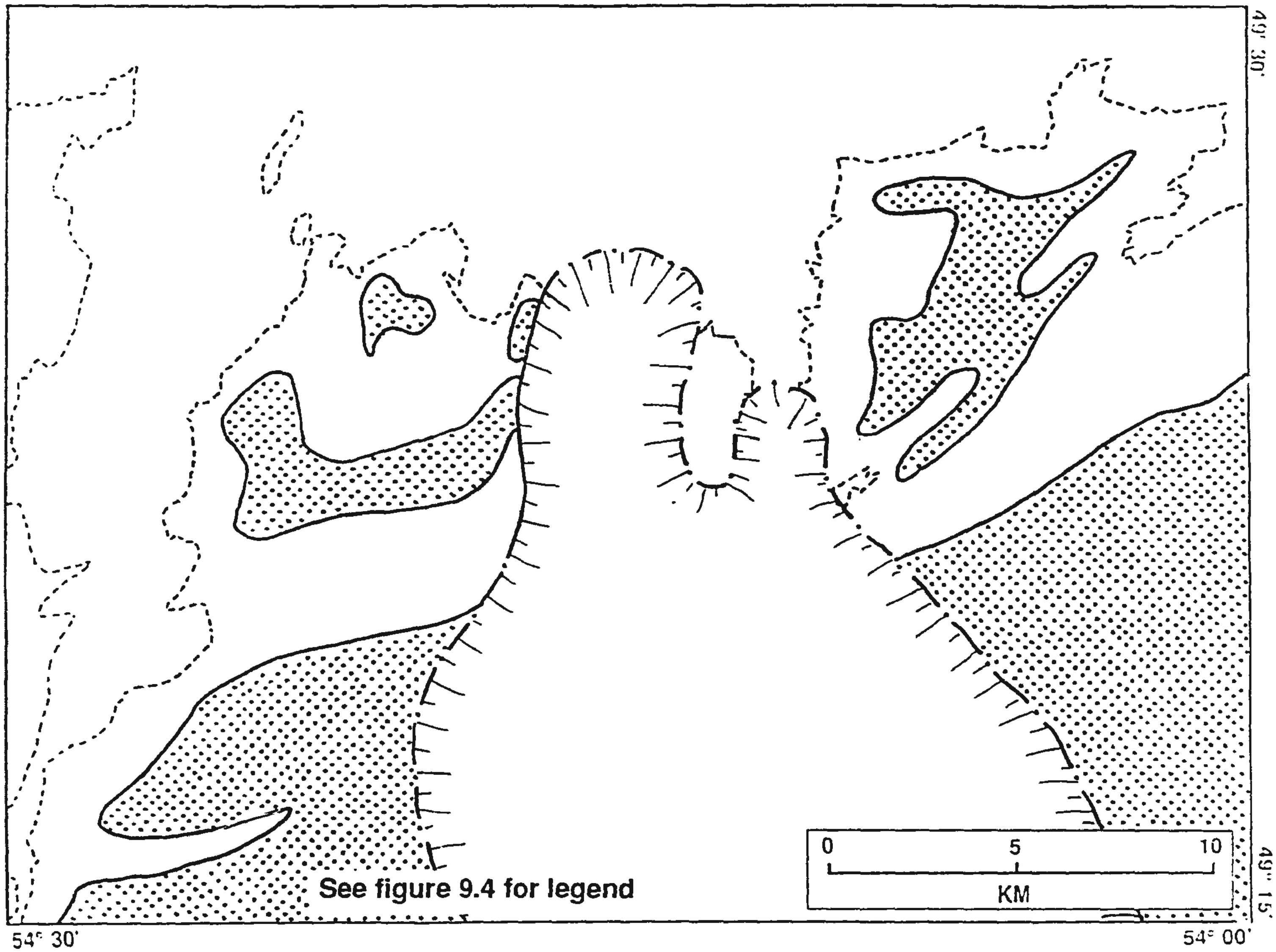


Figure 9.3: Probable ice limits when relative sea level stood between 23m and 29m asl.

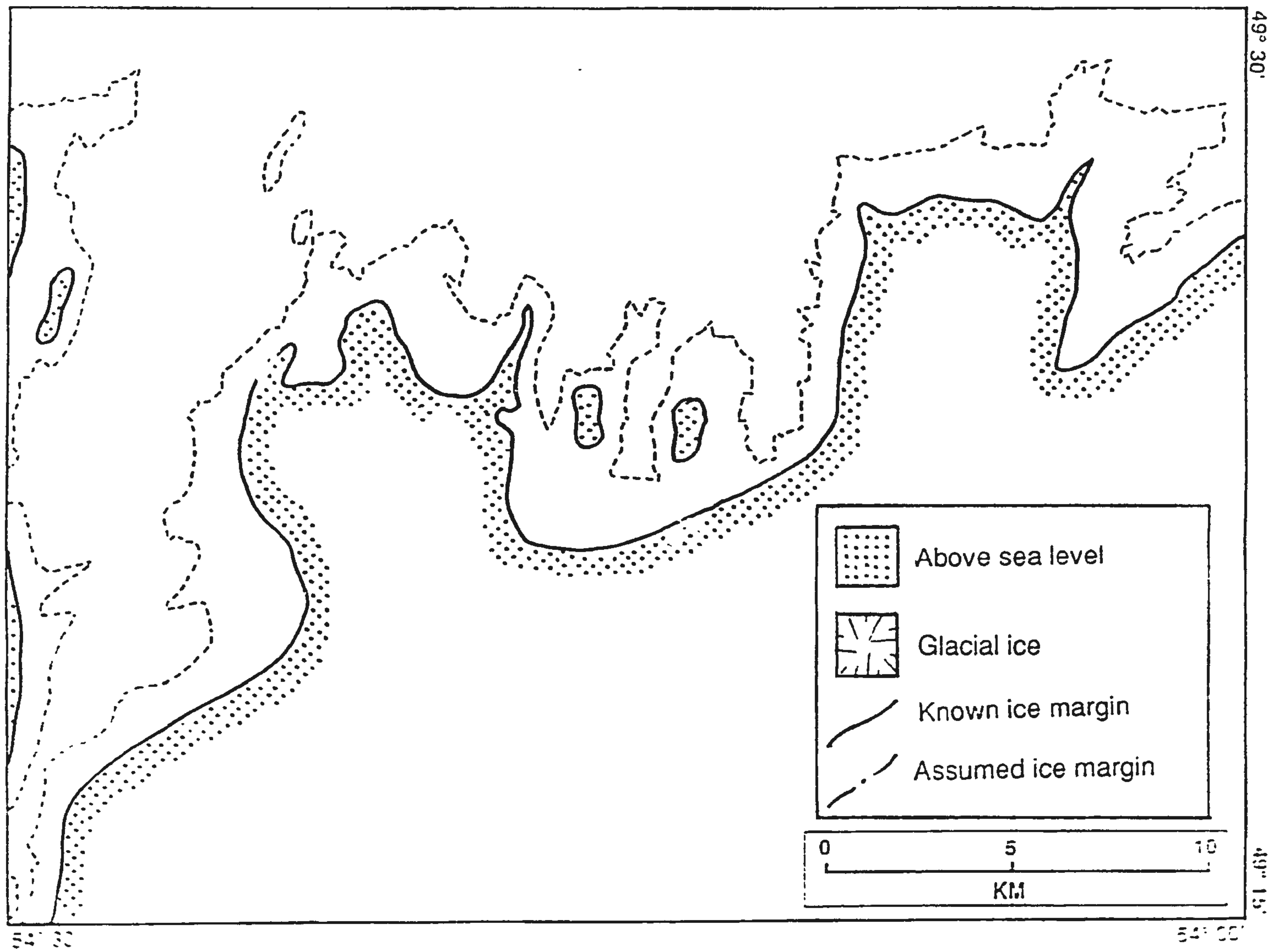


Figure 9.4: Relative sea level stand at 11m asl.

11m asl. Sea-level dropped below present levels at approximately 10,000 BP (Shaw and Forbes, 1990a), to a minimum level of ~-17m asl some time prior to 8600 BP (Shaw *et al.*, *in review*). Sea-levels rose to approximately -4m asl at 5,500 BP and reached -0.7m asl by 3,000 BP (Shaw and Forbes, 1990). Shaw and Forbes suggest that they have risen at a slow but constant rate since 3000 BP until present.

Undercutting and erosion at several locations along the coast, such as Aspen Cove, show that sea-level may be rising and storm activity may be increasing. Rising sea-level is not an immediate threat to the communities in the area, as most are built on terraces several metres above present sea-level.

CHAPTER 10
CONCLUSIONS

10.1 - QUATERNARY HISTORY

There is no firm evidence in the Carmanville area for glaciation prior to the most recent, Late Wisconsinan glaciation, although multiple glaciations prior to this are likely to have occurred. These earlier events may have modified the landscape substantially, forming valleys and large meltwater channels along the main structural trends in the area. These features were subsequently re-used during the Late Wisconsinan glacial phase. Although no numerical ages were obtained from related sediments in the area, the age of all glacigenic sediments evident is assumed to be Late Wisconsinan in the absence of evidence to the contrary. Striation sites are not deeply weathered and there is no differential weathering on bedrock surfaces. Tills also are not deeply weathered. All events discussed in this chapter are listed in Table 10.1.

The earliest event cbserved in the study area involved easterly ice-flow with a mean trend of 075° (Figure 6.4; Chapter 6). This flow affected the entire study area, as indicated by striation orientations. Although one clast fabric on the east coast of Gander Bay indicated easterly ice-flow, there are no large scale landforms associated with this flow, probably because ıne ice was moving transverse to the bedrock strike in the area, and thus across the main topographic trends. It may have, however, aided in sculpting some of the roches moutonnées now oriented NNE. This ice-flow event was probably the result of coalescent flows from the Northern Peninsula and the Central Uplands, and is associated with the maximum extent of ice prior to the onset of coastal

Time (BP)	Event	Evidence/Source
0	Cool conditions Rising sea-levels	Macpherson (1985) Shaw and Forbes (1990a) Shaw *et al. (in review)*
~5000	Climatic optimum followed by climatic deterioration Sea-level stand at -4m asl	Macpherson (1985) Shaw and Forbes (1990a)
8600	Sea-level minimum of ~-17m asl Climatic warming	Shaw *et al. (in review)* Macpherson (1985)
~10,000	Sea-level stands at 17m, 11m, 5m, 2m, and 0m asl.	Terraces at Aspen Cove, Noggin Hill, and Victoria Cove
	Ice-flow 4 (Younger Dryas?)	Till overlying beach deposits at ~46m asl at MM-30
~11,500	Sea-level stand at ~46m asl	Beach deposits at 46m asl at Ragged Harbour and at site MM-30
	Glacial retreat and short stillstand	Formation of Ragged Harbour Moraine
	Falling sea-levels with a stand at 52m asl	Terrace at Aspen Cove, Noggin Hill and Victoria Cove
~12,000	Hamilton Sound ice-free	Shaw *et al. (in review)*
	Marine limit of 57m asl (67m asl?)	Terraces at Noggin Hill and Aspen Cove (eroded platform at Noggin Hill)
	Ice-flow 3 (NW ice-flow)	Striations
	Glacial ablation resulting in multiple ice caps	Grant (1989)
	Ice-flow 2 (NE ice-flow)	Striations, clast fabric, glacial thrusting at MM-29
~13,500	Glacier retreat	
Late Wisconsinan	Onset of coastal deglaciation	Anderson and Macpherson (1994)
	Ice-flow 1 (E ice-flow) Regional sea-levels rising	Striations, clast fabrics Shaw and Forbes (1990a)

Table 10.1: Chronology of Quaternary events in the Carmanville (NTS 2E/8) region in Northeast Newfoundland.

deglaciation (cf. St. Croix and Taylor, 1990). This ice-flow was also detected by Liverman *et al.* (1991) and Liverman and Scott (1990) in Baie Verte, and Batterson and Vatcher (1991) in the Gander region. The time of initiation of this event is uncertain. Regional sea-levels were also rising during this time period, beginning at approximately 18,000 BP (Shaw and Forbes, 1990a).

During subsequent retreat, the glaciers from the central area of the Island may have retreated to the Island Pond area. Ice began to readvance as indicated by the proglacial thrusting related to northeastward ice movement noted at site MM-29 (Island Pond). Northern Peninsula ice did not obstruct the advancing ice, and thus it was able to flow freely to the coast in a northeasterly direction. This is detected in the striation record, and is shown by the trends of roches moutonnées, lineated till, and crag-and-tails (Figure 6.5; Chapter 6). Roches moutonnées throughout the area appear to be associated with this event only. This alignment may, however, be reflective of the bedrock trends rather than of ice-flow conditions. Drumlins and rogen moraines also have north-northeasterly trends but as indicated in Chapter 6, these may represent meltwater flow moving in a northeasterly direction and thus may not be reliable indicators of ice-flow direction. This ice-flow had the most erosive effect on the landscape, and is the most widely recognized flow in Northeast Newfoundland.

Well preserved striations show that this northeastward ice-flow affected the whole study area. They are also consistent in their orientation with trends occurring between 021° and 045° This pattern conforms with other striation evidence across northeast Newfoundland (Baird, 1950; Hayes, 1951; Jenness, 1960; Lundqvist, 1965; Hornbrook *et al.*, 1975; Vanderveer and Sparkes, 1980; Brookes, 1982; Vanderveer, 1983, 1987; Vanderveer and Taylor, 1987; Taylor and St. Croix, 1990; Batterson *et al.*, 1991), suggesting that there may have been an ice divide between the Red Indian Lake area and the Bonavista Peninsula (cf. Jenness, 1960; Grant, 1974; St. Croix and Taylor, 1991).

Meltwater channels associated with this ice-flow event may in part represent successive re-occupation of valleys created by earlier glacial and fluvial events. The trend of the underlying bedrock would influence the flow of meltwater (as it would ice-flow) and consequently pre-existing channels could have been occupied at any time.

As deglaciation continued, ice on the Island split into a series of ice caps (Grant, 1989). This resulted in the third glacial event which was marked by northwestward flow (Figure 6.6; Chapter 6). The major erosional forms associated with this event are striations, which have consistent orientations of approximately 345°. This ice advanced at least as far as the centre of Hamilton Sound, south of Eastern Indian Island (2E/9 map area), an area marked by an extensive sand deposit (Shaw *et al.*, 1990). This may be a moraine similar to the Ragged Harbour moraine. The marine limit occurred just prior to, or contemporaneously with retreat of the ice. The 57m asl terrace is preserved at Noggin Hill and Aspen Cove, showing that no ice occupied these areas while sea-levels were high.

As deglaciation continued, the ice receded rapidly with most of Hamilton Sound being ice free by approximately 12,000 BP (Shaw *et al.*, *in review*). South of Ragged Harbour, the Ragged Harbour Moraine marks a temporary frontal position. Large quantities of meltwater were associated with this ice-flow event, indicated by the presence of drumlins and rogen moraines behind the subaqueous deposits.

The Ragged Harbour Moraine is associated with sea-levels higher than 46m asl, as indicated by a cap of marine gravels overlying the subaqueous deposits. As sea-level is known to have dropped rapidly in the area (Grant, 1989; Shaw and Forbes, 1991; Shaw *et al.*, *in review*) to below 0m asl by 10,000 BP, there was a limited time period during which these deposits could have formed. Maximum and high relative sea-levels are dated to prior to 12,470±470 BP on the basis of radiocarbon ages of 11,600 ±210 (GSC-2134) (Blake, 1983), and 11,500 ±110 BP (GSC-5527)(MacKenzie, *in preparation*), from the Bay of Exploits, and 12,470 ±380 BP (TO-2305) from the Springdale area (Scott *et al.*, 1991). The sea-level stand in which the Ragged Harbour Complex was deposited may, therefore, predate 12,470±380 BP, but this is unlikely as ice did not retreat to this coastline until approximately 12,000 BP (Shaw, *in review*). This suggests that these deposits formed at approximately 12,000 BP and are not necessary marine limit deposits, but deposits associated with high relative sea-levels. MacKenzie (*in preparation*) noted that in the Bay of Exploits region, any marine deposits above 11m asl are older than 11,400 BP.

These sand and gravels form a ridge almost 15km long parallel to the modern shoreline, and thus delineate a stillstand of the northwestward-moving ice sometime between 12,000 BP and 11,400 BP. Meltwater escaping from the front of the ice during the stagnation stage formed a series of fans and subaqueous channels which joined to form the ridge (Figure 10.1). The extremely rapid sedimentation suggests that this feature does not represent a lengthy still-stand.

Similar features described from other regions usually represent the distal parts of a glaciofluvial assemblage, with eskers feeding the complex (Fyfe,

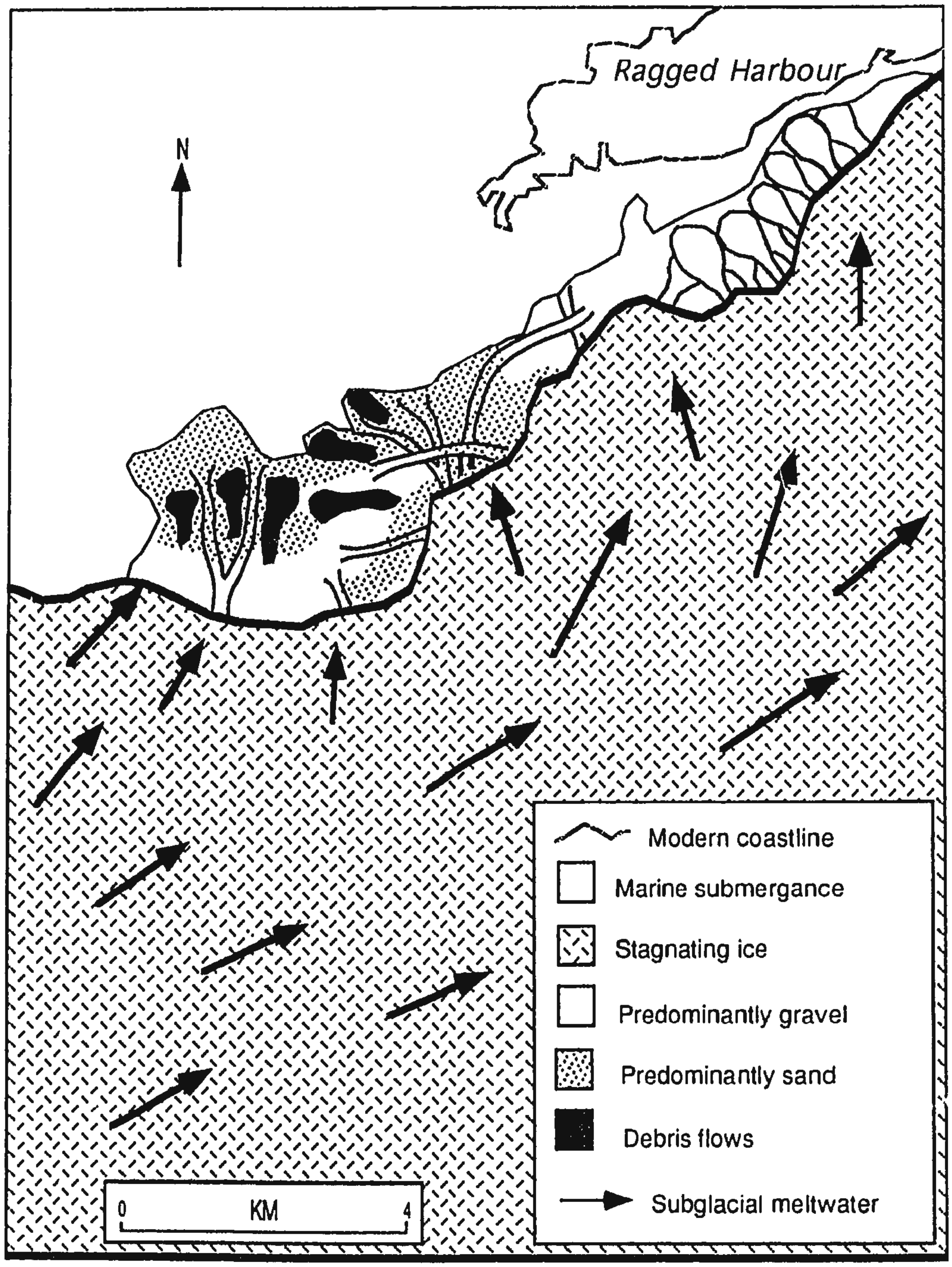


Figure 10.1: Model of meltwater flow and the formation of the Ragged Harbour Moraine during ice stagnation.

