THE LATE QUATERNARY HISTORY OF TERRA NOVA NATIONAL PARK AND VICINITY, NORTHEAST NEWFOUNDLAND

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BY

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Abstract

The Quaternary history of Terra Nova National Park and vicinity, northeast Newfoundland, is dominated by Late Wisconsinan glaciation, glaciofluvial activity, post-glacial sea level fluctuations, and Holocene climate change. The eastern part of the study region is dominated by glacial erosional landforms, comprising thin veneers of fine-textured diamicton. To the west, eskers and glaciofluvial terraces are prominent, and the area is mantled by till and glaciofluvial deposits.

Late Wisconsinan glaciation was marked by flow towards the northeast and glaciogenic sediments contain high proportions of Terra Nova Granite and other locally-derived erratics. Primary and reworked glacial diamictons are present. The primary deposits reflect local ice flow movements rather than the regional ice flow movement towards the northeast. Glaciofluvial deposits include eskers, proximal and distal braided stream sequences, as well as associated pond sediments. Glacial retreat from Terra Nova National Park occurred after 12,000 years BP.

The marine limit is approximately 39 m asl, as indicated by raised deltas at Traytown and Sandy Cove. Marine incursion occurred at St. Chad's about 12,400 years BP. After 12,000 years BP, isostatic recovery resulted in the marine regression to modern sea level about 10,000 years BP, and to a minimum of -17 m by about 8,600 years BP. Ice-wedge casts preserved at 16 m asl at Port Blandford indicate the

existence of a cold climate interval between 11,000 and 10,000 years BP. This is coincident with the Younger Dryas cool event.

Sea level rose throughout the mid- and late-Holocene, reaching its present position about 2,000 years BP. Climate fluctuations during the late Holocene resulted in the expansion of wetlands throughout Terra Nova National Park.

Acknowledgments

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Chapter 1

Introduction

1.0 - Introduction

There has been a limited amount of Quaternary work done within northeast Newfoundland, and much of the recent work has concentrated on the offshore sediments (Cumming 1990; Cumming *et al.* 1992; Jenner and Shaw 1992). Some areas of Newfoundland such as the Northern Peninsula (Grant 1969a; Brookes 1977; Grant 1992), the Baie Verte Peninsula (Liverman and St. Croix 1989; Liverman and Scott 1990), the Bonavista Peninsula (Brookes 1989), and some areas of the Avalon Peninsula, are relatively well-documented. Recent work in northeast Newfoundland has been undertaken in the Carmanville region by Munro (1994), and in the Comfort Cove region by Scott (1993).

Very little research has been carried out in the Terra Nova region regarding the glacial history. Study within Terra Nova National Park and vicinity will aid in the understanding of the glacial history of northeast Newfoundland, and the Island as a whole.

1.1 - Objectives

Terra Nova National Park and its vicinity are characterized by an assemblage of Quaternary landforms and deposits that are the result of glaciation and sea level changes. The area was studied during 1994-1996 with several main objectives:

1) to describe and map landforms and sediments in Terra Nova National Park and its vicinity,

2) to determine the genesis of sediments,

3) to determine the ice flow history using the erosional and depositional landforms and their associated sediments,

4) to determine the post-glacial sea-level history, and

5) to reconstruct the Quaternary geologic history.

1.2 - Location and Access

Research was centred in northeast Newfoundland between latitudes 48°20'N and 48°40'N, and longitudes 53°41'W and 54°05'W (Figure 1.1a and b). Terra Nova National Park lies approximately 35 km north of Clarenville, and the study area extends from Port Blandford in the south to Sandringham in the north and as far as Terra Nova Village to the west. There are no major towns within the region examined, however, there are several smaller ones including Glovertown, Eastport, Sandy Cove and Charlottetown.

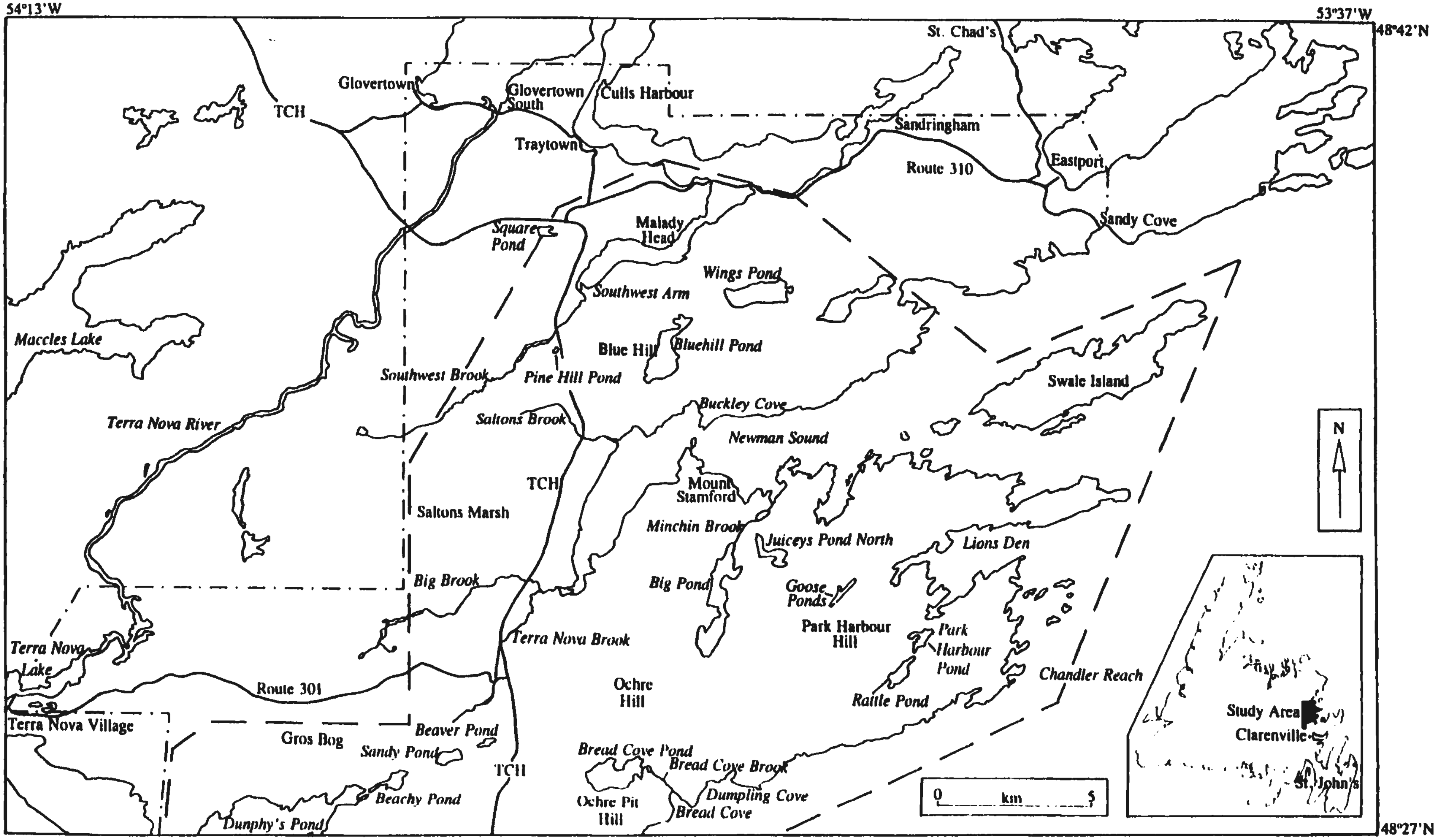

Figure 1.1a - Map showing the northern half of the study area with the location of all places mentioned in the text and a location map of the study area (inset) (see next page for legend)

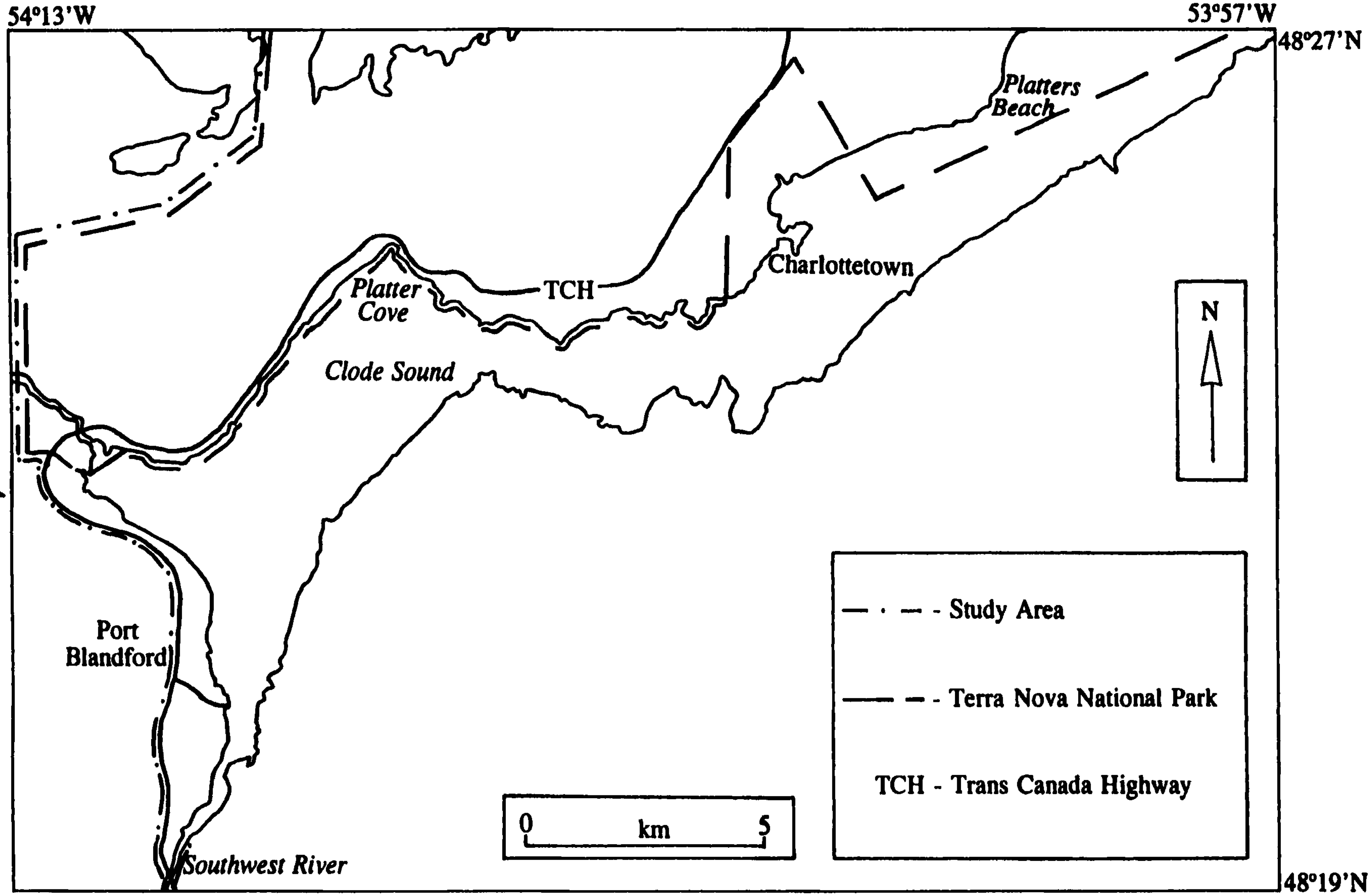

Figure 1.1b - Map showing the southern part of the study area with the location of all places mentioned in the text

Although the Trans Canada Highway runs through Terra Nova National Park, access within the Park by road is generally poor. In contrast, access in areas outside the Park is relatively good. Route 310 is paved and links Sandy Cove, Eastport and Sandringham with the Trans Canada Highway at Square Pond. Terra Nova Village is linked to the Trans Canada Highway by Route 301, and although it is a gravel road, it is well-maintained. There are many gravel logging roads and gravel pit roads within the area, however, many are no longer in use and are overgrown, limiting access to some areas.

1.3 - Bedrock Geology

An understanding of the bedrock geology aids in the analysis of clast lithology, and clast provenance. Clasts found within glacially or fluvially derived material can be associated with a particular bedrock outcrop or unit, and from this the ice source, direction of ice movement and former glacially influenced drainage patterns can be deduced.

The Terra Nova region is contained within the Avalon Zone, one of the four geological zones defined by Williams (1979). There are 4 main geoiogical assemblages within the Terra Nova area (Figure 1.2):

1) the Love Cove Group - deformed volcanic rocks (O'Brien 1987),

2) the Connecting Point Group - altered marine clastic sedimentary rocks (Jenness 1958),

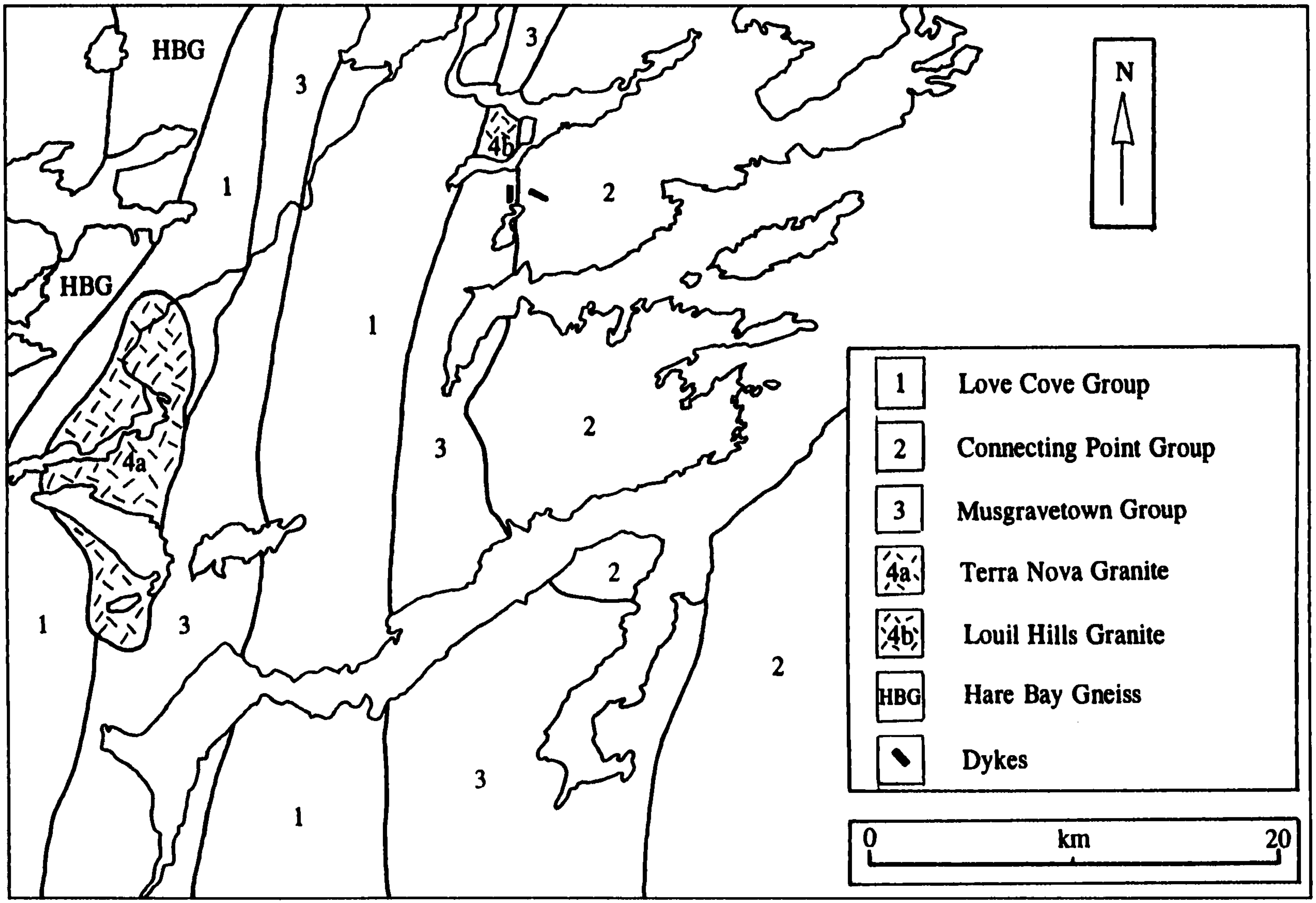

Figure 1.2 - Simplified geology map of Terra Nova National Park and vicinity (after Jenness 1963 and O'Brien 1987).

3) the Musgravetown Group - terrestrial sedimentary and volcanic rocks O'Brien 1987), and

4) igneous intrusions.

The Love Cove Group (Unit 1, Figure 1.2) consists of greyish-green or pale yellow schists. These rocks were originally conglomerates, sandstones, shales and volcanic units which have undergone low grade metamorphism (Jenness 1958; O'Brien 1987). Most of the original bedding within the rocks has been completely destroyed. Based on the degree of deformation and alteration, this group is considered to represent the oldest rock in the area (Baird 1966).

The Connecting Point Group (Unit 2, Figure 1.2) is a marine sedimentary sequence composed of shale, sandstone, and siltstone (O'Brien 1987), which weathers to light grey (Jenness 1958; Baird 1966). Three subunits are recognized within the main unit. Subunit 2a is an assemblage of tuff and black shale interbedded with sandstone and siltstone, with individual beds coarsening and thickening upward, capped by a turbidite assemblage (O'Brien 1987). Subunit 2b is interbedded black shale and siltstone, and Subunit 2c is predominantly black shale with isolated lenses of sandstone. The sandstone within Subunit 2c is cross-stratified, and pebble/granule conglomerate beds are present (O'Brien 1987).

The rocks of the Connecting Point Group are not as deformed and altered as those of the Love Cove Group, and therefore thought to be younger. They are older than the rocks of the Musgravetown Group, however, as pebbles of the Connecting Point Group are found within the Musgravetown conglomerates (Jenness 1958).

The Musgravetown Group (Unit 3, Figure 1.2) is the youngest of the three main groups and represents deposition in terrestrial and marine sedimentary environments (O'Brien 1987), with interbedded volcanic rocks (Jenness 1958). The Group is composed of coarse-grained conglomerates, sandstones, siltstones, basalts and rhyolites (O'Brien 1987). There are four formations (Subunits 3a-d) within the Musgravetown Group.

The Cannings Cove Formation (Subunit 3a) crops out at only a few sites within the Terra Nova Park region. At Ochre Hill it appears as a red/maroon pebble conglomerate which is structureless and clast supported (O'Brien 1987). The conglomerate contains pebbles from a variety of lithologies including rhyolite, shale, siltstone, sandstone and fine grained basalt, some of which were originally part of the Love Cove or Connecting Point Groups (Jenness 1963; Baird 1966). At Clode Sound the conglomerate is characterized by many clasts from the Love Cove sericite schists, but other sedimentary and volcanic clasts are present within the formation (Jenness 1963; O'Brien 1987).

The main outcrop of the Bull Arm Formation (Subunit 3b) within Terra Nova Park is along the north coast of Clode Sound (Jenness 1963). Basalt flows at the base of the unit are interbedded with red conglomerates and sandstones, which are overlain by rhyolite flows. Another basalt flow overlies the rhyolite, and this in turn is capped by a red sandstone bed (O'Brien 1987).

The Rocky Harbour Formation (Subunit 3c) crops out to the northwest and consists of green cross-bedded sandstone and conglomerate (O'Brien 1987). The conglomerate contains individual (rare) clasts of the underlying Bull Arm Formation (O'Brien 1987), and the sandstones appear to have been derived from the Love Cove schists (Jenness 1963).

The youngest formation, the Crown Hill Formation (Subunit 3d), is exposed near Ochre Hill. It consists of red/maroon sandstone, shale, pebble and granule conglomerate, and pebbly sandstone (O'Brien 1987).

A large granite intrusion, the Terra Nova Granite, is found directly to the west of the Park boundary, around Terra Nova Village (Subunit 4a, Figure 1.2), and a smaller intrusion (Subunit 4b, Figure 1.2) is located in the Louil Hills area near the northern boundary of the Park. The Terra Nova Granite (Subunits 4a and b) is part of the Ackley Batholith (Jenness 1963), and is characterized by coarse crystals of pink feldspar, quartz, plagioclase and some ferromagnesian minerals (Jenness 1958; Baird 1966). This granite is thought to be the youngest rock in the area because rocks of the Musgravetown Group have been altered by contact metamorphism where the granite intrudes into them. The granite is easily recognizable and important in the analysis of glacial dispersal of clasts.

The Louil Hills granite is similar in colour to the Terra Nova granite, however, it contains medium crystals of feldspar, quartz, and ferromagnesian minerals, and there are thin milky quartz veins throughout the intrusion (Jenness 1958). It is the only intrusion of this lithology within the area. It intrudes both the Love Cove and Connecting Point Groups, which indicates that it is younger than these groups but older than the Musgravetown Group (Baird 1966).

Gabbroic and andesitic dykes (Figure 1.2) occur throughout the Park, although they are most abundant in the Connecting Point Group. The dykes have intruded following the many structural weaknesses along the coast of Newman Sound (Jenness 1958, 1963).

1.4 - Physiography

Terra Nova National Park and its surrounding area are part of the Atlantic Upland physiographic region defined by Twenhofel and MacClintock (1940), and part of the North East Trough of Roberts (1983). It is characterized by an undulating topography of ridges and valleys that slope towards the northeast, drowned valleys, rocky peninsulas, and numerous islands (Roberts 1983). Most of the western half of the study area lies between 100 and 150 m, whereas the eastern half is characterized by heights of 120 m or less. There are several prominent hills which are distinct in the flat and generally low-lying landscape, all situated within the Park. Blue Hill is the highest at 228 m; the others, Ochre Pit Hill, Ochre Hill, Malady Head, Mount Stamford and Park Harbour Hill, have heights between 168 and 215 m.

As in much of Newfoundland, the valleys, ridges, peninsulas and bays within Terra Nova National Park have a southwest-northeast trend which reflects the underlying geology (Deichmann and Bradshaw 1984). Big Brook and Southwest Brook occupy two of the southwest-northeast trending valleys within the Park, and many of the ponds reflect this orientation, including Dunphy's Pond, Bread Cove Pond, Park Harbour Pond, and Rattle Pond. The peninsulas surrounding Lions Den are oriented in this direction, and Swale Island and Eastport Peninsula reflect the orientation of Newman Sound.

Several authors have suggested that Newfoundland's physiography is characterized by several peneplains at successively lower heights (Twenhofel and MacClintock 1940; Deichmann and Bradshaw 1984). Twenhofel and MacClintock (1940) suggested that some of the hills of Terra Nova National Park were part of a former, low peneplain that was recognized in eastern Newfoundland. Blue Hill, Ochre Hill, Gros Bog and Park Harbour Hill have similar heights, ranging from 207-228 m. These summits are flat-topped, which suggests that they may represent an erosive surface of fluvial origin. Subsequent elevation of the land surface has left the summits as remnants of a preglacial landscape, which has been incised by glacial and fluvial erosion. The coastline is characterized by deeply indented bays, rocky headlands, and numerous islands. Coastal relief is generally high and rises abruptly from the sea to a modal height of approximately 75 m within the first kilometre. Mount Stamford and Malady Head are prominent features, with near vertical cliff faces reaching heights of 195 m and 190 m, respectively. The steep sided valleys continue offshore, reaching depths of 100-450 m below sea level. Both Newman Sound and Chandler Reach have parabolic cross-sections typical of formerly glaciated valleys, and therefore can be termed fjords (Cumming 1990).

The coastline is dominated by exposed bedrock, and beaches are restricted to sheltered bays including Buckley Cove, Platter Cove, Bread Cove, Platters Beach, Dumpling Cove, Eastport Beach, and Sandy Cove. Beaches at Buckley Cove, Bread Cove, Platters Beach, Eastport Beach and Sandy Cove are composed of sand and gravel. The other beaches throughout the region are formed of pebbles or shingle, such as Platter Cove, Dumpling Cove, and those adjacent to Newman Sound. Former marine deltas to 30 m above sea level are prominent at Sandy Cove and Eastport Beach. Several brooks entering into Clode and Newman Sound are currently forming deltas, including Terra Nova Brook, Saltons Brook, and Minchin Brook. Big Brook and Southwest Brook flow through sand and gravel deposits. Their deltas or foreshore flats tend to be fine grained due to the deposition of coarser gravel farther upstream.

The study area is drained by five major rivers: Terra Nova River, Southwest Brook, Big Brook, Bread Cove Brook, and Terra Nova Brook. Most of the rivers follow the southwest-northeast orientation. Bread Cove Brook is the exception, turning south downstream of Sandy Pond, subsequently to the northeast as it flows through Bread Cove Pond, and finally flowing towards the southeast to enter the sea at Bread Cove. Terra Nova River is the largest river in the area and has a gradient of 3.18 m/km. There are three waterfalls and a series of nine rapids along its length. The majority of the falls and rapids are in the upper reaches close to Terra Nova Lake and the middle section of the river. The lower 9 km of the river has no falls and only minor rapids.

The rivers are important for land drainage. Much of Terra Nova National Park and the surrounding area, however, is poorly drained. Western areas of the Park are distinguished by bogs and fens, such as Gros Bog and Saltons Marsh. The eastern part is characterized by ponds and small bogs within bedrock depressions. There are approximately 150 ponds within the Park ranging in size from Dunphy's Pond (708.6 ha) to Pine Hill Pond (2.4 ha). Several form linear successions along valley floors, such as Dunphy's Pond, Beachy Pond, Sandy Pond and Beaver Pond, aligned southwest-northeast. Other ponds within Terra Nova National Park, and between Sandringham and Eastport, are not aligned in a southwest-northeast orientation, such as Juiceys Pond North and Goose Ponds. These areas are characterized by deranged drainage patterns that do not appear to be controlled by the underlying bedrock geology. Jenness (1958, 1963) suggested that the original drainage patterns throughout the area were altered by the deposition of glaciofluvial sediment. This caused natural river valleys to be altered, with rivers finding new routes to the sea and ponds forming in many of the depressions.

1.5 - Climate

Newfoundland's climate is influenced by the cold Labrador Current which flows south along the east coast of the island, its location in terms of the Canadian mainland, and Northern Hemisphere mid-latitude atmospheric circulation (Banfield 1993).

Terra Nova National Park is part of the "East Coast and Hinterlands Zone" defined by Banfield (1981), and has cool, late springs and generally warm and sunny summers (Banfield, 1981, p129). Climate data from Terra Nova National Park Headquarters, reported by AES (1993), shows 1100-1500 mm of precipitation per year, most of which falls throughout the period from October to March. About 75% of the yearly precipitation falls as rain. February (41.1 mm) is considerably drier than November (88.4 mm), although most of the precipitation falls as snow during the former month. The first autumn snow cover (at least 2.5 cm) usually accumulates during November, and the ground is normally snow covered until mid-April or the beginning of May (Banfield 1993). Modal annual snowfall within the Terra Nova National Park area is between 200-300 cm (Banfield 1993).

Freezing precipitation is a common phenomenon throughout Terra Nova National Park and Newfoundland (Deichmann and Bradshaw 1984). It occurs when liquid precipitation falls from a high layer of warm air (above 0 °C) into a lower layer which is at or below 0 °C (Banfield 1993). This results in the build up of 'clear ice' over all surface features. Eastern parts of Newfoundland are more likely to experience this phenomenon due to the influence of milder air off the ocean throughout mid to late winter, and it is common to have 75-150 mean annual hours of freezing rain or drizzle in these eastern areas (Banfield 1993).

The coldest month in Terra Nova National Park is February when temperatures reach a daily mean of -6.6 °C, and July is the warmest month with daily mean temperatures of 16.3 °C (Atmospheric Environment Service 1993). The temperatures experienced within an area are important for vegetation growth. If the temperatures are not high enough or there are not enough days with the required temperature, tree growth is stunted and regeneration may be limited. With only 120-140 frost-free days, this part of Newfoundland has a short growing season (Banfield 1983). Table 1.1 gives climatic data for the Terra Nova National Park region collected at the Park headquarters.

Due to the proximity of the ocean, fog is common throughout the year, becoming more frequent during the spring and summer. Sea ice usually forms in the coastal waters around the Park during the winter, and icebergs are commonly seen floating south on the cold Labrador Current between April and June.

Newfoundland experiences surface winds throughout most of the year and the Terra Nova National Park region is no exception. Winds are generally from the west and the average wind speed is between 22-26 km/hr.

Table 1.1 - Climate data collected at Terra Nova National Park headquarters and is an average of the years 1961-1990 (AES 1993). Frost data from Banfield (1993).

mean annual precipitation	1184.3 mm
mean annual rainfall	886.0 mm
mean annual actual snowfall	297.6 mm
mean February temperature	-6.6 °C
mean July temperature	16.3 °C
prevailing winds	winter W/NW, summer SW
average annual wind speed	22-26 km/hr
average date of last spring frost	June 1st-June 15th
average date of first fall frost	Sept.15th-Oct.1st

1.6 - Soils and Vegetation

Familiarity with the vegetation of Terra Nova National Park aids in the analysis of air photos. If the vegetation can be recognized, the type of soil that underlies the vegetation can be inferred, and hence the type of sediment from which the soil is formed can be proposed.

The type of soil that develops in an area depends mainly on the climate, parent material, and topography (Damman 1983; Roberts 1983; Deichmann and Bradshaw 1984; Fanning and Fanning 1989). Eastern Newfoundland is characterized by warm summers and cold winters, a variety of parent materials e.g. bedrock, diamicton, glaciofluvial sand and gravel, and an undulating topography (Damman 1983).

Deichmann and Bradshaw (1984) identified three soil zones within Terra Nova National Park (Figure 1.3). The main soil type in both Zones 1 and 2 is a Humo-Ferric Podzol which forms on coarse to medium textured glaciofluvial sediments and diamictons (Clayton *et al.* 1977; Roberts 1983). Well drained sites develop Orthic Humo-Ferric Podzols, and poorly drained areas support Gleyed Humo-Ferric Podzols. Within Zone 1 (Figure 1.3) there are pockets of Organic and Gleysolic soils which develop in areas that are poorly drained, often in depressions on the surface of outwash deposits (Deichmann and Bradshaw 1984). Organic soils are located at two main sites within Zone 1, at Gros Bog and Saltons Marsh.

Although much of Zone 2 (Figure 1.3) is dominated by Humo-Ferric Podzols, it also contains large areas of Organic soil, the second most common soil order in the Park. Organic soils are found throughout bogs and fens within the Park and are distributed throughout Zone 2.

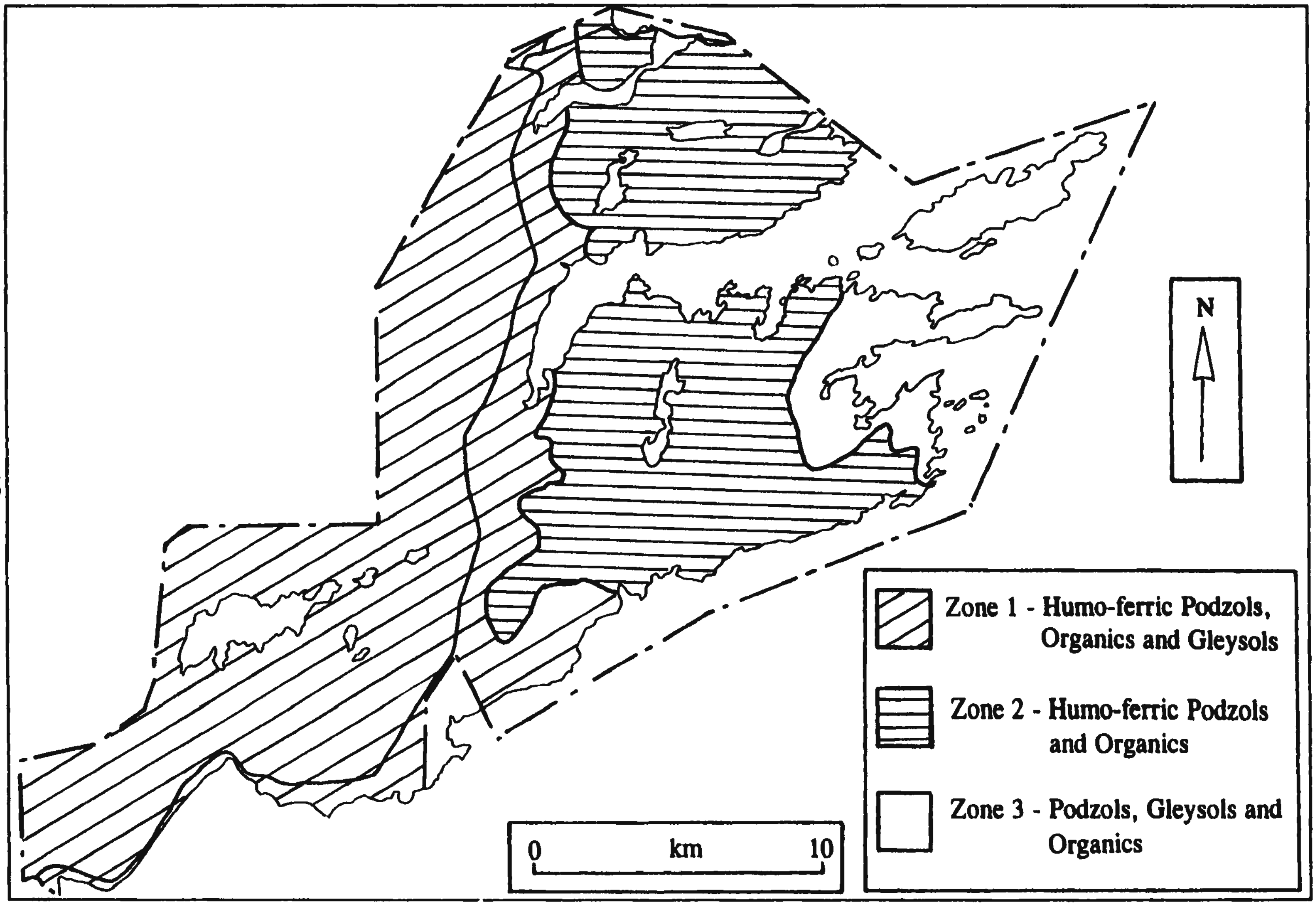

Figure 1.3 - Soil zones of Terra Nova National Park (after Deichmann and Bradshaw 1984)

Zone 3 (Figure 1.3) is characterized by poorly developed soils due to the lack of weathered parent material available for soil development (Deichmann and Bradshaw 1984). Podzols tend to dominate the area with gleysols and organic soils in the depressions.

Regosols are found in small pockets throughout the Park. They have not fully developed due to several limiting factors such as climatic extremes, youthfulness of the parent material, and soil instability (Deichmann and Bradshaw 1984; Strahler and Strahler 1992). Within Terra Nova National Park, unstable slopes and thin parent material are the main limiting factors. They are immature soils which form in areas of shallow diamicton or on scree slopes that tend to be unstable.

Approximately 70% of Terra Nova National Park's area is covered by forest vegetation, and the remainder consists of bog, fen and open water. Three broad vegetation zones are recognized within the Park by Deichmann and Bradshaw (1984), and they appear to be related to the soil zones described above.

Zone 1 encompasses the area west of the Trans Canada Highway. There are dense stands of black spruce (*Picea mariana*) and balsam fir (*Abies balsamea*), and these alternate in dominance depending on the site, microclimate, drainage and fire history. Zone 2 lies east of the Trans Canada Highway. The black spruce forest tends to be more open in this zone, and there is a thick understorey of shrubs and ericades including sheep laurel (*Kalmia angustifolia*), labrador tea (*Ledum groenlandicum*), and mountain alder (*Alnus crispa*). There are dense stands of black spruce and balsam fir in limited areas. The final zone corresponds to Zone 3 of the soils, where rock outcrops are dominant and the vegetation grows on the weathered material (Deichmann and Bradshaw 1984). Small shrubs grow close to the sea where there is a thin cover of sediment covering the bedrock, and in areas where there is a substantial supply of sediment, bushes and trees grow close to the water front.

Approximately 95% of the forests are coniferous, composed of 80% black spruce and 15% balsam fir (Robinson 1989). The remaining 5% of forest is deciduous with white birch (*Betula papyrifera*) and trembling aspen (*Populus tremuloides*) being the dominant species, as well as some larch (*Larix laricina*) and red maple (*Acer rubrum*) scattered throughout the forest (Robinson 1989).

A total of 21% of the Park was classified as wetlands by Deichmann and Bradshaw (1984). Raised bogs such as Gros Bog and Saltons Marsh are the most common type of wetland (Deichmann and Bradshaw 1984). They are treeless and are formed from an accumulation of Sphagnum mosses. The surfaces of the bogs are raised above the surrounding land by about 2 m (Wells and Pollett 1983). Most of the bogs are surrounded by fens which form from the accumulation of grasses and sedges (Wells and Pollett 1983). The water seeps from the bogs into the fens, controlling growth and development (Wells and Pollett 1983; Deichmann and Bradshaw 1984).

Chapter 2

Previous Work

2.0 - Initial Research

Offshore work during the past 20 years has indicated that the Island of Newfoundland may have been glaciated several times during the Wisconsinan (Piper *et al.* 1978; King and Fader 1986; Grant 1989). Research by MacClintock and Twenhofel (1940) found no deposits older than Wisconsinan, and they therefore assumed that all glacial features were of Wisconsinan age. Early workers such as Fairchild (1918), Daly (1921) and Coleman (1926) believed that Newfoundland supported its own ice cap, and suggested a pre-Late Wisconsinan age due to the extent of weathering on glaciated summits (Coleman 1926). Fernald (1911) identified plant species on the summits that were found elsewhere only in the western Cordillera of North America. He proposed that these species could only have survived in glacial refugia, and he therefore suggested that during the Late Wisconsinan glaciation, nunataks protruded above the ice surface. Other researchers such as Flint (1940) and Tanner (1940) argued that Newfoundland was completely overrun by ice from Labrador. Citing evidence from the west and southwest coasts of Newfoundland, Flint (1940) concluded that Labrador ice had influenced Newfoundland throughout the last glaciation. Marine features, such as wave-cut platforms and benches, rise in elevation towards the northwest, suggesting that the ice which influenced Newfoundland thickened in this direction.

Daly (1921) was of the opinion that Newfoundland supported its own ice cap or ice caps during the Late Wisconsinan. MacClintock and Twenhofel (1940) believed that Newfoundland was completely covered by an ice cap during the Wisconsinan. As the ice cap retreated, the Island supported its own ice caps centred on the Avalon Peninsula, Annieopsquotch Mountains and the Red Indian Lake area. Daly's work was corroborated by Murray (1955) and Jenness (1960) who identified several ice caps in south-central and eastern Newfoundland, respectively, based on glacial features which spread radially towards the coast.

Jenness (1960) thought that eastern Newfoundland could be divided into an "outer drift zone" and an "inner drift zone", based on the geomorphology and sedimentology of the area. The two zones were separated by a discontinuous end moraine that lies several kilometres landward of the eastern Newfoundland coast. The outer drift zone was characterized by fjords, cirques, thin till cover of local origin, kames, and outwash that terminated as deltas at the coast. In comparison, the inner drift zone was characterized by thicker till cover of local origin, moraine lakes, fluted terrain, and eskers. The eskers were aligned parallel with glacial striations, and they both indicated that ice movement within the Terra Nova National Park region was towards the northeast and the coast. Jenness did not, however, precisely locate the ice centre within central Newfoundland.

Jenness (1960) also discussed post-glacial coastal movements as a result of glacial isostasy. He compared the data that he collected from raised deltas in eastern Newfoundland with that of Flint (1940). The map Jenness (1960) produced showed that the southern part of the Burin Peninsula and the Avalon Peninsula may be submerging at present. The remainder of the Island, north of the zero isobase, was considered to be emerging towards the northwest.

2.1 - Sources of Glacial Ice and Ice Flow Directions

The majority of workers since Flint (1940) and Tanner (1940) have indicated that Newfoundland supported its own ice cap or caps throughout the Late Wisconsinan (Murray 1955; Jenness 1960; Lundqvist 1965). Grant (1969a, 1989, 1992) studied the glacial landforms and deposits on the extreme northern end of the Northern Peninsula. Streamlined roches moutonnées and Labradorean erratics indicated that Laurentide Ice invaded northern Newfoundland during the Late Wisconsinan. Subsequent retreat of the Laurentide Ice Sheet, however, allowed ice from a separate ice cap on the Long Range plateau to readvance onto the lowlands (Grant 1969a, 1969b, 1992).

Grant (1974, 1977, 1989) discussed the theory of multiple ice caps throughout Newfoundland. Eskers, kames and meltwater channels throughout Newfoundland indicate that there was an ice divide stretching across the island from Port aux Basques to the Avalon Peninsula. The ice divide is thought to have formed by the coalescence of several ice caps within central Newfoundland located at Red Indian Lake, Maelpaeg Lake and Middle Ridge (Rogerson 1982). Ice caps located on many of the peninsulas (e.g. Burin Peninsula, Avalon Peninsula, Bonavista Peninsula) seaward of the ice divide separated from the main ice cap as the ice started to retreat (Grant 1974). Ice sources throughout Newfoundland have been reconstructed using striations and the provenance of clasts within the glacial deposits (e.g. Jenness 1960; Brookes 1989; Liverman and Scott 1990; Grant 1992; Mackenzie and Catto 1993; Klassen 1994). The ice source areas are characterized by thick till which has been streamlined into flutings and drumlins (Grant 1975).

Recent Quaternary investigations in Newfoundland have concentrated on specific areas of the Island (Vanderveer and Taylor 1987; Liverman and St. Croix 1989; Liverman and Scott 1990; St. Croix and Taylor 1990; Liverman *et al.* 1991; Mackenzie and Catto 1993; Munro and Catto 1993; Scott 1993; Batterson *In preparation*). Several ice flow directions were recognized by St. Croix and Taylor (1991) and Liverman (1992) in north-central Newfoundland. The oldest ice flow direction, also recognized by Klassen (1994), was towards the east along Notre Dame Bay. Striations in the Baie Verte-Notre Dame Bay area suggest that the ice source may have been on the Long Range Mountains. The second event was characterized by radially flowing ice, with the source located north of Red Indian Lake (Grant 1975; St. Croix and Taylor 1990). This ice flow was recognized by Jenness (1960) and Brookes (1989) in the Bonavista Bay area. Jenness (1960) deduced from striations, indicator boulders, and glaciated ridges that within the vicinity of Terra Nova National Park, ice moved eastwards away from central Newfoundland towards the coast. The Bonavista Peninsula was characterized by eastward flowing ice from the main Newfoundland Ice Cap, and an ice divide that was located over eastern Bonavista Peninsula (Brookes 1989). Tills on the eastern part of the Bonavista Peninsula contain no erratics from central Newfoundland, indicating that ice from the main Newfoundland Ice Cap did not pass over this area (Brookes 1989). The final ice flow movement recognized by St. Croix and Taylor (1990) was north-northeastward, from a source north of Maelpeag Lake.

2.2 - Extent of Glacial Ice in Newfoundland

The extent of glacial ice over Newfoundland has been debated for many years. Fernald (1911) and Coleman (1926) concluded that the extent of glaciation on Newfoundland was limited throughout the Late Wisconsinan. In comparison, Flint (1940, 1971) argued that Newfoundland was completely covered by ice which extended onto the Grand Banks, seaward to the -200 m isobath during the Late Wisconsinan. A similar view was held by MacClintock and Twenhofel (1940) who suggested that Late Wisconsinan ice initially completely covered Newfoundland and the Grand Banks, but as the ice sheet started to retreat the Island was influenced by local ice caps.

Grant (1969b, 1977) and Brookes (1977) believed that there was a limited ice extent on the west coast of Newfoundland throughout the Late Wisconsinan, based on the identification of three weathering zones within the Long Range Mountains. The oldest zone includes the weathered summits identified by Fernald (1911) and Coleman (1926), and was thought by Grant (1969b, 1977) and Brookes (1977) to have been ice free since the last interglacial. The intermediate zone has clearly been glaciated, although the erratics and glacial erosional features are weathered. Grant (1977) suggested that these features can be assigned to the Early Wisconsinan. The youngest zone retains its freshly glaciated character, and can be traced along the fjord walls until it blends with end moraines on the coastal plain. Nunataks protruded above the ice surface and piedmont glaciers advanced seaward through the troughs onto the coastal plain. Grant (1977) suggested that the Late Wisconsinan ice barely reached the present coast, and was limited to areas currently less than 100 m deep offshore. Research on the Buchans Plateau and the Topsail Hills by Grant (1975) and Grant and Tucker (1976) identified similar weathering zones, with weathered summits and fresh till in the valleys.

The extent of Late Wisconsinan ice on the Burin Peninsula was studied by Tucker and McCann (1980). Throughout the Wisconsinan, the Burin Peninsula was influenced by several ice flow events. During the Early Wisconsinan, ice flowed southwards from the main Newfoundland Ice Cap and covered the entire area. Subsequent retreat of this ice allowed ice which had built up in Placentia Bay to advance onshore and glaciate the southern part of the Peninsula. The final ice flow event during the Late Wisconsinan involved southward moving ice from the main Newfoundland Ice Cap. This event, however, only affected the northern part of the Peninsula, leaving the southern Burin Peninsula ice free.

Recent research in the Long Range Mountains by Gosse *et al.* (1993, 1995), using cosmogenic ^{10}Be and ^{26}Al dating on protruding quartz veins and erratics, attempted to establish the exposure history of the area. The exposure ages of two granite erratics, one from Zone C (Grant 1977) and the other from Zone B, are 18.4 ± 0.7 and 18.3 ± 0.7 ^{10}Be ka years BP respectively. The ages establish the timing of deglaciation in the area, and show that no nunataks existed throughout the last glaciation. Gosse *et al.* (1993, 1995) suggested that the weathered appearance of these higher summits was a result of the presence of thin, cold-based ice throughout the Late Wisconsinan preventing erosion of the underlying surface.

Grant (1989) proposed two models based on the extent of glacial ice over Newfoundland. The "maximum" model is characterized by the entire island and adjacent shelf being covered with ice, with no nunataks in the upland areas. In comparison the "minimum" model identifies several ice centres, nunataks in the west coast uplands, and ice free areas on the Burin, Avalon and Bonavista Peninsulas.

2.3 - Extent of Glacial Ice in Eastern Newfoundland

Relatively little work has been done in Eastern Newfoundland regarding the Late Wisconsinan ice limit in the area (Jenness 1960; Dyke 1972; Cumming 1990; Cumming *et al.* 1992). Jenness (1960) placed the limit of the last significant stand of ice in eastern Newfoundland several kilometres landward of the present coastline, although initially the ice did reach the modern coast. Outwash trains spread radially from the end moraine, and the absence of eskers within the outer drift zone verified the stillstand position of the glacier at this point (Jenness 1960). Dyke (1972) studied the morphology and stratigraphy of the Eastport delta system, and considered its position in relation to the ice limit proposed by Jenness (1960).

The topography between Sandringham and Eastport rises to a height of approximately 55 m above sea level (Dyke 1972). Meltwater from the ice margin proposed by Jenness (1960) would have to flow up the ridge between Sandringham and Eastport before it could continue towards the coast. Dyke (1972) argued that unless the river could flow up steep gradients, or unless large amounts of sediment had been subsequently eroded, the former ice margin was more likely to have been located at the height of land between Sandringham and Eastport. Kettle holes on the delta surface at Eastport suggested that the ice front stood adjacent to the delta during its formation (Dyke 1972). As the ice front retreated westwards, an ice-dammed lake developed between the ice front and the higher land. Dyke (1972) identified lacustrine sediments which were overlain by a thin mantle of diamicton, which he attributed to a small readvance at the ice front.

Surficial sediments, core data and foraminiferal analysis from the continental shelf surrounding Newfoundland have shown that the ice limit during the Late Wisconsinan extended out onto the shelf (Piper *et al.* 1978; King and Fader 1986; Cumming 1990; Scott *et al.* 1991; Cumming *et al.* 1992; Jenner and Shaw 1992; Shaw and Edwardson 1994; Shaw and Forbes 1995).

Sediment cores from the inner continental shelf off northeast Newfoundland indicate that much of the area is covered by bouldery gravel and mud, that was interpreted as glacial diamicton (Jenner and Shaw 1992; Shaw and Edwardson 1994). Sidescan sonar images taken at the mouth of the Gander River show small elliptical mounds which rise up to 15 m above the surrounding seabed (Shaw and Edwardson 1994). These features were interpreted as drumlins formed as the ice moved towards the ENE. This ice flow direction is similar to the on-shore striations and drumlins mapped by St. Croix and Taylor (1991) and Munro (1994), respectively.

Research on the Grand Banks of Newfoundland by King and Fader (1986) showed that glaciers extended out onto the shelf throughout the Wisconsinan. Diamicton, interpreted as till, covered the entire area and extended to the shelf edge. King and Fader (1986) suggested that due to the lack of Early Wisconsinan sediments the till was deposited during the Middle Wisconsinan. In most places the till is overlain by silt that was deposited by meltout beneath the ice shelves as the ice sheet retreated. A readvance during the Late Wisconsinan extended seawards onto the shelf, however the extent of grounded ice was limited. Iceberg furrows carved into glaciomarine sediments have been identified from sidescan sonograms and these were formed by calving ice as the ice retreated onto the land at the end of the last glaciation.

Cumming *et al.* (1992) discussed the glacial and sedimentary history of Bonavista Bay, using high-resolution seismic profiles and core data. They suggested that the Quaternary deposits indicated that during the Late Wisconsinan maximum, at about 20,000 years BP, glaciers advanced into Bonavista Bay from the southwest. The ice was grounded and till was deposited throughout the Bay. There are no major end moraines within the area, and therefore Cumming *et al.* (1992) concluded that the development of an ice shelf, which deposited a thick diamicton from meltout, was fairly rapid and occurred between 20,000 and 14,000 years BP. The ice continued to retreat from the Central Basin of Bonavista Bay towards Clode and Newman Sounds reaching the present coastline about 13,000 years BP. Morainal features formed wherever the ice stagnated and proglacial sediments were rapidly deposited. The inner part of Bonavista Bay continued to be influenced by ice on the Bonavista and Gander Peninsulas (Brookes 1989 and Grant 1989, respectively) until approximately 10,000 years BP.

The offshore evidence suggests that during the Late Wisconsinan maximum, grounded ice from central Newfoundland spread out over the adjacent shelf to an unknown seaward limit (Cumming *et al.* 1992). The Late Wisconsinan ice limit proposed by Jenness (1960) is considered by Cumming *et al.* (1992) to delineate the ice margin at some time between 13,000 and 10,000 years BP.

2.4 - Post Glacial Events

2.4.1 - The Younger Dryas cooling event

Palynological analysis of sediment cores taken from lakes throughout Newfoundland has indicated that a cooling event occurred after the initial deglaciation (Anderson 1983; Macpherson and Anderson 1985; Anderson and Macpherson 1994; Wolfe and Butler 1994). Examination of lake cores from the southern Burin Peninsula and Leading Tickles, northeast Newfoundland, showed that organic sedimentation within both lakes was interrupted by deposition of a layer of inorganic silty clay (Macpherson and Anderson 1985). The number of pollen grains found within the silty clay layer differed from the underlying and overlying organic layers. The assemblage of pollen grains within the silty clay layer represented a vegetation typical of a cooler tundra climate, compared to the warmer low arctic pollen assemblages found within the organic beds. ^{14}C dating of the organic sediments indicated that the silty clay layers formed between 11.3 and 10.7 ka BP (Anderson 1983).

Wolfe and Butler (1994) obtained similar results from Pine Hill Pond, Terra Nova National Park. Organic sedimentation was interrupted by an influx of inorganic sediments, deposited between 12,400±150 years BP (GSC-5335) and 11,400±160 years BP (GSC-5334). The palynological assemblage within this layer is considerably different to those of the enclosing organic layers and represents renewed mineral sedimentation rather than redeposition. A reversion to mineral sedimentation following initial deglaciation occurs only once within these cores, and the stratum is overlain by an uninterrupted sequence of increasingly organic rich sediment deposited throughout the early Holocene (Wolfe and Butler 1994).

Deposition of the silty clay layers within these ponds was correlated with the European Younger Dryas (11-10 ka years BP) by Anderson and Macpherson (1994) and Wolfe and Butler (1994). The Younger Dryas has been recognized in Atlantic Canada by Stea and Mott (1989), Mayle and Cwynar (1991), Mott and Stea (1993), Cwynar *et al.* (1994) and Peteet (1995).

Geomorphic evidence of glacial readvance during the Younger Dryas has also been recognized by Grant (1969a, 1969b, 1989, 1992). After the retreat of the Laurentide Ice Sheet there was a readvance of ice from the Long Range plateau. Marine sediments were ploughed up in front of the ice as it moved into the sea, forming the Ten Mile Lake Moraine. Shells within the moraine have been ^{14}C dated to 10,900 years BP (GSC-1277; Grant 1969a, 1992). The readvance may have been the result of calving causing the glacier to surge. However, the dates obtained from the shells within the moraine suggest that the underlying cause was climate cooling (Grant 1992).

Ice wedge casts identified by Brookes (1971), Eyles (1977), Tucker (1979), Liverman and St. Croix (1989), Liverman *et al.* (1991), Scott (1993) and Batterson (*In preparation*) may be further evidence of a Younger Dryas cooling event. The majority of the ice wedge casts were found in well-sorted sand and gravel deposits that were associated with coastal marine deposition. Brookes (1971) postulated that the ice wedge casts identified on the southwest coast were formed between 12,500 and 11,200 years BP. Eyles (1977) tentatively suggested that ice wedge casts in north-central Newfoundland were formed between 12,000 BP and 10,000 BP. Numerical dates for ice wedge formation cannot be determined in all instances due to the lack of dateable material. Liverman and St. Croix (1989), however, suggest that the ice wedges identified on the Baie Verte Peninsula were formed between 12 and 10 ka BP, after initial deglaciation but before ice had completely disappeared from Newfoundland.

2.4.2 - Relative Sea Level

Relative sea level along the coast of Newfoundland is directly related to the Laurentide Ice Sheet and its influence on the land adjacent to the ice margin (Grant 1969a, 1969b; Liverman 1994). Raised marine features throughout Newfoundland show a progressive increase in elevation towards the northwest, reflecting the influence of the Laurentide Ice Sheet. The maximum recorded marine limit on the Bonavista Peninsula is 15 m asl (Brookes 1989), and in Bonavista Bay it is 38.5 m (Dyke 1972). At Springdale it is 75 m (Tucker 1974), Salmon Point 43 m (Liverman and Batterson 1995), Port au Choix 140 m (Grant 1994), and on the Northern Peninsula in the Bellburns area it is 145 m asl (Proudfoot and St. Croix 1987).

Eighteen raised marine features were identified in the study area by Jenness (1960), Dyke (1972), Deichmann and Bradshaw (1984) and Brookes (1989) (Figure 2.1). The raised deltas occur at elevations between 15 and 39 m asl, and were described as part of several outwash systems that emptied into Newman and Clode Sounds.

Jenness (1960) identified and described raised deltas and their sediments in northeastern Newfoundland. The majority of the delta tops were at approximately 30 m asl, and Jenness (1960) recognized bottomsets, foresets and topsets in many of the sediments, notably those at Eastport and Sandy Cove. The bottomsets were composed of horizontally bedded clay, silt and fine sand, and these were overlain by seaward-dipping sand that constituted the foreset beds. The grain size coarsened upwards and the topsets of the delta were composed of nearly horizontal, partly stratified pebbly gravel. Two lower deltaic deposits, tops at 15 m asl, were also recognized by Jenness (1960) at Eastport and Port Blandford. The sediments at Port Blandford were characterized by bottomsets of clay, silt and sand and seaward-dipping foresets of sand that coarsened upwards.

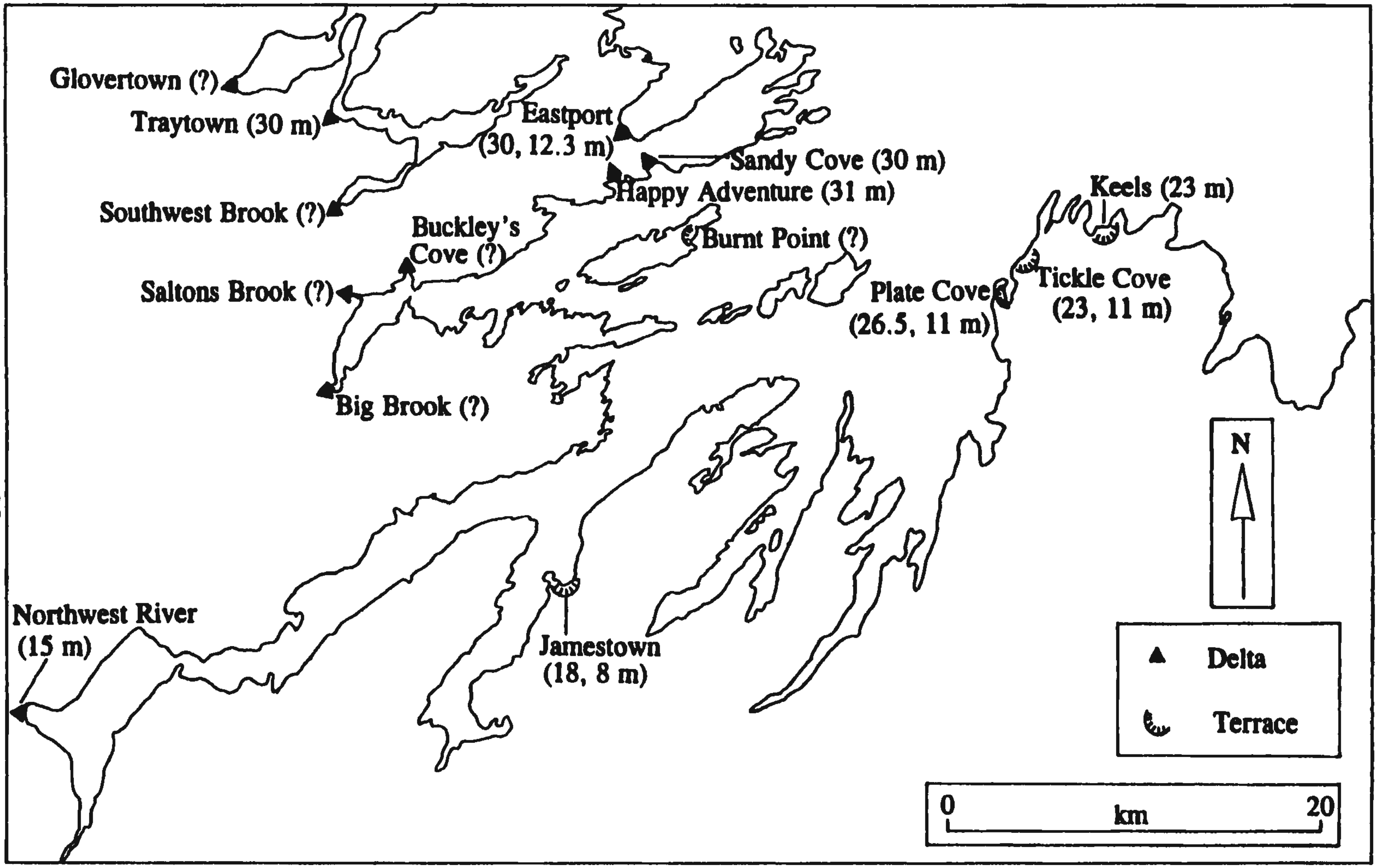

Figure 2.1 - Raised marine features throughout northeast Newfoundland measured by Jenness (1960), Dyke (1972), Deichmann and Bradshaw (1984), and Brookes (1989).

Dyke (1972) described the morphology and stratigraphy of the Eastport and Sandy Cove delta system. Within many vertical sections he recognized bottomsets, foresets and topsets, having a composition similar to that described by Jenness (1960). Dyke (1972) also described the lower deltaic deposits identified by Jenness (1960) at Eastport. The sediments at the southeast end of Eastport beach consisted of nearly horizontally bedded clay, silt and fine sand, whereas those at the northwest end were composed of silt and fine sand. Dyke (1972) suggested that the finer sediments were deposited farther out to sea, and that all the sediments were deposited contemporaneously when sea-level stood at least 30 m higher than present.

Raised beaches were also identified within the area by Deichmann and Bradshaw (1984) and Brookes (1989), and are shown in Figure 2.1. Brookes (1989) measured three levels of terraces at 23-26.5 m, 11 m and 6 m asl, all located close to sediment sources and along the indented coastline.

Quinlan and Beaumont (1981, 1982) developed two geophysical models concerning post-glacial sea levels in Atlantic Canada. The sea level curves produced were based upon the maximum and minimum ice conditions proposed by Flint (1971) and Grant (1980). Regardless of the ice extent over Newfoundland, the Earth's crust below the Laurentide Ice Sheet was isostatically depressed and forced out from underneath the ice sheet producing a forebulge beyond the ice margins (Walcott 1970). As the ice sheet started to melt at the end of the last glaciation the land beneath the ice sheet began to isostatically rebound and the forebulge migrated towards the ice centre. Quinlan and Beaumont (1981) developed four sea level curves that show how the relative sea level changed as the forebulge migrated past several points (Figure 2.2). In the maximum ice model Terra Nova National Park is located in Zone C, which is characterized by an initial emergence followed by a continued and greater submergence. In the minimum ice model the Park is located in Zone D where there has been no emergence, and the land has continually been submerging. They suggest that there is a lack of raised marine features within Zone C, and that none are expected within Zone D. This is a direct contradiction to the geomorphological evidence of raised features throughout northeast Newfoundland described by Jenness (1960), Dyke (1972) and Brookes (1989).

Shaw and Forbes (1990) developed a sea-level curve for the Cape Freels area using in part Grant's (1980) data from the Gander Bay region (Figure 2.3). They suggested that the relative sea level fell from a late-glacial marine limit of +40 m (Grant 1980) to a low of perhaps -25 m between 12,000 and 8,000 years BP. Local sea level has been rising slowly since. ^{14}C dating of organic deposits close to today's shoreline demonstrates the transgressive nature of the shoreline in this area. Organic deposits collected at -3.30 m and -0.15 m below sea level, were dated to 5,490±120 years BP (Beta-27231) and 2980±90 years BP (GSC-4592), respectively. Stabilization of the coastline after this time was demonstrated by the development of sedge peat 0.67 m asl at Deadman's Bay. The peat was dated to 1260±70 years BP (Beta-27234). Shaw and Forbes (1990) suggest that the local sea level has changed

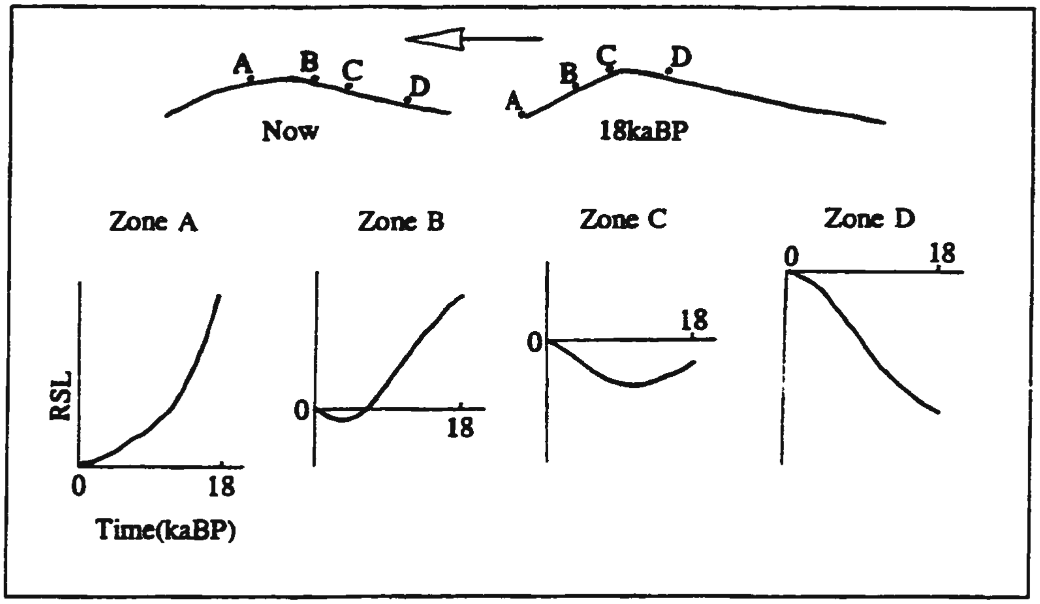

Figure 2.2 - Schematic representation of the migration of the peripheral bulge at two points in time. The bulge migrates in the direction of the arrow. The sea level curves show the relative sea level history at sites A, B, C, and D (after Quinlan and Beaumont 1981).

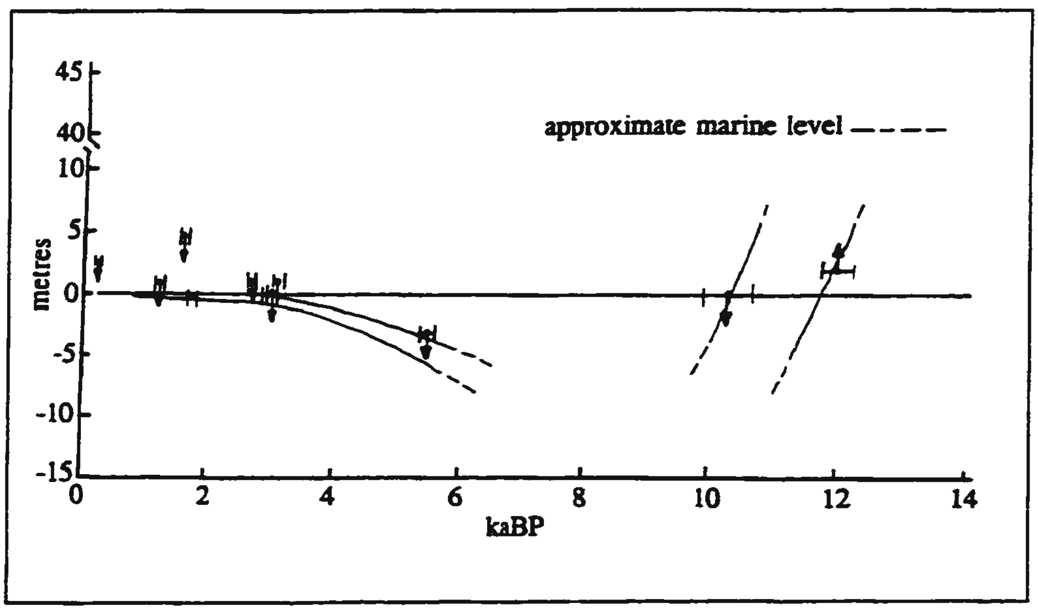

Figure 2.3 - Sea level curve developed for northeast Newfoundland (after Shaw and Forbes 1990).

relatively little since about 2,000 years BP. The sea level curve produced by Shaw and Forbes (1990) for the Cape Freels area is typical of the Zone B curve proposed by Quinlan and Beaumont (1981).