1990). At Ragged Harbour, eskers are not present. Much of the subglacial meltwater from the area more than 2km from the ice front may have been channeled away from the moraine area (Figure 10.1). The presence of rogen moraines directly to the south of the Ragged Harbour Complex may indicate high velocity subglacial meltwater drainage up-ice from the moraine. If this is so, the moraines indicate northeastward flow of subglacial water, bypassing the outwash area and discharging into Hamilton Sound to the northeast. Although these rogen moraines have a northeastward trend they are most probably related to this northwestward ice-flow because they are well preserved and do not appear to have been overridden or eroded by any subsequent ice-flow. During stagnation of the ice, the local geomorphology would constrain water to move northeasterly, parallel to the local slope following the bedrock trends and the large meltwater channels, rather than following the surface profile of the waning ice mass.

These first three ice-flows represent a deglacial sequence related to fluctuating ice margins. Ice did not completely disappear from the area during this time.

Following deposition of the sand and gravel deposits, retreat of the ice mass continued. One final episode of northwestward moving ice and till deposition occurred. It cannot be determined in numerical chronological terms whether this was a readvance related to a pulse of the main Late Wisconsinan ice, or whether ice had retreated from the area and readvanced again during the Younger Dryas cooling event. It is difficult to distinguish between the two northwestward events in the striation record, although they are separated stratigraphically. The first flow extended probably as far north as Eastern Indian

Island and occurred while sea-levels were high. Deposition of the Ragged Harbour Moraine is associated with this event.

The second northwestward ice-flow is marked by till overlying beach deposits at Carmanville South 2 (MM-30). The beach deposits occur at elevations of approximately 45m asl, similar to the highest elevation of the Ragged Harbour Complex, and thus formed at approximately 12,000 BP or just after that date. Till deposition took place subsequent to this. The Younger Dryas climatic episode occurred between 11,000 BP and 10,000 BP on the Island (Anderson and Macpherson, 1994), and it is thus probable that this final phase of glacial activity relates to this time period. In summary, the first northwestward ice-flow occurred before beach formation at 46m asl between 12,000 BP and 11,500 BP, and the second northwestward event occurred subsequent to beach formation at the same elevation. It, therefore, occurred after 11,500 BP.

The Younger Dryas ice did not extend as far north as Ragged Harbour, as the deposits there have not been deformed due to glacial overriding. Consequently, the Aspen Cove and Ladle Cove area was unaffected by this event. Ice extended northwestward to at least as far north as Carmanville, indicated by the till overlying beach gravels at 43m asl at MM-30, and also indicated by the lack of terrace development between 17m and 34m asl on the eastern flank of Noggin Hill. As these terraces are well developed elsewhere in the study area, ice must have been present, preventing terrace development. Ice did not extend to the western side of Gander Bay, as indicated by the presence of well preserved emerged beaches at elevations greater than or equal to 11m asl. MacKenzie (*in preparation*) noted that all stands higher than 11m asl in the Bay of Exploits region are older than 11,400 BP. These have not been subsequently overridden by glacial ice. Also, the development of the marine terraces appears to have been unhindered during the period from 12,500 to 10,000 BP when sea-levels were above present (cf. Shaw and Forbes, 1990a). This ice may have originated on the higher area of land approximately 25km south of Musgrave Harbour in the vicinity of Ten Mile Pond. Grant (1989, p.414) suggested a small remnant ice centre in this area at approximately 12,000 BP. Ice did not completely ablate prior to the Younger Dryas and the remnant ice cap was easily reactivated during the Younger Dryas.

There is evidence for Younger Dryas periglacial activity from the surrounding areas in the form of ice wedge casts (Eyles, 1977; Liverman *et al.*, 1991; Scott, 1993; MacKenzie, *in preparation*), but little evidence was detected in the Carmanville region. Some gravel units, especially in emerged beaches, show evidence of frost heave, with clasts now oriented vertically. Frost disturbance usually only affects the upper 20cm or less of the sections. This may represent modern or late Holocene frost activity rather than Younger Dryas activity. If the last ice-flow in the area is correlative with the Younger Dryas, then much of the area was covered by ice during this time period, preventing the formation of ice wedge casts. However, the most likely deposits to contain ice wedge casts would be the sand and gravel deposits at Ragged Harbour. Ice wedge casts were not observed. This area was not covered by ice during the final glacial episode as the deposits have not been deformed by glacial overriding. Much of the upper portions of the sections have been excavated for aggregate purposes, and it may be that exposure of these parts of the sections is not extensive enough to view ice wedge casts. Other possible areas where these may have occurred are in emerged beach gravels. Exposure of these gravels, however, is not extensive.

There is no evidence of further pauses in retreat or additional glacial activity in the Carmanville area. Retreat occurred at some time after 11,300 BP (the onset of the Younger Dryas), and any numerical chronology of events subsequent to this is speculative. Using the criteria of MacKenzie (*in preparation*), the sea-level stands of 11m asl and below occurred after 11,400 BP, and thus they are contemporary with the Younger Dryas period or later.

Sea-level fell below present at approximately 10,000 BP (Shaw and Forbes, 1990a). Thus all emerged marine features in the region formed between 12,000 and 10,000 BP. It is likely that all the emerged marine beaches above 11m asl formed prior to 11,400 BP (cf. MacKenzie, *in preparation*). Sea-levels continued to fall to a minimum of ~-17m asl some time prior to 8,600 BP, marked by a low elevation marine terrace in Hamilton Sound (Shaw *et al., in review*). The pattern of sea-level change indicates initial isostatic depression, followed by falling levels as isostatic recovery progressed. In the Late Holocene, subsidence due to forebulge migration produced rising sea-levels, a trend which is presently continuing. The pattern is indicative of Quinlan and Beaumont's (1981) zone B, and indicates that the progression of isostatic recovery in the area was controlled by the presence of the Laurentide ice sheet in Labrador during the Late Wisconsinan (cf. Clark *et al.*, 1978).

From 9,500 to 8,300 BP, the Island experienced climatic warming due to the northward movement of the oceanic polar front (Ruddiman and Glover, 1975; Macpherson, 1985). This continued from 8,300 to 5,400 BP with conditions becoming steadily drier, marked by an increase in forest fire frequency (Macpherson, 1985). Climatic optimum was reached between 5,400 and 3,000 BP (Macpherson, 1985).

Between 9,500 and 3,000 BP fluvial and colluvial activity was taking place. There no evidence that the section of Gander River present in the study area was actively downcutting. The upper portions of Ragged Harbour River near New Pond have eroded down in the surrounding sediment by at least 6m.

A period of peat formation began at approximately 3000 BP. Pedogenesis had begun prior to this time as favorable warm and dry climatic conditions prevailed (Macpherson, 1981, 1985). Basal peat at 5.5m asl at Ladle Point yielded an age of 2930±50BP (GSC-5559). This date conforms to the general time of initiation of the most recent phase of peat formation, between 4000 BP and 2500 BP (Davis, 1984). The latest phase of peat formation is indicative of climatic deterioration which started between 3000 and 2500 BP (Macpherson, 1981, 1985). The peat also indicates that sea-levels were below 5.5m asl at 2,930 BP.

A second ^{14}C age from wood (*Larix* sp.) in peat at 41m asl yielded a date of 1040±50 BP (GSC-5558) indicating that during this cooling period woody vegetation prevailed. The wood may have been contaminated by modern rootlets and the date consequently may be unreliable.

In some areas there appears to be evidence of continuing sea-level rise. Undercutting and erosion at several locations, such as Aspen Cove and Ladle Cove, indicate that sea-levels are rising in conjunction with increased storm activity. Rising sea-level is not a major threat to the communities in the area, as most are built on terraces several meters above sea-level and land at the immediate coast is undeveloped.

Soil erosion in the Carmanville area at present is not a major problem. There is evidence, however, that all terrain vehicle use in the area is increasing. As well as old woods roads, ATV-drivers use large flat fen areas at the coast and in the interior of the study area. Increased use of these vehicles in these areas could lead to severe soil erosion.

10.2 - REGIONAL IMPLICATIONS

Reconstructing the Quaternary history of the Carmanville area is important as its adds a vital piece of information to the overall understanding of the Quaternary history of the Island. Small scale studies, such as the one undertaken in the Carmanville region, are imperative if a full understanding of the Quaternary history of the Island is to be gained as they show that locally, events may occur in some areas, but not in others. Thus the glaciation of the Island is not only composed of many large-scale events, but also many regional events.

The Quaternary history of the Carmanville region is comparable to that of other areas in northeastern and eastern Newfoundland. To the west, in the Bay of Exploits, MacKenzie (*in preparation*) noted two ice-flows: an easterly and a northerly flow. These are comparable to the first two flows in the Carmanville area. St. Croix and Taylor (1991) noted all three flows in the Notre Dame Bay area. They also noted a final local ice-flow, which was centred north of Grand Falls. Tentatively, they correlated this with the Younger Dryas cooling event. There is no evidence contradicting the existence of several local ice centres across Newfoundland which were reactivated during the Younger Dryas. One such area is in the region of Ten Mile Pond 30km south of Musgrave Harbour. Ice flowed northwestwards from this centre and produced the tills and striations observed in the Carmanville area.

Tills in the Carmanville region are atypical of tills on the Island of Newfoundland (see Table 10.2). They are generally much finer in texture, and are matrix-supported with up to 88% silt and clay dominating the matrix. The texture is primarily the result of the local bedrock source, dominantly sandstones, basalt, slate, shale, and siltstones. These rocks tend to have their terminal grades concentrated in the fine sand, silt, and clay sized fractions. Consequently, in any area the tills will vary according to the bedrock geology and are thus a local phenomenon dependent on the bedrock source.

Most glaciofluvial deposits on the Island of Newfoundland have been interpreted as eskers, deltas, and kames. Only at two other locations have deposits interpreted as subaqueous outwash been located. In the Bay of Exploits, MacKenzie (*in preparation*) interpreted a large sand and gravel body (15km x 4km) as ice-proximal subaqueous outwash. This has a maximum elevation of 50m asl. In St George's Bay, Western Newfoundland, Shaw and Forbes (1990b) detected deposits that they interpreted as subaqueous outwash. In the Bay of Exploits the deposits must have formed in waters of similar depths as in the Carmanville region based on the elevation of the deposits. Using the criteria of MacKenzie (*in preparation*), this deposit must predate 11,400 BP as all sea-level stands higher than 11m asl in the Bay of Exploits area predate this time.

REGION	TILL COMPOSITION	REFERENCE
Northern Peninsula	sand/abundant clasts	Proudfoot and St. Croix (1987)
Springdale	≤70% clasts	Liverman and Scott (1990)
South coast	sand/boulders	Sparkes (1987)
Deer Lake	sand/cobbles/boulders	Batterson and McGrath (1993)
Avalon Peninsula	sand/≤60% clasts	Catto (1992)

TABLE 10.2: Examples of typical till compositions on the Island of Newfoundland.

Subaqueous outwash deposits are associated with large quantities of meltwater. In the Bay of Exploits and the Carmanville area, these deposits formed at approximately the same time as indicated by their elevation. They must, therefore, represent a regional event of rapid glacier melting which was most likely climatically induced. Anderson and Macpherson (1994) noted a rapid climatic warming between 13,500 BP and 11,000 BP in the pollen record. Most of the deltas around the coast are also related to this time period. It is rather curious, however, that no other subaqueous glaciofluvial deposits have been identified around the coasts of the island that would have formed as sea-levels were high against the ice margins. This may be due to a lack of investigation, or to misinterpretation.

The high elevation of the marine deposits (≤57m asl) and the areal extent of many of the deposits (sometimes up to 10km^2) is rather unusual. This may, in part, be related to the fact that this particular coastline was much more exposed in the past when sea-levels were higher and also because ice had retreated on to the land. Many of the areas in northeast Newfoundland where sea-level studies have been undertaken are in sheltered bays such as in Green Bay and Halls Bay on the Baie Verte Peninsula, the Bay of Exploits, and in Bonavista Bay. These were still occupied by ice when sea-levels were at their highest elevations above present (e.g. Scott *et al.*, 1991; Cumming *et al.*, 1992; MacKenzie, *in preparation*).

The work in the Carmanville region can be regarded as a standard for any terrestrial palaeoenvironmental project to be undertaken in the future on the Island of Newfoundland. When combined with other similar studies, this work adds to the array of palaeoenvironmental information available for the Island, and ultimately will lead to a better understanding of the Quaternary history of the Island.

Importantly, however, the Carmanville region is one of only three areas on the Island that contains deposits that have been recognized as proglacial subaqueous outwash. For reasons outlined in the previous section, these deposits should be relatively common around the coasts of Newfoundland where sea-levels were high during deglaciation. Such deposits, however, have not been recognized. The descriptions and interpretations in this study, along with works from other regions outside Newfoundland, provide adequate criteria for recognizing these deposits.

10.3 - ECONOMIC APPLICATIONS

A thorough understanding of the ice-flow history needs to be established if a mineral exploration programme is set up in the Carmanville area, as it is the single most important factor influencing sediment dispersal. Successful drift prospecting would require recognition that the last northwesterly ice-flow was the dominant dispersal agent in the area, although the earlier flows cannot be disregarded as sediment dispersers.

The possibility of finding gold placers both on and offshore in this area has been considered (Jenner and Shaw, 1992). There is, however, a lack of terrestrial gold occurrences (Andrews, 1980) and sea-levels dropped rapidly following deglaciation. Research in actively exploited placer regions indicates that the formation of economically viable placer deposits requires tens of thousands of years (Debicki, 1983). It is thus unlikely that placer deposits occur in economically viable quantities in the area.

When looking for aggregate sources, the distribution and types of sediments in an area need to be determined. Most sources of aggregate are glaciofluvial deposits that contain large quantities of sand, granules, pebbles, and cobbles. A second source is beach deposits, commonly composed of open-work granules, pebbles, and cobbles. Most of the aggregate sources in the Carmanville region are presently being exploited, or the potential for exploitation has been assessed (Kirby and Ricketts, 1983). One possible new source, previously undetected, lies on the eastern shore of Noggin Hill where there are relatively thick sequences of emerged beach gravels up to elevations of approximately 35m asl, and veneers of beach gravels up to 57m asl. In the Ladle Cove area, residents are removing large quantities of cobbles and pebbles from the local beaches for construction and landscaping purposes. This could lead to increased coastal erosion. A knowledge of terrestrial aggregate sources is, therefore, imperative.

Most sediment in the area is relatively stable, and not subject to colluviation except along the steepest slopes (> 35°). Consequently, most areas are suitable for construction.

More importantly, most of the substrates in the Carmanville area are unsuitable for solid waste disposal. There are six municipal dumps in the area and several areas of unregulated dumping. Almost all of these are situated on higher ground and in some cases within a radius of 1km from settlements. They are placed on tills, and in one case on emerged beach deposits (Victoria Covo). Surface water percolating through solid waste can leach harmful chemicals which are washed downslope from the site in surface and subsurface runoff. No site is totally secure, but more careful placing of these waste sites could reduce the chances of severe water contamination. More ideal locations could include disused gravel pits at lower elevations. If these were lined with an impermeable substance such as clay, this would aid in reducing downward percolation into the water table.

10.4 - SUGGESTIONS FOR FUTURE WORK

Detailed regional studies have to be undertaken across Newfoundland if a thorough understanding of the glacial history and sea-level history for the whole Island is to be established. The history of ice movements cannot be assumed to be similar to nearby areas as ice centres and dispersal directions will change from area to area. For instance, ice centres and flow directions are different in the Carmanville area than in the Baie Verte, Bay of Exploits, and Gander areas. The major flow directions from large ice caps will be similar, but flows from small ice caps may vary considerably. In addition, the study in the Carmanville area showed that sea-level was at least 17m higher than anticipated for the area.

Although Eyles (1977) suggested ice wedge casts in the Birchy Bay area, directly to the west of the Carmanville region, may have been the resu'' of Younger Dryas activity, glacial activity during this time period was previously undetected in the area. Areas which should be targeted for future work are the Musgrave Harbour area, Wesleyville, and Gambo. Very little is known about the ice-flow directions or sequence of ice-flows in these areas. Detailed studies here should lead to a relatively complete understanding of the glacial history of Northeast Newfoundland. Also, these areas could all potentially be affected by Younger Dryas ice dispersing from a small ice cap in the Highland area south of Musgrave Harbour.

It is necessary to undertake detailed sedimentary studies on glaciofluvial deposits on the Island, especially those that occur below the marine limit. There are many deltas around the coasts of Newfoundland dated to approximately the same time period as the subaqueous outwash in the Carmanville region (Blake, 1983; Scott *et al.*, 1991; MacKenzie, *in preparation*). They, therefore, must represent a regional event of increased meltwater activity. There may be more deposits around the coast of Newfoundland that represent this event. Deltas are easy to recognize by their geomorphology, whereas subaqueous outwash has no distinctive geomorphic characteristics. For this reason many deposits relating to this event probably have not been recognized, hence the need for detailed sedimentary work. The descriptions in this study can be used as an analogue for sediments found elsewhere.

A sea-level curve has been developed for this area (Shaw and Forbes, 1990a), although the marine limit used is too low. To improve the curve more ^{14}C age determinations are needed. There are few sea-level curves developed for the Island, and the only way to develop the sea-level history is to undertake several regional surveys.

Ultimately, the information in this study has to be combined with information from several similar studies. This should aid in a better understanding of the overall Quaternary history of the Island of Newfoundland.

REFERENCES

Aber, J.S.
1979: The character of glaciotectonism. Geologie en Mijnbouw, Volume 64, pages 389-395

Agriculture Canada
1976: Manual on soil sampling and methods of analysis. Soil Research Institute, Ottawa, Ontario.

Allen, J.R.L.
1968: The diffusion of grains in the lee of ripples, dunes and sand deltas. Journal of Sedimentary Petrology, Volume 38, pages 621-632.

Allen, J.R.L.
1970: The avalanching of granular solids on dune and similar slopes. Journal of Geology, Volume 78, pages 326-351.

Allen, J.R.L.
1971: Transverse erosional marks of mud and rock: their physical basis and geological significance. Sedimentary Geology, Volume 5, pages 167-385.

Allen, J.R.L.
1982: Developments in Sedimentology. 2 volumes. Elsevier, Amsterdam. 1256 pages.

Allen, J.R.L.
1983: Gravel overpassing on humpback bars supplied with mixed sediment: examples from the Lower Old Red Sandstone, southern Britain. Sedimentology, Volume 30, pages 285-294.

American Society for Testing Materials
1964: Procedure for testing soils. American Society for Testing Materials, Philadelphia, Standard D 422-63, pages 95-106.

Anderson, S.P.
1988: The upfreezing process: experiments with a single clast. Geological Society of America Bulletin, Volume 100, pages 609-621.

Anderson, T.W., and Macpherson, J.B.
1994: Lateglacial environmental change in Newfoundland: a regional review. Journal of Quaternary Science, Volume 9, pages 171-178.

Andrews, K.
1980: Mineral occurrence map, Botwood, Newfoundland. Newfoundland and Labrador Department of Mines and Energy, Mineral Development Association. Map 80-4.

Ashley, G.M., Shaw, J., and Smith, N.D.
1985: Glacial Sedimentary Environments. Short course 16, Society of Economic Palaeontologists and Mineralogists (SEPM), Tulsa, Oklahoma. 246 pages.

Ashley, G.M., Southard, J.B., and Boothroyd, J.C.
1982: Deposition of climbing ripple beds: a flume simulation. Sedimentology, Volume 29, pages 67-79.

Atmospheric Environment Service
1993: Canadian Climate Normals 1961-1990. Volume 6, Atlantic Provinces.

Bagnold, R. A.
1968: Deposition in tho process of hydraulic transport. Sedimentology, Volume 10, pages 45-56.

Baird, D.M.
1950: Fogo Island map-area, Newfoundland. Geological Survey of Canada, paper 50-22, 56 pages.

Banfield, C.E.
1981: The climatic environment of Newfoundland. *In* The Natural Environment of Newfoundland: Past and Present. *Edited by* A.G. Macpherson and J.B. Macpherson. Department of Geography, Memorial University of Newfoundland, pages 83-153.

Batterson, M.J.
In preparation: The Quaternary History and Palaeogeography of the Humber River Valley, Western Newfoundland. PhD thesis, Department of Geography, Memorial University of Newfoundland.

Batterson, M.J, and McGrath, B.
1993: Quaternary geology of the Deer Lake and Pasadena map areas (NTS 12H/3 and 12H/4) *In* Current Research, Newfoundland and Labrador Department of Mines and Energy, Mineral Development Division. Report 93-1, pages 103-112.

Batterson, M.J., St. Croix, I., Taylor, D.M., and Vatcher, S.
1991: Ice-flow indicators on the Gander Lake map sheet (NTS 2D/15). Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Open File 2D/15 (233) Map 91-01.

Batterson, M.J., and Taylor, R.C.
1994: Quaternary Geology of the upper Humber River area, western Newfoundland. *In* Current Research, Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Report 94-1. pages 1-9.

Batterson, M.J. and Vatcher, H.
1991: Quaternary geology of the Gander (NTS 2D /15) map area. *In* Current Research, Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Report 91-1, pages 1-12.

Blake, W.Jr.
1983: Geological Survey of Canada radiocarbon dates XXXIII. Geological Survey of Canada Paper 83-7. 34 pages.

Bouchard, M.A.
1989: Subglacial landforms and deposits in central and northern Québec, Canada, with emphasis on Rogen moraines. Sedimentary Geology, Volume 62, pages 293-308.

Boulton, G.S.
1971: Till genesis and fabric in Svalbard, Spitsbergen. *In* Till: a symposium. *Edited by* R.P. Goldthwait, Ohio State University Press, Columbus, Ohio, pages 41-72.

Boulton, G.S.
1972: Modern Arctic glaciers as depositional models for former ice sheets. Journal of the Geological Society of London. Volume 128, pages 361-393.

Boulton, G.S.
1979: Processes of glacier erosion on different substrata. Journal of Glaciology, Volume 23, pages 15-38.

Boulton, G.S., Cox, F., Hart, J., and Thornton, M.
1984: The glacial geology of Norfolk. Geological Society of Norfolk Bulletin, Volume 34, pages 103-122.

Boulton, G.S., Dent, D.L., and Morris, E.M.
1974: Subglacial shearing and crushing, and the role of water pressures in tills from south-east Iceland. Geografiska Annaler. Volume 56(A), pages 135-145.

Boulton, G.S. and Hindmarsh, R.C.A.
1987: Sediment deformation beneath glaciers: rheology and geological consequences. Journal of Geophysical Research, Volume B92, pages 9059-9082.

Bouma, A.H.
1962: Sedimentology of some Flysch deposits. Elsevier, Amsterdam. 168 pages.

Bourgeois, J. and Leithold, E.L.
1984: Wave worked conglomerate-depositional processes and criteria for recognition. Canadian Society of Petroleum Geologists, Memoir 10, pages 331-343.

Brandal, M.K. and Heder, E.
1991: Stratigraphy and sedimentation of a terminal moraine deposited in a marine environment - two examples from the Ra-ridge in Ostfold, southeast Norway. Norsk Geologisk Tidsskrift, Volume 71, pages 3-14.

Brookes, I.A.
1971: Fossil ice-wedge casts in Western Newfoundland. Maritime Sediments, Volume 7, pages 118-122.

Brookes, I.A.
1982: Ice marks in Newfoundland: a history of Ideas. Géographie physique et Quaternaire, Volume 36, pages 139-163.

Brookes, I.A.
1989: Glaciation of Bonavista Peninsula, Northeast Newfoundland. The Canadian Geographer, Volume 33, pages 2-18.

Burbridge, G.H. and Rust, B.R.
1988: A Champlain Sea subwash fan at St. Lazare, Quebec. *In* The late Quaternary development of the Champlain Sea basin. *Edited by* N.R. Gadd, N.R. Geological Association of Canada, Special Paper 35, pages 47-61.

Catto, N.R.
1989: Sedimentology of two Late Weichselian exposures at Idre, Central Sweden. Geologiska Foreningens i Stockholm Forhandlingar, Volume 111, pages 193-211.

Catto, N.R.
1990: Clast fabric of diamictons associated with some roches moutonnées. Boreas, Volume 19, pages 289-296.

Catto, N.R.
1992: Quaternary geological mapping, southwestern Avalon Peninsula. *In* Current Research, Newfoundland Department of Mines and Energy, Mineral Development Division. Report 92-1, pages 23-26.

Catto, N.R.
1992: Supraglacial sedimentation in continental glacial environments, Dalarna (Sweden) and Avalon, Newfoundland (Canada): a comparative analysis. Sveriges Geologiska Undersökning, Volume 81, pages 81-86.

Catto, N.R. and the Quaternary Research Group, University of Alberta.
1987: Geomorphology: Glacial and periglacial geology. Field and Laboratory Manual. Quaternary Research Group, University of Alberta, Edmonton. 215 pages.

Cheel, R.J. and Rust, B.R.
1986: A sequence of soft-sediment deformation (dewatering) structures in Late Quaternary subaqueous outwash near Ottawa, Canada. Sedimentary Geology, Volume 47, pages 77-93.

Clark, P.U.
1991: Clast pavements in fine grained glacigenic diamicton: indicators of subglacial deforming beds. Geology, Volume 19, pages 530-533.

Clark, J.A., Farrel, W.E., and Peltier, W.R.
1978: Global change in sea-level: a numerical calculation. Quaternary Research, Volume 9, pages 265-287.

Clark, P.U. and Hansel, I.
1989: Till lodgment, clast ploughing and glacier sliding over a deformable glacier bed. Boreas, Volume 18, pages 201-207.

Coleman, A.P.
1926: The Pleistocene of Newfoundland. Journal of Geology, Volume 34, pages 193-222.

Crowley, K.D.
1984: Filtering of depositional events and the completeness of sedimentary sequences. Journal of Sedimentary Petrology, Volume 54, pages 127-136.

Cumming, E.H., Aksu, A.E., and Mudie, P.J.
1992: Late Quaternary glacial and sedimentary history of Bonavista Bay, northeast Newfoundland. Canadian Journal of Earth Sciences, Volume 29, pages 222-235.

Currie, K.L.
1992: A new look at Gander-Dunnage relations in Carmanville map area, Newfoundland. Geological Survey of Canada, Paper 92-1D, pages 27-33.

Currie, K.L. and Pajari, G.E.
1981: Anatectic peraluminous granites from the Carmanville area, northeastern Newfoundland. Canadian Mineralogist, Volume 19, pages 147-158.

Currie, K.L., Pajari, G.E., and Pickerill, R.K.
1980: Carmanville map area, Newfoundland (2E/8), Geological Survey of Canada, Open File 776.

Curry, R..
1966: Observations of alpine mudflows in the Tenmile Range, Central Colorado. Geological Society of America Bulletin. Volume 77, pages 771-776.

Daly, R.A.
1921: Postglacial warping of Newfoundland and Nova Scotia. American Journal of Science, Series 4, volume 1, pages 381-391.

Davenport, P.H., and Nolan, L.W.
1988: Gold and associated elements in lake sediment from regional surveys in the Botwood map area (NTS 2E). Mineral Development Division, Department of Mines and Energy, Government of Newfoundland and Labrador. Open File 2E/563.

Davis, A.M.
1984: Ombrotrophic peatlands in Newfoundland, Canada: their origins, development and trans-Atlantic affinities. Chemical Geology, Volume 44, pages 287-309.

Davis R.A. Jr
1978: Coastal Sedimentary Environments. Springer Verlag, New York.

Davis, R.A.Jr., and Hayes, M.O.
1984: What is a wave-dominated coast? Marine Geology, Volume 60, pages 313-329.

Debicki, R.L.
1983: Placer deposits: their formation, evaluation, and exploitation. *In* Yukon Placer Mining Industry 1978-1982. Exploration and Geological Services, Northern Affairs Program, Indian and Northern Affairs Canada, Whitehorse, Yukon, pages 18-34.

Denton, G.H. and Hughes, J.T.
1981: The Last Great Ice Sheets. John Wiley and Sons, New York. 484 pages

Domack, E.
1984: Rhythmically bedded glaciomarine sediments on Whidbey Island, Washington. Journal of Sedimentary Petrology, Volume 54, pages 589-602.

Doornkamp, J.C. and King, C.A.M.
1971: Numerical analysis in geomorphology, Edward Arnold, London. 372 pages.

Dowdeswell, J.A., and Sharp, M.J.
1986: Characterization of pebble fabrics in modern terrestrial glacigenic sediments. Sedimentology, Volume 33, pages 699-710.

Drake, L.D.
1971: Evidence for ablation and basal till in East-Central New Hampshire. *In* Till: A Symposium. *Edited by* R.P. Goldthwait,Ohio State University Press, Columbus, Ohio, pages 73-91.

Dreimanis, A.
1956: Steep Rock ore boulder train. Geological Association of Canada Proceedings, Volume 8 (part 1), pages 27-70.

Dreimanis, A.
1976: Tills: their origin and properties. *In* Glacial Till. *Edited by* R. Legget. Royal Society of Canada Special Publication12, Ottawa, pages 11-49.

Dreimanis, A.
1982: Work group (1)- Genetic classifications of tills and criteria for their differentiation: Progress report on activities 1977-1982 and definitions on glacigenic terms. *In* INQUA commission on genesis and lithology of Quaternary deposits. *Edited by* C. Schluchter. Report on Activities 1977-1982. ETH, Zurich, pages 12-31.

Dreimanis, A., Hamilton, J.P., and Kelly, P.E.
1987: Complex subglacial sedimentation of Catfish Creek till at Bradtville, Ontario, Canada. *In* Tills and Glaciotectonics, *Edited by* J. Van der Meer. A.A. Balkema, Rotterdam, pages 73-87.

Dreimanis, A., and Lundqvist, J.
What should be called till? Striae, Volume 20, pages 5-10.

Dreimanis, A. and Vagners, U.
1972: The effect of lithology upon the texture of till. *In* Research Methods in Pleistocene Geomorphology. *Edited by* A. Falconer and E. Yatsu. Proceedings of the Second Guelph Symposium of Geomorphology, University of Guelph, Guelph, Ontario, pages 66-82.

Dyke, A.S.
1972: A geomorphological analysis of the elevated glaciofluvial delta system and associated deposits on the Eastport Peninsula, Newfoundland. Honours thesis, Department of Geography, Memorial University of Newfoundland, 78 pages.

Dzulynski, S. and Walton, E.K.
1965: Sedimentary features of Flysch and Greywackes. Elsevier, Amsterdam. 274 pages.

Elliot, T.
1986: Deltas. *In* Sedimentary Environments and Facies, Second Edition. *Edited by* H.G. Reading. Blackwell Scientific, pages 113-154.

Eronen, M. and Vesajoki, H.
1988: Deglaciation pattern indicated by the ice-margin formations in Northern Karelia, eastern Finland. Boreas, Volume 17, pages 317-327.

Evans, D.G.A.
1989: The nature of glaciotectonic structures and sediments at subpolar glacier margins, Northwest Ellesmere Island, Canada. Geografiska Annaler, Volume 71, pages 113-123.

Eyles, N.
1977: Late Wisconsinan glacitectonic structures and evidence of postglacial permafrost in north-central Newfoundland. Canadian Journal of Earth Sciences, Volume 14, pages 2797-2806.

Fader, G. B. H., and Miller, R.O.
1994: An overview of the surficial and shallow bedrock geology of the Grand Banks of Newfoundland. *In* Seabed Processes and Resources, Geological Association of Canada, Newfoundland Section. Annual technical meeting-programme and abstracts, page 9.

Fernald, M.L.
1911: An expedition to Newfoundland and Labrador. Rhodora, Volume 13, No. 151, pages 109-162.

Fisher, T.G. and Shaw, J.
1992: A depositional model for Rogen moraine, with examples from the Avalon Peninsula, Newfoundland. Canadian Journal of Earth Sciences, Volume 29, pages 669-686.

Fitzsimons, S.J.
1992: Sedimentology and depositional model for glaciolacustrine deposits in an ice-dammed tributary valley, western Tasmania, Australia. Sedimentology, Volume 39, pages 393-410.

Flint, R.F.
1940: Late Quaternary changes of level in western and southern Newfoundland. Bulletin of the Geological Society of America, Volume 41, pages 1757-1780.

Flint, R.F
1971: Glacial and Quaternary Geology. John Wiley and Sons, New York. 892 pages.

Foley, M.G.
1977: Gravel-lens formation in antidune-regime flow - a quantitative hydrodynamic indicator. Journal of Sedimentary Petrology, Volume 47, pages 738-746.

Folk, R.L.
1955: Student operator error in determination of roundness, sphericity, and grain size. Journal of Sedimentary Petrology, Volume 25, pages 297-301.

Folk, R.
1966: A review of grain size parameters. Sedimentology, Volume 6, pages 73-93.

Forbes, D.L. and Taylor, R.B.
1987: Coarse-grained beach sedimentation under proglacial conditions, Canadian Atlantic coast. *In* Glaciated Coasts. *Edited by* D. Fitzgerald and P. Rosen, Academic Press, San Diego, pages 51-86.

Fyfe, G.F.
1990: The effect of water depth on ice-proximal glaciolacustrine sedimentation: Salpausselkä I, southern Finland. Boreas, Volume 19, pages 147-164.

Gibling, M. R. and Rust, B.R.
1984: Channel margins in a Pennsylvanian braided, fluvial deposit: the Morien Group near Sydney, Nova Scotia, Canada. Journal of Sedimentary Petrology, Volume 54, pages 773-782.

Gilbert, G.K.
1898: Bowlder-pavement at Wilson, New York. Journal of Geology. Volume 6, pages 771-775.

Gordon, I.
1981: Ice scoured topography and its relationships to bedrock structures and ice movement in parts of Northern Scotland and Greenland. Geografiska Annaler, Volume 63(A), pages 55-65.

Grant, D.R.
1974: Prospecting in Newfoundland and the theory of multiple shrinking ice caps. Geological Survey of Canada, Paper 74-1b, pages 215-216.

Grant, D.R.
1975: Glacial features of the Hermitage-Burin Peninsula area, Newfoundland. *In* Report of Activities, Part C, Geological Survey of Canada, Paper 75-1C, pages 333-334.

Grant, D.R.
1977: Glacial style and ice limits, the Quaternary stratigraphic record, and changes of land and ocean level in the Atlantic Provinces, Canada. Géographie physique et Quaternaire, Volume 31, pages 247-260.

Grant, D.R.
1989: Quaternary geology of the Atlantic Appalachians of Canada. *In* Quaternary Geology of Canada and Greenland. *Edited by* R.J. Fulton. Geological Survey of Canada, No.1, Chapter 5, pages 391-440 (also in Geological Society of America, The Geology of North America, volume K-1).

Grant, D.R.
1991: Quaternary geology of St. Anthony - Blanc-Sablon area, Newfoundland and Quebec. Geological Survey of Canada Memoir 427.

Gravenor, C.P.
1986: Magnetic and pebble fabrics in subaquatic debris flow deposits. Journal of Geology, Volume 94, pages 683-698. 60 pages

Gustavson, T.C., Ashley, G.M., and Boothroyd, J.C.
1975: Depositional sequences in glaciolacustrine deltas. *In* Glaciofluvial and glaciolacustrine sedimentation. *Edited by* A.V. Jopling and B.C. McDonald. Society of Economic Palaeontologists and Mineralogists (SEPM), Tulsa, Oklahoma, Special Publication No. 23, pages 264-280.

Haldorsen, S.
1983: Mineralogy and geochemistry of basal till and their relationship to till forming processes. Norsk Geologisk Tidsskrift, Volume 63, pages 15-25.

Haldorsen, S. and Shaw, J.
1982: The problem of recognising melt-out till. Boreas, Volume 11, pages 261-277.

Harms, J.C., Southard, J.B., and Walker, R.G.
1982: Structures and sequences in clastic rocks. Society of Economic Palaeontologists and Mineralogists, Tulsa, Oklahoma. Short course No. 9. 249 pages.

Hart, J.K.
1990: Proglacial glaciotectonic deformation and the origin of the Cromer Ridge Complex, Northern Norfolk, England. Boreas, Volume 19, pages 165-180.

Hart, J.K., and Boulton, G.S.
1991: The interrelation of glaciotectonic and glaciodepositional processes within the glacial environment. Quaternary Science Reviews, Volume 10, pages 335-350.

Hayes, J.J.
1951: Hodges Hill, Newfoundland. Geological Survey of Canada. Paper 51-5 (preliminary map).

Hicock, S.R.
1991: On subglacial stone pavements in till. Journal of Geology, Volume 99, pages 607-619.

Hicock, S.R., and Dreimanis, A.
1992: Deformation till in the Great Lakes Region: implications for rapid flow along the south-central margin of the Laurentide ice sheet. Canadian Journal of Earth Sciences, Volume 29, pages 1565-1579.

Hicock, S.R., Dreimanis, A., and Broster, B.
1981: Submarine flow tills at Victoria, British Columbia. Canadian Journal of Earth Sciences, Volume 18, pages 71-80.

Hicock, S.R., Kristjansson, F.J., and Sharpe, D.R.
1989: Carbonate till as a soft bed for Pleistocene ice streams on the Canadian Shield north of Lake Superior. Canadian Journal of Earth Sciences, Volume 26, pages 2249-2254.

Hornbrook, E.H.W., Davenport, P.H., and Grant, D.R.
1975: Regional and detailed geochemical exploration studies in glaciated terrain in Newfoundland. Department of Mines and Energy, Mineral Development Division, Province of Newfoundland. Report 75-2. 116 pages.

Hubert, J.F., Reed, A.A., and Carey, P.J.
1976: Palaeogeography of the East Berlin Formation, Newark Group, Connecticut Valley. American Journal of Science, Volume 276, pages 1183-1207.

Jenner, K.A., and Shaw, J.
1992: Inner shelf Quaternary sediments off northeast Newfoundland. Current Research, part D; Geological Survey of Canada, Paper 92-1D, pages 189-198.

Jenness, S.E.
1960: Late Pleistocene glaciation of Eastern Newfoundland. Bulletin of the Geological Society of America, Volume 71, pages 161-180.

Johansson, H.G.
1972: Moraine ridges and till stratigraphy in Vasterbotten, northern Sweden. Sveriges Geologiska Undersökning, Volume 66 (C), 50 pages.

Jopling, A.V. and Walker, R.G.
1968: Morphology and origin of ripple-drift cross lamination, with examples from the Pleistocene of Massachusetts. Journal of Sedimentary Petrology, Volume 38, pages 971-984.

King, L.H., and Fader, G.B.H.
Wisconsinan glaciation of the Atlantic coastal shelf of southeast Canada. Geological Survey of Canada, Bulletin 363, 72 pages.

Kirby, F.T. and Ricketts, R.J.
1983: Aggregate resource map series. Newfoundland Department of Mines and Energy, Mineral Development Division. Open File Nfld 1287.

Kirby, F.T., Ricketts, R.J., and Vanderveer, D.G.
1988: Surficial and glacial geology - gravel resource inventory (NTS 2E/8). Newfoundland Department of Mines and Energy, Geological Survey Branch, Map 88-161.