Liverman (1994) discussed the relative sea level history of Newfoundland using ^{14}C shell dates and geomorphological evidence. Outside of the northern part of the Northern Peninsula no shells dating younger than 8,000 years BP have been found. This was interpreted as indicating that for the past 8,000 years sea level has been lower than the present position around most of the Island. In many of these same areas there is geomorphological evidence for both higher and lower sea level stands (Shaw and Forbes 1995). Liverman (1994) analysed the sea level models of Quinlan and Beaumont (1981) using the ^{14}C shell dates and geomorphological evidence, and suggested that Zone B, characterized by initial emergence followed by subsequent submergence, was far more extensive than previously thought. As designated by Liverman (1994), Zone B includes Terra Nova National Park and vicinity, as shown in Figure 2.4.

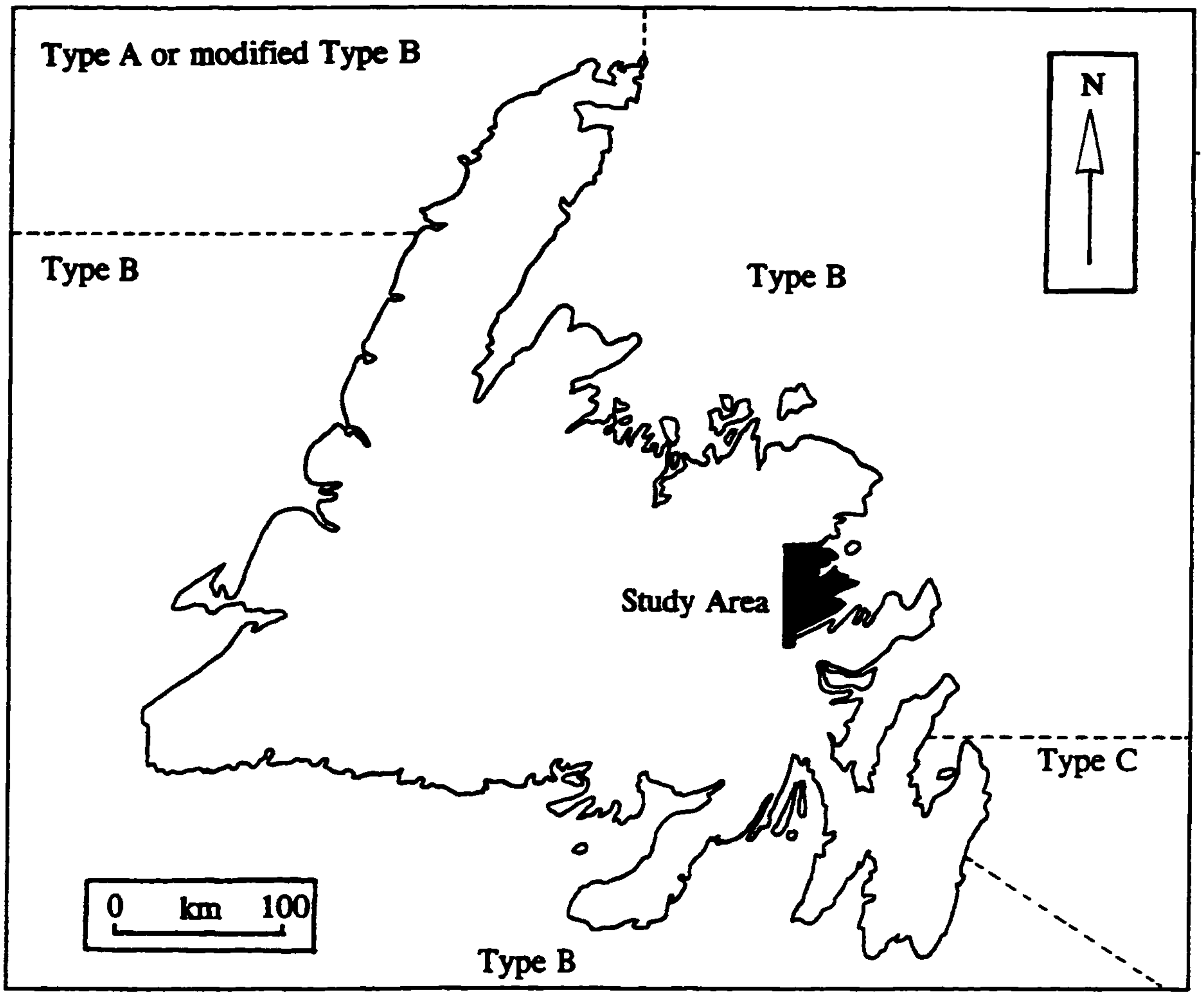

Figure 2.4 - Relative sea level zones of Newfoundland. Terra Nova National Park is placed in Zone B, and this is characterized by initial emergence followed by subsequent submergence (after Liverman 1994)

Chapter 3

Methodology

3.0 - Field Methods

Preliminary fieldwork was undertaken at the beginning of July 1995, and all roads were accessed by vehicle or foot. This reconnaissance allowed natural exposed coastal sections and artificial exposures such as road cuts and gravel pits to be identified, and brief descriptions of the sites were made. Detailed fieldwork was undertaken during July 1995 and additional visits were made to the study area in August 1996 and September 1996. A given exposure was described in detail as a whole if the section was small and composed of only a few units. Larger exposures, such as those in gravel pits, were divided into several vertical transects and each was described in detail. The sections were divided into appropriate units for description and analysis. For each unit the thickness, colour, contacts, texture, sedimentary structures, clast lithology and clast shape were described and recorded. Clast fabric analysis was undertaken on diamicton units and coarse clast-supported gravels. Fabric analysis involved measuring the orientation and dip of the a-axis of at least 25 clasts using a Brunton compass. Clasts with an a-axis to b-axis ratio of 3:2, or greater, were selected for measurement (Catto *et al.* 1989).

Linear landforms were identified throughout the area and their orientations were measured with a Brunton compass. Due to access difficulties within Terra Nova National Park, orientations of roches moutonnées (stoss and lee forms), crag and tails and drumlins were measured from aerial photographs. Bedrock outcrops encountered at the roadside, at the top of several hills within the Park, and at the coast were examined for striations. Between 10 and 30 striations were measured at each site, and nailhead striations were noted (cf. St. Croix and Taylor 1990, 1991).

Heights of marine sedimentary exposures and marker beds within selected sections were measured using an altimeter. Readings were taken throughout July 1995, and the altimeter was sheltered before taking readings on windier days to avoid any inaccuracies. All elevations measured were taken from mean sea level, and altimeter readings were taken at sea level before and after readings on land to ensure that results were as accurate as possible.

3.1 - Laboratory Methods

3.1.1 - Grain Size Analysis

Grain size distributions were described using the Udden-Wentworth grade scale (Udden 1898; Wentworth 1922), and the Phi (ϕ) scale proposed by Krumbein (1934). Both scales are logarithmic (to the base 2), and each grade limit is twice the diameter of the limit directly below it. The grades are related to the grain diameter in millimetres. Krumbein's phi scale is the negative logarithm to the base 2 of the diameter, measured in millimetres.

Grain size analysis was undertaken for units that contained a fine-grained matrix. Sieves and hydrometers were used during the analysis. The sediment was first sieved through a No. 5 (-2ϕ) sieve. Samples that were mostly sand and granules, and had less than 10 g of material left in the sieve pan ($<4\phi$), were subjected only to sieving. Sieve sizes used in the analysis ranged from a No. 5 (-2ϕ) to a No. 230 (4ϕ). Samples that had more than 10 g of sieved sediment were analysed by suspension settling using a hydrometer. The hydrometer method used is that described by ASTM (1964) and Catto *et al.* (1989).

Suspension settling analysis within a cylinder is based on the theory that larger, heavier particles settle out of the water column faster than smaller, lighter particles. The rate of grains settling out of suspension is quantified by Stokes' Law which states that a spherical particle settling at a uniform velocity will encounter the resisting force of the liquid. If the settling velocity can be determined then the particle diameter can be calculated. Hydrometers measure the density changes within the liquid as suspension settling occurs (Lindholm 1987). These density changes are directly related to the size of clasts still in suspension.

Each sample that had more than 10 g of sediment remaining in the sieve pan was soaked overnight in a beaker containing 125 ml of 4% sodium hexametaphosphate (Calgon) solution that dispersed the sediment. The sample and solution were thoroughly mixed in a blender and then transferred to a 1000 ml graduated cylinder that was filled to the 1000 ml mark with distilled water. The cylinder was then agitated by hand until all of the sediment was in suspension, and the cylinder was then placed on a flat surface. A hydrometer was added to the cylinder and readings were taken at 15 and 30 seconds; 1, 2, 5, 10, 15, 30, 60, 90, 120 minutes and 4, 8 and 24 hours.

At each time interval, temperature and hydrometer readings were recorded from the test and control cylinders. The corrected hydrometer reading (R) was calculated by subtracting the control reading from the test reading. The effective depth (L), composite correction (K) and specific gravity factor (a), were obtained from predetermined standard tables (ASTM 1964). These parameters allow the particle diameter in suspension to be calculated from the formula:

$$d=K(L/t)^{1/2}$$

and the percentage of material still in suspension can be calculated using the formula:

$$\%=(aR/\text{mass of sample})\times100\%$$

A cumulative curve was plotted for each sample on log-probability paper (Folk 1966). The particle diameter was plotted on the x-axis in millimetres, and the percentage finer than a given particle diameter on the y-axis. Both the sieve data and the hydrometer data are plotted on the same graph to account for the entire matrix texture. There is often a gap between the two data sets due to the behaviour of very fine sand in a sieve and in the hydrometer, and mechanical errors brought about by measurement during the initial stages of hydrometer analysis. This is significant when dealing with clay and silt-rich sediments, however the sediments tested were very sandy, and therefore there was little difference between the combined percentages of fines as determined through sieving and hydrometry.

3.1.2 - Clast Analysis

Boulton (1978) and Dowdeswell *et al.* (1985) suggested that clast shape can be important for the interpretation of sediments. Shape is partly controlled by the mineralogy of the clasts (Drake 1970). It has, however, been recognized that specific environments can impart a certain amount and style of abrasion on the clasts during transport (Boulton 1978; Dowdeswell *et al.* 1985). Clast shapes were determined using the classification of Zingg (1935) that grouped clasts into blades, rollers, discs and spheres. Roundness was assessed using the classification scheme of Folk (1955) that placed clasts into groups between 0 (very angular) and 6 (well rounded).

3.2 - Clast Fabric Analysis

Clast fabric analysis involves the measurement of the dip and orientation of the long axis (*ab* plane) of 25 clasts within a unit. Field measurements were analysed using a Stereo™ program designed for the Macintosh computer by MacEachern (1989). The program plots the 25 clast orientations and dips onto an equal area projection net, and this allows three-dimensional data to be plotted in two dimensions.

The mean orientation and strength of the fabrics were calculated using the eigenvalue method of Woodcock (1977). This involves determining the strength and orientation of the principal normalized eigenvalue (S_1). Every observation is regarded as a unit vector. There are three vectors calculated that are oriented at 90° to each other in three dimensions (v_1, v_2, v_3), and these are referred to as eigenvectors. The eigenvector v_1 is considered as the direction about which the "moment of inertia" of the distribution is minimized (Watson 1966) and refers to the direction of maximum clustering (Dowdeswell and Sharp 1986). Eigenvector v_3 is associated with the largest moment of inertia and the direction of minimum clustering. Each eigenvector has a corresponding eigenvalue that is calculated by dividing the strength of each by the total strength of all three. This produces the normalized eigenvalues S_1, S_2 and S_3, and the sum of all three equals 1.

The principal normalized eigenvalue (S_1) measures the degree to which the clasts are aligned. A fabric pattern with a large S_1 value close to 1, and low S_2 and S_3 values, indicates that most of the clasts are aligned in the same direction. When all three eigenvalues have similar values the fabric is random and there is no preferred orientation of the clasts.

The K value is calculated by dividing the natural logarithm of (S_1/S_2) by the natural logarithm of (S_2/S_3), and describes the tendency for the fabric to show a girdle or cluster pattern. Where K is equal to 1 the distributions have equal girdle and cluster tendencies. Where $0 \leq K < 1$, the fabric patterns are represented by girdle plots, and unimodal fabric patterns are represented by cluster plots where $K > 1$ (Woodcock 1977).

3.3 - Aerial Photograph Interpretation

Aerial photograph interpretation was undertaken on vertical black and white photographs taken in 1966. Due to access difficulty within Terra Nova National Park, aerial photograph interpretation was important for identifying geomorphological features and interpreting the surficial geology.

During the field season (July 1995) ground-truthing of areas that were accessible by vehicle or foot were identified on the aerial photographs, and the surficial geology of these areas was verified. The tone of the areas on the aerial photographs allowed the surficial geology of inaccessible areas to be determined. Terrain that reflects little light, for example water or fine sediment, appears dark grey on the photographs. Well-drained coarse sediments and bedrock are highly reflective and appear light grey or white. Vegetation can conceal the underlying sediment, however if the type of trees growing within an area can be identified, the vegetation can aid in the interpretation of the sediment.

Chapter 4

Geomorphology

4.0 - Introduction

Geomorphological features are commonly used in reconstructing the glacial and postglacial history of an area. The geomorphological features identified in Terra Nova National Park and vicinity are shown in Figure 4.1 (in the back pocket). Many of the features were located and described throughout the field season of July 1995. However, large areas of Terra Nova National Park are inaccessible, and aerial photographs were also used to map the geomorphology of these areas.

4.1 - Flyggbergs

There are several prominent hills within Terra Nova National Park that protrude from the low-lying undulating landscape, such as Malady Head, Blue Hill, Mount Stamford, Ochre Hill, Ochre Pit Hill and Park Harbour Hill. Their shape resembles that of roches moutonnées, but their size suggests that they are similar to the "flyggbergs" first described by Rudberg (1970, 1973). The flyggberg at Malady Head is shown in Plate 4.1. These features have moderate up-glacier slopes with a gradient of approximately 70 m/km, and are generally 1-2.5 km long.

Plate 4.1 - Flyggberg at Malady Head. The steep face is 190 m high and ice flowed down the valley towards the northeast.

The down-glacier slopes are very steep, and several of the hills rise 180-200 m vertically out of the water. These features are generally restricted to eastern areas of Terra Nova National Park that are dominated by bedrock.

The flyggbergs described by Rudberg (1970, 1973) were formed by the plucking and abrasion of bedrock as the ice moved down the valleys and over the bedrock knobs. Truncated spurs are commonly associated with this process, and the valley is widened and often straightened (Rudberg 1988). Most of the flyggbergs in Terra Nova National Park are oriented with the steep lee-side slopes facing the northeast. It is suggested that these features were formed in a similar manner to roches moutonnées with bedrock being plucked from the lee-side as ice moved to the northeast. Malady Head and the northernmost hill of the Blue Hills range are exceptions. Their steepest slopes are parallel to the orientation of Southwest Arm, and they emphasize the U-shaped nature of a glaciated valley (Plate 4.1). It is likely that ice flowing through Southwest Arm truncated the spurs protruding into the valley, producing the steep slopes evident today.

4.2 - Roches Moutonnées

As with the flyggbergs, roches moutonnées are restricted to eastern areas of the study area where they are relatively easy to identify from aerial photographs, due to the lack of vegetation. Many of the best examples within the area are evident from the Trans Canada Highway which commonly cuts through these bedrock features.

Roches moutonnées are much smaller than flyggbergs but have the same moderate up-glacier slope and a steep, craggy down-glacier slope (Plate 4.2). The examples in Terra Nova National Park and vicinity range in size and are commonly between 8 and 500 m long and between 4 and 30 m high. There are approximately 35 well-defined roches moutonnées in Terra Nova National Park and vicinity, and many are formed from the shale, siltstone and sandstone of the Connecting Point Group.

Roches moutonnées are formed subglacially and commonly at the margins of glaciers where the ice tends to be thinner (Sugden *et al.* 1992). The thin ice creates small cavities in the lee of bedrock knobs, and the low pressure within these cavities enhances fracturing of the bedrock (Sugden *et al.* 1992). Subglacial meltwater is also involved in the formation of roches moutonnées (Sugden and John 1976). The higher ice velocities associated with a thin film of meltwater between the ice and the bedrock enhances stoss side abrasion and lee-side plucking (Sugden and John 1976; Röthlisberger and Iken 1981; Sugden *et al.* 1992). The removal of the loose material occurs when ice velocities are low and the lee-side cavities are closed as the ice comes into contact with the bedrock below (Sugden and John 1976).

Plate 4.2 - Roches moutonnées near Glovertown. The stoss slope is 7.5 m long and the lee face is 4 m high. Ice flowed towards the east.

4.3 - Crag and Tails

There are approximately 20 well-defined crag and tail features throughout Terra Nova National Park and vicinity. They are typically characterized by a resistant bedrock knob and a tail of glacial or glaciofluvial material, as shown in Plate 4.3 (cf. Embleton and King 1975; Lundqvist 1989; Czechówna 1994). The features in the Park are between 300 and 2,000 m long, and between 30 and 50 m high. They have a broader and higher proximal end, and the sediment tails are generally oriented towards the northeast.

Crag and tails form subglacially within cavities that are opened up as the ice flows over bedrock knobs (Sugden and John 1976). Boulton (1982) suggested that the tails are formed by sediment falling into the cavities from the basal layers of ice, and are subsequently streamlined by the overlying glacier as the base of the ice comes into contact with the upper surface of the sediment. A second proposal for their formation was given by Hoppe and Schytt (1953) and Boulton (1976), who suggested that the crag and tails are formed in a similar way to flutes. A cavity forms in the lee of the bedrock and due to the low pressures experienced in these cavities, the underlying till is squeezed up into the cavity and streamlined by the ice above.

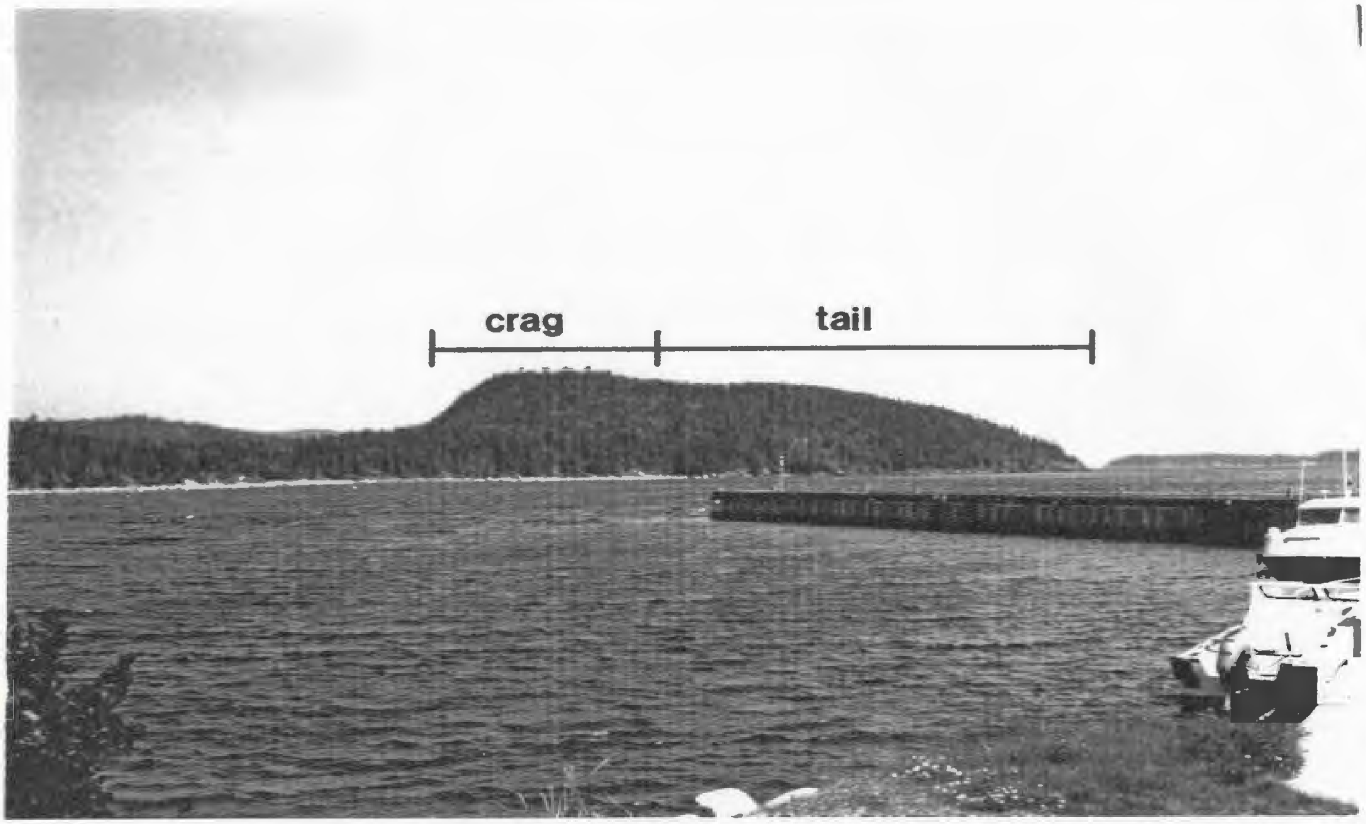

Plate 4.3 - Crag and tail landform 45 m long opposite the wharf at Saltons Brook, Terra Nova National Park. Ice flowed towards the east.

4.4 - Drumlins/Till Ridges

There are approximately 50 drumlins and till ridges throughout the study area that are between 200 and 1,600 m long, and between 7 and 45 m high. Some of the landforms have a blunted proximal end, and a relatively long tapering tail similar to the classical drumlin shape described by Chorley (1959). Others have a spindle form with rounded ends and their highest point is located in the middle of the feature. All the drumlins and till ridges are oriented southwest-northeast.

Drumlins and till ridges are formed subglacially, relatively close to the ice margin (Lundqvist 1970; Patterson and Hooke 1995). There are several hypotheses for their formation, ranging from entirely erosional to entirely depositional processes (Shaw 1994). Lundqvist (1970) suggested that drumlins in Sweden were formed by an initial stage of lodgement or melt-out till accumulation below the glacier, followed by erosion of this sediment by the moving glacier. In contrast, Shaw and Kvill (1984) suggested that subglacial meltwater initially eroded into the overlying ice forming cavities that were subsequently infilled as the meltwater flow started to decrease. A third hypothesis proposed by Shaw *et al.* (1989) was that drumlins were the result of sediment laden meltwater that flowed under the ice and eroded into the underlying bedrock or sediment. Finally, Aario and Peuraniemi (1992) proposed that drumlins were composed of two till units. The lower or central till was deposited by lodgement and meltout, whereas the upper till was deposited during deglaciation by meltout and flow processes.

It is not possible to speculate which process or processes formed the drumlins and till ridges in Terra Nova National Park and vicinity. None of the landforms identified in the field or from the aerial photographs provided an exposure showing the internal structure of the features. McCabe and Dardis (1989) discussed the internal structure of drumlins and concluded that due to the different types of depositional structures in drumlins, the method of drumlin formation cannot be determined unless the internal structures are studied in detail. The different streamlined features within Terra Nova National Park may indicate that several different processes were involved in their formation.

4.5 - Eskers

There are six discontinuous eskers within the study area and all are located west of the Park boundary and adjacent to Route 301 that leads to Terra Nova Village. Three of the eskers are located in Big Brook valley. The longest esker is oriented towards the northwest-southeast and runs parallel with the orientation of the valley. The other two eskers are aligned parallel with Big Brook which flows northeast towards the coast. Two of the remaining three eskers are oriented parallel with the southern shore of Terra Nova Lake towards the east, and the final esker is located on the southern margin of the Terra Nova River floodplain and is oriented towards the southwest-northeast. The eskers are between 0.3 and 1.25 km long, roughly 50 m wide and 10 m high, and they are relatively steep sided with a flat crest. They are composed of alternating beds of sand and gravel. Sedimentary structures throughout the eskers indicate deposition by flowing water, and beds close to the top and sides of the ridges slope downwards, producing an arch.

Eskers are long, linear and sinuous ridges that have relatively sharp crests, and are closely oriented with the regional ice flow direction (Price 1969; Sugden and John 1976; Reineck and Singh 1980). They are formed sub, en or supraglacially by the movement of meltwater in tunnels or open channels often close to the ice margin, and are commonly composed of stratified sediments (Price 1969; Reineck and Singh 1980; Syverson *et al.* 1994; Warren and Ashley 1994). The beds within an esker tend to be sub-horizontal in the centre and arched close to the surface, mimicking the geomorphology of the ridge (Warren and Ashley 1994). This is likely due to the collapse of the sediments as the adjacent ice melts (Price 1973; Embleton and King 1975; Banerjee and McDonald 1975; Gorrell and Shaw 1991; Warren and Ashley 1994).

There are two main depositional models for the formation of eskers. The most common type of esker formation is by sedimentation in sub or englacial tunnels (Embleton and King 1975; Reineck and Singh 1980; Syverson *et al.* 1994). The sediment may be deposited either at the mouth of the tunnel or farther back within the tunnel under the ice (Banerjee and McDonald 1975). The other method of formation involves deposition in a subglacial tunnel as the meltwater stream emerges into a lake or pond (Embleton and King 1975; Syverson *et al.* 1994; Warren and Ashley 1994).

Lacustrine deposits or fans are commonly associated with these types of eskers, and the eskers themselves are frequently beaded recording the ice front position as the ice retreats.

The eskers in the vicinity of Terra Nova National Park are typical of sub-glacial deposition, and were formed close to the ice margin of an ice sheet that was steadily retreating towards the west/southwest. The internal structures are well-preserved and the eskers appear to lie directly on the underlying bedrock. There are some ponds adjacent to several eskers. However, it is thought that these were formed by the eskers damming small depressions on the valley floor or against the valley slopes.

4.6 - Glaciofluvial Terraces

Several glaciofluvial or fluvial terraces were recognized within Big Brook valley adjacent to an esker that is located in the bottom of the valley (see Section 4.5). There are three large flat-topped terraces 150-300 m north of the esker that are between 0.85 and 1.7 km long, 0.25-0.75 km wide, and all are roughly 30 m high. The terraces are composed of sand and gravel and they are separated by small channels approximately 40-100 m wide. South of the esker there are eight ridges that are also separated by channels 20-65 m wide (Figure 4.2). The ridges are 0.15-0.6 km long and have an approximate width of 50 m. Although the ridges appear to have

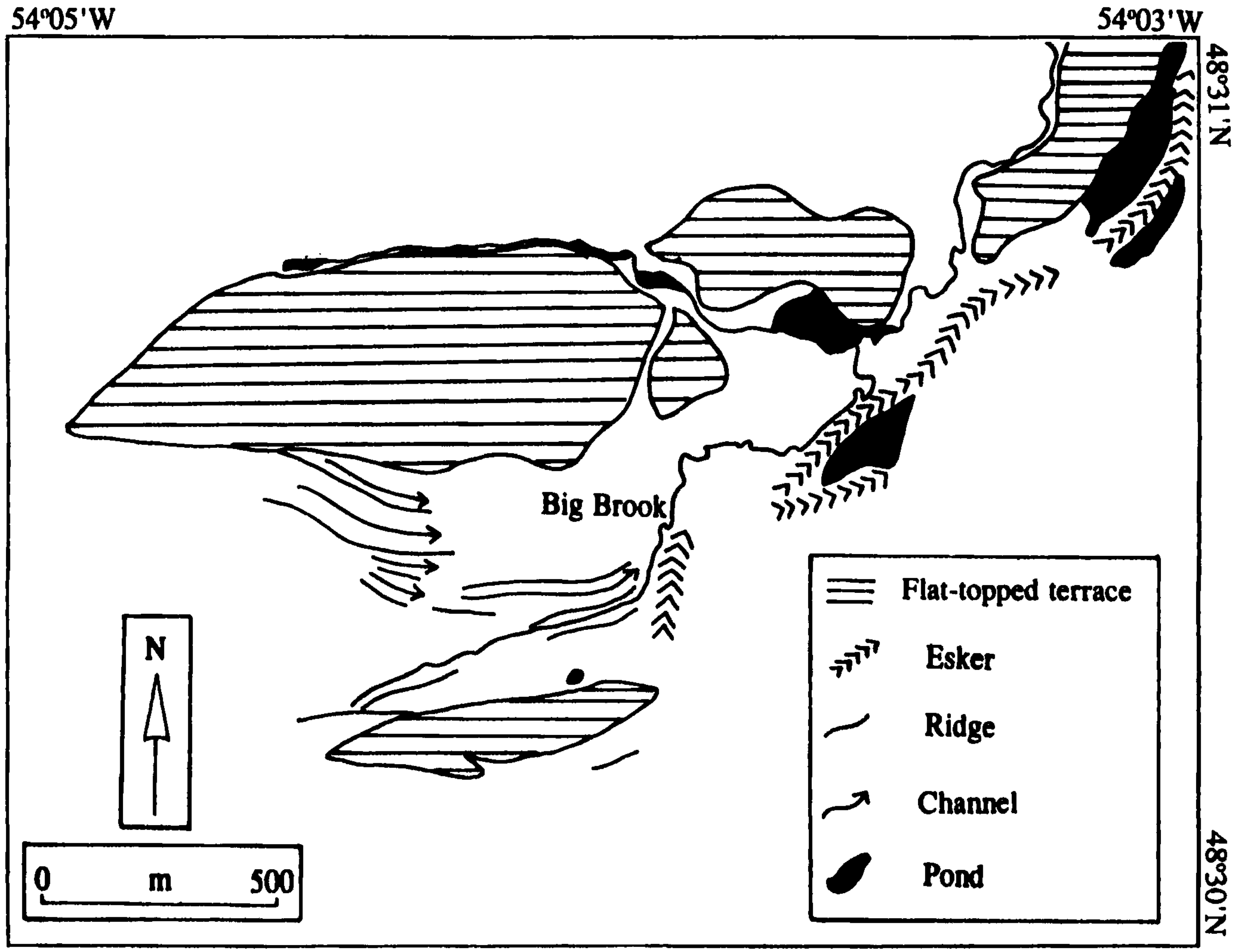

Figure 4.2 - Terraces and ridges in Big Brook valley.

been separated by channels, only two of the ridges have a river currently flowing between them.

Roughly 500 m south of Terra Nova Lake there is another large flat-topped terrace 1.9 km long, 250 m wide and 20 m high. The terrace is composed of medium to fine sand, and very little coarse material. A similar structure was identified 3 km east of the terrace at Terra Nova Lake. The internal structure of this terrace was studied in detail and is discussed in Section 6.1.2, Chapter 6.

4.7 - Raised Marine Deltas and Beaches

Throughout the study area there are a number of sheltered bays, and many of these are backed by a raised marine delta or beach. These raised landforms range from 5 m asl at Charlottetown to 30 m asl at Sandy Cove and Traytown. They are generally flat-topped, and the surfaces of the larger features at Sandy Cove/Eastport/Sandringham and Traytown/Glovertown are frequently used for agriculture. Only one raised marine deposit containing clay was found within the study area (see Chapter 7, Section A18). The remainder of the deposits are composed of sand and gravel (see Chapter 7, Section A19).

4.8 - Raised Lake Shorelines

There are approximately 150 lakes or ponds within Terra Nova National Park. Five of the largest ponds, such as Dunphy's Pond, Bluehill Pond, Wings Pond, Broad Cove Pond and Big Pond, are partially surrounded by raised lake shorelines. These shorelines are typically very flat, commonly covered by organic deposits and their size often depends on the surrounding topography. They are between 100 and 450 m wide, lie approximately 5 m above the present lake levels, and between 15 and 105 m asl. None of the lakes has multiple shorelines.

The existence of these landforms indicates that throughout the past these lakes were slightly larger than today. This may be the result of a cooler and wetter climate in the past, or beavers may have dammed the downstream end of ponds causing the ponds to increase in size (cf. Butler and Malanson 1994). The ponds may also have been larger due to human intervention in the form of logging in the area throughout the early part of this century. Logs may have blocked the outlets causing the water levels to rise.

It is not possible to determine without more research when the ponds in Terra Nova National Park decreased in size. Terra Nova became a National Park in 1957 (Deichmann and Bradshaw 1984) and since this time there has been no logging in the area. The pond levels are more likely to have been lowered as a result of the destruction of beaver dams close to the ponds. Approximately 30-35% of the ponds within Terra Nova National Park are currently occupied by beavers (K. Robinson, Terra Nova National Park, pers. comm. 1997), and almost all the ponds in the Park have been occupied by beavers at some point in the past (Deichmann and Bradshaw 1984). Another factor that may have influenced the pond levels is that the climate has steadily deteriorated since 4,000 years BP becoming colder and wetter (Macpherson 1995). Due to the wetter conditions, the rivers may have increased their discharge and cut down through the sediment or bedrock at the pond outlets improving the drainage throughout the area and lowering the levels of the ponds.

4.9 - Fluvial Terraces

Adjacent to several of the larger rivers in the study area, notably Terra Nova Brook, Big Brook, Southwest Brook, Terra Nova River and Bread Cove Brook, are flat topped fluvial terraces. These range in width between 50 and 650 m and their width is usually restricted by the surrounding topography. The terraces are paired and lie at similar heights on either side of the rivers, approximately 1-3 m above the present river level, and between 15 and 45 m asl. All of the terraces are laterally continuous and composed of sand, gravel and fine grained sediment. Some of the terraces are topped by organic deposits indicating that there is poor drainage close to the rivers. The organic deposits and fine grained sediments suggest that during floods the terraces are, or were, often covered by water.

River terraces commonly represent the upper level of aggradation before erosion and downcutting take place (Lowe and Walker 1984). There are several factors influencing river aggradation and erosion including sediment supply, changing base levels, and climatic change (Lowe and Walker 1984; Rice 1988). During deglaciation a large quantity of coarse and fine grained sediment was transported by the meltwater streams towards the coast. Much of this sediment was deposited along the way due to changes in discharge and gradient and this caused the river to quickly aggrade. When there is a reduction in the amount of sediment available for transportation the river quickly starts to incise into its bed forming river terraces.

Base level is usually regarded as sea level, however, there may be local base levels in the form of ponds or bedrock obstructions (Lowe and Walker 1984). Since the end of the last glaciation there have been several changes in base level due to isostatic recovery and input of glacial meltwater to the ocean. As sea level rises, the rivers develop shallower gradients, and therefore they drop their loads causing aggradation. Once sea level starts to fall, the rivers must adjust to a lower base level, resulting in incision and downcutting, and the formation of river terraces. A change in local base levels, such as pond formation, may result in aggradation and incision in a river farther upstream. Base level in the study area has risen and fallen in response to sea level changes and isostatic recovery. When sea level was higher than present, forming the deltas at Sandy Cove and Traytown, the rivers had a shallower gradient and aggraded their beds. As sea level dropped due to isostatic recovery the rivers incised into the previously deposited sediments.

Finally, river terraces may be formed due to climatic changes. During colder periods river discharges tend to be large, flashy and laden with coarse and fine grained sediment, leading to aggradation. Warmer periods tend to be characterized by a more uniform discharge and reduced sediment transport, resulting in river incision and terrace development (Lowe and Walker 1984). Low river terraces may also be the result of "normal" variations due to flooding.

It is suggested that the low river terraces in Terra Nova National Park and vicinity were formed due to flooding throughout the Holocene. Higher river terraces are thought to have formed due to changes in the climate and sediment supply during the Holocene. As deglaciation commenced the rivers discharged massive amounts of meltwater that transported and deposited large quantities of sediment. River incision and the formation of the terraces began as the climate ameliorated when there was a reduction in the amount of sediment being transported by the rivers. There is only one river that appears to have been influenced by local base levels. Bread Cove Brook is located upstream of Bread Cove Pond, and it may be that changes in the pond level may have influenced river aggradation and incision.

4.10 - Ox-bow Lakes

Four small ox-bow lakes were identified from aerial photographs adjacent to Bread Cove Brook (Figure 4.3). The lakes are approximately 50-150 m long and all are dry indicating that sediment has completely infilled the abandoned lakes. The sediments infilling the ox-bow lakes are typically fine-grained.

Ox-bow lakes are associated with meandering rivers and form when the river cuts a new course across the neck of a meander abandoning the meander loop (Reineck and Singh 1980; Strahler and Strahler 1992). The lakes are rapidly plugged at either end and slowly infilled with fine grained flood deposits and organic debris (Collinson 1986; Strahler and Strahler 1992).

4.11 - Organic Deposits

Approximately 25% of the study area is covered by organic deposits. Western areas are characterized by large deposits, whereas eastern sections of the Park have smaller accumulations located in small depressions and in bedrock hollows. Gros Bog and Saltons Marsh are two large wetlands close to the western boundary of the Park, and both are located on relatively flat land 150 m above sea level. They have areas of approximately 14 and 8 km^2, respectively, and are typical of the raised or domed bogs described by Wells and Pollett (1983). Smaller bogs and fens have developed throughout the region where there is a lack of drainage, causing the water table to be close to the surface (Deichmann and Bradshaw 1984). Organic deposits are also

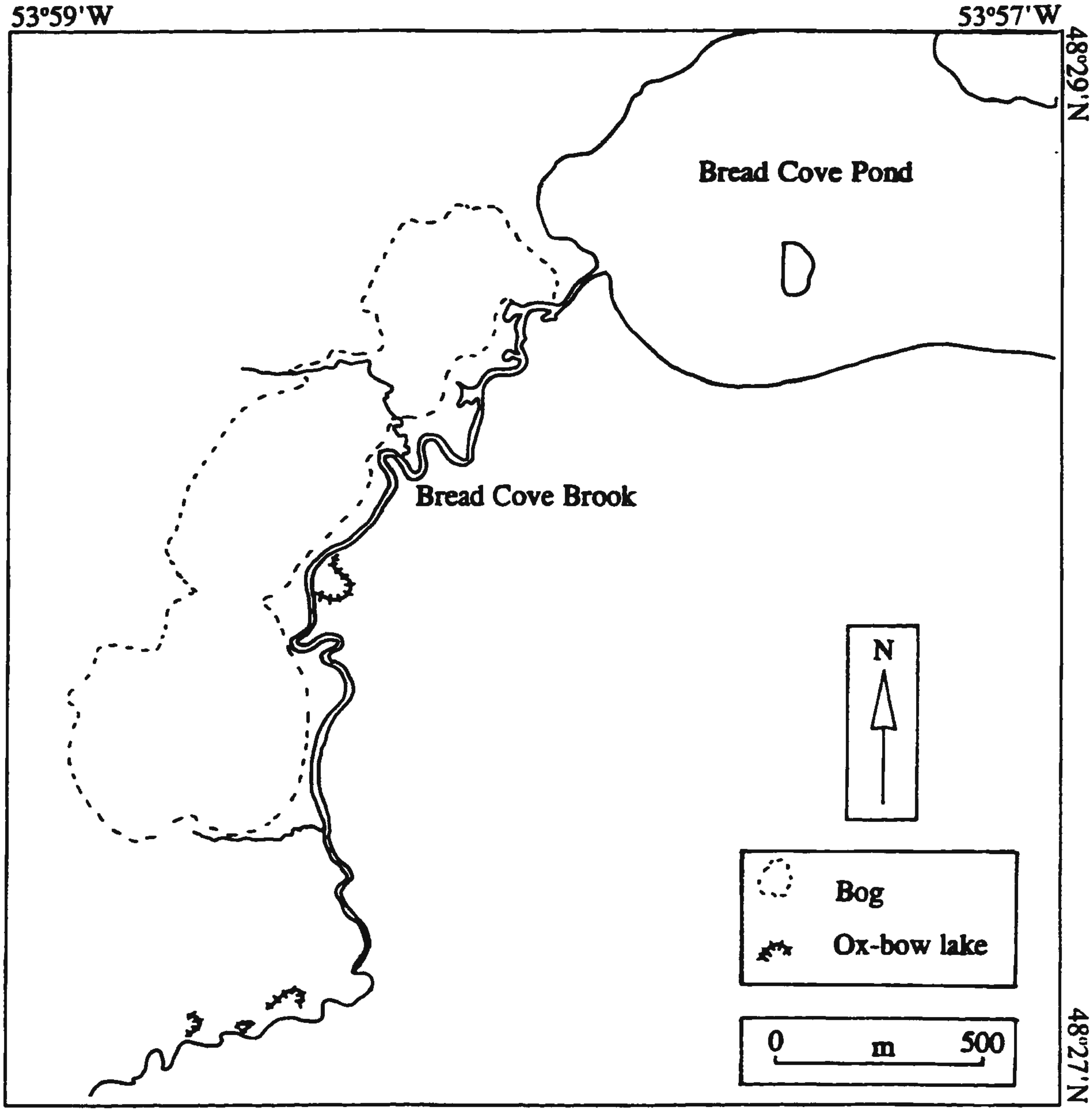

Figure 4.3 - Ox-bow lakes adjacent to Bread Cove Brook

relatively common adjacent to some of the ponds within the Park. Their location suggests that the ponds may originally have been larger than they are today and that they have reduced in size due to improved drainage, or that the bogs are encroaching into the ponds.

4.12 - Modern Rivers

The modern river channel patterns within Terra Nova National Park and vicinity have been influenced by changes in the base levels, discharge and sediment supply throughout the Holocene. Several of the rivers, such as Southwest Brook, Saltons Brook and Big Brook, have incised into the glaciofluvial or fluvial deposits especially in their lower reaches resulting in a braided or sinuous channel pattern.

Terra Nova River has an ample supply of coarse-grained sediment and is characterized by a braided channel pattern in its upper reaches. I propose that the sediment was initially deposited by meltwater streams during deglaciation as the ice retreated. Farther downstream, the river becomes more sinuous and appears to be contained possibly by bedrock. There are rapids and falls along the length of the river indicating that bedrock is relatively close to the ground surface.

Southwest Brook displays a meandering channel pattern in its upper reaches. The river has a floodplain roughly 300 m wide and a gentle gradient of 2.2 m/km. Two other rivers, Terra Nova Brook and Bread Cove Brook, meander throughout their full length. Terra Nova Brook occupies a floodplain approximately 150 m wide, and the floodplain is bordered on either side by terraces of sand and gravel 15 m high. The terraces suggest that the river has incised into the glaciofluvial or fluvial sediments. Bread Cove Brook meanders down a gentle gradient of 4.5 m/km through relatively fine grained sediments. Ox-bow lakes were recognized adjacent to Bread Cove Brook and these have been described in Section 4.10.

4.13 - The Modern Coastline

Approximately 80% of the study area coastline is dominated by bedrock that commonly rises steeply out of the sea, forming cliffs from a few metres up to 200 m in height. There are, however, a few sandy and/or pebbly beaches in sheltered bays and along the shoreline of Newman and Clode Sounds. Many of these are backed by raised deltas or beaches, that have been discussed in Section 4.7.

Modern deltas are currently being formed at the mouths of Southwest River, Big Brook, and Southwest Brook. They extend for 2.5, 1.5 and 0.15 km, respectively, beyond the river mouths. The water is fairly shallow because at low tide large areas of these deltas are subaerially exposed. Some of the rivers, such as Terra Nova River and Cobblers Brook, transport large quantities of sand and gravel into the sea. Sediment spits have built up at the mouths of these rivers and are aligned parallel with the coastline due to long shore drift.

Chapter 5

Glacial Landforms and Sediments

5.0 - Glacial Landforms

Where possible, glacial landforms were identified, described and measured in the field. However, given access limitations within Terra Nova National Park, descriptions of some of the glacial landforms are based solely on aerial photograph interpretation. The landforms provide evidence for glacial erosion and deposition, and they are used to deduce regional ice flow directions throughout the study area.

5.1 - Glacial Erosional Landforms

Flyggbergs, roches moutonnées, and crag and tails are all typical glacial erosional landforms, and they have been discussed in Chapter 4. The majority of these landforms indicate ice flow directions towards the northeast. A few of the coastal landforms are oriented directly towards the coast. It is suggested that these landforms represent local ice flow directions that existed during deglaciation due to the underlying topography having an increasing effect on the ice flow directions.

Glacial striations are erosive scratches on bedrock that are formed sub-glacially. They are the result of clasts contained within the basal layers of the ice being dragged across bedrock surfaces (Lowe and Walker 1984). The striations record regional and local ice movements. During deglaciation the underlying topography has an increasing effect on ice movements, and therefore the orientation of the striations may vary considerably from the regional ice flow direction. Nailhead striations are commonly used as definitive evidence for the direction of ice movement due to their characteristic shape (Iverson 1991). However, other ice flow indicators such as roches moutonnées should also be used when reconstructing regional ice movements.

Much of the exposed bedrock within Terra Nova National Park and vicinity provided little, if any, striation data due to the weathering patterns of the granites and metasediments. In areas where the bedrock has recently been exposed by human excavation, striations were distinct and easy to measure. The type of bedrock is an important factor in the preservation of striations, and within the Terra Nova National Park area they are best preserved on the volcanic and sedimentary rocks of the Musgravetown Group. The rocks are relatively hard and are not readily eroded by subaerial weathering.

Striations were measured at twenty sites within Terra Nova National Park and vicinity, and their orientations are shown in Figure 4.1 and detailed in Appendix 1. Data from other sites around the Park and on the Bonavista Peninsula were recorded by Jenness (1963), Brookes (1989), St. Croix and Taylor (1990), and Taylor *et al.* (1994), and were used as an aid in the overall interpretation of ice movement in the area. The striations measured in July 1995 vary from 4.5 to 12 cm long, although several reach lengths of approximately 75 cm. Most of the striations are 1-2 mm deep, however the longer striations tend to be slightly deeper (2-5 mm).

The majority of striations throughout northeast Newfoundland are oriented towards the east and the northeast. Striation sites indicating multiple flow directions are rare throughout Terra Nova National Park and vicinity, although there are several sites adjacent to Eastport Beach, and on several bedrock outcrops along the Trans Canada Highway. Cross-cutting striations indicate that the glacial conditions have changed over time and that the ice was flowing in different directions (Lowe and Walker 1984). This may be a result of the underlying topography having an increasing effect on ice flow directions as the ice thins during deglaciation, or the cross-cutting striations may represent a separate advance from a different direction (Lowe and Walker 1984; Syverson 1995). Superimposed striations commonly cut directly across striations below and the unbroken striations are taken to be the younger of the two (cf. Gray and Lowe 1982). The older striations in the study area are oriented towards the northeast, whereas those cross-cutting them, and representing a younger event, are oriented between north and northwest.

5.2 - Glacial Depositional Landforms

Drumlins and till ridges are scattered throughout western areas of the Park, and relate to some of the depositional processes that occurred subglacially. These features have previously been described in Chapter 4. The drumlins and till ridges are concentrated on the floors of the main river valleys, and are generally oriented towards the northeast reflecting the regional ice flow directions.

Erratics are clasts or boulders that are located in an area that has an entirely different lithology from the clast (Price 1973). They are commonly used to reconstruct the movement of glaciers across the landscape, especially in areas where the source is known (Lowe and Walker 1984). The most useful erratics in this region are clasts of Terra Nova Granite. The granite is characterized by coarse crystals of pink feldspar, quartz, plagioclase and some ferromagnesian minerals (Jenness 1958; Baird 1966). It is a distinctive rock and easy to identify on the ground surface or within sedimentary sections.

Clasts and boulders of Terra Nova Granite are found throughout the area. The shoreline of Newman Sound is littered with sub-rounded clasts and boulders of Terra Nova Granite, and erratics are also found at the top of some of the hills within the Park. Their location at the top of these hills indicates that glacial ice covered the entire landscape, leaving no nunataks protruding above the ice surface. Smaller erratics are also observed on top of roches moutonnées. Clasts of Terra Nova Granite are also found within all the glacial and glaciofluvial sediments throughout the study area (see below and Chapter 6).

The Terra Nova Granite intrusion is located west of the Park boundary around Terra Nova Village. The erratics identified indicate that they were entrained into the glacial ice and transported across the study area in a northeasterly direction. They were then deposited either on the ground surface or within the glacial or glaciofluvial sediments.

5.3 - Glacial Sediments

Throughout the western part of the region, a thin (approximately 1 m) veneer of diamicton covers much of the undulating uplands. Deposits 1 to 4 m thick are restricted to local depressions or valleys, such as Big Brook valley and the Terra Nova River valley. Eastern areas of the Park are dominated by bedrock, and glaciogenic sediment is restricted to bedrock depressions that commonly contain ponds. Typically, the thickness of till is less than 2 metres.

Tills throughout northeast Newfoundland are dominated by sandy matrices, with silt and clay contents of 10%. Coarse clasts form 40% of the diamictons. Clasts within the tills tend to be of local origin, with more than 75% derived from within 5 km of the depositional site, and the matrices generally reflect the local bedrock geology. The most conspicuous erratics are derived from the Terra Nova Granite outcrop to the west/southwest.

Four exposures containing diamictons with a minimum thickness of 1 m were analysed in detail (Figure 5.1, Appendix 2). All of these diamictons have been interpreted to have been initially deposited directly from glacial ice. Several deposits, however, have undergone post-depositional remobilization as a result of mass movement processes.

5.3.1 - Terra Nova Village Road (Route 301) - A11a

Description

Section A11a is located approximately 7 km west of the Trans Canada Highway on Route 301. The section is roughly 500 m north of Route 301 along a gravel road, and situated on the south side of Big Brook valley. It is 4 m high and 20 m wide. The basal contact with the underlying bedrock was not observed, and therefore the total thickness of the unit is not known. Colluvial debris mantles the base of the exposure. Although the basal contact was not observed, the presence of exposed bedrock within 600 m of the section suggests that the section consists of a single unit of diamicton.

The diamicton is matrix-dominated. Approximately 75% of the matrix is sand, 10% is silt and clay, and the remaining 15% is granules. The matrix has a reddish grey colour when dry and fresh (Munsell: 5YR 5/2). A high degree of fissility is apparent where the matrix has dried due to subaerial weathering. The fissile structure commonly drapes underlying clasts or follows the convex shape of

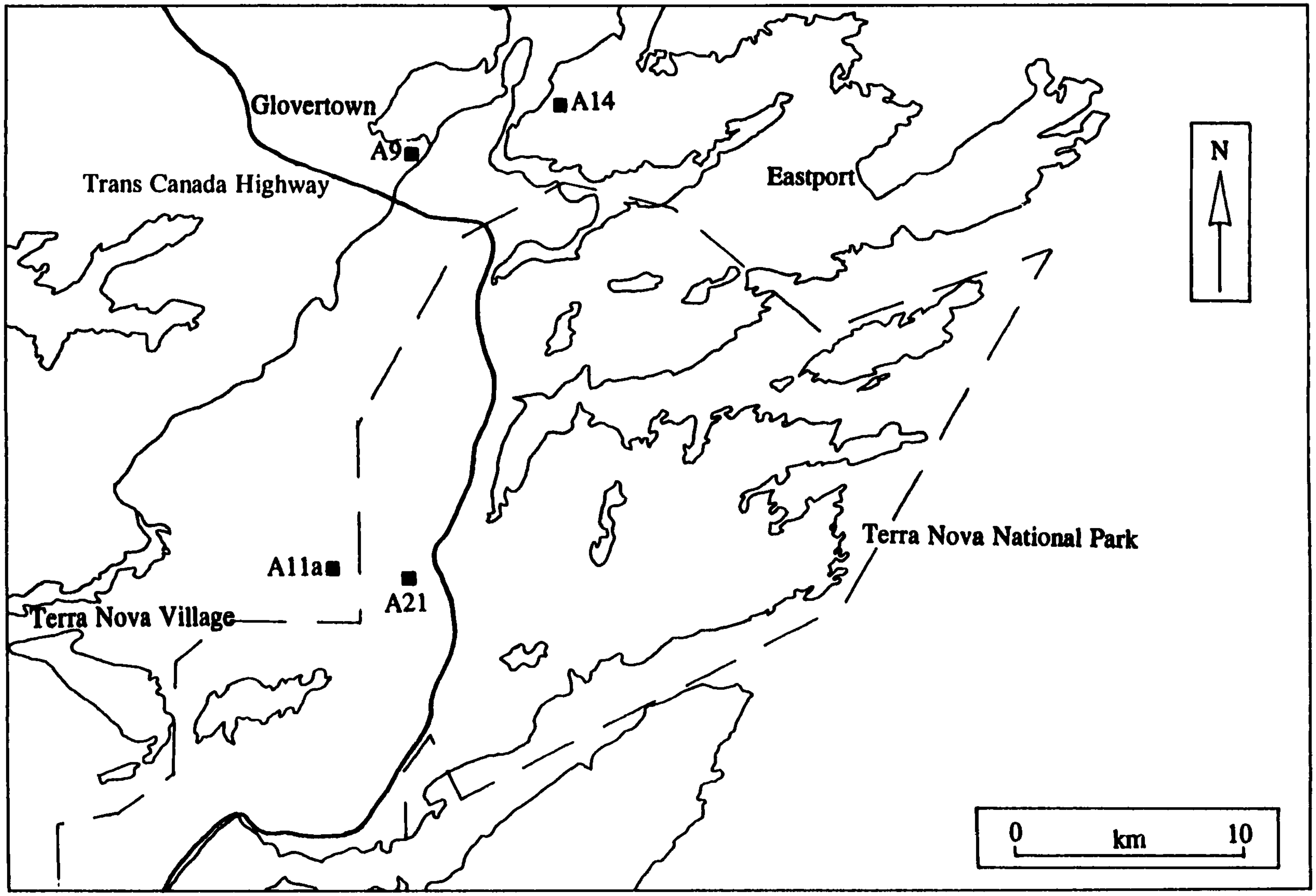

Figure 5.1 - Map showing the location of glacial deposits examined in detail.

lenses throughout the exposure. Beneath the weathered surface layer, the matrix is hard and difficult to excavate.

Approximately 30% of the unit is composed of clasts that are generally subrounded and are distributed randomly throughout the exposure. Clasts are dominated by Terra Nova Granite (50%) and Hare Bay and Square Pond Gneiss (40%); 10% of the clasts are sandstones, metasediments and rhyolites from the Musgravetown Group. The shape of clasts is commonly controlled by their lithology (Drake 1970). The majority of the clasts in this exposure are discs (45%) or blades (29%), and most of the granite clasts have a disc shape. Many of the clasts are capped by silt layers approximately 0.5 cm thick forming hard layers over the surface of the clasts.

There are sparse striations on the surfaces of some of the clasts within the diamicton. Granite is a coarse-grained rock that is easily weathered, and gneiss is very resistant to weathering and erosion. Both rock types are, therefore, rarely suitable for the preservation of striations. Sandstone and basalt from the Musgravetown Group are softer rocks and therefore can be striated easily. Most of the striated clasts within the section are sandstone. The low number of clasts from the Musgravetown Group within the diamicton accounts for the low number of striated clasts observed.

Clast fabric analysis was undertaken at 4 points within the exposure (Figure 5.2). The fabrics were taken at the same height approximately 0.8 m below the

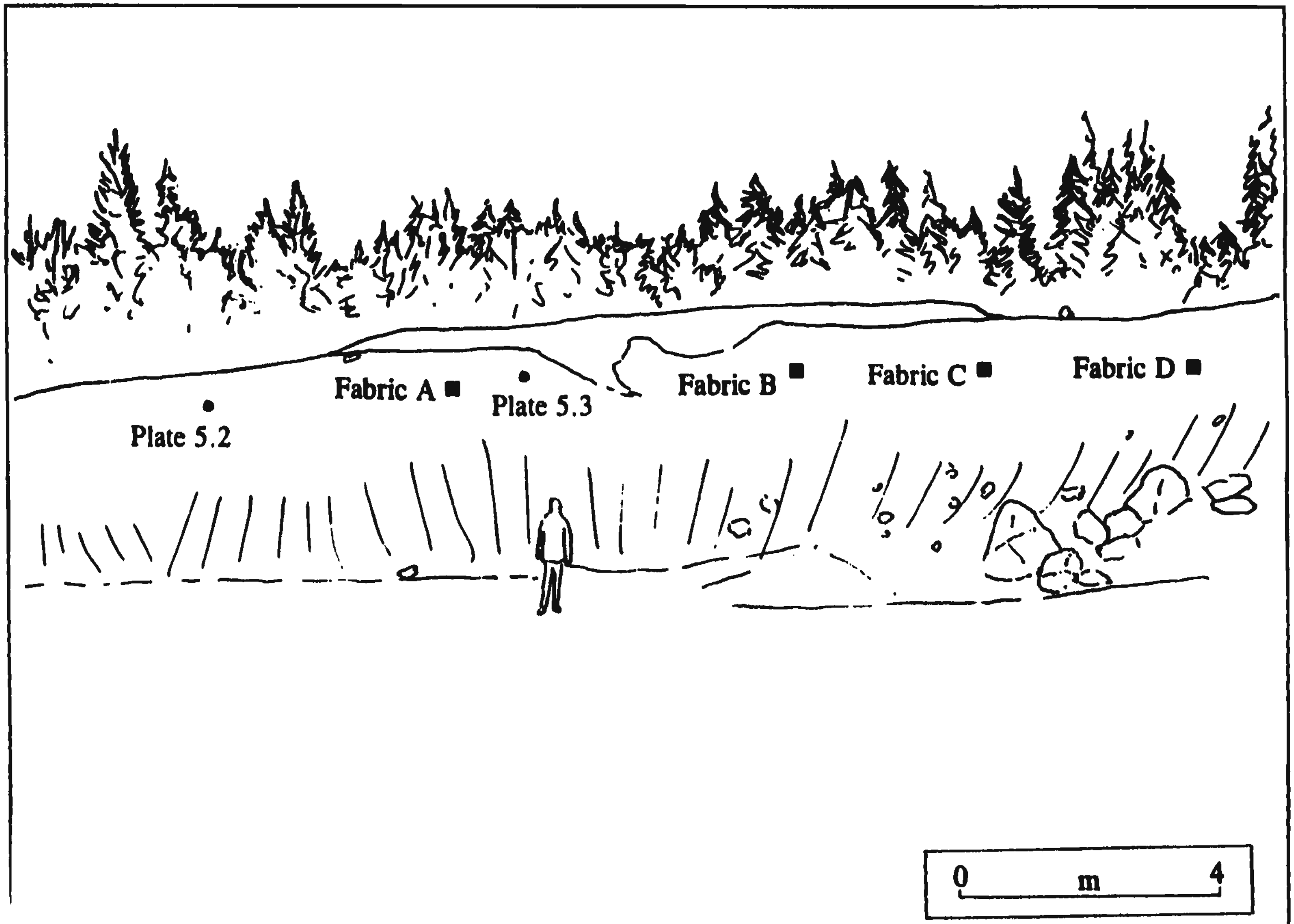

Figure 5.2 - Clast fabric analysis sites in Section A11a and the location of Plates 5.2 and 5.3.

surface to avoid any post-depositional reworking, and roughly 15 m apart. The fabrics have S_1 values ranging between 0.523 and 0.609, and K values of 0.22-0.31. Three of the fabrics have mean orientation vectors of 292.9°, 288.9° and 197.1°, and mean plunges of 21.8°, 19.7° and 16.9°, respectively. The fourth fabric has a mean orientation vector of 115.8° and a mean plunge of 6.1°. The fabrics are moderately oriented girdle distributions (Figure 5.3) as defined by Woodcock (1977). The clasts are oriented parallel and transverse to the local ice flow direction which is towards the northeast, as determined by roches moutonnées and striations throughout the area. Some are aligned parallel with the trend of the local slope which is towards the north-northwest.

Throughout the exposure there are discontinuous lenses of sand that vary in length (15-70 cm) and thickness (1.5-30 cm). The upper and lower contacts of the lenses are generally sharp, although the lenses commonly grade laterally into the diamicton. There is some fining upwards within each lens from medium sand to coarse silt. Several lenses are horizontal and extend for 0.7 m. The majority of well-exposed lenses, however, are sinusoidal and commonly drape the clasts (Plate 5.1).

There are several lenses that are not influenced by underlying clasts. One of the lenses has a clear convexo-concave shape with a flattened base, that protrudes into the underlying sediment. The lens is 15 cm wide and 1.5 cm thick and grades laterally into the diamicton. There is some fining upwards within the lens.

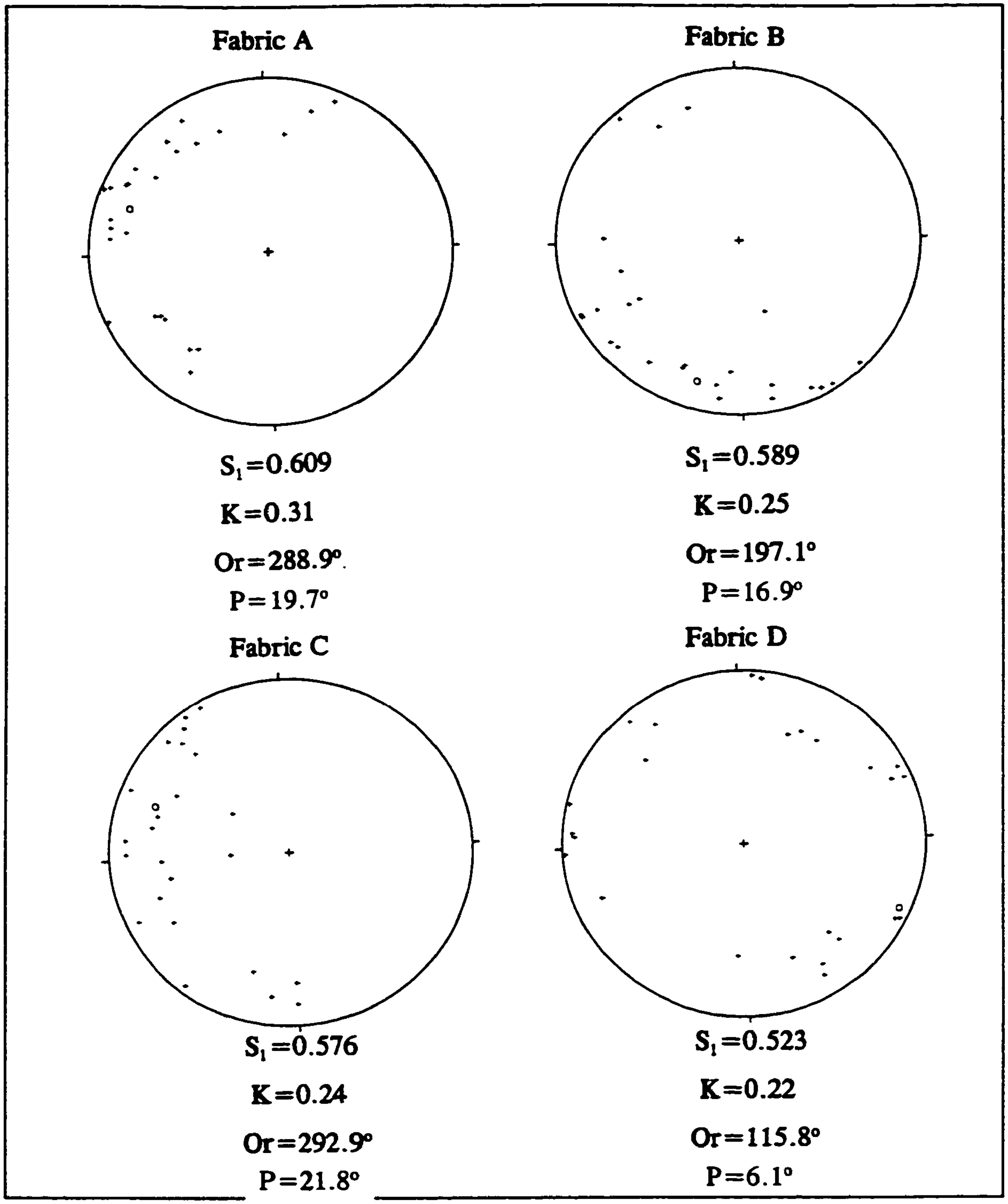

Figure 5.3 - Stereoplots of fabrics used to determine ice flow direction at Section A11a.

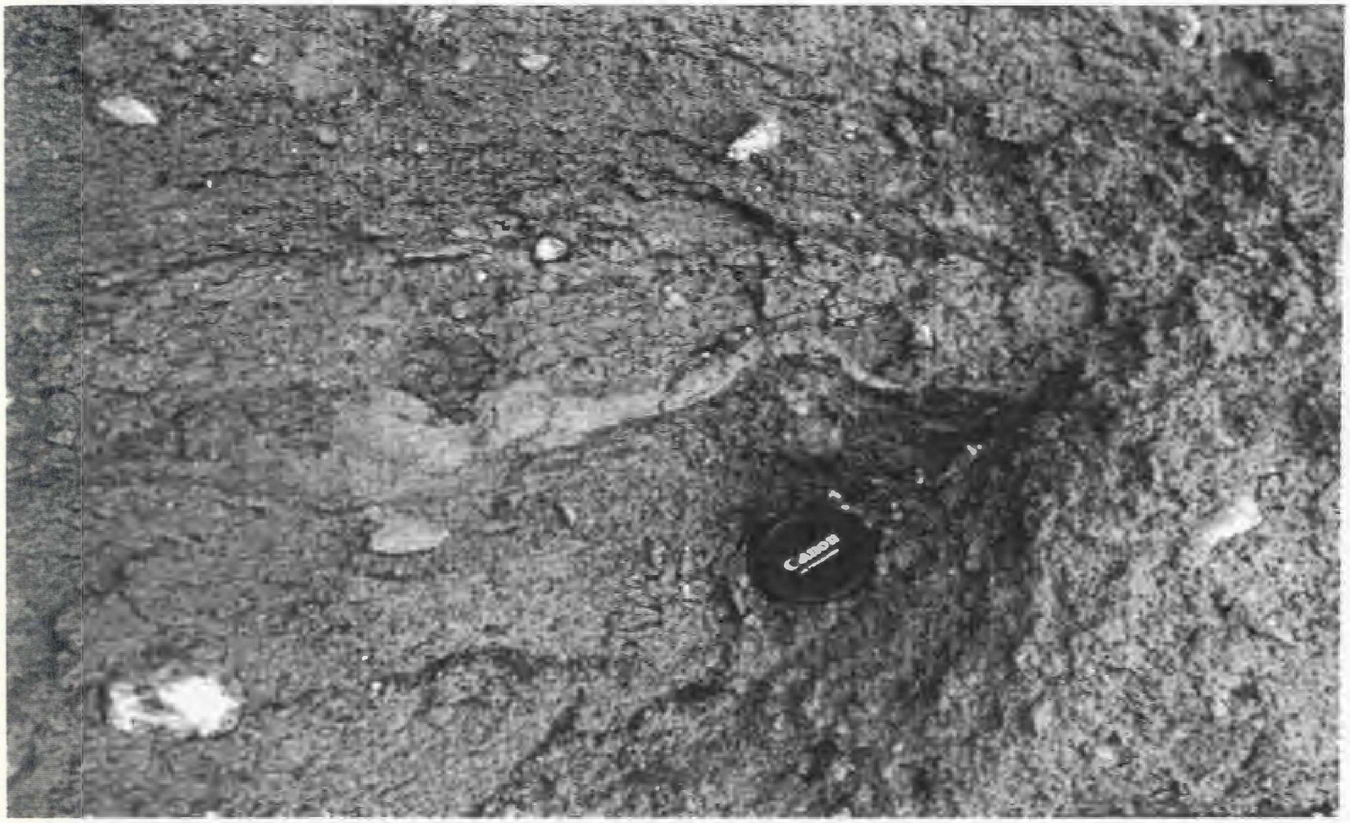

Plate 5.1 - Sinusoidal sand lens within Section A11a draping the underlying clast. The lens is 17 cm wide and 1 cm thick.