Klassen, R.A. and Henderson, P.J.
1992: Quaternary geological studies, Buchans area of Central Newfoundland. *In* Current Research, Part D. Geological Survey of Canada Paper 92-1D, pages 11-19.

Komar, P.D.
1976: Beach processes and sedimentation. Prentice-Hall Inc., Englewood Cliffs, New Jersey. 429 pages.

Krüger, J.
1979: Structures in till indicating subglacial deposition. Boreas. Volume 8, pages 323-340.

Krumbein, W.
1934: Size frequency distribution of sediments. Journal of Sedimentary Petrology, Volume 4, pages 65-77.

Kuenen, P.H.
1949: The formation of beach cusps. Journal of Geology, Volume 56, pages 34-40.

Kuenen, P.H.
1965: Value of experiments in geology. Geologie en Mijnbouw, Volume 44, pages 22-36.

Kuenen, P.H.
1967: Emplacement of Flysch-type sand beds. Sedimentology, Volume 9, pages 203-243.

Kuenen, P.H.
1969: Grain size of turbidite ripples. Sedimentology, Volume 8, pages 253-261.

Kuenen, P. H., and Humbert, F.L.
1969: Grain size of turbidite ripples. Sedimentology, Volume 13, pages 253-261

Lawson, D.E.
1979: A comparison of pebble orientation in ice and deposits of the Matanuska Glacier, Alaska. Journal of Geology, Volume 90, pages 78-84.

Lawson, D.E.
1981: Distinguishing characteristics of diamictons at the margin of the Matanuska Glacier, Alaska. Journal of Geology, Volume 90, pages 279-300.

Lindholm, R.
1987: A practical guide to sedimentology. Allen and Unwin, London. 270 pages.

Lindström, E.
1988: Are roches moutonnées mainly preglacial forms? Geografiska Annaler, Volume 70A, pages 323-331.

Liverman, D.G.E.
1992: Application of regional Quaternary mapping to mineral exploration, Northeastern Newfoundland, Canada. Transactions of the Institution of Mining and Metallurgy (B), Volume 101, pages B89-B98.

Liverman, D.G.E.
1993: Postglacial sea-level change in Newfoundland as deduced from the distribution of radiocarbon dated marine shells. *In* The scientific challenge of out changing environment. *Edited by* J. Hall and M. Wadleigh. Canadian Global Change Programme. Incidental Report Series, IR93-2, pages 38-39.

Liverman, D.G.E., and Scott, S.
1990: Quaternary geology of the King's Point map sheet (NTS 12H / 9). *In* Current Research, Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Report 90-1, pages 27-38.

Liverman, D.G.E., Scott, S., and Vatcher, H.
1991: Quaternary geology of the Springdale map area (12H / 8). *In* Current Research, Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Report 91-1, pages 29-44.

Liverman, D.G.E., and St. Croix, 1989
1989: Quaternary geology of the Baie Verte Peninsula. *In* Current Research, Newfoundland and Labrador Department of Mines and Energy, Report 89-1, pages 237-247.

Liverman, D.G.E., and Taylor, D.
1990: Surficial geology of Insular Newfoundland, preliminary version. Newfoundland Department of MInes and Energy, Geological Survey Branch, Map 90-08.

Lowe, D.R.
1975: Water escape structures in coarse grained sediments. Sedimentology, Volume 22, pages 157-204.

Lowe, D.R.
1979: Sediment gravity flows: their classification and some problems of application to natural flows and deposits. *In* Geology of Continental Slopes. *Edited by* L.J. Doyle, and V.H. Pilkey. Society of Economic Palaeontologists and Mineralogists, Tulsa, Oklahoma, Special Publication No. 27, pages 75-84.

Lowe, D.R.
1982: Sediment gravity flows II: Depositional models with special reference to the deposits of high-density turbidity currents. Journal of Sedimentary Petrology, Volume 52, pages 279-297.

Lowe, D.R., and LoPiccolo, R.D.
1974: The characteristics and origins of dish and pillar structures. Journal of Sedimentary Petrology, Volume 44, pages 484-501.

Lundqvist, G.
1959: Description to accompany the map of the Quaternary deposits of Sweden. Sveriges Geologiska Undersökning, Volume 17. 116 pages.

Lundqvist, J.
1965: Glacial geology in Northeastern Newfoundland. Geologiska Foreningens i Stockholm Forhandlingar, Volume 87, pages 285-306.

Lundqvist, J.
1969: Problems of the so-called rogen moraine. Sveriges Geologiska Undersökning, Volume 64, 32 pages.

Lundqvist, J., Clayton, L., and Mickelson, D.M.
1993: Deposition of the Late Wisconsinan Johnstown Moraine, South Central Wisconsin. Quaternary International, Volume 18, pages 53-59.

McBride, E.F., Shepherd, R.G., Crawley, R.A.
1975: Origin of parallel, near horizontal laminae by migration of bedforms in a small flume. Journal of Sedimentary Petrology, Volume 45, pages 132-139.

McCabe, A.M., Dardis, G.F., and Hanvey, P.M.
1984: Sedimentology of a late Pleistocene submarine-moraine complex, County Down, Northern Ireland. Journal of Sedimentary Petrology, Volume 54, pages 716-730.

MacClintock, P. and Twenhofel, W.H.
1940: Wisconsin glaciation of Newfoundland. Geological Society of America Bulletin, Volume 51, pages 1729-1756.

McDonald, B.C., and Shilts, W.W.
1975: Interpretation of faults in glaciofluvial sediments. *In* Glaciofluvial and glaciolacustrine sedimentation. *Edited by* A.V. Jopling and B.C. McDonald. Society of Economic Palaeontologists and Mineralogists, Tulsa, Oklahoma, Special Publication No. 23, pages 123-131.

McDonald, B.C. and Vincent, J.S.
1972: Fluvial sedimentary structures formed experimentally in a pipe and the interpretation of subglacial sedimentary environments. Geological Survey of Canada, paper 72-27.

MacEachern, D.B.
1989: Stereo^{TM}, the stereographic projection program for the MacIntosh. Distributed by Rockware Inc. Wheat Ridge Colorado, U.S.A.

Mackay, J.R.
1984: The frost heave of stones in the active layer above permafrost with downward and upward freezing. Arctic and Alpine Research, Volume 16, pages 439-446.

MacKenzie, C.
In Preparation: The Quaternary History of the Botwood (NTS 2E/3) map area. MSc thesis, Department of Geography, Memorial University of Newfoundland.

Macpherson, J.B.
1981: The development of vegetation of Newfoundland and climatic change during the Holocene. *In* The natural environment of Newfoundland, past and present. *Edited by* Macpherson, A.G. and Macpherson, J.B. Memorial University of Newfoundland, St. John's, Newfoundland, pages 184-217.

Macpherson, J.B.
1985: The postglacial development of vegetation in Newfoundland and Eastern Labrador-Ungava: synthesis and climatic implications. *In* Climatic change in Canada 5: critical periods in the Quaternary climatic history of Northern North America. Syllogeus, Volume 55, pages 267-280.

Macpherson J.B., and Anderson, T.W.
1985: Further evidence of late glacial climatic fluctuations from Newfoundland: pollen stratigraphy from a north coast site. *In* Current Research (Part B), Geological Survey of Canada, Paper 85-1B, pages 383-390.

Massari, F., and Parea, G.C.
1988: Progradational gravel beach sequences in a moderate to high energy, microtidal marine environment. Sedimentology, Volume 35, pages 881-913.

May, R.W., Dreimanis, A., and Stankowski, W.
1980: Quantitative evaluation of clast fabrics within the Catfish Creek till, Bradtville, Ontario. Canadian Journal of Earth Sciences. Volume 17, pages 1064-1074.

Mayewski, P.A., Denton, G.H., and Hughes, T.J.
1981: Late Wisconsinan ice sheets in North America. *In* The last great ice sheets. *Edited by* G.H. Denton and J.T. Hughes. John Wiley and Sons, New York, pages 67-238.

Menzies, J.
1979: A review of the literature on the formation and location of drumlins. Earth Science Reviews, Volume 14, pages 315-359.

Menzies, J.
1990: Brecciated diamictons from Mohawk Bay, South Ontario, Canada. Sedimentology, Volume 37, pages 481-493.

Miall, A.D.
1977: A review of the braided river depositional environment. Earth Science Reviews, Volume 13, pages 1-62.

Miall, A.D.
1978: Lithofacies types and vertical profile models in braided river deposits: a summary. *In* Fluvial Sedimentology, *Edited by* A.D. Miall. Canadian Society of Petroleum Geologists Memoir 5, Calgary, Alberta, pages 597-604.

Middleton, G.V.
1969: Experimental studies related to problems of Flysch sedimentation. *In* Flysch Sedimentology in North America. *Edited by* J. Lajoie. Geological Survey of Canada, Special Paper 7, pages 253-272.

Middleton, G.V.
1970: Turbidity currents, grain flows and other mass movements downslope. *In* The New Concepts of Continental Margin Sedimentation. *Edited by* D.J. Stanley. American Geological Institution Short Course notes, pages GM-A-1 to GM-B-14.

Middleton, G.V., and Hampton, M.A.
1973: Sediment gravity flows: Mechanics of flow and deposition. *In* Turbidites and deep water sedimentation. *Edited by* G.V. Middleton and A.H. Bouma. Society of Economic Palaeontologists and Mineralogists, Tulsa, Oklahoma, Short Course. Pages 1-38.

Mills, P.C.
1983: Genesis and diagnostic value of soft sediment deformation structures - a review. Sedimentary Geology, Volume 35, pages 83-104.

Minnel, H.
1977: Transverse moraine ridges of basal origin in Härjedalen. Geologiska Foreningens i Stockholm Forhandlingar, Volume 99, pages 271-277.

Mitchell, G.F.
1992: Notes on a raised beach between two diamicts, Beginish Island, Valencia Harbour, County Kerry. Irish Journal of Earth Sciences, Volume 11, pages 151-163.

Moran, S.R.
1971: Glaciotectonic structures in drift. *In* Till: A Symposium. *Edited by* R.P. Goldthwait. Ohio State University Press, Columbus, Ohio, pages 127-148.

Moran, S.R., Clayton, L., Hooke, R.LeB., Fenton, M.M., and Andrashek, L.D.
1980: Glacier bed landforms of the Prairie region of North America. Journal of Glaciology, Volume 25, pages 457-476.

Munro, M.
1993: Surficial geology and landform classification of the Carmanville map sheet (NTS 2E/8). Newfoundland Department of Mines and Energy, Geological Survey Branch, Open File 2E/8 (1844), map number 93-13.

Murray, R.C.
1955: Directions of glacier ice motion in south-central Newfoundland. Journal of Geology, Volume 63, pages 268-274.

Pajari, G.E., Pickerill, R.K., and Currie, K.L.
1979: The nature, origin and significance of the Carmanville ophiolitic melange, northeastern Newfoundland. Canadian Journal of Earth Sciences, Volume 16, pages 1439-1451.

Pederson, G.K. and Surlyk, F.
1977: Dish structures in Eocene volcanic ash layers, Denmark. Sedimentology, Volume 24, pages 581-590.

Pickerill, R.K., Currie, K.L., and Pajari, G.E.
1981: Resedimented volcaniclastics in the Carmanville area, northeastern Newfoundland. Canadian Journal of Earth Sciences, Volume 18, pages 55-71.

Pickerill, R.K., Pajari, G.E., and Currie, K.L.
1979: Evidence of Caradocian glaciation in the Davidsville Group of northeastern Newfoundland. Geological Survey of Canada Paper 79-1C, pages 67-72.

Piper, D.J.D., Mudie, P.J., and Fader, G.B.
1990: Quaternary Geology. *In* Geology of the continental margin of Eastern Canada. *Edited by* M.J. Keen and G.L. Williams, Geological Survey of Canada, Geology of Canada, Volume 2, Chapter 10, pages 475-607.

Postma, G.
1984: Slumps and their deposits in fan delta front and slope. Geology, Volume 12, pages 27-30.

Powell, R.D.
1984: Glacimarine processes and inductive lithofacies modelling of ice-shelf and tidewater glacier sediments based on Quaternary examples. Marine Geology, Volume 57, pages 1-52.

Proudfoot, D.N. and St. Croix, L.
1987: Quaternary geology of the Bellburns (12I/5 and 6) map area. *In* Current Research, Newfoundland and Labrador Department of Mines and Energy, Mineral Development Division. Report 87-1, pages 11-21.

Quinlan, G. and Beaumont, C.
1981: A comparison of observed and theoretical postglacial relative sea level in Atlantic Canada. Canadian Journal of Earth Sciences, Volume 18, pages 1146-1163.

Rappol, M.
1985: Clast fabric strength in tills and debris flows compared for different environments. Geologie en Mijnbouw. Volume 64, pages 327-332.

Rastas, J. and Seppälä, M.
1981: Rock jointing and abrasion forms on roches moutonnées, southwest Finland. Annals of Glaciology, Volume 2, pages 57-62.

Roberts, B.A.
1983: Soils. *In* Biogeography and Ecology of the Island of Newfoundland. *Edited by* G.R. South. D. W. Junk Publishers, The Hague. pages 107-161.

Rogerson, R.J.
1982: The glaciation of Newfoundland and Labrador. *In* Prospecting in areas of glaciated terrain. *Edited by* P.H. Davenport. Canadian Institute of Mining and Metallurgy, Geology Division Publication, pages 37-56.

Ruddiman, W.P. and Glover, L.K.
1975: Subpolar North Atlantic circulation at 9300 yr. BP; faunal evidence. Quaternary Research, Volume 5, pages 361-389.

Ruegg, G.H.J.
1983: Glaciofluvial and glaciolacustrine deposits in the Netherlands. *In* Glacial deposits in Northwest Europe. *Edited by* J. Ehlers. A.A. Balkema, Rotterdam, pages 379-184.

Rust, B.R.
1977: Mass flow deposits in a Quaternary succession near Ottawa, Canada: diagnostic criteria for subaqueous outwash. Canadian Journal of Earth Sciences, Volume 14, pages 175-184.

Rust, B.R.
1988: Ice proximal deposits of the Champlain Sea at South Gloucester, near Ottawa, Canada. *In* The late Quaternary development of the Champlain Sea basin. *Edited by* N. R. Gadd. Geological Association of Canada, Special Paper 35, pages 37-45.

Rust, B.R. and Jones, B.G.
1987: The Hawkesbury sandstone south of Sydney, Australia: Triassic analogue for the deposits of a large, braided river. Journal of Sedimentary Petrology, Volume 57, pages 222-233.

Rust, B.R. and Romanelli, R.
1975: Late Quaternary Subaqueous outwash deposits near Ottawa, Canada. *In* Glaciofluvial and Glaciolacustrine Sedimentation. *Edited by* A.V. Jopling, and B.C. McDonald. Society of Economic Paleontologists and Mineralogists, Tulsa, Oklahoma, Special Publication 23, pages 177-192.

Sauer, E.K.
1974: Geotechnical implications of Pleistocene deposits in southern Saskatchewan. Canadian Geotechnical Journal, Volume 11, pages 359-373.

Scott, J.S.
1976: Geology of Canadian tills. *In* Glacial Till: an interdisciplinary study. *Edited by* R.F. Legget. Royal Society of Canada Special Publication No. 12, pages 11-49.

Scott, S.
1993: Placer gold in Quaternary glaciofluvial and raised marine deposits of the Comfort Cove map area (NTS 2E/7). Current Research, Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Report 93-1, pages 387-400.

Scott, S., Catto, N. and Liverman, D.
1991: Quaternary marine deposits of the Springdale-Hall's Bay area, Newfoundland. Atlantic Geology, Volume 27, pages 181-191.

Sharpe, D.R. and Cowan, W.R.
1990: Moraine formation in Northwestern Ontario: product of subglacial fluvial and glaciolacustrine sedimentation. Canadian Journal of Earth Sciences, Volume 27, pages 1478-1486.

Shaw, J.
1975: Sedimentary successions in Pleistocene ice marginal lakes. *In* Glaciofluvial and glaciolacustrine sedimentation, *Edited by* A.V. Jopling and B.C. McDonald. Society of Economic Palaeontologists and Mineralogists, Tulsa, Oklahoma, Special Publication No. 23, pages 281-303.

Shaw, J.
1979: Genesis of the Sveg tills and Rogen moraines of Central Sweden: a model of basal meltout. Boreas, Volume 8, pages 409-426.

Shaw, J.
1982: Melt-out till in the Edmonton area, Alberta, Canada. Canadian Journal of Earth Sciences, Volume 19, pages 1548-1569.

Shaw, J.
1983: Drumlin formation related to inverted melt-water erosional marks. Journal of Geology, Volume 29, pages 461-479.

Shaw, J.
1987: Glacial sedimentary processes and environmental reconstructions based on lithofacies. Sedimentology, volume 34, pages 103-116.

Shaw, J.
1989. Geochemical exploration in areas of glaciated terrain: geological processes. *In* Proceedings of Exploration '87. *Edited by* G.D. Garland. Ontario Geological Survey, Special Volume 3, 1989. Page 335.

Shaw, J. and Gorrel, G.
1990: Subglacially formed dunes with bimodal and graded gravel in the Trenton drumlin field, Ontario. Géographie physique et Quaternaire, Volume 45, pages 21-34.

Shaw, J. and Kvill, D.
1984: A glaciofluvial origin for drumlins in the Livingston Lake area, Saskatchewan. Canadian Journal of Earth Sciences, Volume 21, pages 1442-1459.

Shaw, J., Kvill, D. and Rains, B.
1989: Drumlins and catastrophic subglacial floods. Sedimentary Geology, Volume 62, pages 177-202.

Shaw, J. and Sharpe, D.R.
1987: Drumlin formation by subglacial meltwater erosion. Canadian Journal of Earth Sciences, Volume 24, pages 2316-2322.

Shaw, J.
1991: Quaternary Sediments and seabed conditions offshore from La Scie, Newfoundland. Geological Survey of Canada, Open File 2385, 9 pages.

Shaw, J., Beaver, D.E. and Wile, E.
1990: Marine geological surveys in northeast Newfoundland coastal waters: Hamilton Sound, Baie Verte, La Scie, Halls Bay, Little Bay, Sunday Cove Island. Cruise Report 90-035. Geological Survey of Canada, Open File 2333. 18 pages.

Shaw, J. and Forbes, D.L.
1990a: Relative sea level change and coastal response, Northeast Newfoundland. Journal of Coastal Research, Volume 6, pages 641-660.

Shaw, J. and Forbes, D.L.
1990b: Late Quaternary sedimentation in St George's Bay, southwest Newfoundland: acoustic stratigraphy and sea bed deposits. Canadian Journal of Earth Sciences, Volume 27, pages 964-983.

Shaw, J. Edwardson, K.A., and Russel, H.A.
In Review: Surficial sediments of Hamilton Sound, Newfoundland: Evidence constraining the postglacial relative sea level low-stand.

Shepard, F.P.
1954: Nomenclature based on sand-silt-clay ratios. Journal of Sedimentary Petrology, Volume 24, pages 151-158.

Shepard, F.P..
1955: Delta-front valleys bordering the Mississippi distributaries. Geological Society of America Bulletin. Volume 66, pages 1489-1498.

Shilts, W.W.
1976: Glacial till and mineral exploration. *In* Glacial Till: An Interdisciplinary Study. *Edited by*, R. F. Legget. Royal Society of Canada Special Publication Number 12, pages 205-224.

Shilts, W.W.
1982: Glacial dispersal-principles and practical applications. Geoscience Canada, Volume 9, pages 42-47.

Shvetsov, M.
1954: Concerning some additional aids in studying sedimentary formations, Academy of Sciences of the USSR, Doklady Earth Sciences, Volume 29, pages 61-66.

Simpson, C. and Schmidt, S.M.
1983: An evaluation of criteria to deduce the sense of movement in sheared rocks. Geological Society of America Bulletin, Volume 94, pages 1281-1288.

Sissons, J.B.
1961: Some aspects of glacial drainage channels in Britain, Part II. Scottish Geographical Magazine, Volume 77, pages 15-36.