Small flame structures are associated with one of the horizontal lenses (Plate 5.2). The lens is 37 cm wide and 4 cm thick and has sharp upper and lower contacts. The upper contact undulates slightly but is generally horizontal. In contrast, the lower contact is irregular and outlines a small flame structure and a finger of diamicton that protrudes laterally into the lens. The flame structure is filled with a finer grained diamicton that grades down into the main diamicton below.

One of the horizontal lenses drapes over a large block of sediment which is approximately 40 cm wide and 30 cm thick. The matrix is similar to that of the surrounding diamicton but has a slightly lighter grey colour when dry and weathered (Munsell: 5YR 6/1). There appears to be no fissility and there is no internal structure within the block. The lower contact grades into the underlying sediment, whereas the upper contact is sharp and is marked by the silty-sand lens above.

Plate 5.2 - Small flame structure associated with one of the sand lenses in Section A11a. The lens is 37 cm wide and 4 cm thick. Clasts above the lens have a high angle of dip as a result of being dropped from the base of the ice into the underlying cavity.

Interpretation

Clasts within the diamicton at A11a are dominantly local, and have travelled a distance of approximately 20 km or less. The local derivation, roundness of the clasts, and the striated nature of the sandstones and basalts, suggests that the clasts were transported in a basal ice position (Krüger 1979; Dreimanis 1989). Approximately 50% of the clasts within the diamicton are derived from Terra Nova Granite. There is a large outcrop of Terra Nova Granite roughly 10 km to the southwest of the exposure, suggesting that ice flowed northeastwards across the study area.

The matrix has a high degree of fissility once it has dried by subaerial processes. This type of structure is often observed in lodgement or basal meltout tills and is associated with the initial lodgement process or by subsequent overriding of the glacier (Dreimanis 1976, 1989; Muller 1983). Both processes generate large amounts of pressure on the underlying sediments (Johansson 1983). Virkkala (1952) suggested that the fissile structure formed by the accretion of layers of sediment underneath the glacier. The base of the overlying moving glacier then shifted upwards depositing the sediment with interstitial ice. As the interstitial ice melted, the fissile structure was formed. Boulton (1970) suggested that the sediment may appear fissile due to post-depositional wetting and drying, freezing and thawing. The fissile structure within A11a commonly follows the sinusoidal nature of sand lenses below, and therefore the fissility is interpreted to have a primary origin.

The fissile structure at A11a thus was initially formed beneath active ice by primary depositional processes. Subsequent stagnation of overlying active ice allowed the interstitial ice to melt and the fissility to form. The lack of any obvious fissile structure in freshly exposed sediment suggests that post-depositional weathering processes are required to enhance the fissile structure.

Silt caps on the upper surfaces of clasts have been reported from central and western Newfoundland by M. Batterson (Newfoundland and Labrador Department of Mines and Energy, pers. comm.) and D. Liverman (Newfoundland and Labrador Department of Mines and Energy, pers. comm.), however there is no reference to these features in the Newfoundland literature. Boulton and Dent (1974) observed silt caps on the upper surfaces of clasts in lodgement till from Breidamerkurjökull, Iceland. The features were formed by the translocation of water down through the till from an upper silt-rich horizon within the till soon after the till had been exposed from beneath the glacier (Boulton and Dent 1974). Silt caps may also form during meltout of the diamicton. As the interstitial ice melts the meltwater moves down through the sediment and deposits the silt on the clasts. Translocation of groundwater through the diamicton may result in the formation of silt caps (Liverman, pers. comm.), however there is no other evidence of groundwater movement, such as oxidation, mottling or a fragipan, within the section. It is therefore suggested that the silt caps at A11a were formed by the movement of meltwater down through the sediment during meltout of the diamicton or soon after the till had been exposed by the retreating glacier.

Lenses of sand or gravel are commonly found in basal meltout tills, supraglacial tills and debris flows (Shaw 1979; Lawson 1981; Catto 1992; Krzyszkowski 1994). Sand lenses within the section show some internal structure, but none of the lenses has been deformed. In some locations, the lenses drape the underlying clasts. Shaw (1979) and Lawson (1981) suggested that such lenses are typical of a meltout till. As the ice melts cavities develop within the till, and fine grained sorted sediments are deposited in the cavities by meltwater (Shaw 1979). Once the lenses have formed they are commonly deformed by the melting out of underlying ice causing them to drape clasts below (Shaw 1979; Lawson 1981).

Boulton (1972), Lawson (1981) and Catto (1992) discussed the formation of fine grained lenses in supraglacial tills and debris flows. The surface of a glacier is commonly characterized by coarse, angular clasts and very little fine grained material due to the movement of meltwater across the sediment (Lawson 1981). In areas where there is poor drainage, small streams may develop on the surface of the supraglacial till or debris flow. The streams erode small depressions into the sediment forming small pools into which fine grained sediment is deposited (Boulton 1972; Lawson 1981). Subsequent burial by more supraglacial sediment or by another debris flow forms the lenses which may be deformed as a result of loading (Boulton 1972; Lawson 1981; Catto 1992).

The lenses in the diamicton at A11a are interpreted to have been formed in a basal ice position by meltout. Clasts within the diamicton are subrounded indicating basal ice transport and deposition (Dreimanis 1989). Others have suggested that such lenses are initially deposited horizontally, however due to melting of the underlying sediment the lenses are draped over clasts or deformed by differential settling (Lawson 1981). The supraglacial till or debris flow pools are generally characterized by clay and silt and therefore this method of lens formation is rejected due to the coarse nature of the lenses in A11a.

One of the lenses drapes a large cluster of clasts and matrix that has no internal structure and is significantly different to the surrounding sediment. This is interpreted to be a frozen mass of sediment that was entrained by the glacier and deposited farther down-glacier still in its frozen state. The block subsequently melted out beneath the stagnant ice mass and a cavity developed above the block depositing the fine grained sediment. Deposition by lodgement would have destroyed the block, however the block would be preserved during basal meltout (Ashley *et al.* 1985).

The clast fabrics are represented by moderately oriented girdle plots. The S_1 values are reasonably high and are typical of a meltout till or a debris flow (Lawson 1979; Dowdeswell *et al.* 1985; Rappol 1985). Clasts within the section are oriented transverse and parallel to flow. Lawson (1979) suggested that this is typical of resedimented till or sediment flows. The amount of water within the matrix affects the mechanics of a sediment flow and hence the strength of the fabric. When the water content is low, shear will only take place at the base of the section, and very few clasts will be realigned. As the amount of water within the matrix increases a more preferential orientation will develop and the glacial signature may be completely destroyed.

The mean plunge of the clasts is relatively low (16.1°) and this is characteristic of lodgement tills or basal meltout tills that experience some settling after deposition (Ashley *et al.* 1985). Clasts located above sand lenses commonly have a high angle of dip as a result of being dropped from the base of the ice into the underlying cavity (Plate 5.2).

The sediment exposed at A11a thus is interpreted as a basal meltout till. The local clast provenance, striations and fissility of the matrix indicate that the sediment was transported and deposited at the base of the ice. It is suggested that as the interstitial ice melted the fissile structure was formed, and the meltwater produced translocated down through the sediment forming the silt caps observed on the upper surfaces of many of the clasts. Sand lenses throughout the section also were likely formed by the movement of meltwater through the till. They may have been deformed by the melting of the underlying sediment, which would indicate that there was very little reworking during their formation. The clast fabric may indicate a meltout till or a debris flow, and although some of the clasts have a similar trend to the local slope, the preservation of sedimentary structures indicates that there was little, if any, post-depositional modification.

5.3.2 - Terra Nova Village Road (Route 301) - A21

Description

The section exposed by road construction at A21 is located approximately 2.5 km from the Trans Canada Highway on Route 301. It is 1 m high and is confined to approximately 1 m wide. The basal contact with the underlying bedrock was not observed, and no bedrock was observed close to the section.

The diamicton is matrix-dominated and is brown when dry and fresh (Munsell: 7.5YR 5/2). Clasts within the diamicton are generally subrounded and the majority are blade or disc shaped. The clasts (40%) are distributed randomly throughout the exposure. The dominant clast lithology is Hare Bay and Square Pond Gneiss (50%); 30% of the clasts are Terra Nova Granite, and the remaining 20% are sandstones, basalts and siltstones of the Musgravetown Group. Striae preservation is poor due to the resistance of gneiss clasts to weathering and erosion and to the coarse-grained texture of granite which is easily weathered. Striations were, therefore, only observed on softer sandstones, basalts and siltstones. The diamicton is structureless.

One clast fabric was measured at the section. The section was excavated to ensure that the clasts were measured *in situ* and had not been moved during road construction. The fabric has an S_1 value of 0.803 and a K value of 1.17. The mean trend of the clasts is 159.2° and the mean plunge is 22.4°. The values plot as a cluster (Figure 5.4), as defined by Woodcock (1977).

Interpretation

The diamicton analysed at A21 is interpreted as being deposited either directly by glacial ice or by a debris flow that formed after the initial deposition by ice. Clasts are dominantly local and have travelled a maximum distance of 25 km or less. The local origin of the clasts and the striated nature of the sandstones, basalts and siltstones suggests that the clasts were transported and deposited in a basal ice position (Krüger 1979; Dreimanis 1989).

The unit is structureless and poorly sorted, and is typical of a lodgement till, meltout till or debris flow deposit (Dreimanis 1989). Sedimentary structures within these deposits often provide an indication of flow direction, and the extent of deformation of these structures may distinguish meltout tills from lodgement tills and debris flows. Structures within meltout tills are rarely deformed and often drape the underlying clasts (Boulton 1970; Lawson 1981; Dreimanis 1989; Krzyszkowski 1994). Lodgement tills are, however, commonly characterized by sheared lenses and overturned folds, whereas debris flows exhibit overturned folds, slump folds, diapirs, stringers, and other deformational structures (Eyles and Kocsis 1988; Dreimanis 1989; Krzyszkowski 1994). Sedimentary structures were not observed within the exposure at A21, and therefore the sediment may be interpreted as a lodgement till, meltout till or debris flow.

The clast fabric has a unimodal distribution, and indicates that flow was towards the north-northwest. The S_1 value is high (0.803), and is typical of a lodgement or basal meltout till (Dowdeswell and Sharp 1986; Ham and Mickelson 1994). The mean plunge of the clasts is low (22.4°) and this suggests that there was some clast settling as the till was deposited. Clasts within meltout tills are characterized by low angles of dip due to rotation about the vertical plane as the clasts settle (Ashley *et al.* 1985). Dowdeswell and Sharp (1986) suggest, however, that the plunge of clasts in undeformed lodgement tills are similar to those in meltout tills.

The sediment exposed at A21 is thus interpreted as a primary basal till. Striated clasts and local provenance indicate that the sediment was transported in a basal ice position. The clast fabric is typical of a lodgement or meltout till with a strong unimodal distribution, indicating that ice flowed towards the north-northwest. However, striations and roches moutonnées throughout the Terra Nova National Park region indicate that the regional ice flow direction was towards the northeast, and I conclude that the till at A21 represents a local ice flow direction. The lack of sedimentary structures within the exposure limits the choice between lodgement or meltout till, and therefore the sediment is interpreted as a primary basal till.

5.3.3 - Glovertown - A9

Description

A9 is located within an abandoned gravel pit 1.6 km south of Glovertown South, near the Terra Nova River. The section is approximately 5 m high and although extensive, only a 1 m swath was cleared for analysis. Two units were recognized within the section (Plate 5.3). The lower unit is a diamicton, and is mostly covered by colluvial debris. It is unconformably overlain by sands and gravels (see Section 6.1.1, Chapter 6).

The lower unit is approximately 2 m thick and the basal contact with bedrock was observed close to the section. The diamicton is matrix-dominated, and has a brown colour when dry and fresh (Munsell: 7.5YR 5/2). Approximately 85% of the matrix is sand, 13% is granules and the remaining 2% is silt and clay.

The clasts (40%) are subrounded and are randomly distributed throughout the unit. A total of 60% of the clasts are Terra Nova Granite, 20% are Square Pond and Hare Bay Gneiss, and the remaining 20% are sandstones and rhyolites from the Musgravetown Group. Roughly 50% of the clasts have a disc shape and 30% have an equant shape. None of the clasts analysed was striated.

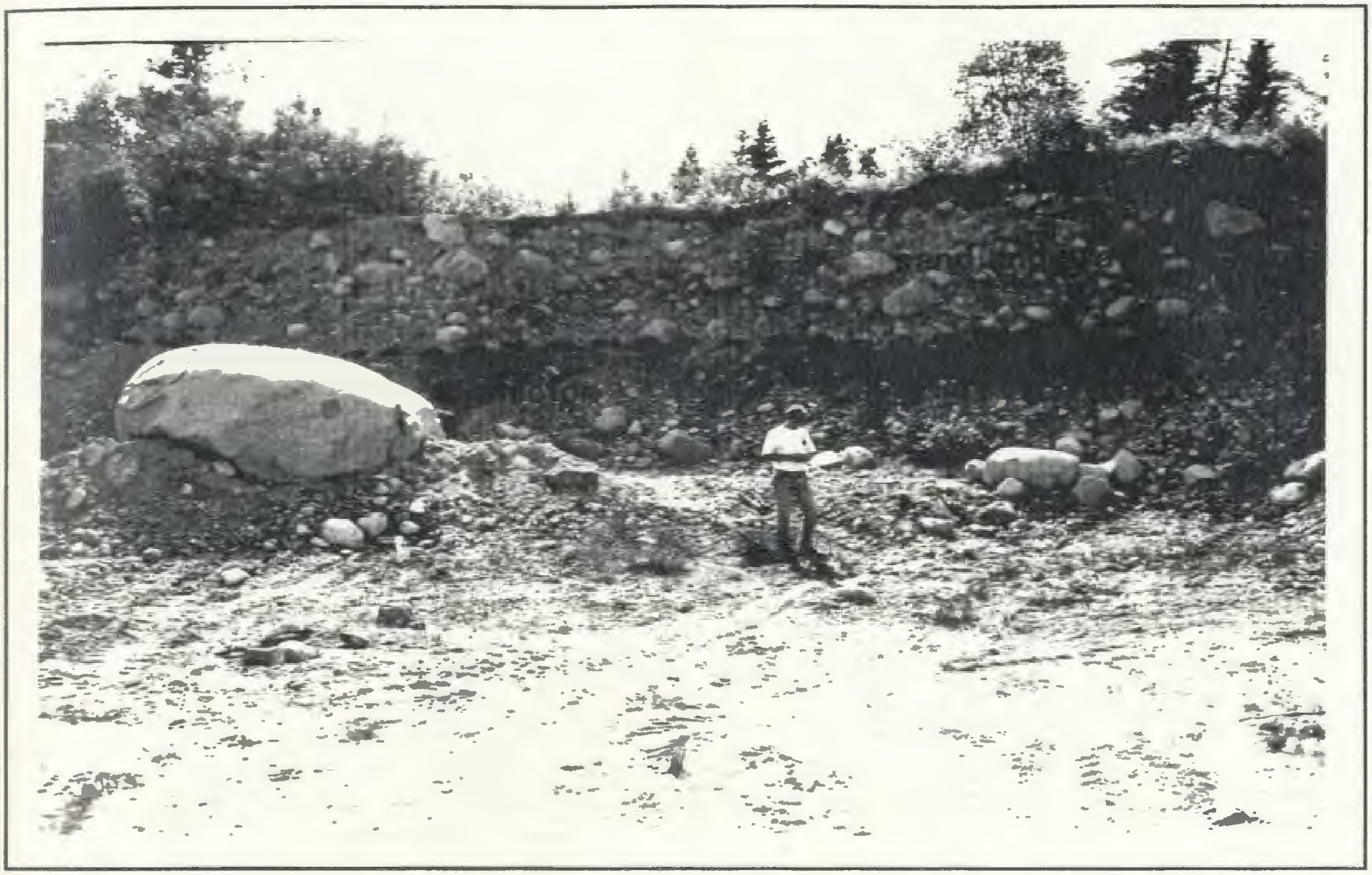

Plate 5.3 - Two units recognized in Section A9. The lower unit is 2 m thick and composed of a fine-textured diamicton. It is unconformably overlain by 3 m of glaciofluvial sand and gravel.

Only one clast fabric analysis was carried out at this site since much of the diamicton is covered by colluvium. The S_1 value is 0.495, and the fabric has a K value of 0.02. The mean trend of the clasts is 316.6°· indicating flow towards the northwest, and the mean plunge is 14.2°. The values are representative of a very weak girdle plot (Figure 5.4), as defined by Woodcock (1977).

Throughout the unit there are coarse sand lenses that have gradational upper and lower contacts. The lenses are 12-18 cm long, 2-6 cm thick, and have a sinusoidal shape. Most of the lenses grade laterally into the diamicton. One of the lenses is deformed due to loading from the diamicton above.

Interpretation

The clasts within the diamicton at A9 have a local origin and have travelled a distance of 14 km or less towards the northeast. Their local provenance indicates that they were initially transported in a basal ice position (cf. Dowdeswell *et al.* 1985; Dreimanis 1989; Krüger 1979). The clasts have a disc or equant shape which is typical of lodgement or basal meltout tills (Boulton 1978; Dowdeswell *et al.* 1985). Clasts that come into contact with the glacier bed tend to be more rounded due to the increased effects of abrasion and crushing at the ice-bed interface (Boulton 1978).

Lenses of sorted sediment have been recognized in debris flows and meltout tills by Shaw (1979), Lawson (1981) and Krzyszkowski (1994), and their formation is discussed in Section 5.3.1. The lenses in the section at A9 consist of sand and thus are too coarse grained to have accumulated at the surface of debris flows. It is, therefore, suggested that the lenses in A9 were deposited by the movement of meltwater through the diamicton as it melted. Deformation of some of the lenses probably occurred as a result of the sediment above the lenses melting out and loading the underlying structures.

The clast fabric is typical of supraglacial sedimentation or debris flow deposits, as described by Lawson (1979), Dowdeswell *et al.* (1985), Eyles and Kocsis (1988), and Brodzikowski and Van Loon (1991). Supraglacial sediments are generally carried along on the surface of the glacier with no preferred clast orientation and do not experience any reorientation until they are deposited. Clasts that are deposited passively as the underlying ice melts have no preferred orientation and therefore show very weak fabrics (Brodzikowski and Van Loon 1991). Some of the deposits experience post-depositional movement due to saturation of the sediment by meltwater. The clasts in these deposits have stronger fabrics compared to the sediments that were deposited passively, however the preferred orientation of these deposits may be different to the direction of ice movement (Brodzikowski and Van Loon 1991).

The strength of the clast fabric within a debris flow is largely dependent on the water content within the flow (Lawson 1979). When there is very little water only the base of the flow undergoes shear and clasts in this thin zone become aligned parallel to the flow direction (Lawson 1979). This produces a fabric that has either no preferred orientation or a very weak orientation (Dowdeswell *et al.* 1985; Eyles and Kocsis 1988). As the water content increases, the zone of shear increases, and more clasts within the sediment become preferentially aligned with the flow direction.

High angles of dip are commonly associated with supraglacial sedimentation and debris flow deposits that have a high percentage of clasts. As the sediment is transported and deposited collisions between the clasts causes some of the clasts to dip at unusually high angles (Lawson 1979; Gravenor 1986). Lodgement and meltout tills tend to have lower percentages of clasts and therefore have lower angles of dip (Lawson 1979; Brodzikowski and Van Loon 1991). The low angles of dip at A9 suggests that the diamicton was deposited by lodgement or meltout, however debris flows or supraglacial sedimentation may produce low angles of dip if there is a low percentage of clasts within the sediment.

The diamicton exposed at A9 is interpreted as a glaciogenic sediment flow. The local clast lithologies and the roundness of the clasts indicate that the sediment was transported in a basal ice position (Krüger 1979). I propose that the sediment was initially deposited as a meltout till, and subsequent movement of meltwater through the sediment formed sorted lenses of coarse sand (Haldorsen and Shaw 1982; Krzyszkowski 1994). The clast fabric is typical of a sediment flow, even though the clasts have a low mean angle of dip. I, therefore, conclude that after the initial deposition of the sediment as a basal meltout till, the till became saturated and moved downslope as a sediment flow. This caused some of the clasts to be reoriented, and some of the sand lenses to be slightly deformed.

5.3.4 - Culls Harbour - A14

Description

Section A14 is located in a gravel pit at the end of the road that passes through the village of Culls Harbour, 2 km north of Traytown. The exposure is 10 m high and 140 m wide. The basal contact with bedrock was not observed. Colluvial debris mantles the base of the exposure and therefore only the upper 4 m could be examined. Two units were recognized within the section (Plate 5.4). The lower unit is a diamicton and it is unconformably overlain by sand and silty-clay. Only the lower unit will be described and interpreted here. The upper unit is similar to the sediments described in Section 6.1.3, Chapter 6.

The diamicton is matrix dominated and the matrix is dark grey when dry and weathered (Munsell: 5YR 4/1). Approximately 45% of the diamicton is composed of clasts; these are generally subrounded to subangular, and are distributed randomly throughout the unit. Clast lithology is dominated by Musgravetown sandstones and metasediments (50%) and Terra Nova Granite (40%); 10% of the clasts are Hare Bay and Square Pond Gneiss. The majority of the clasts have a disc (40%) or equant shape (30%), and most of the sandstone clasts have a disc or roller shape. Striae preservation is generally poor due to the coarse-grained texture of granite, however striations were observed on the surface of some sandstone clasts.

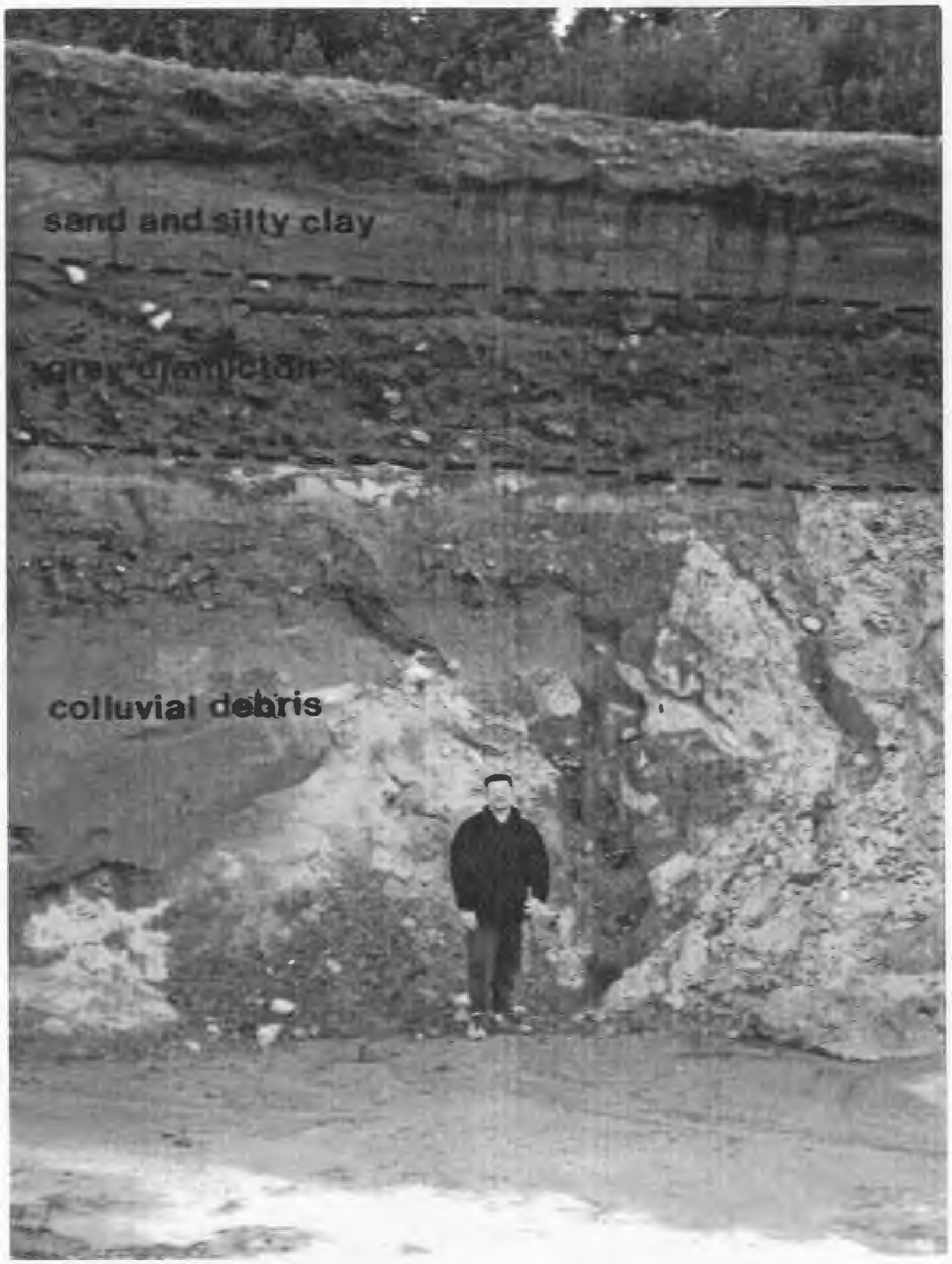

Plate 5.4 - The exposed face at Section A14. The base of the section is 6 m thick and covered by colluvial debris. Above this there is a lower grey unit that is 2 m thick and composed of diamicton. This unit is unconformably overlain by 2 m of sand and silty-clay.

Only one clast fabric was measured at the section since much of the section is covered by colluvium. The fabric has an S_1 value of 0.625 and a K value of 0.40. The mean trend orientation of the clasts is 134.2°, indicating flow towards the northwest, and the mean plunge is 29.3°. The values are representative of a moderately oriented girdle plot (Figure 5.4), as defined by Woodcock (1977).

Clasts throughout the exposure range in size from 1.3 m to 0.45 m, and some of the clasts are bullet-shaped and striated. Accumulations of smaller clasts and sand lenses are found on either the left or right side of some of the larger boulders (Plate 5.5). One of the boulders has a cluster of smaller clasts below it (Plate 5.5). The smaller clasts form clast-supported lenses with a sandy matrix. The lenses are 0.8-1.1 m wide, 40 cm thick, and thin laterally. There is no internal structure, and the clasts are clustered together with no preferred orientation. The sand lenses are 50-60 cm wide and 35-69 cm thick. They are composed of medium sand with one or two small clasts and they have no internal structure.

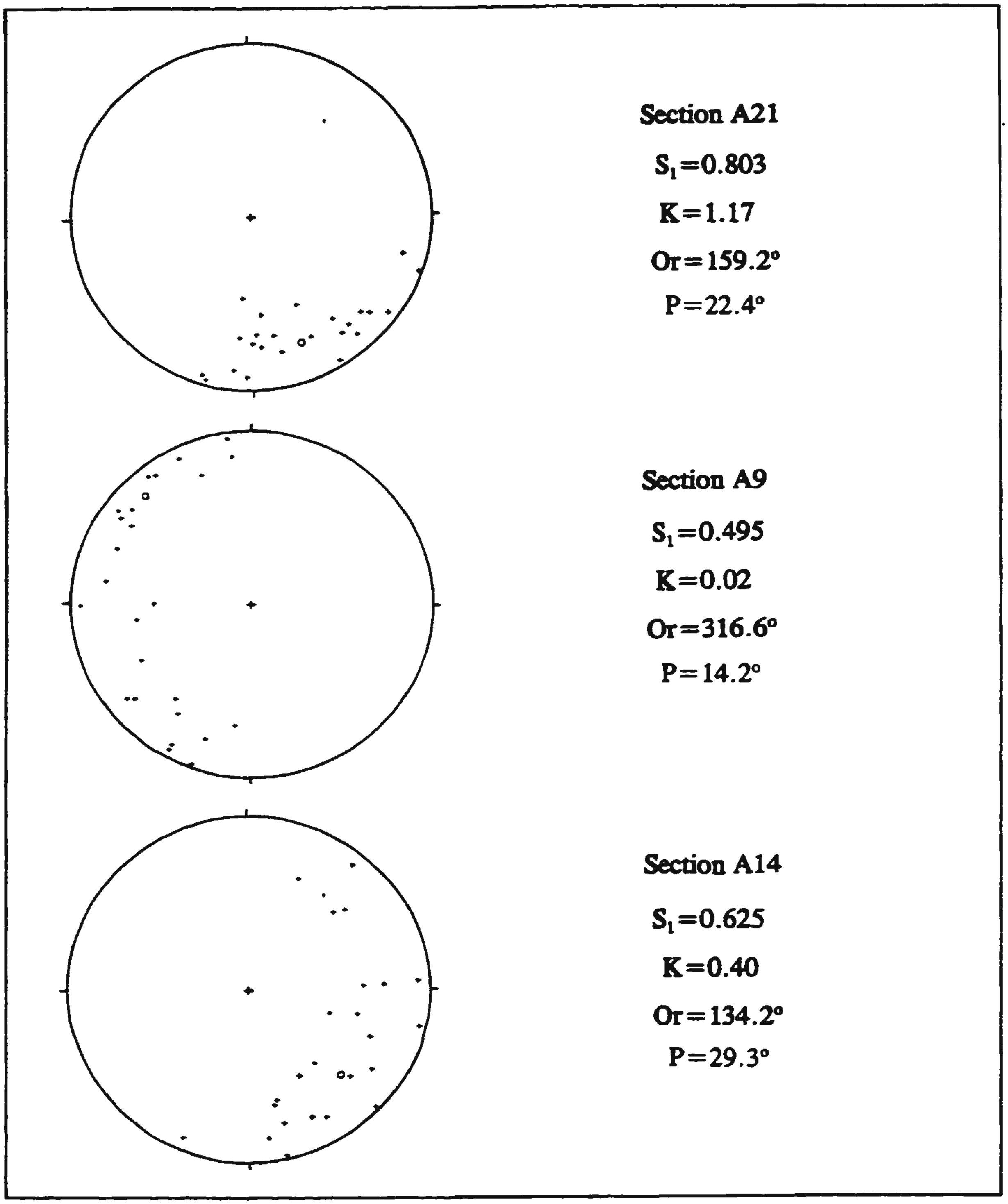

Figure 5.4 - Stereoplots of fabrics used to determine ice flow direction at Sections A21, A9, and A14.

Plate 5.5 - Clast-supported lens and sand lens associated with a boulder in Section A14. The lens is 1 m wide and 40 cm thick, and clasts within the lens are dominantly shale.

Interpretation

Clasts within A14 are relatively local and have travelled a distance of 22 km or less. The local lithology, roundness of the clasts and the striated nature of the sandstones indicate that the clasts were transported in a basal ice position (Krüger 1979; Dreimanis 1989).

The clast fabric is typical of a basal meltout till or debris flow (Lawson 1979; Dowdeswell *et al.* 1985; Eyles and Kocsis 1988; Ham and Mickelson 1994). Basal meltout tills have moderate to strong fabric strengths reflecting transportation in a basal ice position (Lawson 1979, 1981). As the ice melts some of the clasts are slightly reoriented due to settling, causing the fabric strength to decrease (Boulton 1970; Lawson 1979).

Lawson (1979, 1981, 1982) discussed the mechanics of debris flows and concluded that weak, moderate or strong fabrics will develop depending on the amount of water in the flow. Debris flows that have a low water content will shear only at the base and therefore only the clasts at the base of the flow will become oriented parallel to the flow direction. The upper sediments are not deformed and this reduces the strength of the fabric. As the water content increases, an increasing number of the clasts are affected by the flow until the majority of the clasts are oriented parallel to the flow direction, thus producing a strong fabric (Lawson 1979, 1982). Debris flows commonly form at the front or the sides of active or stagnant ice from previously deposited glacial sediments (Lawson 1981). Fabrics measured within debris flow deposits are generally distinct from those deposited by basal ice, and commonly have significantly different preferred flow directions to that of the overall ice flow direction (Lawson 1979).

Lenses of coarse or fine grained sediment are commonly formed in basal meltout tills, supraglacial tills and debris flows (Shaw 1979; Lawson 1981; Catto 1992; Krzyszkowski 1994). These processes have previously been discussed in Sections A11a and A9 and need not be repeated here. The lenses here, however, are much larger and coarser than those described previously.

Shaw (1983) suggested that lenses were formed by the action of meltwater scouring sediment either from around the base of a boulder that was suspended in the base of the ice, or in a cavity that formed at the junction of the till and the overlying debris-rich ice. As the meltwater velocity waned sediment was deposited in the scours either from the meltwater itself, or by mass flows that formed as the surrounding till melted (Shaw 1983).

Sharpe (1988) and McCabe and Dardis (1994) described similar structures to those seen here from the Scarborough Bluffs, Ontario and County Galway, Ireland, respectively. They interpreted the features to have been formed by meltwater either at the ice/substrate interface or within the diamicton. Sharpe (1988) observed that there was no evidence to suggest that the diamicton and lens structures were formed by an overriding active glacier, and therefore concluded that they were deposited in a cavity system by stagnant ice. McCabe and Dardis (1994) described disorganized coarse gravels and indicated that it was difficult to determine if the gravels had been deposited as a lag, and the fines had subsequently been winnowed out, or if the coarse material had actually moved through the diamicton. The gravel lenses commonly terminated abruptly against large clasts, and McCabe and Dardis (1994) compared the formation of the lenses to the development of phreatic tube systems in a subsurface karst environment. They concluded that the structures were formed under high pressure conditions by meltwater eroding sinuous passages through the diamicton or at the ice/substrate interface. The meltwater had enough energy to transport larger clasts that were subsequently deposited as the flow waned (McCabe and Dardis 1994).

The lenses in the diamicton at A14 are interpreted to have been formed under stagnant ice by meltwater erosion and deposition. The lenses are formed preferentially on one side of the boulders and it is suggested that meltwater flowing at the ice/substrate interface was forced to erode sediment on one side of the boulders as it came into contact with them. I suggest that the gravel lens under the boulder in Plate 5.5 was formed by the boulder protruding from the base of the ice causing the meltwater to increase in turbulence and erode sediment from underneath the boulder (cf. Shaw 1983). Sand lenses associated with the gravel lenses indicate that there was a decrease in the flow energy causing the coarser clasts to be deposited. The cluttered and unoriented nature of the clasts suggests that they were deposited very quickly. The gravel lenses have a sandy matrix suggesting that sand infiltrated the gravel as the sand lenses were deposited. Subsequent deposition from sediment in the ice above may have closed off the cavity causing the meltwater to find another route through the diamicton, or the meltwater may have abandoned the channel causing the overlying ice to close the cavity (Haldorsen and Shaw 1982).

The diamicton at A14 is thus interpreted to be a basal meltout till. The fabric is relatively strong, however, the preferred orientation is towards the northwest. This is in direct contrast to the regional ice flow direction that is towards the northeast and has been inferred from roches moutonnées and striations. I conclude that the sediment was initially brought to the area by ice moving towards the northeast, and as the ice retreated the underlying topography had an increasing effect on the local ice flow directions. This may have caused the ice to flow down the local slope which at A14 is towards the northwest, or a subglacial cavity may have formed causing the basal till to slide down the local slope. As the ice stagnated meltwater flowing along the ice/substrate interface created small cavities at the sides of boulders. These were subsequently infilled and lenses were formed as sediment was deposited from the overlying ice.

5.4 - Summary of Glacial Deposits in Terra Nova National Park and vicinity

The glacial deposits studied in detail generally do not reflect the regional ice flow direction towards the northeast. The primary basal till at A21 indicates an ice flow direction towards the north. I propose that this represents a local ice flow direction, possibly during deglaciation when the underlying topography has an increasing effect on ice flow directions. The basal meltout tills at A11a and A14 contain undeformed lenses suggesting that they have experienced little post-depositional modification. The clasts within the deposits are generally oriented towards the northwest, reflecting the orientation of the local slopes. I suggest that these too represent local ice flow directions. The final glacial deposit at A9 is a glaciogenic sediment flow. The clasts within this deposit are oriented towards the northwest, but the sediment likely experienced some post-depositional modification and therefore the fabric is not expected to parallel the regional ice flow directions.

5.5 - Summary of Ice Flow History

The area of Terra Nova National Park and vicinity was subjected to one main ice flow event during the Late Wisconsinan. The majority of flyggbergs, roches moutonnées, crag and tails, striations and drumlins/till ridges indicate a regional ice flow direction towards the northeast. The larger erosional features, such as flyggbergs and roches moutonnées, are oriented obliquely with the strike of the underlying bedrock and their steep, craggy slopes reflect the structural landscape. There are several striation sites throughout the study area indicating multiple ice flow directions. Although it has not been possible to numerically date the striations, there is no indication of a significant time lag between the events. The cross-cutting striations may represent the increasing effect of the underlying topography on ice flow directions during deglaciation, or the area may have experienced a small readvance of ice that was constrained by the surrounding topography.

Erratics located on the summits of some hills and roches moutonnées within the Park indicate that ice covered the entire landscape, leaving no nunataks protruding above the ice. Cumming *et al.* (1992) noted that glacial ice extended at least 50 km offshore in Bonavista Bay. Depositional landforms, such as crag and tails and drumlinoid ridges are oriented towards the coast and I suggest that these were deposited during deglaciation when ice was drawn down into the bays and coves.

Due to the lack of sedimentological data, the orientation of striations and clast provenance data are important for the overall understanding of the glacial history of the area. The striations reflect the regional ice flow direction and they also provide evidence of local ice flow directions that occurred throughout deglaciation. These local ice flows are not recorded in the sedimentological data. Erratics within the glacial sediments and throughout the landscape are also important for determining regional ice flow directions. Terra Nova Granite is a very distinctive rock, and erratics of this lithology have been recognized over the entire region. The source is located to the southwest, and the erratics indicate the movement of glacial ice over Terra Nova National Park towards the northeast.

Similar ice flow directions towards the east/northeast were recognized by Brookes (1989) on the inner part of the Bonavista Peninsula, and towards the north by Vanderveer and Taylor (1987) in the Gander region. The orientation of striations and landforms throughout northeast Newfoundland reflect radial flow away from an ice centre located over Middle Ridge (Rogerson 1982; St. Croix and Taylor 1991).

Chapter 6

Glaciofluvial Landforms and Sediments

6.0 - Glaciofluvial Landforms

The surficial sediments of Terra Nova National Park and vicinity are dominated by glaciofluvial sands and gravels. The glaciofluvial landforms are commonly located on the floors of the main valleys throughout the study area. The largest and most complex landforms are located along Big Brook valley adjacent to Route 301. The landforms record the movement of meltwater through the glacial system and in proglacial areas. Where possible, descriptions of glaciofluvial landforms were made whilst in the field. Other landforms could not be reached by road and were recognized and described from aerial photographs.

Eskers are formed in a sub, en, or supraglacial positions by the movement of meltwater through tunnels at the margin of the ice. Six eskers recognized in Terra Nova National Park and vicinity were discussed in Chapter 4. Their structure and internal sediments indicate deposition in subglacial tunnels close to the ice margin as it retreated towards the west/southwest.

There are three large flat-topped terraces and eight smaller ridges in Big Brook valley that are composed of sand and gravel. Although no exposures were found within these features, the terraces appear to have originally been part of one flat-topped landform that may have been incised by glaciofluvial or fluvial flow. An esker lies approximately 350 m southwest of the large flat-topped terraces, suggesting that the ice margin may have been adjacent to the depositional site of the sand and gravel. The esker appears to be unaffected by terrace formation. The original terrace was deposited in a proglacial position by rivers close to the ice margin. Subsequent incision of these rivers, possibly by an increase in discharge or more likely a change in the base level, formed the individual terraces seen today. The esker may have been protected from the subsequent glaciofluvial deposition and incision by a tongue of stagnant ice that remained along the valley wall. Two of the larger flat-topped terraces are separated by Big Brook, and two of the smaller ridges are separated by a tributary of Big Brook. The location of these rivers provides further evidence that the ridges were separated by glaciofluvial or fluvial incision.

Adjacent to Terra Nova Lake there are two flat-topped, relatively steep-sided terraces that are approximately 20 m high and are composed of medium to fine grained sand. There is little coarse material within the deposits. The internal structures of these features were studied in detail and are discussed in Section 6.1.2.

6.1 - Glaciofluvial Sediments

There are several large accumulations of glaciofluvial sediments throughout the area and a total of 20 sections were described. Five exposures will be described in detail here (Figure 6.1). These sections are representative of the glaciofluvial exposures within the area.

6.1.1 - Sections A11b and A9

Units of clast-supported gravel were recognized at two locations (A9 and A11b) in the vicinity of Terra Nova National Park. The section at A11b is laterally and vertically more extensive and will therefore be described in more detail.

Section A11b

Description

A11b is located 6 km west of the Trans Canada Highway, and roughly 1 km north of Route 301 on the south side of Big Brook valley. The section is 13 m thick and approximately 20 m wide. The base of the section is covered by colluvial debris, and therefore only the upper 4 m could be analysed in detail (Plate 6.1). Four beds of coarse-textured clast-supported gravel were recognized in the upper 4 m and all beds dip 11° towards the south (Figure 6.2). Contacts between each bed are gradational. The clast-supported beds unconformably overlie approximately 2.5 m of pebbly sand, which in turn overlies 6.5 m of fine-grained sand.

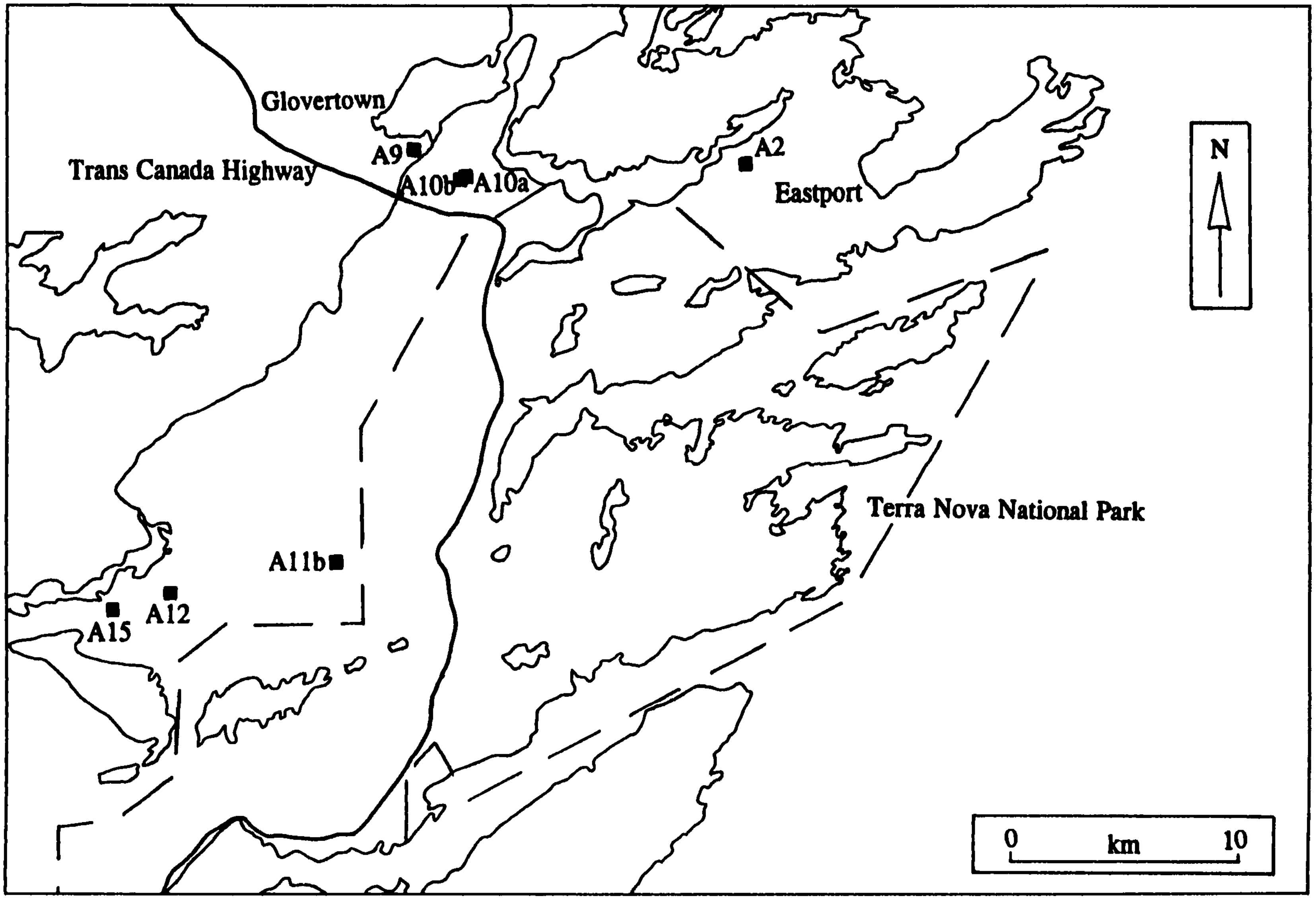

Figure 6.1 - Map showing the location of all glaciofluvial deposits discussed in the text.

Plate 6.1 - The coarse clast-supported gravel beds in Section A11b. The well-exposed section is 4 m thick.

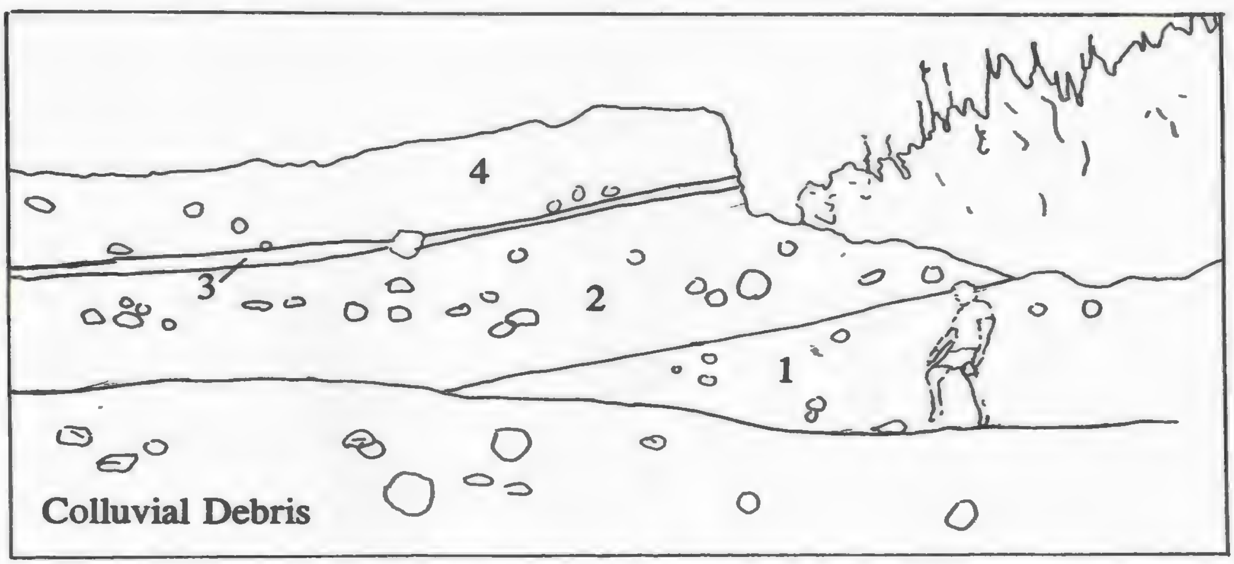

Figure 6.2 - Clast-supported gravel beds in section A11b. Clast fabric analysis was undertaken on beds 1, 2, and 4.

The lowermost bed is 1.65 m thick and its lateral exposure is restricted to approximately 5 m due to the build up of colluvial debris to the west of the exposure. The matrix is mainly coarse sand and has a reddish brown colour when dry and weathered (Munsell: 5YR 5/3). The unit is dominated by cobbles and the clasts have a modal long axis length of 15 cm. The clasts are subrounded and lithology is dominated by Terra Nova Granite. One clast fabric was measured 0.55 m from the top, and 0.5 m from the bottom of the bed. The fabric has an S_1 value of 0.659 and a K value of 0.40, typical of a weak girdle plot (Figure 6.3, Woodcock 1977). The mean trend orientation is 279.7° and the mean plunge is 18.7°.

The second bed is 1.4 m thick and only 5 m of the basal contact can be observed. The remainder of the basal contact is concealed by colluvial debris, however the upper 0.9 m of the bed is laterally exposed. The matrix is composed of coarse sand and fine granules, and has a dark reddish brown colour when dry and weathered (Munsell: 5YR 3/2). The clasts are subrounded and of local provenance. They are slightly larger than the clasts in the bed below, with a modal clast long axis length of 20 cm. One clast fabric was measured within the bed, 0.8 m and 0.1 m from the top and bottom of the bed respectively. The fabric provided an S_1 value of 0.731 and a K value of 1.08. The mean trend orientation was 291.7° and the mean plunge was 19.5°. The values are representative of a weak cluster plot (Figure 6.3), as defined by Woodcock (1977).

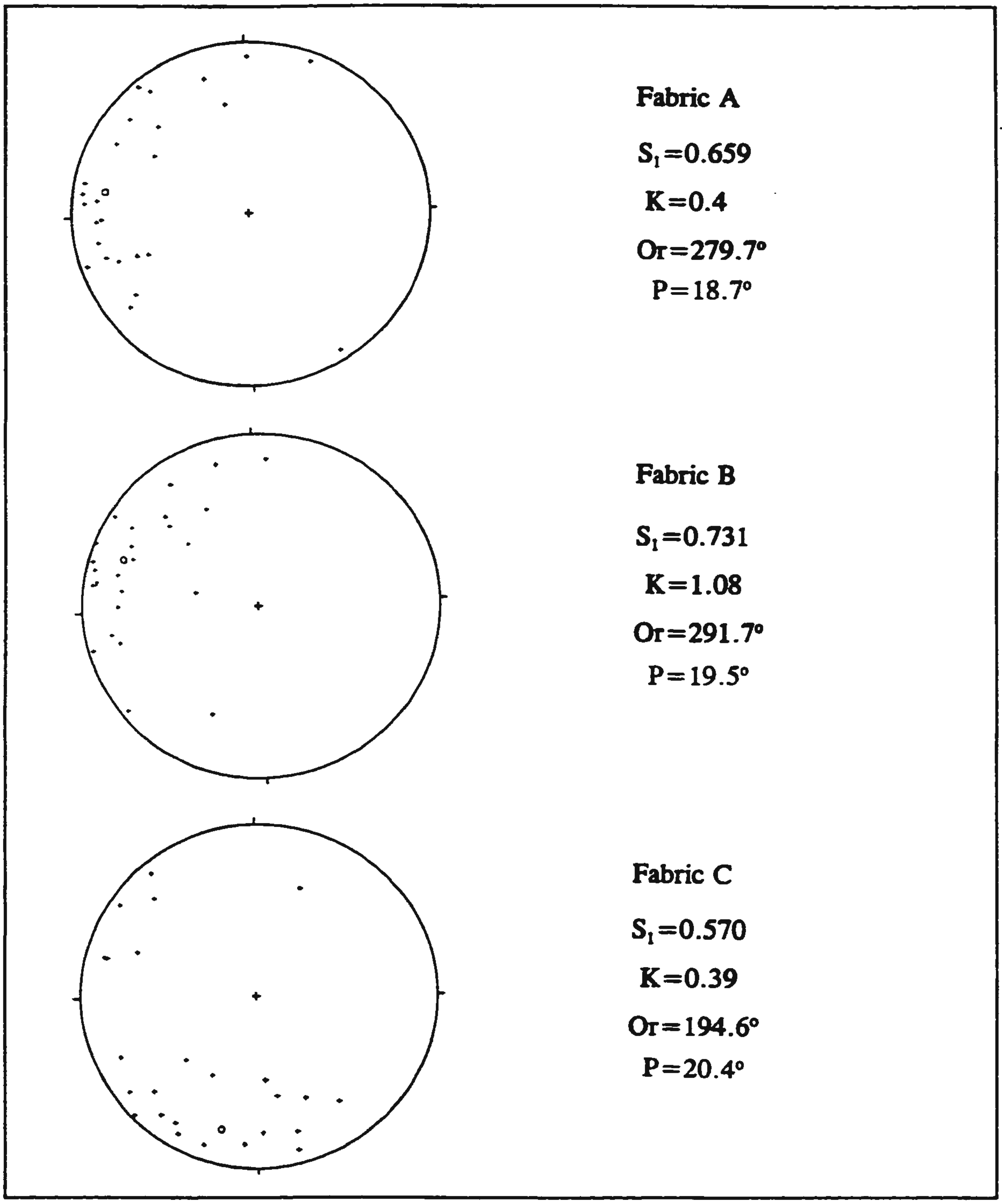

Figure 6.3 - Stereoplots of fabrics measured at Section A11b.

This unit is overlain by a 0.1 m thick clast-supported pebble gravel. The clasts have a modal long axis length of 4 cm. The matrix is sandy and has a reddish brown colour when weathered and dry (Munsell: 5YR 4/3). No internal structure was observed within the bed. Clast fabric analysis was not undertaken due to the thinness of the bed and the small size of the clasts.

The upper unit is 1 m thick and due to its base being unobscured by colluvial debris it is exposed over approximately 20 m. The matrix is sandy and has a reddish brown colour when dry and weathered (Munsell: 5YR 5/4). Clast lithologies are dominantly local and the clasts have an average length of 15 cm, similar to clasts of the lowermost unit. One clast fabric was measured 0.9 m from the top of the bed. The fabric has an S_1 value of 0.570 and a K value of 0.39, and is typical of a girdle plot, as shown in Figure 6.3 (Woodcock 1977). The mean trend orientation is 194.6° and the mean plunge is 20.4°.

Section A9

Description

The sand and gravel pit at A9 is located 1.6 km south of Glovertown South, adjacent to the Terra Nova River. The section is approximately 5 m thick and it is laterally continuous over 100 m. Two units were recognized within the section, separated by a sharp and abrupt contact. The lower 2 m is composed of a fine-textured sandy matrix-supported diamicton that has been interpreted as a glaciogenic sediment flow (see Section 5.3.3, Plate 5.3, Chapter 5). The upper 3 m unit of the section is characterized by a coarse-textured clast-supported cobble gravel. The matrix is sandy and has a light brown colour when weathered (Munsell: 7.5YR 6/4). The clasts are subangular, have an average length of 20 cm, and are mostly Terra Nova Granite. Three clast fabrics were measured at the section. All the fabrics were measured approximately 0.75 m from the base of the unit. Fabric B was located 10 m west of Fabric A on a part of the section oriented transverse to the Fabric A exposure, and Fabric C was obtained approximately 15 m east of Fabric A. The S_1 values are 0.710, 0.770 and 0.692, and the K values are 0.67, 0.78 and 1.19, respectively. The fabrics have girdle plots (Figure 6.4), although the last fabric is close to the girdle/cluster transition, as defined by Woodcock (1977). The mean trends are 54.9°, 118.8° and 67.2°, and the mean plunges are 16.6°, 16.1° and 19.6°, respectively.

Interpretation of Sections A11b and A9

The coarse clast-supported gravels in sections A11b and A9 are interpreted to represent the remnants of braided river bars. Proximal braided river deposits are dominated by coarse gravel and very little fine sediment (McDonald and Banerjee 1971; Rust 1975; Cant 1982; Rust and Koster 1984; Smith 1985). Boothroyd and Ashley (1975) and Collinson (1986) discussed the internal structures associated with longitudinal and transverse bars. Longitudinal bars tend to be coarse-grained and

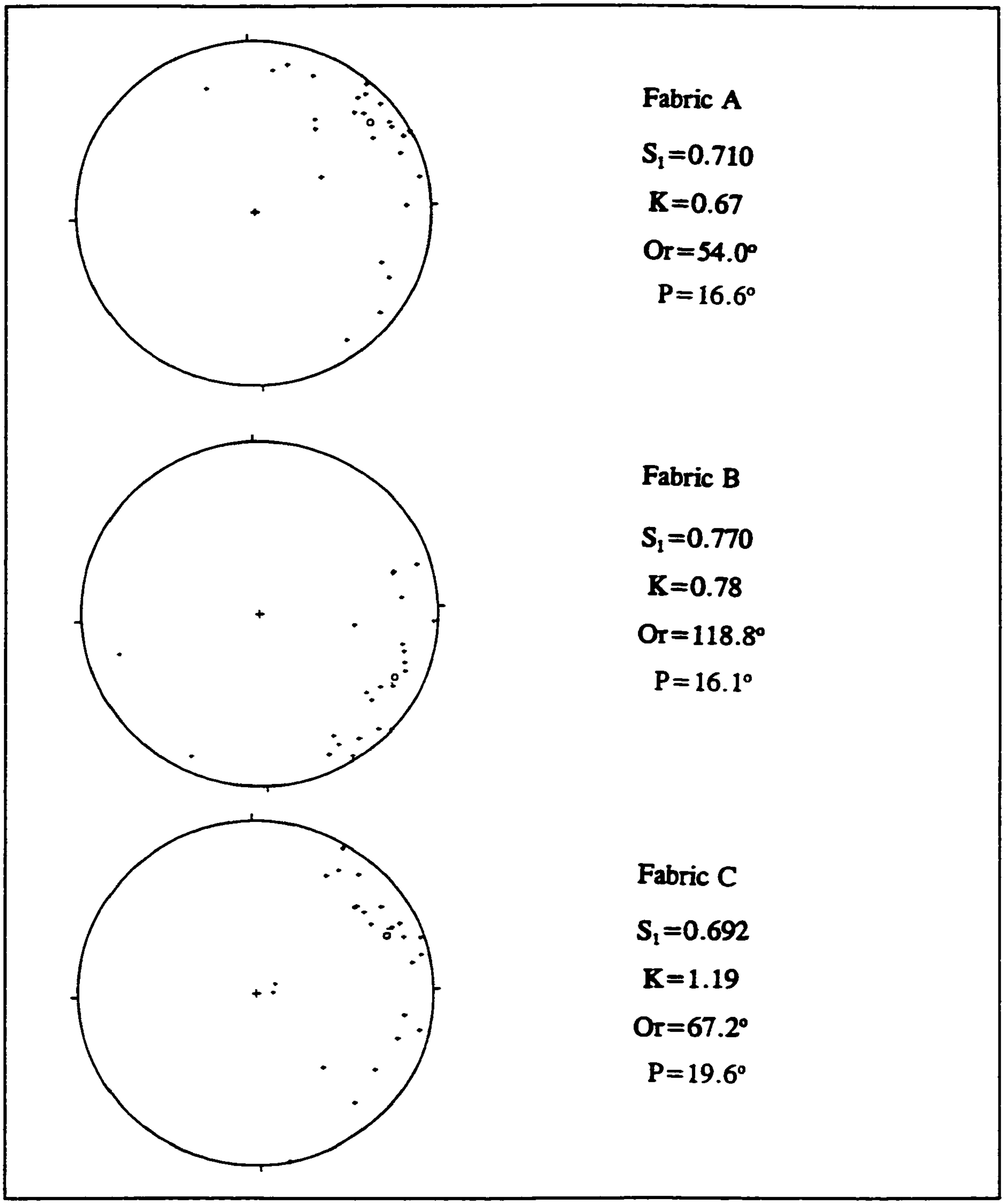

Figure 6.4 - Stereoplots of fabrics measured at Section A9.

poorly sorted, and although they are generally structureless or crudely bedded, they show well defined clast imbrication. They are typically formed in shallow braided rivers where sediment aggrades quickly, and consequently tend to accrete vertically rather than horizontally (Rust 1984; Fraser 1993). No cross-stratification develops and therefore longitudinal bar deposits have very little internal structure. Although transverse bars are commonly formed in fine-grained sediments, they have been observed in gravel-bed streams (Smith 1974, 1985; Collinson 1986). The bars migrate laterally downstream by sediment avalanching over the slip-face and accumulation in front of the bar. This produces cross-stratification that is mainly tabular and dips at angles between 20°-30° (Rust 1972). Due to the migration of channels in braided streams, bars are continuously eroded and deposited, and therefore braided stream deposits are commonly characterized by reactivation surfaces (Smith 1985; Fraser 1993).

Clasts within braided rivers are generally deposited transverse to the flow direction, reflecting the rolling nature of clast transportation in gravel-bed rivers (Johansson 1965; Rust 1975; Cant 1982; Smith 1985; Collinson 1986; Morison and Hein 1987). Longitudinal bars are assumed to have strong fabrics transverse to the flow direction. Tandon and Kumar (1981) and Morison and Hein (1987) found that fabrics from braided stream deposits showed relatively large deviations from the channel orientation. Tandon and Kumar (1981) concluded that some of the clasts were reoriented transverse to new flow directions as the discharge decreased after a flood event. Morison and Hein (1987) reached similar conclusions, although they suggested that crudely bedded gravel may be more characteristic of sediment gravity flows. The deposits at A11b and A9 have relatively high S_1 values indicating that the clasts have a strong preferred orientation. This is typical of braided river deposits (Cant 1982; Harms *et al.* 1982; Brennand and Shaw 1996). Clasts within the lower two units of A11b have typical longitudinal bar fabrics, oriented transverse to the postulated flow direction, towards the northeast. The upper unit fabric is oriented parallel with the northeast trend of the valley. It is suggested that the upper unit was deposited in a longitudinal bar by water flowing in a different direction than that which deposited the lower units.

I therefore propose that the section at A11b represents the deposits of a longitudinal bar with the lower units preserving the transverse nature of the clasts that were deposited during a flood event, and the upper unit representing deposition on the surface of the bar as the flood waned and the channel changed directions.

The fabrics measured at A9 were all taken at approximately the same height, however, Fabrics A and B are oriented parallel with the valley, and Fabric C is oriented transverse to the valley. It was suggested by Jenness (1963) and Dyke (1972) that during the Late Wisconsinan the mouth of the Terra Nova River was situated near Traytown, forming the delta observed today. This suggests that the river may have been flowing across rather than parallel with the modern valley.

The modern Terra Nova River has considerable variation in channel orientation. Thus, by analogy the fabrics at A9 may represent a shift or a bend in the main channel orientation as the river made its way towards the coast.

Boothroyd and Ashley (1975) discussed grain sizes on the Scott and Yana outwash plains, Alaska. The first 5 km from the source were characterized by the largest clasts that had a modal long axis greater than 10 cm. A braided river developed in the next 5 km, and erosion of the clasts resulted in a large reduction in grain size. At 16 km from the ice front, sand was the dominant sediment size. The clasts in sections A11b and A9 are of comparable lithology with the braided outwash deposits as described by Boothroyd and Ashley (1975). Scott *et al.* (1991) concluded that deltaic sediments in the Springdale-Hall's Bay area were deposited 1-2 km from the ice front. The clasts varied in size from 1-12 cm and the majority of the clasts were subrounded. The clasts in sections A11b and A9 have a modal long axis length of 17.5 cm, and most are subrounded.

I therefore propose that the sediments at A11b and A9 were deposited by braided rivers in sandur environments relatively close to the ice front. Due to the coarse texture, absence of internal structure, and the fabrics of the sediments, they are interpreted as the erosional remnants of longitudinal bars that were formed during periods of high flow. There is no cross-stratification at either of the sections, indicating that vertical aggradation was more important than horizontal aggradation, characteristic of a shallow braided river system.

6.1.2 - Sections A12 and A15

A12 and A15 are located in gravel pits approximately 13.5 and 16 km west, respectively, from the Trans Canada Highway and roughly 250 m north of Route 301. The pits are dominated by sand with very little coarse material. The pit at A15 is 150 m wide and 21 m high. The sand is actively being eroded by ATV and motorbike activity. Much of the exposed face of the pit is covered by an eroded layer of sand that conceals the underlying structures (Plate 6.2). The smaller pit at A12 is approximately 8.5 m high and 20 m wide. There are good exposures throughout the pit (Plate 6.3) and therefore it was analysed in detail. Due to soil formation at the top of the exposure the upper 1 m of the section was not analysed. Twelve units were recognized within the remaining section (Figure 6.5).

Unit 1

Unit 1 is 2 m thick, although the base of the unit was not observed, and is composed of medium to fine grained, well-sorted sand. The unit is characterized by faint horizontal bedding that is identified by differences in grain size.

Unit 2

Unit 2 is 0.34 m thick and consists of well-sorted medium grained sand. Climbing ripples are found throughout the unit. Each bed of climbing ripples is approximately 1.5 cm thick, and they climb at an angle of 14°. The ripples are asymmetrical and only the lee side of the ripples is preserved. The bounding planes

Plate 6.2 - The exposed and eroded face at Section A15 showing horizontal bedding and trough cross-bedding. The section is 21 m high, 100 m wide and faces towards the east. Beds within the exposure dip at an angle of 15° towards the northeast.

Plate 6.3 - The well-exposed face at Section A12 showing trough cross-bedding, horizontal bedding and climbing ripples. The exposed section is 8.5 m high, 15 m wide and faces towards the northeast. Beds within the exposure dip 24° towards the northeast. The section in Figure 6.5 is located 5 m west of the exposure shown above.