Skinner. R.G.
1973: Quaternary stratigraphy of the Moose River Basin, Ontario. Geological Survey of Canada, Bulletin 225. 77 pages.

Sparkes, B.G.
1987: Quaternary mapping - La Poile (110/9) and La Poile River (110/16) map area, southwestern Newfoundland. *In* Current Research, Newfoundland and Labrador Department of Mines and Energy, Mineral Development Division, Report 87-1, pages

Stauffer, P.H.
1967: Grain flow deposits and their implications, Santa Ynes Mountains, California. Journal of Sedimentary Petrology, Volume 37, pages 487-508.

St. Croix, L. and Taylor, D.M.
1990: Ice flow in North Central Newfoundland. *In* Current Research. Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Report 90-1, pages 85-88.

St. Croix, L. and Taylor, D.M.
1991: Regional striation survey and deglacial history of the Notre Dame Bay area, Newfoundland. *In* Current Reasearch. Newfoundland Department of Mines and Energy, Geological Survey Branch. Report 91-1, pages 61-68.

Stea, R.R.
1984: The sequence of glacier movements in northern mainland Nova Scotia determined through mapping and till provenance studies. *In* Correlation of Quaternary Chronologies, *Edited by* W. Mahaney. Geo Books, Norwich, England, pages 279-297.

Stea, R.R. and Brown, Y.
1989: Variation in drumlin orientation, form , and stratigraphy relating to successive ice flows in southern and central Nova Scotia. Sedimentary Petrology, Volume 62, pages 223-240.

Stea, R.R., and Mott, R.J.
1989: Deglaciation environments and evidence for glaciers of Younger Dryas age in Nova Scotia, Canada. Boreas, Volume 18, pages 169-187.

Stone, R.O., and Summers, H.J.
1972: Study of subaqueous and subglacial sand ripples. University of Southern California, Department of Geological Science, Final Report 72-1.

Sudom, M.D. and Van de Hulst, J.W.
1985: Soils of the Botwood-Wesleyville area, Newfoundland. Report No. 6. Newfoundland Soil Survey. Land Resource Research Institute, Publication 82-37.

Sugden, D.E., and John, B.E.
1976: Glaciers and Landscape. Edward Arnold, London. 376 pages.

Tanner, V.
1940: The glaciation of the Long Range of Western Newfoundland: a brief contribution. Geologiska Foreningens Stockholm Forhandlinger, Volume 62, pages 361-368.

Taylor, D.M. and St. Croix, L.
1989: Glacial striations in north-central Newfoundland. Newfoundland and Labrador Department of Mines and Energy, Geological Survey Branch, Map 89-108, Open File Nfld (1875).

Taylor, D.M., St. Croix, L., and Vatcher, S.V.
1991: Newfoundland Striation Database. Newfoundland and LabradorDepartment of Mines and Energy, Mineral Development Division, Open File Nfld 2155, 73 pages.

Thomas, G.S.P.
1984: Sedimentation of a sub-aqueous esker-delta at Strathbathie, Aberdeenshire. Scottish Journal of Geology, Volume 20, pages 9-20.

Thorliefson, H.
1989: Quaternary stratigraphy of the Central Hudson Bay Lowland, Northern Ontario, Canada. Unpublished PhD thesis, Department of Geology, University of Colorado, Boulder, Colorado, 363 pages.

Thornburg, T.M. and Kulm, L.D.
1987: Sedimentation in the Chile Trench: depositional morphologies, lithofacies, and stratigraphy. Geological Society of America Bulletin, Volume 62, pages 932-966.

Tucker, C.M.
1973: The glacial geomorphology of west -central Newfoundland: Halls Bay to Topsails. Unpublished M.Sc thesis, Department of Geography, Memorial University, St. John's, Newfoundland.

Tucker, C.M.
1976: Quaternary studies in Newfoundland: a short review. Maritime Sediments, Volume 12, pages 63-73.

Twenhofel, W.H., and MacClintock P.
1940: Surface of Newfoundland. Geological Society of America Bulletin, Volume 51, pages 1665-1728.

Udden, J.
1898: Mechanical composition of wind deposits. Augustana Library Publication No. 1.

Van der Knapp, W., and Eijpe, R.
1969: Some experiments on the genesis of turbidity currents. Sedimentology, Volume 11, pages 115-124.

Vanderveer, D.B.
1985: Quaternary mapping Moran Heights, Labrador and Gander-Gambo area, Newfoundalnd. *In* Current Research, Newfoundland Department of Mines and Energy, Mineral Development Division, Report 85-1, pages 55-58.

Vanderveer, D.G., and Sparkes, B.G.
1980: Glacial flow features-Red Indian Lake to Grand Falls. Newfoundland Department of Mines and Energy, Mineral Development Division, Open File (NFLD 93), Map Number 80-20.

Vanderveer, D.G., and Taylor, D.M.
1987: Quaternary mapping in the Gander River area, Newfoundland. *In* Current Research. Newfoundland Department of Mines and Energy, Mineral Development Division. Report 87-1, pages 39-43.

Visser, J.N.J., Colliston, W.P., and Terblanche, J.C.
1984: The origin of soft-sediment deformation structures in Permocarboniferous glacial and proglacial beds, South Africa. Journal of Sedimentary Petrology, Volume 54, pages 1183-1196.

Walker, R.G.
1965: The origin and significance of the internal sedimentary structures of turbidites. Proceedings of the Yorkshire Geological Society, Volume 35, pages 1-32.

Watson, G.S.
1986: The statistics of orientation data. Journal of Geology, Volume 74, pages 786-797.

Weimer, R.J.
1970: Rates of deltaic sedimentation and intrabasin deformation, Upper Cretaceous of Rocky Mountain Region. *In* Deltaic Sedimentation: Modern and Ancient. *Edited by* J.P. Morgen and R.H. Shaver. Society of Economic Palaeontologists and Mineralogists, Tulsa, Oklahoma. Special Publication 15. Pages 270-292.

Wells E.D., and Pollett, F.C.
1983: Peatlands. *In* Biogeography and Ecology of the Island of Newfoundland. *Edited by* G.R. South. D. W. Junk Publishers, The Hague, pages 207-265.

Wentworth, C.
1922: A scale of grade and class terms for clastic sediments. Journal of Geology, Volume 30, pages 377-392.

Wentworth, C.M.
1967: Dish structures, a primary sedimentary structure in coarse turbidites (abstract). American Association of Petrologists and Geologists Bulletin 51, page 485.

Williams, E.P.
1971: Experimental studies of ablation surface patterns and resulting roll tongues. AIAA Journal, Volume 9, pages 1315-1321.

Williams, H., and Flint, S.
1990: Anatomy of a channel-bank collapse structure in Tertiary fluvio-lacustrine sediments of the lower Rhine basin, Germany. Geological Magazine, Volume 127, pages 445-451.

Williams, H., Piasecki, M.A.J., and Johnston, D.
1991: The Carmanville melange and Dunnage-Gander relationships in northeast Newfoundland. Geological Survey of Canada, Paper 91-1D, pages 15-24.

Wolfe, A.P., and Butler, D.L.
1994: Late-glacial and early Holocene environments at Pine Hill Pond, Newfoundland, Canada: evidence from pollen and diatoms. Boreas, Volume 23, pages 53-65.

Woodcock, N.H.
1977: Specification of fabric shapes using an eigenvalue method. Geological Society of America Bulletin, Volume 88, pages 1231-1236.

Wright, H.E. Jr., Matsch, C.L., Cushing, E.J.
1973: Superior and Des Moines lobes. *In* The Wisconsinan Stage: Geological Society of America Memoir 136. *Edited by* R. Black, R. Goldthwait, and H. Willman, pages 153-185.

Zingg, T.
1935: Beitrage zur Schatteranalyse. Schweizerische Mineralogische und Petrographische Mitteilungen, Volume 15, pages 39-140.

Appendix 1

Textural and lithological data from tills in the Carmanville region

		% FINER THAN														
SITE	UNIT[1]	-2Ø	-1Ø	0Ø	1Ø	2Ø	3Ø	4Ø	5Ø	6Ø	7Ø	8Ø	9Ø	10Ø	11Ø	COLOUR
11	290-340	-	-	84	74	65	55	49	43	34	25	19	16	14	-	5Y 4/3
11	120-200	-	-	85	86	79	74	69	68	59	51	32	26	22	-	5Y 3/2
13	Unit 1	-	-	87	72	56	33	30	22	17	14	12	12	12	-	5Y 3/2
13	Unit 2	-	-	83	73	62	39	33	28	20	11	7	2	-	-	5Y 3/2
15	100-170	-	-	-	-	-	49	32	22	15	10	5	3	-	-	5Y 6/6
15	170-200	-	-	84	68	55	41	39	37	28	20	16	14	12	-	-
27	-	-	-	77	41	50	40	40	38	32	27	20	14	14	-	5Y 4/4
28	-	-	-	87	76	67	52	52	48	42	33	21	16	-	-	5Y 6/2
29	Unit 1	-	-	-	-	-	-	-	-	-	-	-	-	-	-	10YR 5/8
29	Unit 2	-	-	79	71	67	59	51	45	39	30	24	20	18	-	2.5Y 4/4
29	Unit 3	-	-	83	71	68	60	52	48	41	37	32	24	22	-	2.5Y 4/2
30	Unit 4	-	-	79	76	72	69	68	68	52	49	33	20	20	-	-
30	Unit 3	-	-	-	-	-	-	-	-	-	-	-	-	-	-	5Y 3/2
30	Unit 2	-	-	74	69	52	41	32	20	20	19	18	16	15	-	-
30	Unit 1	-	-	-	-	-	-	-	-	-	-	-	-	-	-	5Y 4/4
31	185-120	-	-	77	64	51	42	40	34	28	19	10	7	-	-	5Y 5/3
31	145-200	-	-	69	41	30	26	26	26	22	20	16	13	-	-	5Y 4/2
32	Unit 1	-	-	62	50	35	29	19	15	13	11	7	4	-	-	5Y 5/2
32	Unit 2	97	89	82	71	54	49	44	39	27	16	15	11	-	-	5Y 5/2
32	Unit 3	91	89	87	82	69	62	54	49	41	32	18	16	-	-	5Y 3/2

List of grain size data and colour determinations for all till sites mentioned in the text.

[1]Where there is more than one unit and unit numbers are not indicated in the text, the depth (cm) from the surface is indicated for each sample.

SITE	UNIT[2]	-2Ø	-1Ø	0Ø	1Ø	2Ø	3Ø	4Ø	5Ø	6Ø	7Ø	8Ø	9Ø	10Ø	11Ø	COLOUR
		% FINER THAN														
32	Unit 4	93	87	81	69	65	60	57	55	32	29	18	18	-	-	5Y 4/4
35	Unit 1	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
35	Unit 2	-	-	-	-	-	-	89	88	72	51	32	26	-	-	5Y 4/4
37	-	-	85	85	60	49	49	42	41	36	29	19	16	13	-	5Y 4/4
41	-	-	-	86	70	57	43	32	28	21	17	12	7	7	-	5Y 3/2
47	-	-	-	89	79	74	73	60	53	48	35	24	12	-	-	2.5Y 3/2
48	-	-	-	76	58	41	42	23	19	14	11	9	6	-	-	2.5Y 4/4
50	-	-	-	87	72	64	54	50	50	48	42	33	23	20	18	2.5Y 5/2
51	-	-	-	-	-	-	-	54	47	42	29	20	14	-	-	5Y 5/4
52	-	-	-	86	79	70	62	50	46	35	21	16	9	-	-	5Y 3/2

List of grain size data and colour determinations for all till sites mentioned in the text.

[2]Where there is more than one unit and unit numbers are not indicated in the text, the depth (cm) from the surface is indicated for each sample.

	LITHOLOGY					SHAPE					FABRIC				
SITE	GRC	GG	DG	HS	G	S	R	D	B	Ro	S_1	S_3	K	Or(°)	P (°)
11	23	0	65	C	12	6	62	8	24	3	0.712	0.017	1.22	163.1	10
11	30	12	34	0	24	-	-	-	-	-	0.717	0.019	1.78	171.3	9
13 (2)	68	0	24	0	8	12	48	8	36	3-4	0.509	0.136	0.38	319.3	5
13 (1)	42	0	36	0	22	8	84	8	4	3	0.526	0.065	0.14	289.1	10
15	48	0	50	0	2	8	72	8	12	3	0.718	0.038	0.58	146.3	4
27	0	0	76	0	24	-	-	-	-	-	0.814	0.022	0.79	028.8	4
28	-	-	-	-	-	-	-	-	-	-	0.543	0.064	0.64	104.5	13
29	0	0	81	7	12	-	-	-	-	-	-	-	-	-	-
30 (4)	72	0	12	6	10	8	72	8	12	3-4	0.723	0.088	1.76	143.2	6
30 (4)	-	-	-	-	-	-	-	-	-	-	0.772	0.059	1.44	137.2	14
30 (2)	60	0	24	4	12	8	6	62	24	4	0.683	-	0.78	107.0	-
30 (2)	-	-	-	-	-	-	-	-	-	-	0.653	-	0.66	166.0	-

CLAST LITHOLOGY, SHAPE, AND FABRIC DETERMINATIONS FOR ALL THE TILL SITES OBSERVED IN THE CARMANVILLE AREA

GRC = Gander River Complex
GG = Gander Group
DG = Davidsville Group
HS = Hamilton Sound Sequence
G = Granites

S = spheres
R = rollers
D = discs
B = blades
Ro = roundness
Or = orientation

1 = very angular
2 = angular
3 = subangular
4 = subrounded
5 = rounded
6 = very rounded

SITE	LITHOLOGY					SHAPE					FABRIC				
	GRC	GG	DG	HS	G	S	R	D	B	Ro	S_1	S_3	K	Or(°)	P(°)
31	18	0	18	30	12	0	64	24	12	3-4	0.752	0.089	0.39	310.1	5
32	28	0	13	34	25	8	62	28	2	3	0.503	0.147	0.42	227.3	4
35	24	0	0	60	16	0	72	6	22	3	0.853	0.070	10.69	65.7	7
37	50	0	6	35	9	4	58	8	30	3	0.558	0.157	0.98	340.3	45
41	5	0	65	24	6	-	-	-	-	-	-	-	-	-	-
47	-	-	-	-	-	2	72	6	20	3	0.543	0.132	0.53	229.1	7
48	8	0	92	0	0	0	68	8	24	3-4	0.556	0.079	0.28	25.4	1
50	25	0	68	0	7	6	64	6	24	3	0.748	0.097	3.39	280.6	3
51	32	0	48	12	8	8	62	2	28	3	0.595	0.114	0.77	272.3	15
51	-	-	-	-	-	-	-	-	-	-	0.434	0.165	0.09	149.7	8
52	52	0	28	12	8	8	52	24	16	3-4	0.709	0.022	0.39	211.3	3

CLAST LITHOLOGY, SHAPE, AND FABRIC DETERMINATIONS FOR ALL THE TILL SITES OBSERVED IN THE CARMANVILLE AREA

GRC = Gander River Complex
GG = Gander Group
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HS = Hamilton Sound Sequence
G = Granites

S = spheres
R = rollers
D = discs
B = blades
Ro = roundness
Or = orientation

1 = very angula
2 = angular
3 = subangular
4 = subrounded
5 = rounded
6 = very rounded

Appendix 2

Textural data from glaciofluvial deposits in the Carmanville region

SITE	SECN	UNIT[1]	% FINER THAN -2Ø	-1Ø	0Ø	1Ø	2Ø	3Ø	4Ø	5Ø	6Ø	7Ø	8Ø	9Ø	10Ø
1	-	-	89	78	25	12	6	3	2	-	-	-	-	-	-
3	1	1a	-	-	98	97	94	82	73	43	24	15	9	5	-
3	1	1c	-	-	97	97	92	75	70	48	29	20	14	11	-
3	1	1d	-	-	100	97	94	78	69	39	28	18	15	11	-
3	1	1e	-	-	-	-	-	-	62	47	30	20	9	5	-
3	6	2	54	32	10	2	2	1	1	-	-	-	-	-	-
3	6	2	-	-	-	-	-	-	86	52	32	18	9	5	-
3	6	3	-	-	-	-	-	-	79	30	20	15	13	12	-
3	6	5	-	-	-	-	-	-	16	7	2	2	2	1	-
4	1	LGr	95	80	64	12	4	2	1	-	-	-	-	-	-
4	1	LGr	-	-	96	36	6	1	1	-	-	-	-	-	-
4	1	LGr	36	20	12	6	4	3	1	-	-	-	-	-	-
4	2	SS	-	-	-	99	98	64	21	-	-	-	-	-	-
4	2	UGr	-	-	-	98	76	35	11	-	-	-	-	-	-
4	2	SS	-	-	-	95	95	86	29	-	-	-	-	-	-
4	3	SS	-	-	-	-	-	-	54	22	15	11	10	10	5
4	3	SS	-	-	-	-	-	92	68	48	28	19	10	7	-
4	4	UGr	-	-	-	99	96	62	21	-	-	-	-	-	-
4	4	UGr	98	96	92	77	36		1	-	-	-	-	-	-
4	4	SS	-	-	-	-	-	-	80	74	58	36	21	14	4

GRAIN-SIZE DETERMINATIONS FROM GLACIOFLUVIAL SAND AND GRAVELS IN THE CARMANVILLE REGION

[1] LGr = Lower Gravel Unit. UGr = Upper Gravel Unit. SS = Sands and Silts. D = Diamicton

			% FINER THAN												
SITE	SECN	UNIT[1]	-2Ø	-1Ø	0Ø	1Ø	2Ø	3Ø	4Ø	5Ø	6Ø	7Ø	8Ø	9Ø	10Ø
4	7	SS	-	-	-	99	84	32	8	-	-	-	-	-	-
4	8	D	-	-	-	-	-	-	64	51	42	26	20	14	14
4	8	UGr	94	82	50	12	2	1	-	-	-	-	-	-	-
4	8	UGr	-	-	98	92	52	16	1	-	-	-	-	-	-
4	8	SS	-	-	-	98	86	10	2	-	-	-	-	-	-
4	8	SS	-	-	-	99	92	28	1	-	-	-	-	-	-
4	8	SS	-	-	-	98	86	22	3	-	-	-	-	-	-
5	-	SS	87	67	32	7	2	2	1	-	-	-	-	-	-
5	-	UGr	62	44	32	22	11	1	1	-	-	-	-	-	-
5	-	UGr	-	-	-	99	60	7	1	-	-	-	-	-	-
5	-	UGr	-	-	-	92	48	18	7	-	-	-	-	-	-
5	-	UGr	-	-	97	77	24	7	4	-	-	-	-	-	-
5	-	UGr	60	44	24	6	1	1	1	-	-	-	-	-	-
5	-	SS	-	-	-	99	96	68	16	-	-	-	-	-	-
5	-	SS	-	-	-	-	-	28	18	15	10	6	6	5	-

GRAIN-SIZE DETERMINATIONS FROM GLACIOFLUVIAL SAND AND GRAVELS IN THE CARMANVILLE REGION

[1] LGr = Lower Gravel Unit, UGr = Upper Gravel Unit, SS = Sands and Silts, D = Diamicton

Appendix 3

Location of striation sites and striation orientations

SITE	LOCATION	ELEVATION	ORIENTATION
1	688700, 5462300 (UTM)	65(°)	344(°)
2	688700, 5461300	65	265
			314
3	687800, 5460250	85	044
			322
4	686650, 5458650	85	342
5	692950, 5462900	75	098
			045
6	690550, 5461900	50	125
			055
7	689850, 5458750	90	105
			048
			347
			309
8	699200, 5467350	65	320
9	709250, 5474900	55	069
			021
10	695800, 5473750	70	023
			122
11	708650, 5475050	60	038
			355
12	692950, 5466200	70	025
			348
13	690800, 5464750	60	068
			358
14	692800, 5464750	60	114
			074
			041
15	703100, 5465800	70	024
			049

STRIATION ORIENTATIONS FROM ALL SITES IN THE CARMANVILLE AREA

SITE	LOCATION	ELEVATION	ORIENTATION
16	700000, 5463800	55	082*
			042*
17	700100, 5463950	60	082*
			039*
18	699350, 5461650	110	116
19	700000, 5463800	70	039*
20	710800, 5477650	50	056*
			026*
			344*
21	710300, 5479350	90	074
22	710325, 5479375	90	070*

STRIATION ORIENTATIONS FROM ALL SITES IN THE CARMANVILLE AREA

* Striation measurements which have a specific orientation. Measurements without an asterix are trends rather than orientations

FIGURE 4.2: Detailed Surficial Geology Map of the Carmanville Area

GEOLOGICAL SURVEY BRANCH
DEPARTMENT OF MINES AND ENERGY
GOVERNMENT OF NEWFOUNDLAND AND LABRADOR

SURFICIAL GEOLOGY AND LANDFORM CLASSIFICATION OF THE CARMANVILLE MAP SHEET (NTS 2E/8)

LANDFORM CLASSIFICATION

Each outlined area is assigned a landform classification consisting of up to three genetic categories and modifiers that designate the types of deposits within each area. Each category within a landform classification is listed in order of dominance and is separated from the other categories by a slash (e.g., Tv/R). The areas are divided so that generally three landforms or deposit types are identified within a given area. The landform classification system is also used to denote the approximate percentage of landforms occurring within an outlined area, but those which compose less than 5 percent of the area are not included in the classification. Five variations of the landform system are as follows:

1. Where three different landforms are included in a single map unit they are each separated by a single slash (/) and their relative percentages are (60 – 85), (15 – 35) and (5 – 15) respectively.
2. Where two landforms are included in a single map unit, a double slash (//) or single slash (/) is used to separate them, and their relative percentages are (85 – 95) and (5 – 15) for double slash, or (60 – 85) and (15 – 40) for a single slash.
3. A horizontal line is used to show that one material overlies another. For example, $\frac{O}{T}$ indicates that there is a 95 – 100 percent organic veneer overlying till.
4. A hyphen between two landform types indicates that they are approximately equal in area. For example, Tv-Rc indicates that till veneer and rock concealed by vegetation or a thin regolith are equal in area.
5. A composite symbol is used to show combinations of the above cases. For example, $\frac{AG}{T}$ indicates that about 60 – 85 percent of the area is covered by alluvium, 15 – 40 percent by glaciofluvial sediments, and is underlain by till.