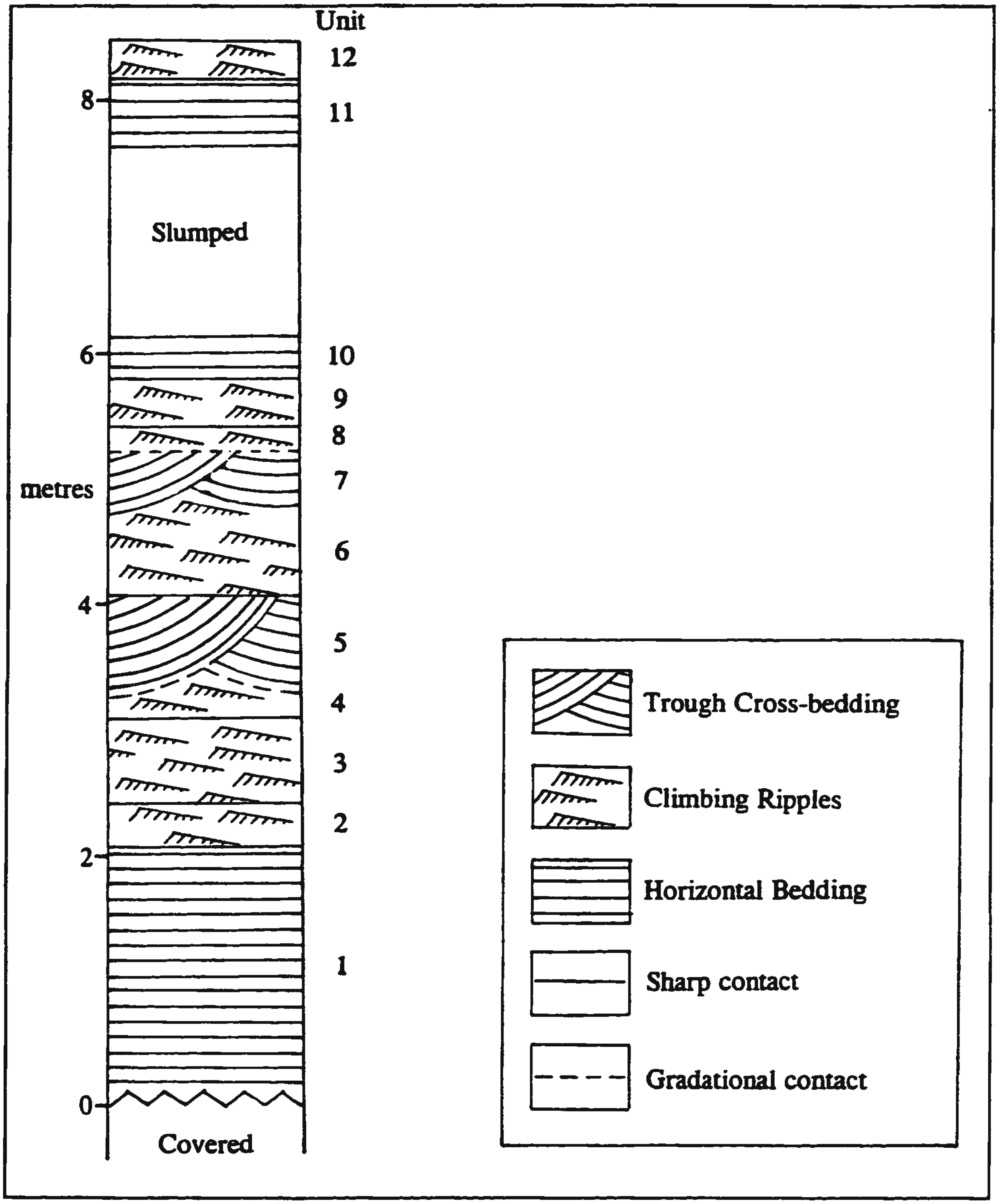

Figure 6.5 - Stratigraphic log of Section A12.

between the beds and the lee side of the ripples are defined by a thin layer of silty sand. The lee faces dip at an angle of 36° and flow directions are towards the northeast.

Unit 3

Unit 3 is 0.67 m thick and composed of medium to coarse textured, well-sorted sand. Individual beds throughout the unit are between 2 and 4 cm thick and most have a sinusoidal structure. The bounding planes between the individual beds are accentuated by a thin layer of silt. Each bed is characterized by climbing ripples. Although the majority of the climbing ripples only have lee side preservation dipping at 30°, some are gently curved at the top of the bed. A few of the beds have a convexo-planar contact with the beds below, and the climbing ripples within these underlying beds have been truncated. The convexo-planar shape suggests that flow directions may have been towards the north or north-northeast.

Unit 4

Unit 4 is 0.23 m thick and is characterized by medium textured, moderately sorted sand. Climbing ripples are found throughout the unit, climbing at an angle of 10°, and lee faces dip at 38° towards the northeast. The structures within this unit are very similar to those described in Unit 2.

Unit 5

Unit 5 is 0.77 m thick and composed of moderately sorted, medium to coarse grained sand. The unit is trough-shaped and trough cross-bedding dominates the internal structure of the unit. At the base of the unit there is a thin layer (0.01 m) of fine gravel that truncates the climbing ripple lamination of Unit 4. The trough cross-beds are approximately 0.5 cm thick and they dip at an angle of 35° towards the north or north-northeast.

Unit 6

Unit 6 is 0.71 m thick and consists of well sorted, fine to medium grained sand. Individual beds throughout the unit are approximately 1.5 cm thick and are defined by thin layers of silt along the upper and lower contacts (Plate 6.4). The beds are characterized by climbing ripples with lee side preservation. The lee faces dip 28° and are highlighted by changing grain sizes. The climbing ripples in this unit are similar to those in Unit 2 and suggest that flow was towards the northeast.

Unit 7

Unit 7 is 0.43 m thick and composed of moderately sorted, medium to coarse sand. Two sets of trough cross-beds were observed in the unit dipping 10° and 33°. The higher angled trough cross-beds truncate the lower angled trough cross-beds and form a distinct trough shape within the unit. Both sets of trough cross-beds indicate that the current flow was towards the north-northeast. Individual beds within the unit are about 2 cm thick and coarsen upwards from medium to coarse sand.

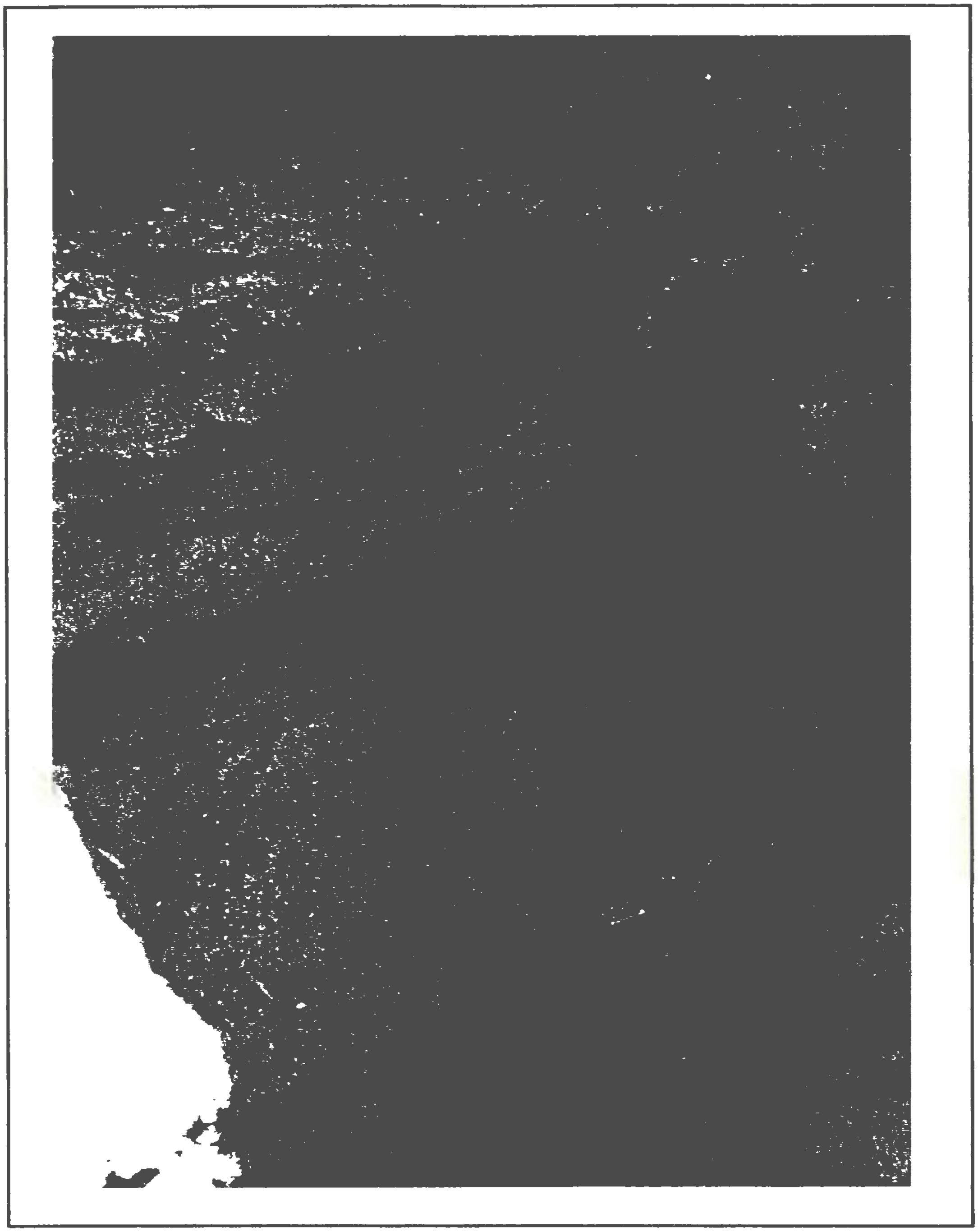

Plate 6.4 - Climbing ripples accentuated by thin layers of silt along the upper and lower contacts, Unit 6 (A12). The ripples were developed by water flowing towards the northeast (right to left). The trough handle is approximately 12 cm long.

Unit 8

Unit 8 is 0.20 m thick, moderately sorted and fines upwards from medium sand to silt. There are climbing ripples throughout the unit climbing at 12°, and the individual beds are between 0.5 and 3 cm thick. Thin silt layers drape the individual beds and occasionally define lee side preservation. The lee faces dip 35° and suggest that the current flow direction was towards the northeast. Many of the beds are truncated by overlying beds indicating that the sediment below was reworked to form the overlying beds. The upper 6 cm of the unit is composed of silt and has little or no internal structure.

Unit 9

Unit 9 is 0.38 m thick and is composed of moderately sorted, medium grained sand. Throughout the unit there are poorly defined climbing ripples. The climbing ripples are interrupted by a layer of rippled silt 1 cm thick.

Unit 10

Unit 10 is 0.33 m thick and is characterized by moderately sorted, medium to coarse sand. Beds within the unit are horizontal and are defined by textural differences. Although this unit is coarser, it is otherwise similar to Unit 1.

Unit 11

Obscuring the contact between units 10 and 11 there is approximately 1.5 m of colluvial debris. Unit 11 lies above the colluvial debris. The unit is 0.5 m thick and it is composed of moderately sorted, medium to coarse sand. It is horizontally bedded and faint individual beds are recognized by differences in grain sizes. This unit is similar to Unit 1.

Unit 12

Unit 12 is 0.34 m thick and characterized by well sorted, fine to medium sand. Throughout the unit there are several beds of climbing ripples. The beds are approximately 0.05 m thick and climb at an angle of 12°. Only the lee face is preserved within the beds and these dip 30° towards the northeast. The climbing ripples in Unit 12 are similar to those described for Unit 2.

Interpretation

The twelve units described above can be placed into three groups based on the sedimentary structures within the units. The groups are (1) horizontal bedding, (2) climbing ripples, and (3) trough cross-bedding. Each group of sedimentary structures will be interpreted rather than the individual units to prevent repetition.

Horizontal Bedding

Horizontal bedding was observed in Units 1, 10 and 11, and may have been deposited by suspension settling from turbidity currents, by low relief bedforms in very shallow rivers, or by upper or lower flow regime currents (cf. Reineck and Singh 1980).

Bouma (1962) developed a facies model for sediments and structures that area associated with turbidity currents. Within the Bouma sequence, plane laminations tend to form above structureless or graded beds that are deposited by very rapid deposition from suspension settling preventing the formation of bedforms. Experimental work by Kuenen (1966) concerning turbidite lamination found that alternating fine and coarse laminae were produced by traction carpets when there was a constant flow and a constant supply of sediment. As the flow velocity decreased the rate of suspension settling also decreased and upper flow regime planar beds were then formed (Kuenen 1966).

Smith (1971) discussed the internal structure of low amplitude sand waves that formed in very shallow flows. He suggested that due to the planar nature of the stoss slopes, horizontal lamination was preserved as the next sand wave migrated across the bed. Although these were formed in a lower flow regime they may be confused with upper flow regime planar beds. Smith (1971) recognized alternating coarse and fine layers in these deposits, however within the upper flow regime there is no such sorting and the size of the sediment being transported and the bedload material are essentially the same, producing a more homogenous deposit. I suggest that the horizontal beds in Units 1, 10 and 11 were not formed by this process due to the almost homogenous nature of the units with relatively little sorting.

Horizontal bedding can also be produced in lower or upper flow regime environments (Lindholm 1987). A limited amount of sediment is transported in the lower flow regime due to a large resistance to flow (Middleton and Hampton 1976). Small sand particles are light enough to be entrained by the low current velocities and small ripples are commonly formed. Coarse sand, however, is too heavy to be entrained and can only produce horizontal bedding (Lindholm 1987). In the upper flow regime environment flow velocities are high and there is a large amount of sediment in transport (Middleton and Hampton 1976). Cheel and Middleton (1986) discussed the formation of fining and coarsening upward horizontal laminae in an upper flow regime environment by burst and sweep processes. Bursting involves the slow ejection of fluid away from the bed, increasing the amount of sediment in suspension, and leading to the formation of fining upwards laminae as a result of suspension settling. Sweeping involves the rapid movement of fluid towards the bed creating higher shear stresses on the bed surface. The coarsening upward laminae are formed due to the larger grains experiencing greater pressures during a sweep and migrating upwards to areas of lower pressure (Cheel and Middleton 1986).

The horizontal bedding in Units 1, 10 and 11 is interpreted to have been deposited in an upper flow regime environment. Due to the fine to medium texture of the sand it is thought that in a lower flow regime environment ripples would have formed and there would be no horizontal bedding. An origin as low amplitude sand waves is rejected on the basis that alternating beds of fine and coarse sand are not readily apparent within Units 1, 10 and 11. The sediment may have been deposited by suspension settling from a turbidity current or by burst and sweep processes in an upper flow regime current. Based on the sorting, grain size, faint bedding and other sedimentary structures throughout the exposure I conclude that the horizontal bedding was deposited by burst and sweep processes that formed in an upper flow regime environment.

Climbing Ripples

Climbing ripples were observed in Units 2, 3, 4, 6, 8, 9, and 12, and they are interpreted to have been formed in flowing water where there was an ample supply of suspended sediment (Reineck and Singh 1980; Fritz and Moore 1988). They are formed by the migration and vertical accretion of ripples, and flow direction can be inferred from the nearly parallel bounding planes that dip upstream (Reineck and Singh 1980; Ashley *et al.* 1982). The climbing ripples observed in all units of A12 are typical of the Type A ripples of Jopling and Walker (1968). Type A ripples are characterized by lee side preservation only, indicating that there was either not enough sediment in suspension to preserve the entire ripple structure, causing the stoss side of the ripples to be eroded, or higher velocities caused the stoss side of the ripples to be eroded (Jopling and Walker 1968; Reineck and Singh 1980; Ashley *et al.* 1982).

The normal grading observed in Unit 8 indicates that either the source of sediment changed or there was a decrease in the flow energy and only the finer sediment could be transported. Due to the lack of internal structure within the upper part of the unit I propose that the flow velocity decreased allowing the finer sediment to settle out of suspension.

Trough Cross-Bedding

Trough cross-bedding was observed in Units 5 and 7, and is interpreted to have been deposited by migrating dunes or ripples, or by the infilling of eroded channels (Reineck and Singh 1980).

Harms and Fahnestock (1965) and Williams (1968) suggested that trough cross-bedding formed as a result of dunes migrating downstream into pre-eroded scours. The scours were formed by jets that started at the dune crests and impinged on the bottom of the channel farther downstream (Harms and Fahnestock 1965), or by re-entrants at the front of the dunes that formed an erosional hollow (Williams 1968). As the dunes migrated the sediment that avalanched down the lee face of the dunes infilled the erosional hollows, forming curved strata that conformed to the eroded scour shape.

McKee (1957) and Sutton and Watson (1960) discussed the formation of trough cross-bedding by the infilling of eroded channels. The flume experiments of Mckee (1957) identified two types of trough cross-stratification. Those formed by the erosion of a channel in a stream tended to be flat based and the channel fills were horizontal. In contrast, channels formed by submerged currents were infilled with curving strata that followed the outline of the erosional contact. Sutton and Watson

(1960), however, discussed the formation of "festoon-bedded grits" (p 109) in a river environment, and suggested that the channels would have to be developed before the sediment could be deposited. Based on the eroded remnants of the trough structures, they concluded that the water was relatively shallow and that erosion and deposition was by sheet flood or shallow braided rivers.

Harms *et al.* (1963) discussed the orientation of large and small scale trough cross-stratification from a point bar on the Red River, Louisiana. They concluded that the trough axes were oriented parallel to the flow direction and that as the river curved and changed its orientation so too did the trough axes. Using this information from modern rivers, Harms *et al.* (1963) suggested that flow directions can be inferred from trough cross-stratification within ancient records.

Trough cross-stratification within Units 5 and 7 are interpreted to have been deposited by the infilling of eroded channels. The thin layer of gravel at the base of Unit 5 and the coarser sediment within each unit indicates that there was an increase in the flow energy that initially formed an eroded channel before sediment was deposited. Although the entire trough structures were not observed within the section, the palaeoflow can be inferred to have been towards the north or north-northeast.

Discussion

On the basis of the sediment textures and structures within the exposure at Section A12, the succession has been interpreted as the product of a unidirectional current that fluctuated between an upper and lower flow regime, forming the horizontal beds and climbing ripples, respectively. The sediments at the base of the trough cross-bedding are coarser than those in other beds of the exposure, indicating that these currents had sufficient velocity to transport slightly coarser sediment.

Both sections at A12 and A15 are dominated by medium to fine grained sand and there is a distinct lack of coarse sediment. This suggests that the flow velocity was either too low to move coarse sediment, there was no coarse sediment available for transportation, or that the coarse sediment was deposited farther upstream.

The exposures are part of flat-topped terraces with relatively steep sides that are 250 m wide, 1.9 km long and approximately 20 m high. Structures of this kind may have formed in a sub-glacial, ice marginal, proglacial or sandy braided river environment. Deposition in each of these environments is discussed below, and shown in Figure 6.6.

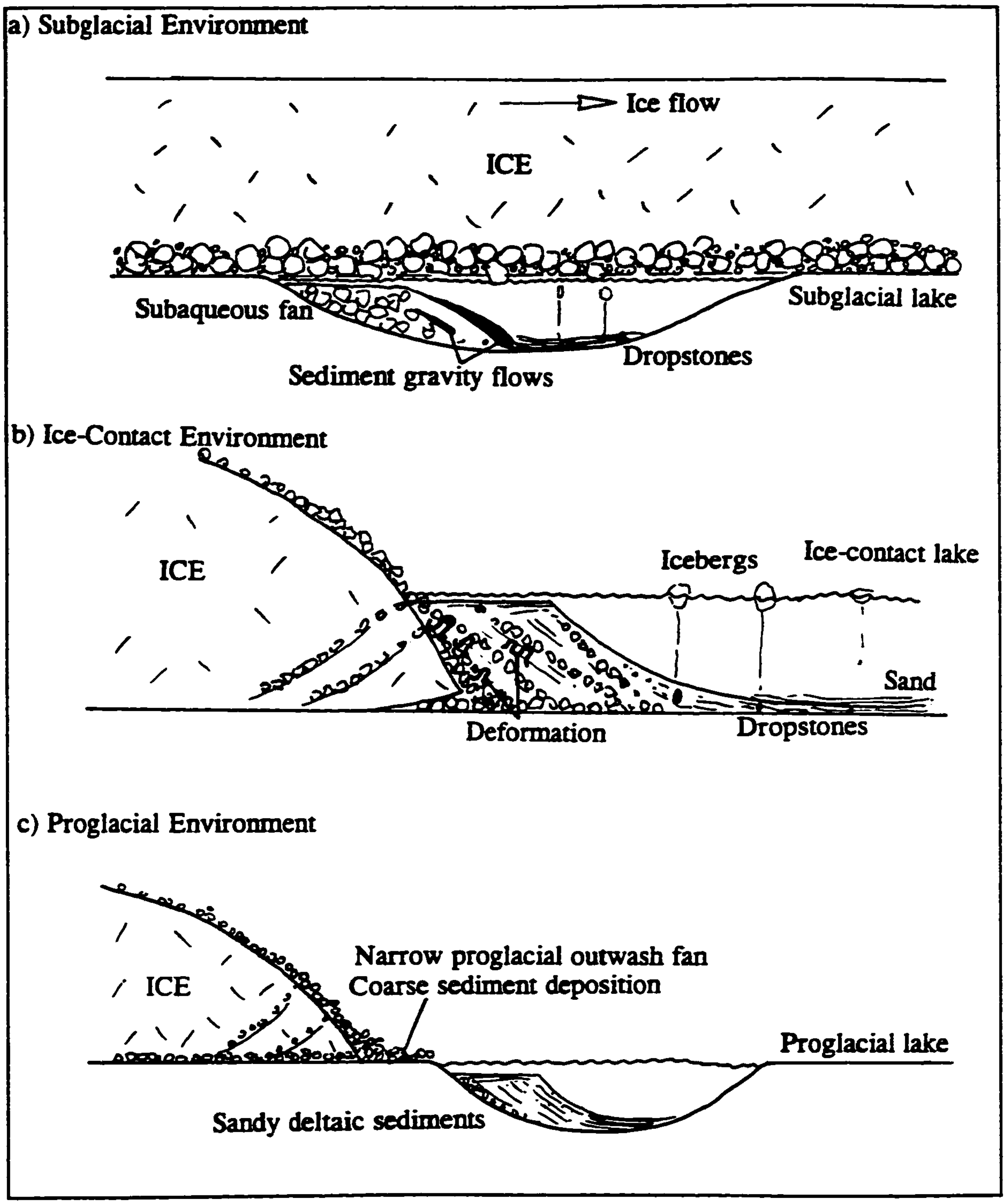

Figure 6.6 - Proposed subglacial, ice-contact and proglacial depositional environments for the large flat-topped sand deposits at A12 and A15

Subglacial Environment

Subglacial lakes are known to exist beneath the ice masses of Vatnajökull, Iceland (Thorarinsson 1953; Björnsson 1992), and Antarctica (Oswald and Robin 1973; Ridley *et al.* 1993). The development of a lake in a subglacial position requires a temperate ice mass producing large quantities of meltwater that pond within a depression in the substrate commonly under thin ice in the ablation zone (Brodzikowski and van Loon 1991). Subglacial streams continue to debouch into the lakes once they have formed depositing large quantities of sediment (Figure 6.6a).

McCabe and Cofaigh (1994) discussed sedimentation in a subglacial lake that developed in the lee of a bedrock ridge at Enniskerry, eastern Ireland. The subaqueous fans were composed of coarse boulder gravel overlain by cross-bedded sand and foreset gravel. Occasional cut and fill structures indicated erosion and deposition by subglacial meltwaters. At the top of one of the exposures there was glaciotectonised gravel suggesting that the upper beds were deformed by the drag of the ice above. There was a lack of fine grained sediment in the exposures at Enniskerry so McCabe and Cofaigh (1994) suggested that the finer sediment was transported through the lake. Brodzikowski and van Loon (1991) discussed the finer sediments associated with distal deposits in subglacial lakes. They suggested that the sediments are dominated by silt but they are commonly disturbed by dropstones falling in from the ice roof above, and by sediment gravity flows that originate on the delta slope.

Ice-Contact Environment

At the ice terminus the sediment may be deposited in an ice-contact position into a glaciolacustrine (Figure 6.6b) or glaciofluvial environment, or by mass movements off the glacier snout.

Orombelli and Gnaccolini (1978) discussed the sediments deposited in an ice-contact delta in the Italian Alps. The proximal delta slopes were composed of poorly sorted sand and gravel, commonly interbedded with lodgement till that was deposited on the ice-contact slope of the delta as the ice front oscillated. Farther down the delta slope the sand and gravel alternated with beds of medium to fine sand that showed some cross-lamination and climbing ripples. Occasional lenses of diamicton within these medium sands were interpreted to be flow tills. The distal delta slope was characterized by climbing ripples of fine sand, and these beds graded into fine grained rhythmites deposited by suspension settling onto the lake floor.

Similar sediments were described by Weddle (1992) and Hart (1996) who identified coarse grained ice-proximal deposits that were deposited by gravity flows, subaqueous sediment discharge and fluvial deposition. These deposits were occasionally interbedded with lodgement till. Distal sediments were much finer and deposited by suspension settling, and grain flows. Deformation structures were found throughout both sections, and were attributed to have been formed by sediment gravity flows and/or direct action by the ice readvancing and pushing into the deltaic sediments (Weddle 1992; Hart 1996).

Dropstones and coarse grained dump structures interpreted to have been deposited from icebergs are commonly found within the intermediate and distal fine grained glaciolacustrine deposits (Thomas and Connell 1985; Benn 1989; Weddell 1992; Hart 1996). These deposits are significantly coarser and commonly deform the surrounding sediment (Thomas and Connell 1985).

Ice-contact fluvial deposits are deposited directly off the glacier terminus by meltwater streams. The sediments are characterized by coarse clast-supported gravel that is deposited in longitudinal bars (McDonald and Banerjee 1971; Miall 1983; Fraser 1993; Maizels 1993). Finer sediments are deposited during periods of low or waning flow commonly within the core of a bar, as a channel fill or within pools adjacent to the main channel (McDonald and Banerjee 1971; Miall 1983; Maizels 1993). The sand units are interbedded with the coarse fluvial deposits and are characterized by planar and trough cross-bedding, climbing ripples and horizontal laminations (Maizels 1993).

Ice-contact mass movement sediments can be deposited in a subaerial or subaqueous environment. Subaerial mass movement sediments are deposited by slumping and sliding off the glacier snout forming irregular bodies within glaciolacustrine and glaciofluvial sediments (Brodzikowski and van Loon 1991). The sediments are poorly sorted, structureless, matrix-supported diamictons with the clasts having no preferred orientation. Subaqueous mass movement deposits form by slumping, sliding and debris flows off the glacier into an ice-contact lake. As with the subaerial deposits these sediments are often poorly sorted and the clasts have no preferred orientation. Throughout the sediments there are many fluidization structures, water escape structures and load casts that are formed as the sediment moves down the ice-contact delta slope (Brodzikowski and van Loon 1991). The deposits form lens-shaped bodies and are interbedded with glaciolacustrine deposits.

Proglacial Environment

The proglacial environment is characterized by meltwater streams and proglacial lakes that are unaffected by the direct action of glacial ice (Brodzikowski and van Loon 1991). There are fewer lakes in the proglacial area compared to the ice-contact environment, however they do exist and glacial meltwater streams debouch into them (Figure 6.6c). Brodzikowski and van Loon (1991) discussed sediments associated with proglacial lacustrine deposition. They suggested that a large proportion of the coarse sediment is deposited by the meltwater streams before they reach the proglacial lakes. This results in only gravel, sand and fines being available for deposition in the proglacial lakes.

Shaw (1975) discussed the sediments deposited into Pleistocene ice-marginal lakes close to the ice front. The coarsest sediment in the sections was gravel that was horizontally bedded and commonly overlying scoured surfaces. The majority of the lake deposits were composed of sand that were either structureless, horizontally bedded or displayed Type A and Type B climbing ripple structures, as classified by

Jopling and Walker (1968). With distance from the incoming meltwater source the sediments became finer and exhibited parallel lamination. Shaw (1975) recognized some fining upwards sequences within the deposits and suggested that this was due to the ice front retreating causing deposition in the lake to become more and more distal.

Similar sediments were described by Gustavson *et al.* (1975) who described glaciolacustrine deltas that were composed mainly of ripple-drift cross-laminated sand. Foresets within the deltas dipped at angles less than 15°.

Sandy Braided River Environment

Due to the proglacial lakes storing large amounts of water and sediment the rivers may enter the lake as braided rivers and leave as meandering rivers. This is the result of a regulated outflow and deposition of the coarse sediment within the lake (Brodzikowski and van Loon 1991). Distal outwash rivers are commonly composed of cross-bedded or pebbly sand (Boothroyd and Nummedal 1978; Miall 1983).

Sandy braided and meandering rivers were recognized on the distal Scott and Yana outwash fans, Alaska by Boothroyd and Ashley (1975). The lower outwash fans were characterized by low gradients and medium to coarse sand with little or no gravel. The braided reaches of the rivers were dominated by longitudinal and linguoid bars that formed planar cross-beds, and thick units of climbing ripples were formed on the surface of the bars. The meandering reaches were composed of trough and planar cross-beds that formed in the channels and on the bars, respectively. The deposits described by Boothroyd and Ashley were similar to the sandy braided rivers described by Cant and Walker (1978) and Haszeldine (1983).

Summary

Based on the sediment size and structures described for each of the environments above, it is suggested that the Sections at A12 and A15 were deposited in a proglacial lacustrine or fluvial environment. There are no coarse sediments within the exposures and the lack of deformation structures indicates that glacial ice was not close to the site of deposition. The subglacial lake and ice-contact depositional environments are rejected on this basis.

The exposures at A12 and A15 are part of flat-topped terraces with relatively steep sides. Flat-topped features are typical of fluvial deposition suggesting that the deposits at A12 and A15 were deposited in a proglacial fluvial environment (Collinson 1986). The structures within the sections are typical of sandy braided rivers, however the terraces are steep sided suggesting that they are the erosional remnants of a larger feature. Terra Nova Lake is located approximately 600 m northwest of the terraces, and it is suggested that the lake provided a sink for the coarse grained sediments. The sand deposits may represent distal lacustrine deposits that were deposited in a proglacial lake much larger than the lake today. For this to occur drainage farther downstream would have to be impeded preventing drainage of the lake water. The Terra Nova River valley and/or Big Brook valley may have been blocked by ice causing the lake level to rise. The processes and sediments associated with ice-contact environments have been discussed above and there is no evidence of an ice dam, or dams, in the area.

I suggest that the terraces were deposited in a subaerial sandy braided river beyond the margin of a proglacial lake (Terra Nova Lake) that was a sink for the coarse sediment. The sandy sediments are thought to have been deposited when sea level was higher than present (see Chapter 7) causing the river to aggrade. As sea level fell due to isostatic recovery the river was forced to grade to a lower base level. This resulted in the incision of the sandy braided sediments leaving the terraces as remnants of a much larger landform. Similar terraced sediments were described for the Assiniboine and Qu'Appelle valleys by Klassen (1975), and the Chalk River region by Catto *et al.* (1982).

6.1.3 - Section A2

The exposure at A2 is located in Sandringham, approximately 12 km east of the Trans Canada Highway, and roughly 150 m south of Route 310. The section is 4 m high and 25 m wide and is composed of sand and gravel. The upper part of the section exhibits some internal structure, whereas the lower part is more homogenous. Eleven units were recognized in the section and are described below (Figure 6.7).

Unit 1

Unit 1 is 0.7 m thick and is composed of coarse to medium grained, moderately sorted sand that has a yellowish red colour when fresh (Munsell: 5YR 4/6). Throughout the unit there are small clasts with a long axis length of approximately 5 cm, and these along with other textural changes in the sand define faint horizontal bedding. Contacts between individual beds are gradational.

Unit 2

Unit 2 is 0.32 m thick and is characterized by clast-supported cross-bedded gravel. The cross-beds are planar and dip at an angle of 24° towards the northeast. The clasts have an modal long axis length of 7 cm and are primarily Musgravetown sandstones (50%) and Terra Nova Granite (40%); 10% of the clasts are Hare Bay and Square Pond Gneiss. The coarse sand matrix is moderately sorted and has a light reddish brown colour (Munsell: 5YR 6/3). One clast fabric was measured from this bed. The fabric has an S_1 value of 0.642 and a K value of 0.81, typical of a girdle

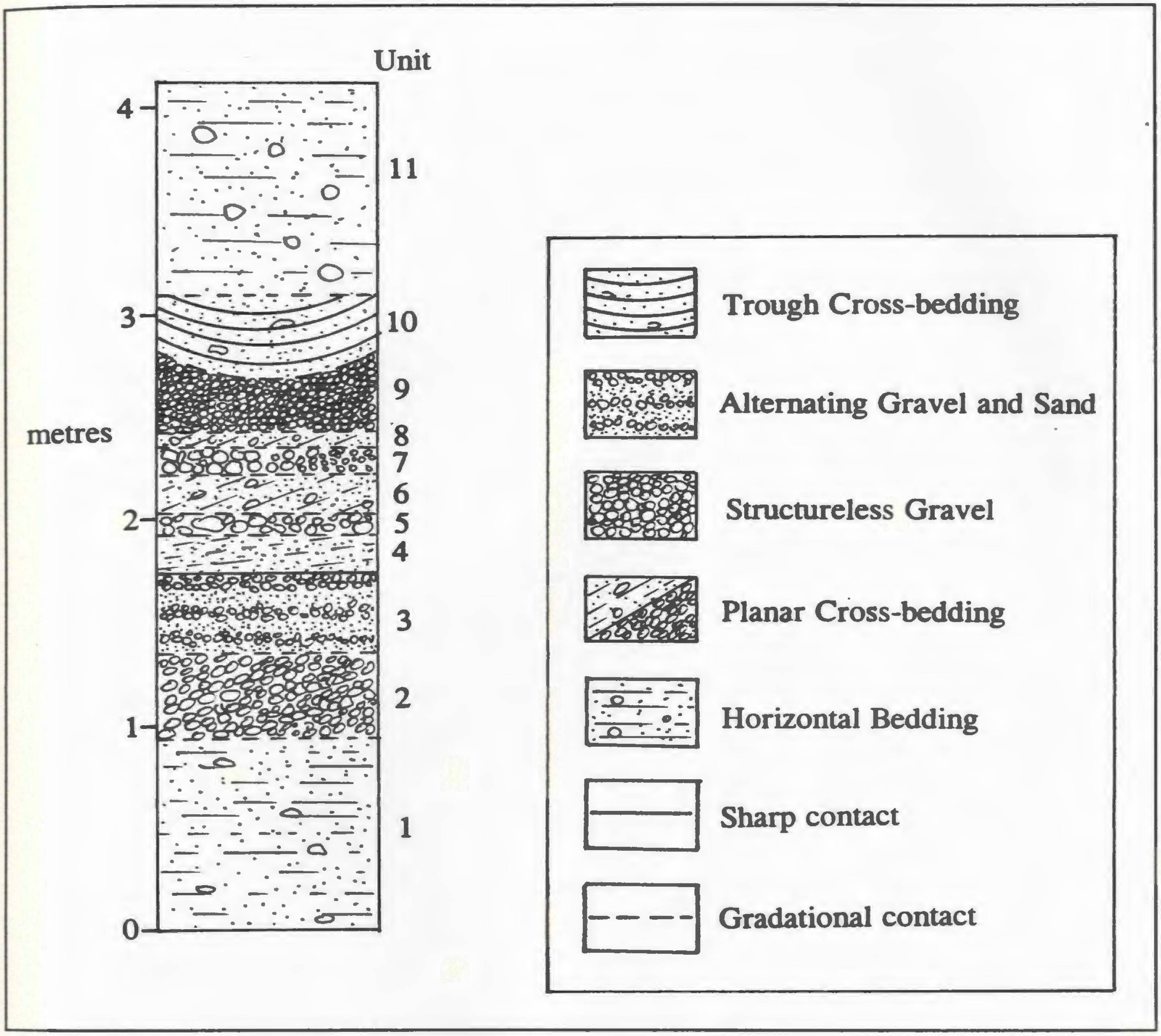

Figure 6.7 - Stratigraphic log of Section A2.

distribution, as defined by Woodcock (1977). The mean trend orientation is 180.4° and the mean plunge is 0.6°.