LANDFORM CLASSIFICATION: GENETIC CATEGORY

Symbol	Depositional Environment	Origin and Characteristics of Materials
F	Fluvial	Alluvium, consisting of silt and clay to bouldery gravel, forms terraces and plains associated with modern stream channels, their floodplains and deltas; usually less than 1 m thick; deposited by fluvial action at or below maximum flood levels
C	Colluvial	Colluvium, generally consisting of coarse grained bedrock-derived materials, but may include sand, silt or clay, accumulates on the lower parts, or at the base, of steep rock faces; transported by gravity
E	Aeolian	Medium to fine grained sand and silt, well sorted, poorly compacted; commonly occur as dunes up to 10 m high; transported and deposited by wind

LANDFORM CLASSIFICATION

	Genetic Category								
Morphology	Fluvial (F)	Colluvial (C)	Aeolian (E)	Glaciofluvial (G)	Lacustrine (L)	Marine (M)	Glacial (T)	Organic (O)	Rock (R)
apron (a)		Ca							
concealed by vegetation (c)		Cc							Rc
drumlinoid (d)							Td		
eroded and dissected (e)		Ce	Ee	Ge	Le	Me	Te		
fan (f)		Cf		Gf					
hummock (h)			Eh	Gh			Th		
kettle (k)				Gk			Tk		
lineated (l)			El				Tl	Ol	Rl
plain (p)	Fp			Gp	Lp	Mp	Tp	Op	
ridge (r)			Er	Gr		Mr	Tr		Rr
terrace (t)	Ft			Gt	Lt	Mt	Tt		Rt
veneer (v)	Fv	Cv	Ev	Gv	Lv	Mv	Tv	Ov	
blanket (b)							Tb		
weathered (w)									Rw
complex (x)				Gx	Lx	Mx	Tx		Rx

Geological boundary (defined, gradational, assumed)

Scarp face at edge of fluvial terrace

Esker (flow direction known or assumed, unknown)

Meltwater channel or ice-marginal channel showing interpreted flow direction

Trend of ribbed or minor moraine ridges

Crestline of major moraine ridges

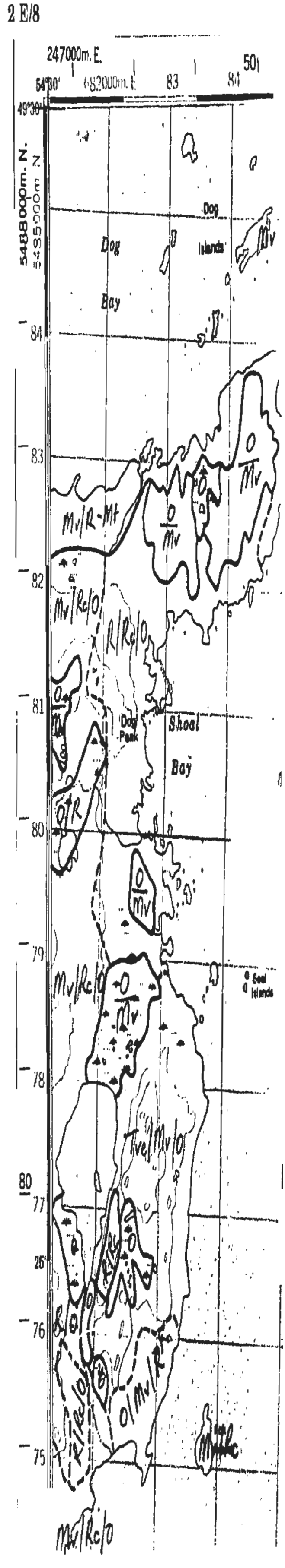
Fluting

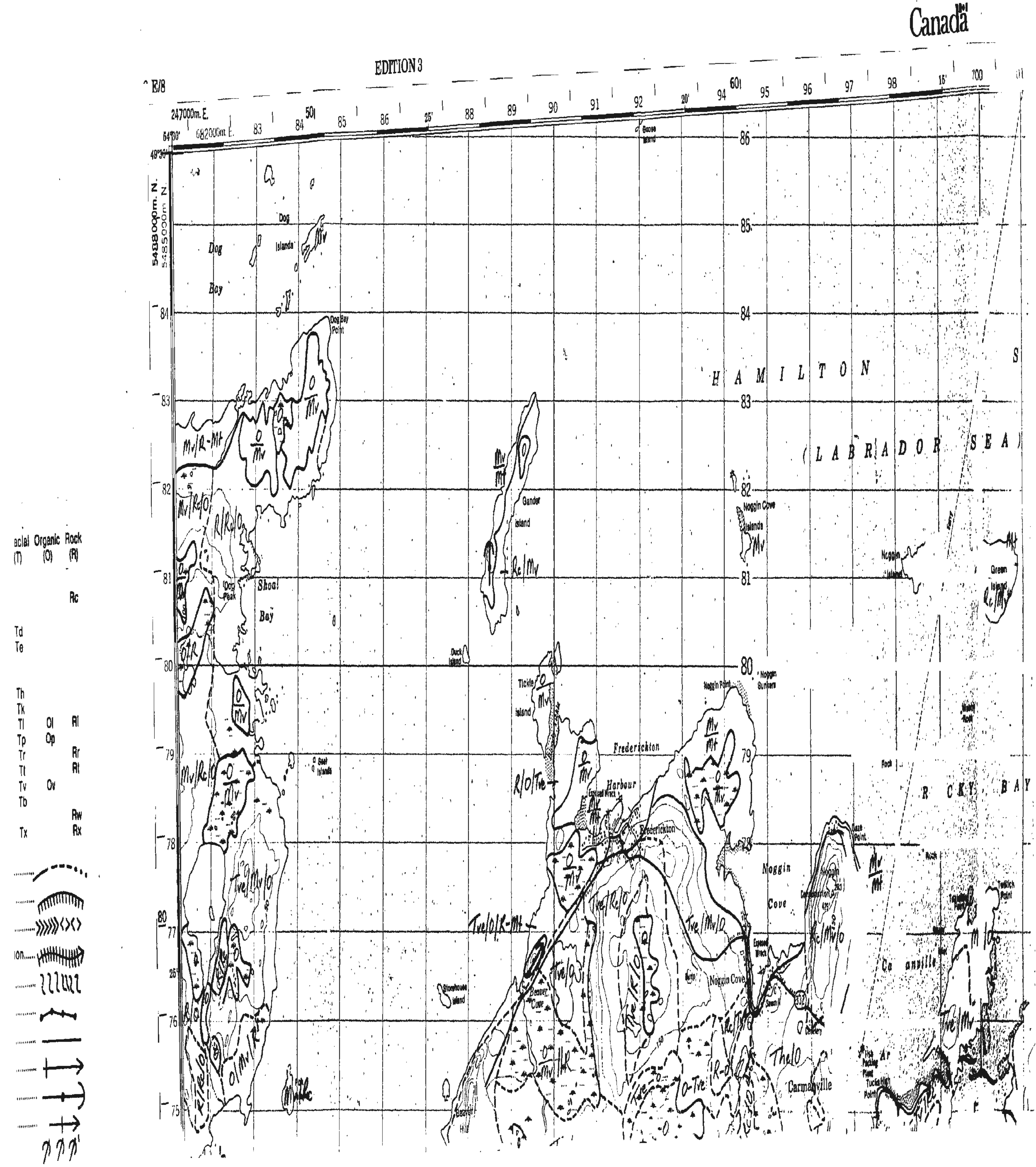
Crag and tail hill

Drumlin or drumlinoid feature

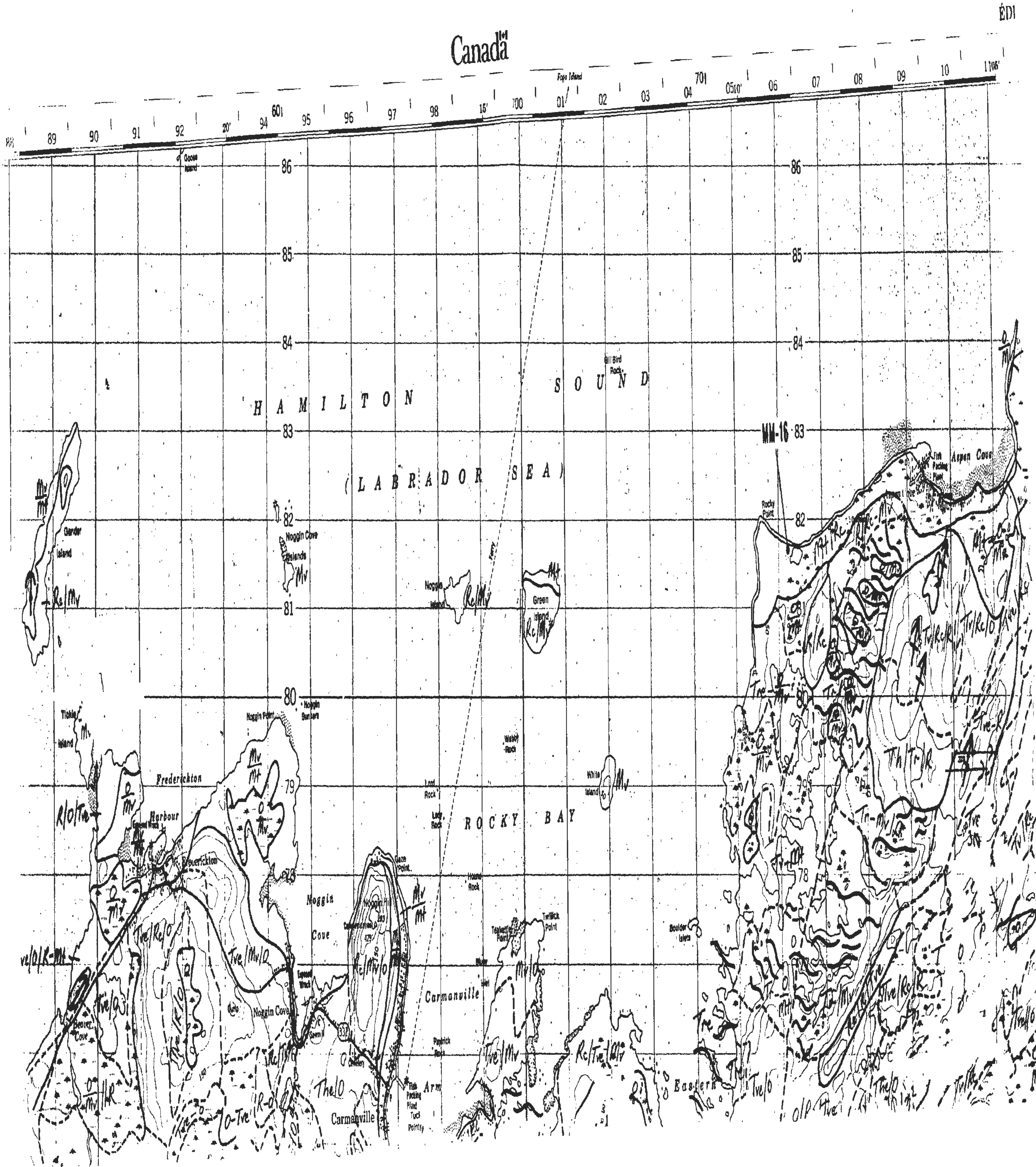
Rôche moutonée

Striations (directional, non directional). Numbers indicate age relationships (1 oldest, 2 younger)

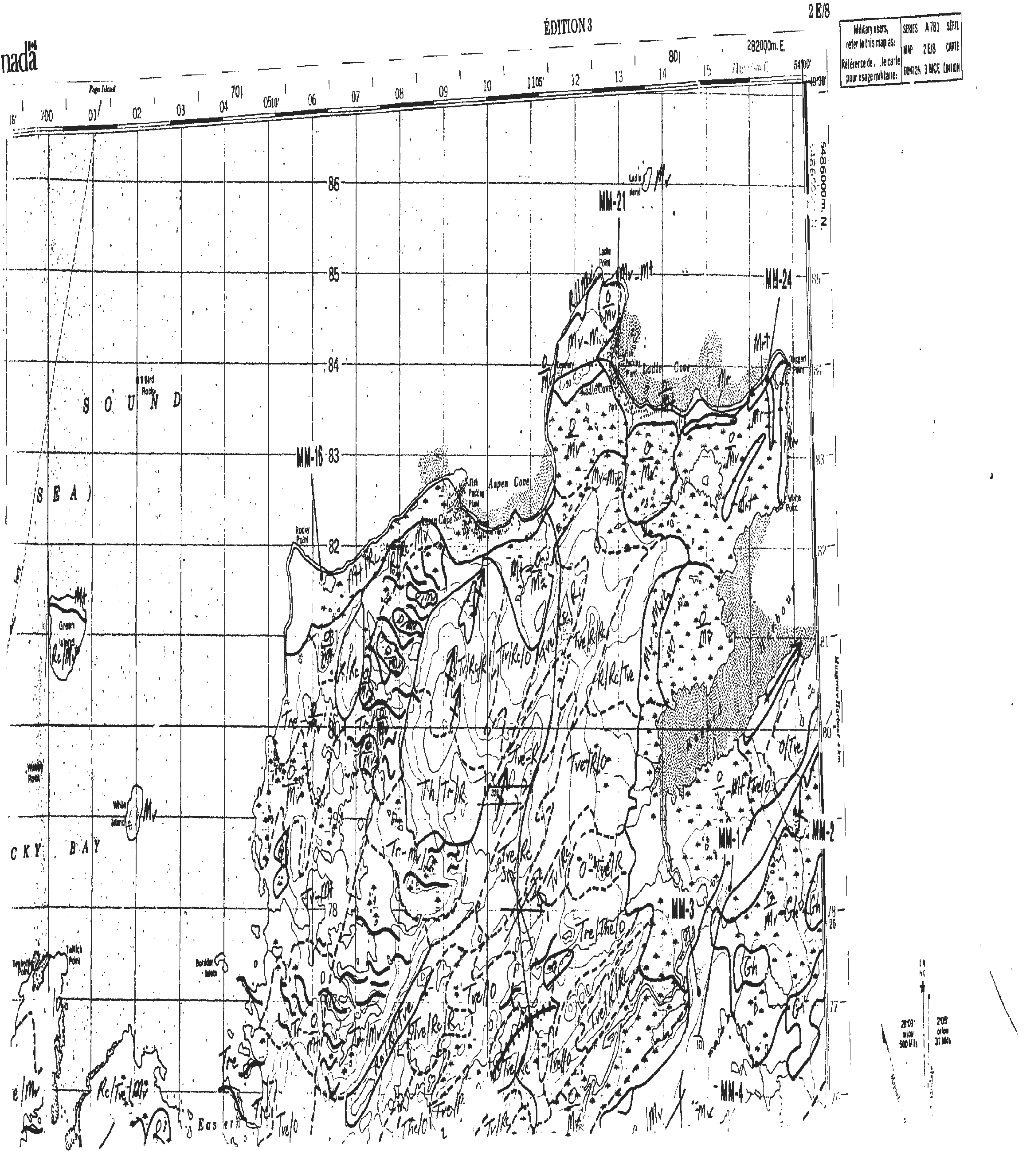




Canada
EDITION 3
247000m. E.
682000m. E.
5488000m. N.
5485000m. N.
49°30'
Dog Bay
Dog Islands
Dog Bay Point
Shoal Bay
Dog Peak
Goose Island
Gander Island
Duck Island
Tickle Island
Seal Islands
Storehouse Island
H A M I L T O N
(L A B R A D O R S E A)
R O C K Y B A Y
Noggin Cove Islands
Noggin Island
Green Island
Noggin Point
Noggin Sunkers
Fredericton Harbour
Fredericton
Noggin Cove
Beaver Cove
Carmanville
Glacial (T)
Organic (O)
Rock (R)
Rc
Td
Te
Th
Tk
Tl
Ol
Rl
Tp
Op
Tr
Rr
Tt
Rt
Tv
Ov
Tb
Rw
Tx
Rx



Canada
ÉDI
Fogo Island
HAMILTON SOUND
(LABRADOR SEA)
Goose Island
Gander Island
Noggin Cove Islands
Noggin Island
Green Island
MM-16
Rocky Point
Aspen Cove
Fish Packing Plant
Tickle Island
Noggin Point
Noggin Bunkers
Wabby Rock
White Island
Fredrickton
Harbour
Fredrickton
Gaze Point
Noggin Cove
Noggin Hill
Carmanville
Tapper Point
Twillick Point
Boulder Islets
Beaver Cove
Noggin Cove
Paddock Rock
Carmanville
Arm
Fish Packing Plant
Tuck Pointry
Eastern
ROCKY BAY


nadä
ÉDITION 3
2 E/8
Military users, refer to this map as:
Référence de la carte pour usage militaire:
SERIES A781 SÉRIE
MAP 2 E/8 CARTE
EDITION 3 MCE ÉDITION
282000m. E.
5486000m. N.
S O U N D
S E A
B A Y
MM-21
MM-24
MM-16
MM-1
MM-2
MM-3
MM-4
Ladle Point
Ladle Cove
Aspen Cove
Fish Packing Plant
Rocky Point
White Point
Green Island
Boulder Islets
Musgrave Harbour 4 km
Eastern

4. A hyphen between two landform types indicates that they are approximately equal in area. For example, Tv-Rc indicates that till veneer and rock concealed by vegetation or a thin regolith are equal in area.

5. A composite symbol is used to show combinations of the above cases. For example, $\frac{AG}{T}$ indicates that about 60 – 85 percent of the area is covered by alluvium, 15 – 40 percent by glaciofluvial sediments, and is underlain by till.

LANDFORM CLASSIFICATION: GENETIC CATEGORY

Symbol	Depositional Environment	Origin and Characteristics of Materials
F	Fluvial	Alluvium, consisting of silt and clay to bouldery gravel, forms terraces and plains associated with modern stream channels, their floodplains and deltas; usually less than 1 m thick; deposited by fluvial action at or below maximum flood levels
C	Colluvial	Colluvium, generally consisting of coarse grained bedrock-derived materials, but may include sand, silt or clay, accumulated on the lower parts, or at the base, of steep rock faces; transported by gravity
E	Aeolian	Medium to fine grained sand and silt, well sorted, poorly compacted; commonly occur as dunes up to 10 m high; transported and deposited by wind
G	Glaciofluvial	Fine grained sand to coarse grained cobbly gravel occur as plains, ridges (eskers), hummocks, terraces and deltas; generally greater than 1 m thick; deposited as outwash in an ice-contact position or proglacially
L	Lacustrine	Silt, clay, gravel and sand occur as plains and blankets; silt and clay deposited in fresh-water lakes from suspension, sand and silt by lake-floor currents, gravel and sand by shoreline wave action
M	Marine	Clay, silt, gravel and diamicton; sand is present in some places, generally moderately to well sorted and commonly stratified, but may be massive; occur as beach ridges, deltas, terraces and bars deposited in a marine environment; gravel and sand by shoreline wave action; may include shells, clay and silt deposited from suspension and turbidity currents; gravel is generally a wave-washed lag
T	Glacial	Includes all types of till; composed of diamicton (a sediment with particle size range from boulder to silt/clay); transported and subsequently deposited by or from glacier ice with no significant sorting by water
O	Bog	Poorly drained accumulations of peat, peat moss and other organics; developed in areas of poor drainage
R	Rock	Bedrock

LANDFORM CLASSIFICATION: MORPHOLOGY

Symbol	Morphology	Description
a	apron	A relatively gentler slope at the foot of a steeper slope – commonly used to describe colluvium at the base of a rock escarpment; consists of materials derived from the usually steeper upper slope
c	concealed by vegetation	Vegetation mat developed on either colluvium surfaces or a thin layer of angular frost-shattered and frost-heaved rock fragments overlying bedrock; includes areas of shallow (less than 1 m) discontinuous overburden
d	drumlinoid	Elongate ridge(s) between 6 and 60 m high, 75 and 300 m wide, and 100 to 2000 m long; ridges have a rounded end pointing in the up-ice direction and gently curving sides that taper in the down-ice direction; exhibit a convex longitudinal profile, commonly with a steeper slope in the up-ice direction; consist of subglacially formed deposits shaped in a streamlined form parallel to the direction of glacial flow; commonly consists of till, although some may contain stratified drift; may have a rock core
e	eroded and dissected	Series of closely spaced gullies or deeply incised channels; can have a dendritic pattern or may be a single straight or arcuate channel; gullies and channels may contain underfit streams
f	fan	A gently sloping accumulation of debris deposited by a stream issuing from a valley onto a lowland; has its apex at the mouth of the valley from which the stream issues; the fan shape results from the deposition of material as the stream swings back and forth across the lowland; alluvial fans are usually derived from eroded glacial and glaciofluvial deposits; glaciofluvial fans (deltas) are deposited in standing water rather than a terrestrial environment; colluvial fans are derived from bedrock and are usually steeper (i.e., cone shaped)
h	hummock	An apparently random assemblage of knobs, mounds, ridges and depressions without any pronounced parallelism, significant form or orientation

[illegible]				Mv	Tv	Ov	
blanket (b)					Tb		
weathered (w)							Rw
complex (x)		Gx	Lx	Mx	Tx		Rx

Geological boundary (defined, gradational, assumed)

Scarp face at edge of fluvial terrace

Esker (flow direction known or assumed, unknown)

Meltwater channel or ice-marginal channel showing interpreted flow direction

Trend of ribbed or minor moraine ridges

Crestline of major moraine ridges

Fluting

Crag and tail hill

Drumlin or drumlinoid feature

Rôche moutonée

Striations (directional, non directional). Numbers indicate age relationships (1 oldest, 2 younger)

Elevation in feet above mean sea level. Contour interval 50 feet.

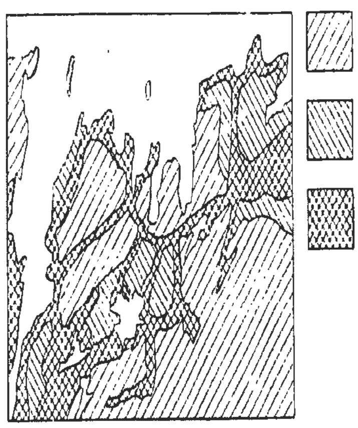
Geology by M.J. Munro, Department of Geography, Memorial University.

This map is subject to review and revision. Comments concerning errors or omissions are invited.

Copies of this map may be obtained from the Publications and Information Section, Geological Survey Branch, Department of Mines and Energy, P.O. Box 8700, St. John's, Newfoundland, A1B 4J6.

This project is funded by the Canada – Newfoundland Cooperation Agreement on Mineral Development (1990-1994); project carried out by Geological Survey Branch, Department of Mines and Energy, Government of Newfoundland and Labrador.

Scale 1:50 000
Open File 2E/08 (0844)
Map Number 93-13
Published 1993

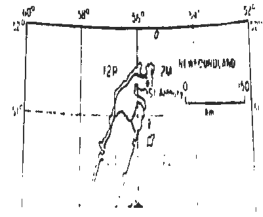


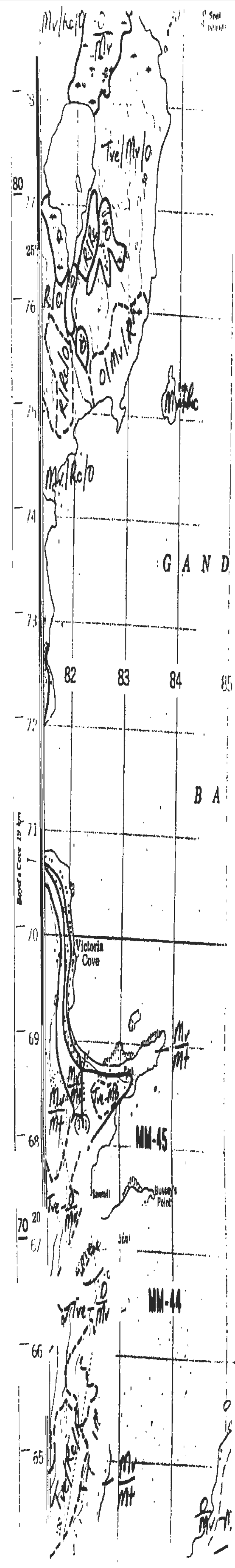
air photo interpretation only

limited ground checking (and airphoto interpretation)

detailed ground checking (and airphoto interpretation)

RELIABILITY DIAGRAM





Er Gr Mr Rr
F Gt Lt Mt Tt Rt
F Cv Ev Gv Lv Mv Tv Ov
Tb
Rw
Gx Lx Mx Tx Rx

oundary (defined, gradational, assumed)

t edge of fluvial terrace

direction known or assumed, unknown)

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major moraine ridges

I hill

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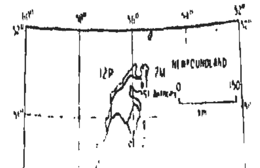
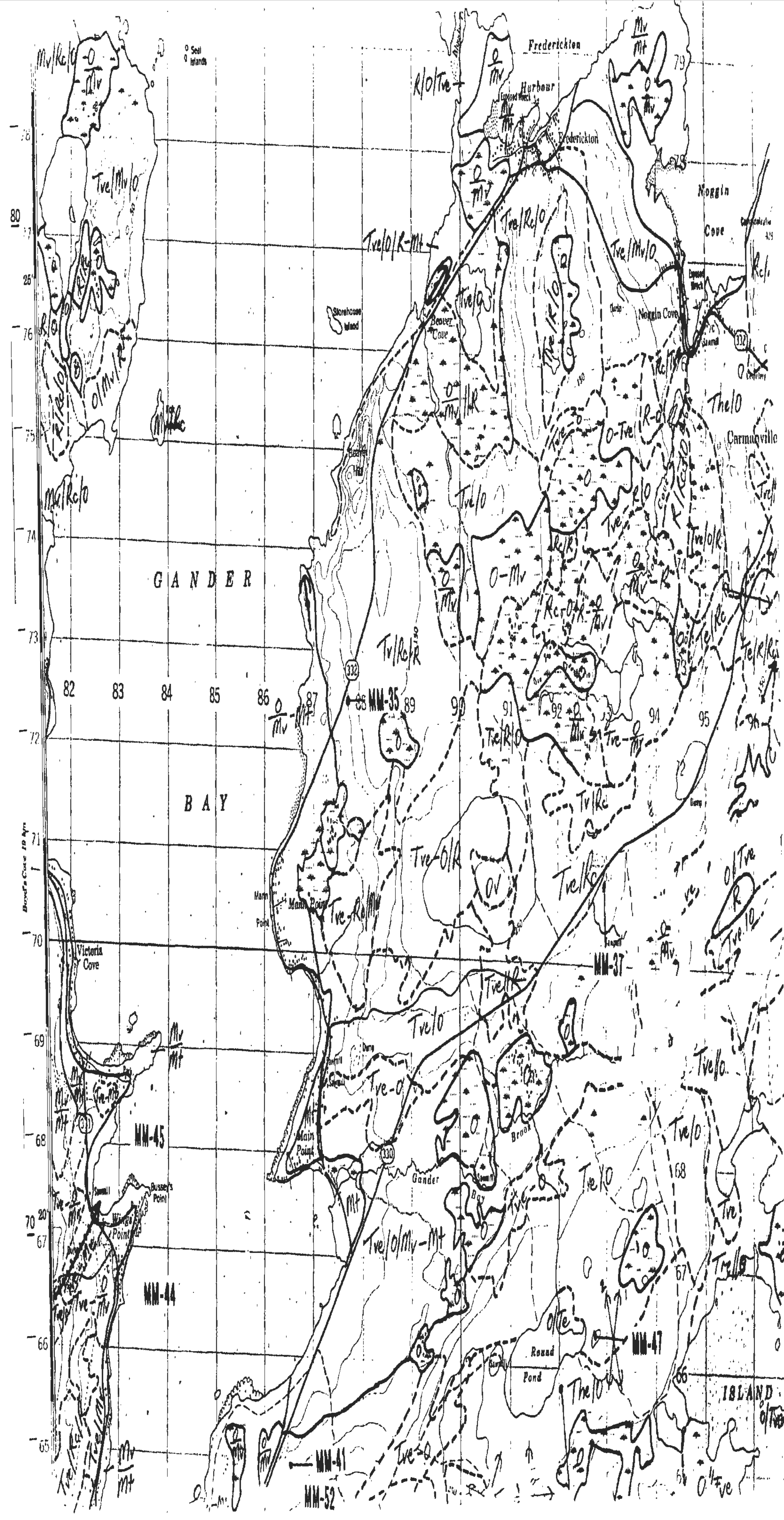
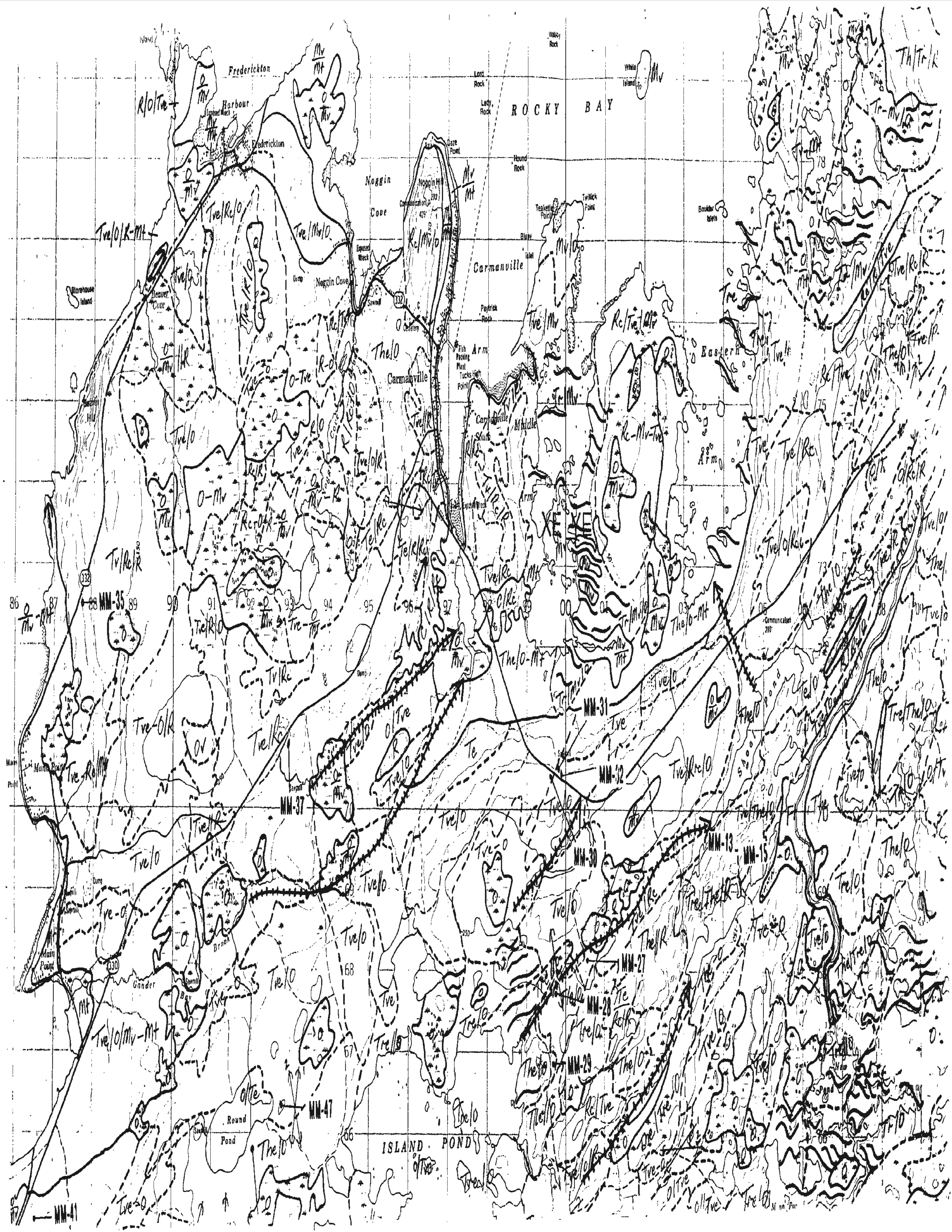
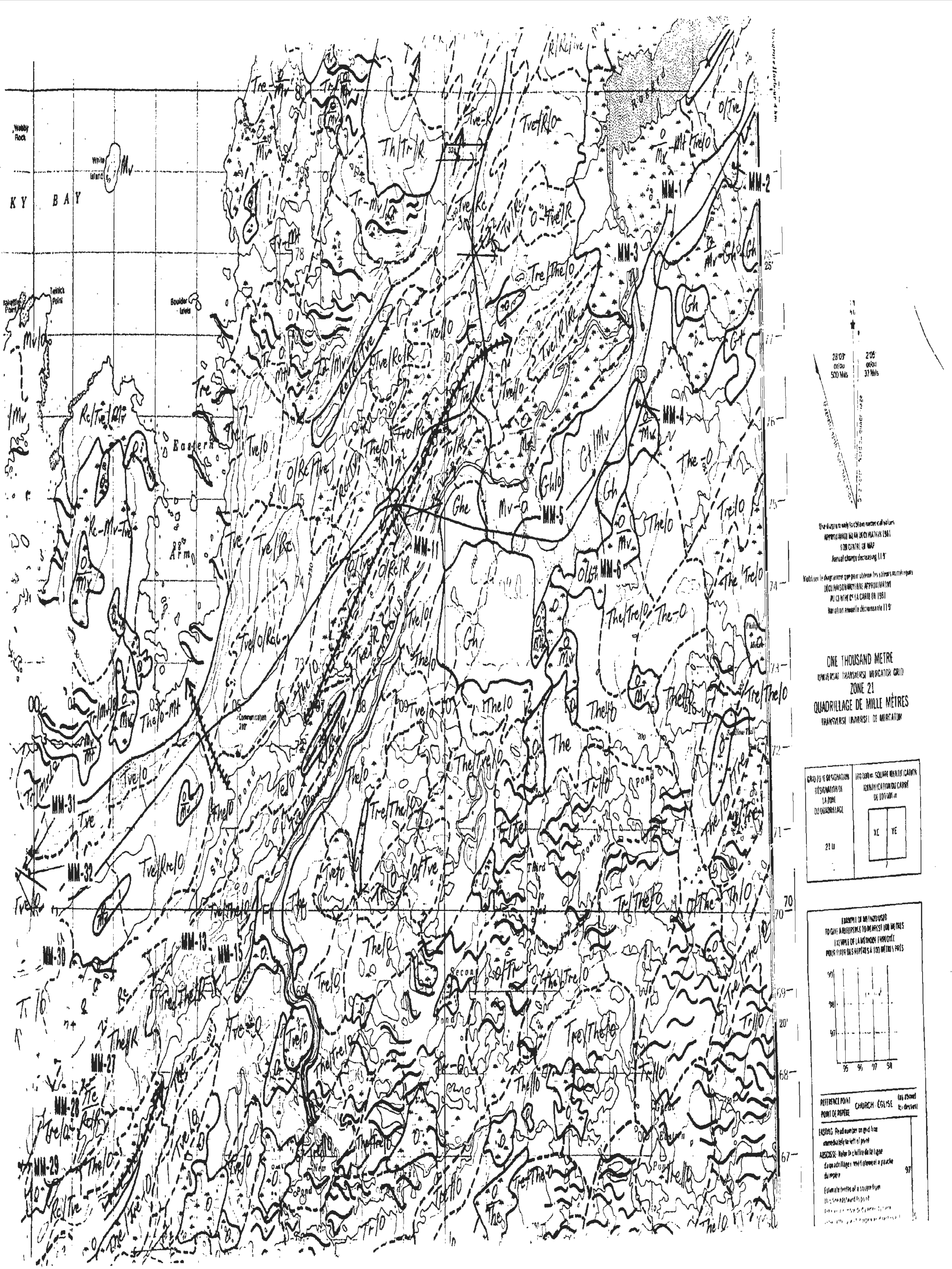






Fredrickton
Harbour
Fredrickton
ROCKY BAY
Lord Rock
Lady Rock
White Island
Hound Rock
Gaze Point
Noggin Cove
Noggin Cove
Teakettle Point
Twillick Point
Boulder Islets
Paytrick Rock
Carmanville
Carmanville
Arm
Eastern
Arm
Middle
Storehouse Island
Beaver Cove
Main Point
Main Point
Gander
Round Pond
ISLAND POND
MM-35
MM-37
MM-31
MM-32
MM-30
MM-13
MM-15
MM-27
MM-28
MM-29
MM-47
MM-41


MM-1
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MM-6
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MM-15
MM-27
MM-28
MM-29
MM-30
MM-31
MM-32
White Island
Boulder Islets
ONE THOUSAND METRE
UNIVERSAL TRANSVERSE MERCATOR GRID
ZONE 21
QUADRILLAGE DE MILLE MÈTRES
TRANSVERSE UNIVERSEL DE MERCATOR
REFERENCE POINT
POINT DE REPÈRE
CHURCH ÉGLISE

a	apron	[illegible] to describe colluvium at the base of a [illegible] materials derived from the usually steeper upper slope
c	concealed by vegetation	Vegetation mat developed on either colluvium surfaces or a thin layer of angular frost-shattered and frost-heaved rock fragme .ts overlying bedrock; includes areas of shallow (less than 1 m) discontinuous overburden
d	drumlinoid	Elongate ridge(s) between 6 and 60 m high, 75 and 300 m wide, and 100 to 2000 m long; ridges have a rounded end pointing in the up-ice direction and gently curving sides that taper in the down-ice direction; exhibit a convex longitudinal profile, commonly with a steeper slope in the up-ice direction; consist of subglacially formed deposits shaped in a streamlined form parallel to the direction of glacial flow; commonly consists of till, although some may contain stratified drift; may have a rock core
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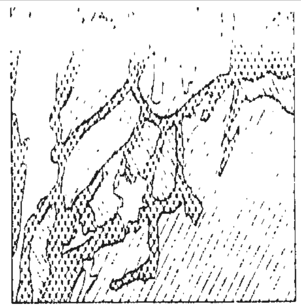
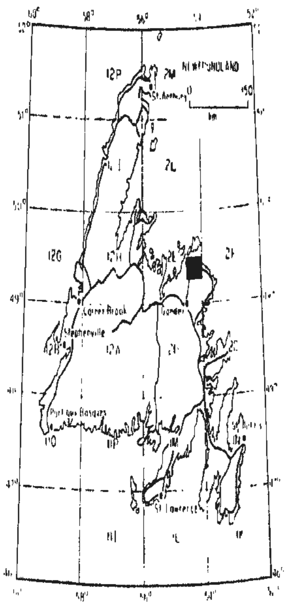
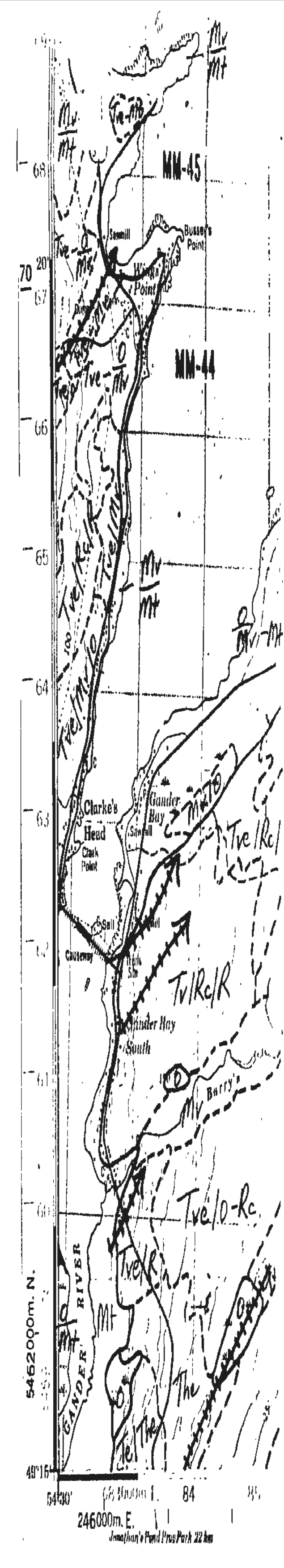
h	hummock	An apparently random assemblage of knobs, mounds, ridges and depressions without any pronouced parallelism, significant form or orientation; formed by glacial melting during ice stagnation and disintegration; includes subglacial, englacial, supraglacial and stratified materials
k	kettle	A basin or bowl-shaped closed depression or hollow in glacial drift; results from the melting of a buried or partly buried detached block or lens of glacier ice; commonly occurs in association with hummocks
l	lineated	Elongate spindle-shaped ridge(s) between 6 and 60 m high, 75 and 300 m wide and up to 4000 m long; ridges are commonly straight sided, taper at one or both ends, and have a flat longitudinal profile; consist of subglacially formed deposits shaped in a streamlined form parallel to the direction of glacial flow; commonly consist of till, although some may contain stratified drift; may have a rock core. Includes slope lineated bogs (Ol)
p	plain	A comparatively flat, level, or slightly undulating tract of land; materials are either till, glaciofluvial, alluvial, marine, lacustrine or organic sediments; bedrock features are commonly masked by the overlying sediments
r	ridge	Narrow, elongated and commonly steep-sided feature that rises above the surrounding terrain; materials are either rock, till, glaciofluvial, marine, lacustrine, aeolian, or organic sediments. Includes string bogs (Or)
t	terrace	Long, narrow, level or gently inclined step-like surface, bounded along one edge by a steeper descending slope or scarp and along the other by a steeper ascending slope or scarp; materials are either till, glaciofluvial, alluvial or lacustrine sediments; generally formed by fluvial and glaciofluvial erosion and marine wave action
v	veneer	Any deposit less than 2 m thick; morphology of the underlying unit is evident
b	blanket	Any deposit generally greater than 2 m thick; irregularities of the underlying unit are masked but the major topographic form is still evident
w	weathered	A thin layer, generally less than 1 m thick, of frost-heaved and frost-shattered bedrock fragments
x	complex	Commonly used to indicate numerous esker ridges that are closely spaced; can be used where any genetic catagory exhibits numerous surface expressions in a small area, and in which no single element can be defined



RELIABILITY DIAGRAM



INDEX MAP



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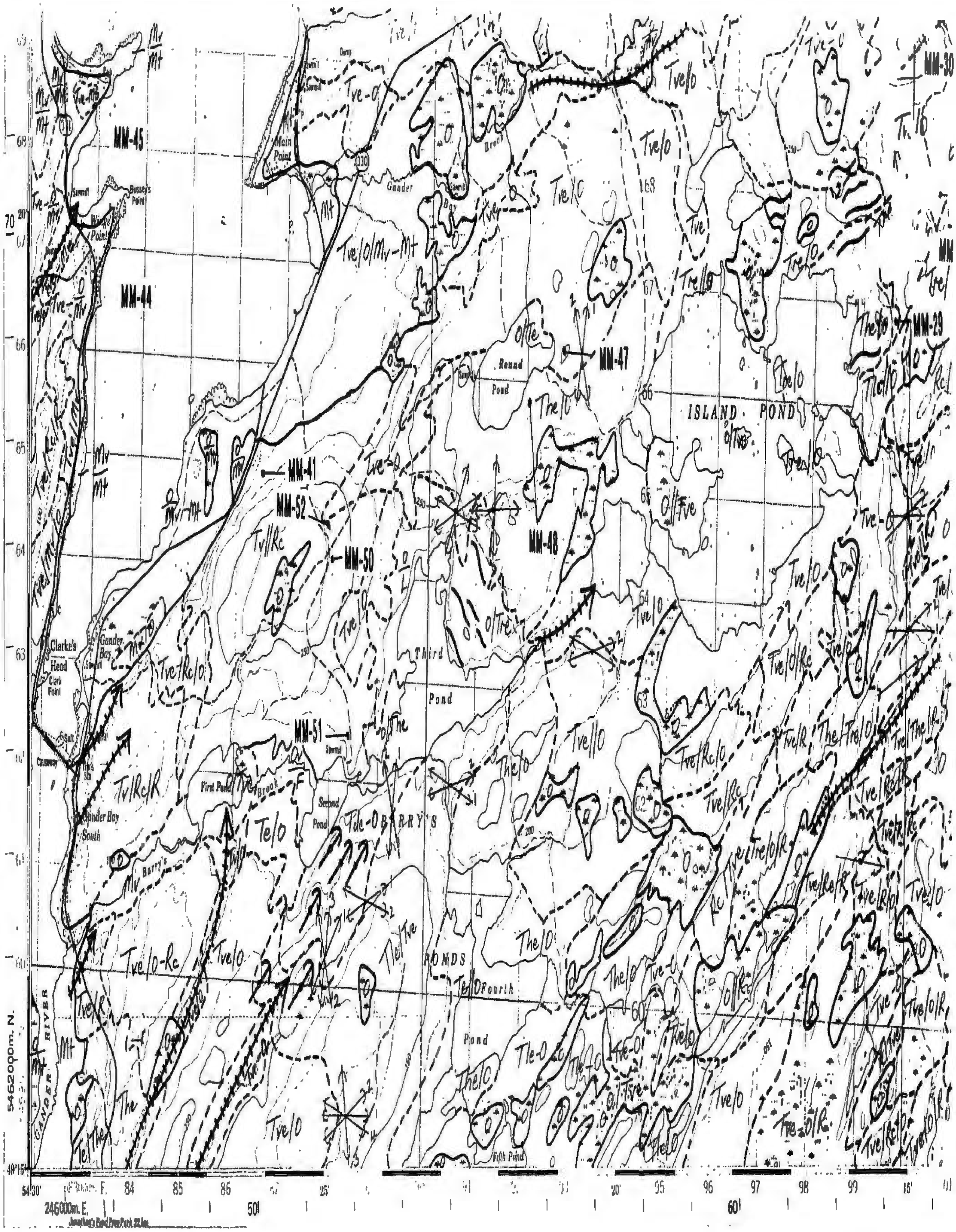

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FOR COMPLETE REFERENCE SEE REVERSE SIDE

POUR UNE LISTE COMPLÈTE DES SIGNES VOIR AU VERSO

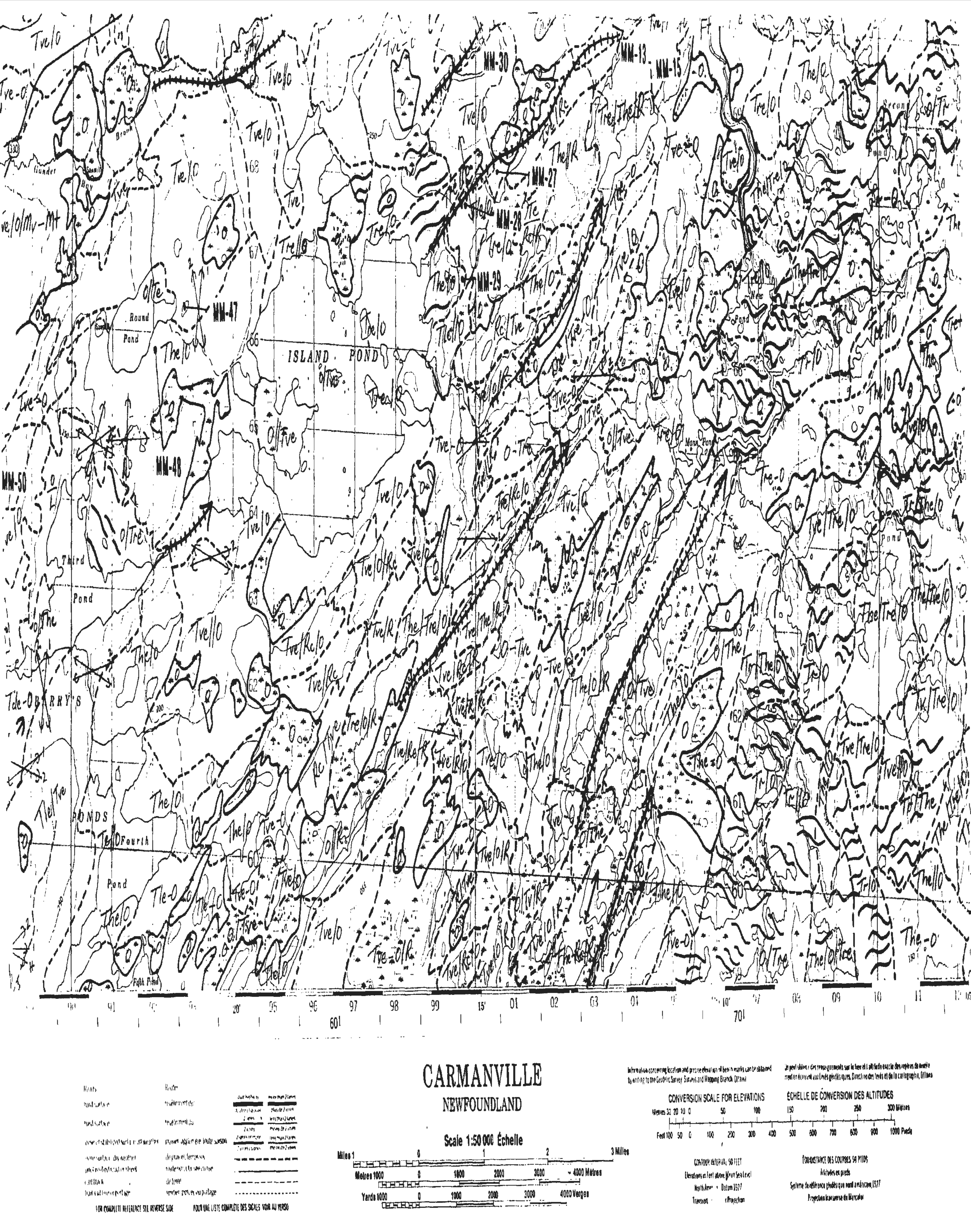
CARMANVI

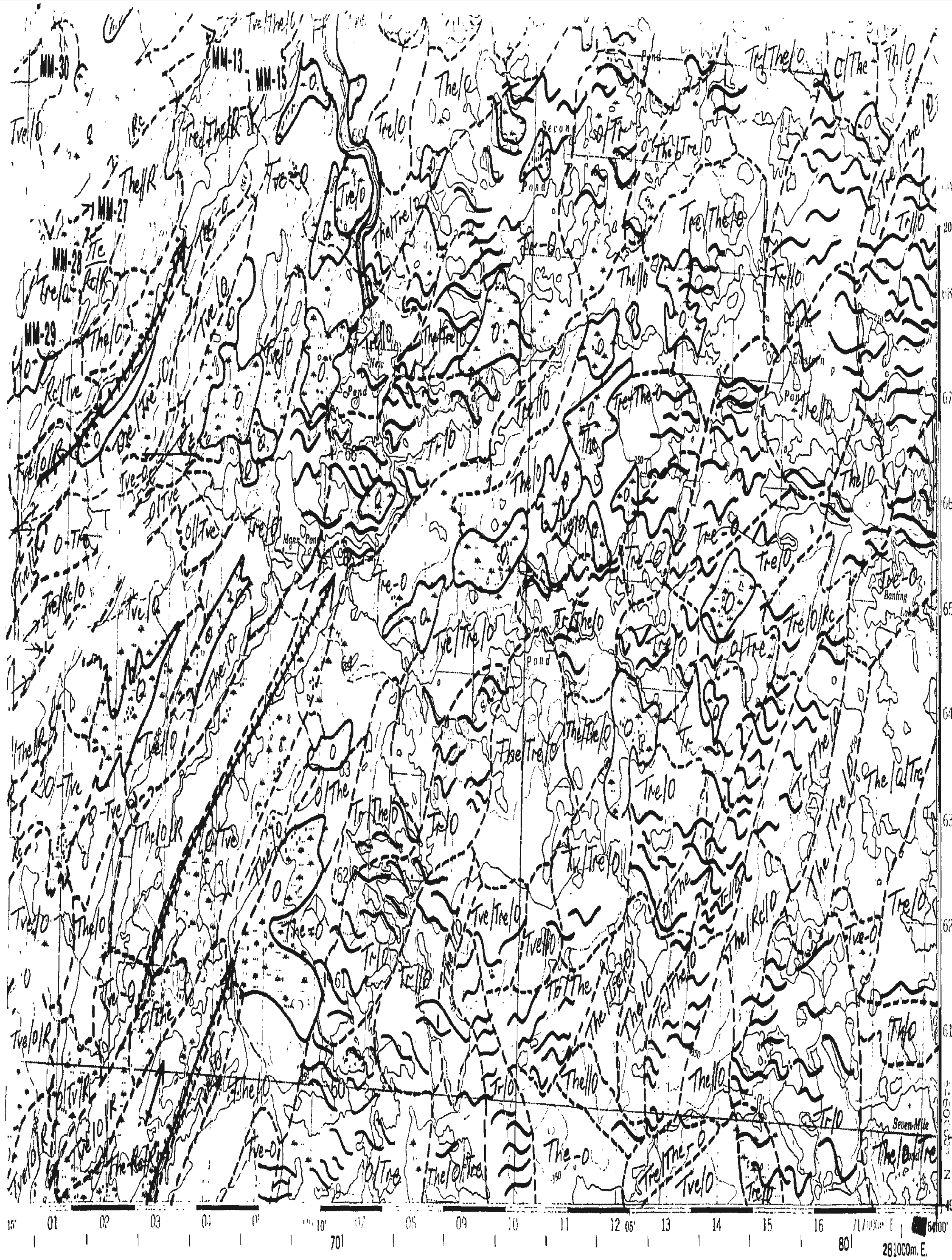
NEWFOUNDLANI

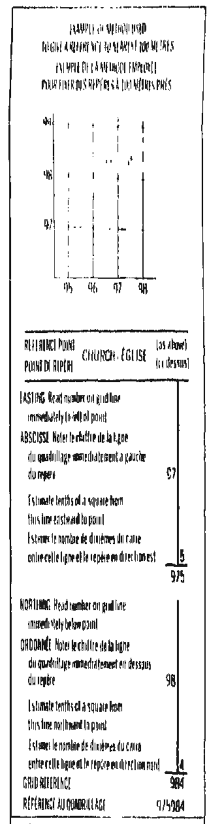
Scale 1:50 000 Échel

Metres 1000 0 1000 2000

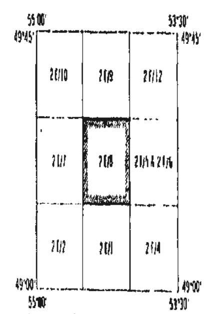
Yards 1000 0 1000 2000


MM-13
MM-15
MM-30
MM-27
MM-28
MM-29
MM-47
MM-48
MM-50
Round Pond
ISLAND POND
Third Pond
Fourth Pond
Fifth Pond
PONDS
Gander Bay
Brook
601
701
CARMANVILLE
NEWFOUNDLAND
Scale 1:50 000 Échelle
Miles 1 0 1 2 3 Milles
Metres 1000 0 1000 2000 3000 4000 Mètres
Yards 1000 0 1000 2000 3000 4000 Verges
CONVERSION SCALE FOR ELEVATIONS
ÉCHELLE DE CONVERSION DES ALTITUDES
CONTOUR INTERVAL 50 FEET
Elevations in Feet above Mean Sea Level
ÉQUIDISTANCE DES COURBES 50 PIEDS
Altitudes en pieds
FOR COMPLETE REFERENCE SEE REVERSE SIDE
POUR UNE LISTE COMPLÈTE DES SIGNES VOIR AU VERSO





BROWN NUMBERED TICKS INDICATE
THE 1000 METRE U.T.M. GRID
ZONE 22
LES AMORCES BRUNES NUMÉROTÉES REPRÉSENTENT
LE QUADRILLAGE DE 1000 MÈTRES T.U.M.



Index to adjoining Maps of the National Topographic System
Tableau d'assemblage du Système national de référence cartographique

NVILLE

UNDLAND

000 Échelle

1 2 3 Milles

2000 3000 4000 Mètres

2000 3000 4000 Verges

Information concerning location and precise elevation of bench marks can be obtained by writing to the Geodetic Survey, Surveys and Mapping Branch, Ottawa

On peut obtenir des renseignements sur le lieu et l'altitude exacte des repères de nivellement en écrivant aux Levés géodésiques, Direction des levés et de la cartographie, Ottawa

CONVERSION SCALE FOR ELEVATIONS — ÉCHELLE DE CONVERSION DES ALTITUDES

Mètres 30 20 10 0 50 100 150 200 250 300 Mètres

Feet 100 50 0 100 200 300 400 500 600 700 800 900 1000 Pieds

CONTOUR INTERVAL 50 FEET
Elevations in Feet above Mean Sea Level
North American Datum 1927
Transverse Mercator Projection

ÉQUIDISTANCE DES COURBES 50 PIEDS
Altitudes en pieds
Système de référence géodésique nord américain 1927
Projection transverse de Mercator

Établie par la DIRECTION DES LEVÉS ET DE LA CARTOGRAPHIE, MINISTÈRE DE L'ÉNERGIE, DES MINES ET DES RESSOURCES. Mise à jour à l'aide de photographies aériennes prises en 1978. Vérification des ouvrages en 1979. Publiée en 1982.

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CARMANVILLE
2 E/8
EDITION 3 ÉDITION

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