Unit 3

Unit 3 is 0.29 m thick and consists of alternating beds of medium sand and clast-supported gravel. Six horizontal and planar beds were recognized in the unit and all the contacts are gradational. The sand beds are moderately sorted and generally structureless, however there are occasional small clasts, approximately 5 cm long, throughout the sand. The gravel beds are structureless and have a sandy matrix similar to Unit 2.

Unit 4

Unit 4 is 0.15 m thick and is composed of moderately sorted, coarse sand. Planar cross-bedding dominates the unit and the beds dip at an angle of 5° towards the north. Although this unit is composed of finer sediment it has a similar structure to Unit 2.

Unit 5

Unit 5 is 0.07 m thick and is characterized by clast-supported structureless gravel. The largest clast recovered from the unit has a long axis length of 12 cm, and the matrix is composed of coarse sand with a light reddish brown colour when weathered (Munsell: 5YR 6/3).

Unit 6

Unit 6 is 0.14 m thick and is composed of moderately sorted coarse sand with occasional small pebbles throughout the unit. The unit is dominated by planar cross-beds that dip at an angle of 22° towards the north. Some of the pebbles are imbricated and dip at a similar angle, defining the cross-bedded structure. The structures in this unit are similar to those described in Unit 2.

Unit 7

Unit 7 is 0.1 m thick and characterized by clast-supported gravel with a light reddish brown sandy matrix, similar to that described in Unit 5. The clasts are approximately 6 cm long, and although the unit is structureless, the clasts become smaller towards the east.

Unit 8

Unit 8 is 0.05 m thick and is composed of moderately sorted coarse sand with a few pebbles scattered through the unit. The unit is characterized by planar cross-beds that dip at an angle of 22° towards the north. This unit is similar to Unit 2.

Unit 9

Unit 9 is 0.2 m thick and consists of clast-supported openwork gravel. The pebbles are approximately 2 cm long. There is very little matrix in the unit and therefore the pebbles are loose and easily fall from the section.

Unit 10

Unit 10 is 0.3 m thick and is composed of moderately sorted trough cross-bedded coarse sand. Throughout the unit there are a few small pebbles with long axis lengths of approximately 5 cm. The base of the unit has a shallow trough shape, and truncates the underlying unit. Individual beds within the unit conform to the basal contact and the beds are defined by textural differences. The small pebbles are generally concentrated at the base of the unit.

Unit 11

Unit 11 is 0.78 m thick and is characterized by structureless coarse grained sand with a few small clasts throughout the unit. The clasts have a maximum long axis length of 6 cm. This unit is similar to Unit 1.

Interpretation

The eleven units described above can be placed into five groups based on the sedimentary structures within the units. The groups are (1) horizontal bedding, (2) planar cross-bedding, (3) structureless gravel, (4) alternating gravel and sand, and (5) trough cross-bedding. Each group will be interpreted rather than the individual units to avoid repetition.

Horizontal Bedding

Horizontal bedding was observed in Units 1 and 11, and may have been deposited by suspension settling from turbidity currents, by upper or lower flow regime currents, or by the migration of low relief bedforms in very shallow rivers (Reineck and Singh 1980). These processes have already been outlined in the discussion of Sections A12 and A15, and need not be repeated here. The horizontal bedding observed in Units 1 and 11 is interpreted to have been deposited by upper flow regime currents as described by Middleton and Hampton (1976) and Cheel and Middleton (1986). The almost structureless nature of the units indicates that the sediment was deposited quickly from suspension, preventing the formation of bedforms (Kuenen 1966; Middleton and Hampton 1976). Textural differences throughout the units define the horizontal bedding suggesting that the sediment was crudely sorted before deposition. The small clasts throughout the units are interpreted to have been entrained due to pulses in the flow energy. Cheel and Middleton (1986) discussed the formation of fining and coarsening upward horizontal laminae and attributed their formation to a burst and sweep cycle, respectively. Upper flow regime planar beds have been recognized in turbidite sequences as a result of suspension settling and in upper flow regime unidirectional currents. Due to the grain size, sorting and other structures in the exposure, I conclude that the horizontal bedding in Units 1 and 11 was deposited by upper flow regime currents.

Planar Cross-Bedding

Planar cross-bedding was observed in Units 2, 4, 6, and 8, and is interpreted to have formed by the migration of bars in a fluvial environment (Reineck and Singh 1980). Smith (1970, 1972), Rust (1975), Miall (1977) and Collinson (1986) discussed the formation and internal structure of fluvial longitudinal and transverse bars. Longitudinal bars are composed of relatively coarse gravel, and their internal structure is dominated by massive or crudely bedded units (Miall 1977). Large scale planar cross-beds were observed in outwash gravels by Eynon and Walker (1974) and Boothroyd and Ashley (1975), and they concluded that they were formed during a high river stage with gravel being swept across the bar top and avalanching over the lee side of the bar. Transverse bars are typical of sandy braided rivers although they are also observed in gravel bed rivers (Smith 1972; Boothroyd and Ashley 1975; Rust 1975). Planar tabular cross-bedding is the main sedimentary structure in transverse bars and forms by sediment moving across the surface of the bar as ripples or dunes, and then avalanching over the slip face of the bar (Smith 1970, 1972). The cross-beds commonly alternate between coarse and fine strata, due to the sediment being sorted as it moves across the bar surface (Smith 1972).

The planar tabular cross-bedded units in Section A2 are interpreted to have been deposited by sediment avalanching down the slip face of longitudinal and transverse bars. Unit 2 is significantly coarser than the other units displaying planar cross-bedding and it is interpreted to have been deposited on the slip face of a longitudinal bar during a high flood stage. The unit has a moderately oriented fabric and I propose that some of the clasts were reoriented by changing flow directions as the flood waned (cf. Tandon and Kumar 1981). Clasts in a fluvial environment are generally transported and deposited transverse to the flow direction (Rust 1975; Cant 1982; Smith 1985; Collinson 1986), and therefore the clasts in Unit 2 indicate that they were deposited by a river flowing coastward towards the northwest. The remaining units are sand dominated and are interpreted to have been deposited by sediment avalanching down the slip face of transverse bars.

Structureless Gravel

Structureless gravel was observed in Units 5, 7, and 9, and is interpreted to have been deposited as gravel sheets in longitudinal bars in a shallow braided river environment. Such bars are characterized by coarse grained sediment that is commonly structureless or crudely bedded (Rust 1975; Miall 1977). The gravel units in Section A2 are horizontally bedded and structureless, and there is no evidence of scouring at the base of the beds. I suggest that the sediment was deposited as gravel sheets in a minor channel that was only occupied during flood events.

The openwork gravel in Unit 9 is thought to have formed by the winnowing of interstitial finer sediment. I suggest that the pebbles were initially deposited in a thin longitudinal bar in a shallow braided river during a flood event, and that a decrease in the flow velocity caused the sand matrix to be winnowed from the bed. This could only occur under fairly low velocities given the relatively fine nature of the gravel, indicating that there was no deposition above the gravel at this time.

Alternating Gravel and Sand

Alternating horizontal beds of gravel and sand were only observed in Unit 3. The gravel beds are interpreted to have been deposited by gravel sheets in minor channels during floods, as discussed above. I suggest that the sand beds in Unit 3 were deposited as the discharge and flow velocity decreased from these flood events, possibly as upper flow regime planar beds. There are gradational contacts between all the beds in Unit 3 and this indicates that the discharge and flow velocity increased or decreased steadily preventing erosion of the underlying sediment.

Trough Cross-Bedding

Trough cross-bedding was observed in Unit 10, and is interpreted to have been deposited by migrating dunes or ripples along the floor of a river channel, or by the infilling of a scoured channel (Reineck and Singh 1980). These processes have previously been discussed in Section 6.1.2, and therefore will not be further discussed here.

The trough cross-bedding in Unit 10 is interpreted to have been deposited by the infilling of a scoured channel. The basal contact truncates the underlying unit suggesting that water eroded into the underlying unit before the channel was infilled with coarse grained sand (Reineck and Singh 1980). Small pebbles at the base of the unit indicate that flow energies were greater allowing the clasts to be transported. As the flow waned the clasts and coarse sand were deposited in thin beds that conform to the basal trough-shaped contact. The orientation of trough cross-beds was discussed by Harms *et al.* (1963), who concluded that trough axes are generally aligned parallel to the flow direction. The trough cross-bedded unit in Section A2 suggests that the sediment was deposited by water flowing towards the northwest.

I conclude that the section at A2 represents a shallow braided fluvial environment. The lack of distinct channelized structures in the section suggests that the section cuts through sediments of a minor channel within the main braided system. The majority of the section is composed of sand indicating that either flow velocities were relatively low, or that coarse sediment was not being transported into the minor channel. Occasional horizontal gravel beds throughout the section are interpreted to represent flood sediments that were deposited when the river discharge was greater causing the main channel to increase in width and depth. This allowed coarser sediment to be transported to the minor channels. During a normal stage, water did flow through the minor channel transporting sand and forming transverse bars.

6.1.4 - Sections A10a and A10b

Sections A10a and A10b are located in a gravel pit approximately 2 km southwest of Traytown. The gravel road leading to the pit extends southwest from the top of the Traytown delta and passes close to a cemetery. The area is dominated by sand and gravel (Section A10a), although there is an exposure of fine sand and silt roughly 80 m southwest of the sand and gravel deposits. These finer deposits are described in Section A10b.

6.1.4.1 - Section A10a

The sand and gravel section at A10a is 6 m high and 50 m wide. Much of the base of the section was covered by colluvial debris and therefore only the upper 3 m could be analysed (Plate 6.5). Fourteen units of trough cross-bedding were recognized in the section (Figure 6.8). Two of the units that are typical of the exposure will be described in detail.

Unit 1

Unit 1 (Figure 6.8) is composed of clast-supported gravel with a sandy matrix that has a reddish brown colour when weathered and dry (Munsell: 5YR 4/3). The unit is 1.12 m thick and extends laterally throughout the section. The sediment is poorly sorted, with clasts ranging in size from 5 cm to 75 cm. Although the unit is generally structureless, coarser clasts tend to accumulate at the base of the unit. The

Plate 6.5 - Well-defined trough cross-bedding in Section A10a.

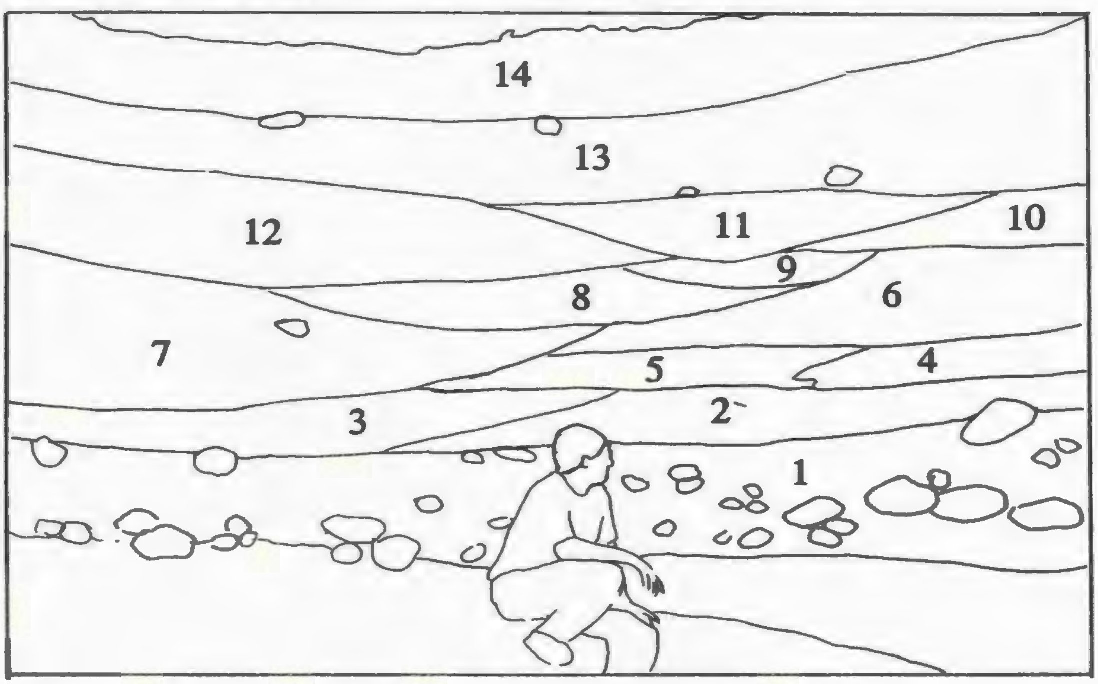

Figure 6.8 - Fourteen trough cross-bedded units in Section A10a.

upper and lower contacts of the unit are sharp and erosional, and the lower contact truncates the underlying beds.

Unit 7

Unit 7 (Figure 6.8) is 1 m thick, extends laterally over 9 m, and is composed of coarse sand and gravel. The basal 15 cm consists of pebble gravel with the clasts having long axis lengths of 1-9 cm. Faint trough cross-bedding can be seen in these gravel deposits, and the beds conform to the trough-shaped basal contact. Trough cross-bedding is well defined in the central and upper parts of the unit. The central area is composed mainly of coarse sand and the beds are defined by granules and pebbles along the basal contacts. All the beds conform to the basal contact in orientation. Scattered larger clasts, up to 9 cm long, were observed in the sand, commonly protruding into the underlying trough cross-beds. The top 30 cm of the unit is similar in texture to the basal gravel bed, although there is more sand. Therefore, the trough cross-bedding is well defined due to textural differences.

The upper and lower contacts of all the units in the section are sharp and erosional. Units 7, 8, 9 and 11 have very distinct trough shapes and the basal contacts truncate the underlying beds. The units suggest a flow direction towards the northeast and the coast. Many of the basal contacts are defined by a bed of medium to coarse gravel. Unit 8 is a good example of this and has several large clasts, 30 cm long, lying along the basal contact.

Interpretation

Trough cross-bedding is formed in subaerial fluvial environments either by the migration of ripples or dunes along the channel floor, or by the infilling of scoured channels, and these processes were discussed in Section 6.1.2.

Singh (1972) discussed the formation of stacked trough cross-bedded units on natural levees, cut across by crevasse splays. The channels were cut during flood conditions and the sediment was deposited mainly by suspension as the flow receded. A series of trough cross-bedded units were formed as the following floods eroded and deposited more trough cross-bedded units (Singh 1972). Similar series of truncated trough cross-bedded units were recognized in a braided outwash stream by Costello and Walker (1972), and in a low sinuosity gravel stream by Ramos and Sopeña (1983).

Coarser material is commonly observed at the base of channels and is referred to as a gravel lag (Bluck 1967; Reineck and Singh 1980; Ramos and Sopeña 1983; Olsen 1993; Kraus 1996). Bluck (1967) and Costello and Walker (1972) identified gravel lags in river channel deposits and concluded that they were formed as migrating channels repeatedly winnowed fine grained sediment from the base of previously deposited channels.

I suggest that the trough cross-beds in A10a were formed by the infilling of channels in a braided stream environment. Sediment is continually being eroded and deposited in this type of environment, and due to the constant migration of channels, only remnants of the bedforms and channels are preserved (Bluck 1967).

The sediment in Unit 1 is significantly coarser than the rest of the section, indicating that flow velocities were greater during transportation and deposition of this sediment. The remainder of the units are dominated by coarse sand suggesting that flow velocities were lower. However, coarser lag deposits at the base of some of the channels, and clasts in the trough cross-bedded sand indicate that the flow occasionally pulsated, allowing coarser sediment to be entrained.

All of the units in the section have sharp and erosional contacts and most of the trough cross-beds are truncated by an overlying or laterally adjacent unit. As the river channels migrated they eroded into previously deposited channel sediments, truncating the trough cross-bedding, and forming new trough cross-bedded units (cf. Williams and Rust 1969; Costello and Walker 1972). This created a vertical section characterized by multi-storey trough cross-bedded units, similar to that described by Ramos and Sopeña (1983).

The deposits at A10a are located on the relatively flat upper surface of the Traytown delta at 39 m asl. Moore (1966) suggested that trough cross-bedding was typical of channel sediments that were deposited in a subaerial delta environment. The location of the deposits at A10a, and their internal structure suggests that they were deposited by a braided river that flowed northeastwards over the delta surface towards the sea. There is no modern fluvial deposition on the surface of the Traytown delta, and the deposits identified at A10a thus were formed when sea level was higher than present.

6.1.4.2 - Section A10b

Two exposures of fine grained, gently dipping sediments were recognized at Section A10b. Due to the similar sedimentary characteristics in both exposures, only one exposure will be described in detail.

The section is 5 m high, 20 m wide and is characterized by beds of sand and silt that dip at 19° towards the north. Approximately 60% of the section consists of sand, 30% is silt, 5% is clay and the remainder is granules. Twenty-seven units were recognized within the section, and these can be assigned to seven facies (Figure 6.9).

Facies A1 and A2

Facies A1 is characterized by structureless silty-clay and silt. Throughout the section there are seven units assigned to Facies A1 (1, 9, 17, 19, 21, 23, and 26). Unit 1 is approximately 0.8 m thick and is formed of homogenous silty-clay that is very hard and difficult to excavate. The base of the unit was not observed and therefore the actual thickness of the unit is not known.

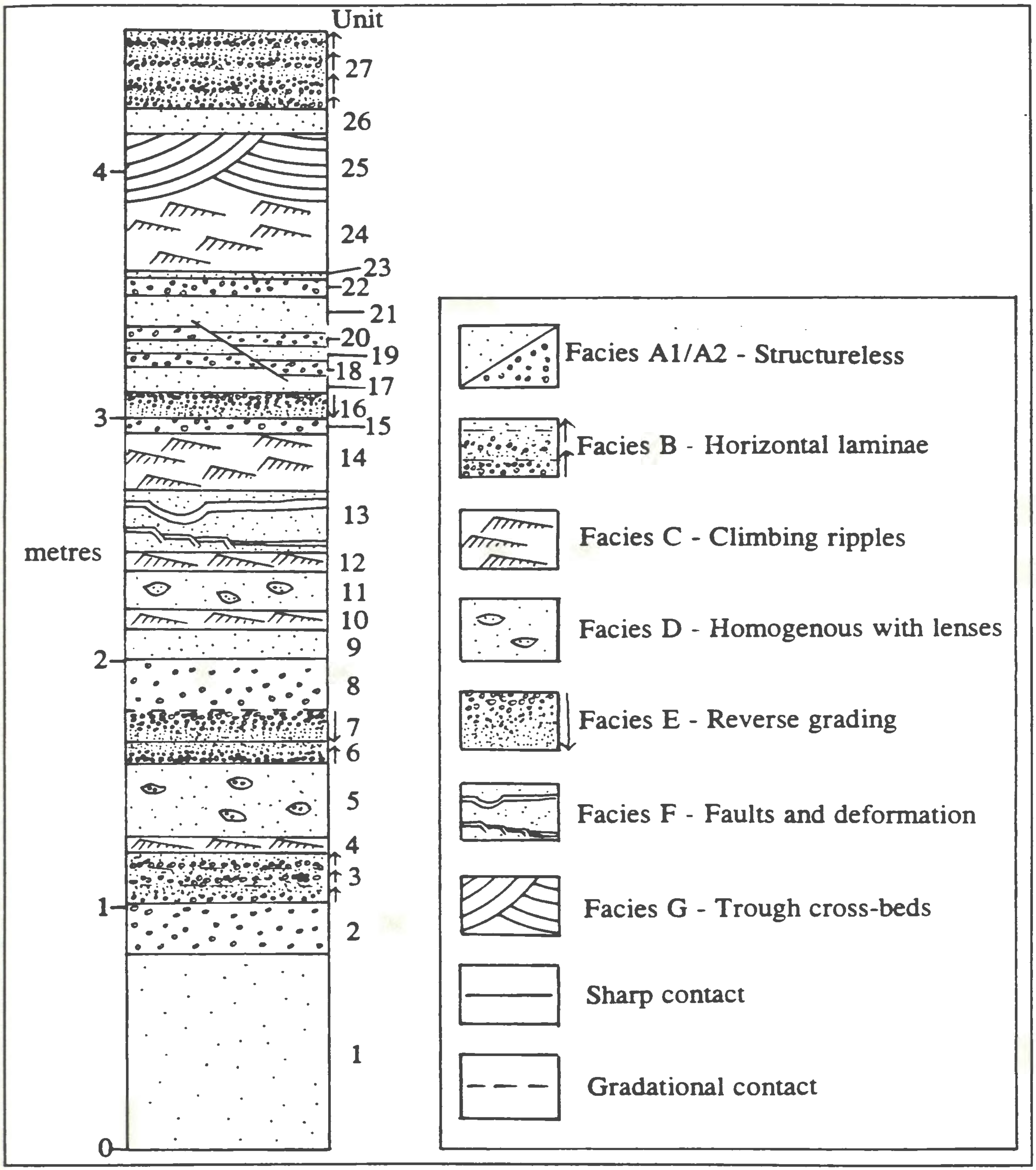

Figure 6.9 - Stratigraphic log of Section A10b.

Facies A2 is composed of structureless sand and Units 2, 8, 15, 18, 20 and 22 can be assigned to this facies. Unit 2 is 21 cm thick and formed of homogenous coarse sand. The contact between Unit 1 and 2 is abrupt and sharp.

Interpretation

Although the sediments in Facies A1 and A2 are both homogenous, the different grain size indicates that they were formed by different processes. The fine grained sediments in Facies A1 are interpreted to have been deposited by suspension settling over a relatively long period of time. These deposits may have been introduced into the system by interflows or overflows that allowed sediment to slowly settle out of suspension.

The coarser sediments of Facies A2 indicate more rapid sedimentation. Beds with little or no internal structure commonly form by the introduction of large influxes of sediment over a short period of time (Reineck and Singh 1980). The sediment is deposited quickly out of suspension from an interflow or overflow and due to continual rapid sedimentation, bedforms are unable to form.

Gilbert and Shaw (1981) and Catto (1987) described similar medium to fine grained structureless sediments within proglacial lacustrine sediments. They attributed the formation of the structureless beds to suspension settling within small shallow lakes. Catto (1987) suggested that this process can only dominate lacustrine sedimentation where there are very high concentrations of suspended sediment preventing the formation of underflows and forcing the meltwater to enter the lake as an overflow or interflow.

Facies B

Facies B is composed of normally graded thin parallel laminae of sand and silt, or coarse and fine sand, and is found in Units 3, 6 and 27. Unit 3 is 21.5 cm thick and approximately 40 laminae of silt and fine sand were observed within the unit. The laminae are about 0.5 cm thick, laterally continuous and the upper contacts commonly have a wavy structure. Although the laminae are normally graded they appear as almost two distinct beds, or couplets, with a relatively sharp contact between the fine sand and silt, or coarse and fine sand.

Interpretation

Normal grading is typically formed by suspension settling from overflows, interflows or turbidity currents (Kuenen 1951; Ashley 1975; Reineck and Singh 1980; Lowe 1982; Brodzikowski and van Loon 1991; Liverman 1991). Relatively sharp contacts between different grain sizes in proglacial lacustrine rhythmites and varves were observed by Kuenen (1951) and Ashley (1975). Kuenen (1951) suggested that the coarser sediment at the base of each rhythmite was deposited by turbidity currents, and that the finer sediment was deposited out of suspension during the winter months. Ashley (1975) came to a similar conclusion and suggested that ripples observed within the proximal lacustrine rhythmites were formed by the turbidity currents that were depositing the sediment.

In contrast, Shaw (1977) and Shaw and Archer (1978), suggested that coarser sediments in winter clay laminae were the result of turbidity currents that occurred throughout the winter months, and were not representative of summer deposition. This was a result of slumping on the delta front due to a decrease in the internal strength of the sediment as the lake levels fell. Some of the turbidity currents were strong enough to sweep across the lake bed to distal areas, and the resulting deposits were typical of turbidity current deposition under waning flow.

I conclude that the normal grading observed in Facies B formed by deposition from turbidity currents. As the currents moved across the lake bed they eroded some of the underlying sediments forming wavy contacts with the beds below. The coarser sediments were deposited by the initial movement of the turbidity current and the finer sediments by suspension settling after the turbidity current had passed. Liverman (1991) suggested that the normal grading within individual beds may indicate that each turbidity current is characterized by small pulses that slowly decline in intensity. Some of the couplets in Facies B are dominated by coarse and fine sand, rather than sand and silt and this is possibly due to an increase in size and volume of suspended sediment due to stronger turbidity currents. Although it is not possible to determine if the turbidity currents formed in summer or winter, the coarse nature of some of the couplets suggests that deposition took place during the summer due to an increase in flow energy.

Facies C

Facies C is characterized by fine, medium and coarse sand that displays ripple structures. Units 4, 10, 12, 14, and 24 are typical of this facies, however Unit 24 shows the best development and will be described in detail. The unit is 32 cm thick and is composed of silt and fine to coarse sand. Climbing ripples are observed throughout the unit and individual beds are defined by a layer of silt capping the ripples below. Beds of coarse sand approximately 1.5 cm thick drape some of the silt layers, and have little or no internal structure. The climbing ripples are characterized by stoss and lee side preservation identified by textural variations throughout the beds. The ripples have a typical wavelength of 12 cm, and an amplitude of 1.5 cm. Many of the climbing ripples within the beds have been truncated.

Interpretation

The climbing ripple laminae of Facies C are interpreted to have been deposited within flowing water where there was an ample supply of sediment. Climbing ripples are commonly formed when there is a continual supply of suspended sediment that constantly buries the underlying ripple structures preserving their form and internal structure (Fritz and Moore 1988). The climbing ripples identified in Facies C are similar to the Type B ripple-drift cross-lamination of Jopling and Walker (1968). Formation of this type of cross-lamination requires bedload movement and a continual supply of sediment from suspension (Jopling and Walker 1968).

Stanley (1974) discussed the formation and hydraulic significance of climbing ripples in lake silts. Due to the discontinuous nature of some of the ripples and the preservation of coarse and fine layers on the lee faces of the ripples, he concluded that the climbing ripples were formed by traction currents along the surface of the bed, and by suspension settling. Gustavson *et al.* (1975) reached a similar conclusion after studying sequences of draped lamination, overlain by the Type B and Type A climbing ripple lamination of Jopling and Walker (1968). They suggested that the sequence represented an increase in the density underflow velocity throughout the melt season, causing the amount of bedload in transport to increase. The layers of silt capping many of the rippled beds suggests that there were periods when the flow velocity decreased sufficiently for the finer sediments to settle out of suspension (Stanley 1974; Gustavson *et al.* 1975).

The base of individual rippled beds within Facies C are commonly characterized by structureless sediment, indicating that rapid deposition from suspension was followed by the formation of climbing ripples. This, and the truncated beds throughout Facies C indicate that turbidity currents from the delta front eroded some of the underlying sediment before depositing large quantities of sand in a short period of time. As the current velocity began to decrease, beds of climbing ripples were formed. Many of the beds are interrupted by thin beds of silt suggesting that throughout the formation of the climbing rippled units, there were periods when the flow velocity decreased sufficiently for the finer sediments to settle from suspension.

Facies D

Facies D is represented by two units. Units 5 and 11 are 30 and 16 cm thick, respectively, and they are composed of homogenous silty-clay that contains lenses of medium sand. The lenses in Unit 5 are the largest and will be described here, however the lenses in Unit 11 have a similar composition and similar contacts with the surrounding sediment. The lens within Unit 5 is 25 cm wide and 13 cm thick. The lens has an irregular shape and the contact with the surrounding sediment is generally sharp. There is no internal structure within the lens and it does not deform the upper and lower contacts of Unit 5. Smaller lenses with irregular shapes surround the top of the main lens. There is a sand unit directly above Unit 5 and the texture of the sand within both units is comparable.

Interpretation

The lenses within Units 5 and 11 are interpreted as frozen balls of sand that were dropped into homogenous silty-clay that was being deposited by suspension settling. I suggest that the blocks originated from a bank close to the site of deposition. Ice forming on top on the water may have incorporated the blocks and dropped them into the sediment upon melting, or the ice may have eroded the bank causing it to collapse and drop the frozen sediment into the homogenous silty-clay. The irregular shape of the lenses suggests that the frozen blocks experienced some deformation after deposition. This may be due to compaction of the surrounding silty-clay by continual sedimentation above the bed. Deformation of the surrounding sediment may be expected, however due to the lack of internal structure within the primary sediments of Units 5 and 11, no deformation was observed (cf. Gilbert 1990).

Facies E

Facies E is characterized by reverse grading of beds from fine sand to medium/coarse sand. Two units throughout the section showed these characteristics (Units 7 and 16). Unit 7 is 13 cm thick and is composed of fine sand that coarsens upwards to medium sand. The upper and lower contacts of Unit 7 are sharp and abrupt. Unit 16 has a sharp upper contact and the lower contact is gradational.

Interpretation

The reverse graded beds of Facies E are typical of high density turbidity currents or grain flows within a sand dominated environment (Reineck and Singh 1980; Lowe 1982). Coarser particles start to collect at the base of turbidity currents and transport within this layer is increasingly dominated by grain collisions. The form of this basal layer is maintained by dispersive pressures, causing the coarser particles to rise above the finer grained traction carpet below (Bagnold 1954; Lowe 1982). As sediment continues to fall out of suspension the traction carpet becomes loaded and freezes almost instantaneously, causing a new traction carpet to develop above the reversely graded bed (Lowe 1982). Reverse grading may also form in grain flows due to the dispersive pressures mentioned above, or by a kinetic sieve mechanism whereby the finer grains fall through the coarser grains and collect at the bottom of the flow (Middleton 1970; Reineck and Singh 1980). Subaqueous grain flows commonly require a static angle of repose between 18° and 28° for their initiation and continuation (Reineck and Singh 1980; Lowe 1982). They are generally less than 5 cm thick due to the inability of grains at the base of the flow to maintain the dispersive pressures required against the force of gravity (Lowe 1976, 1982).

The reversely graded units in Facies E are 10 and 13 cm thick, and according to Lowe (1976, 1982), this suggests that they cannot have been deposited by grain flows. The units in Facies E are therefore interpreted to have been deposited by traction carpets that formed at the base of high density turbidity currents in a sand dominated environment.

Facies F

There is only one unit in Facies F and it is characterized by faulted beds and deformation. Unit 13 is 26 cm thick and composed of beds of silty clay and sand. The base of the unit consists of 6 cm of homogeneous silty clay that is interrupted by a bed of medium to fine sand approximately 1 cm from the base of the unit. The bed of sand is faulted and three small normal faults were observed with relatively little displacement. Above the silty clay bed there is a sand bed that is 20 cm thick. Roughly 3 cm from the base of the sand bed there is a thin bed of silty clay that has a clear convexo-concave structure with a flattened base that is 2.5 cm thick. The silty clay bed appears to remain horizontal on either side of the deformed structure, and the horizontal parts of the bed are roughly 7 cm thick.

Interpretation

Small normal faults and distorted beds are commonly associated with glaciolacustrine and glaciofluvial sediments that have been deposited over blocks of stagnant ice (McDonald and Shilts 1975; Sugden and John 1976; Reineck and Singh 1980; Van der Meer *et al.* 1992). As the ice melts the sediment above starts to slump down into the void left by the ice, causing the sediment above to develop normal faults or be deformed (Reineck and Singh 1980). The faults in the lowest 6 cm of Unit 13 are interpreted to have been formed in this way, with a small block of ice located under the sediment slowly melting and causing the sediment to slump and fault. The convexo-concave structure is interpreted to have formed in a similar manner, with the silty clay sediment being deposited over a small block of ice that subsequently melted. No faults are associated with this silty clay bed and I therefore suggest that this ice block melted relatively slowly, and that deposition of the silty clay may have continued as the ice below was melting.

Facies G

Trough cross-beds were observed in Unit 25 and they appear to be laterally equivalent to alternating beds of fine and coarse sand (Facies B). The unit is 25 cm thick and the beds are composed of fine, medium and coarse sand. There is some coarsening towards the east. Two sets of trough cross-beds were observed separated by an erosional contact. The lower cross-beds dip 17° and the upper cross-beds dip 30°. The lower contact is gradational, and the upper contact is sharp, and the high angled coarse sand trough cross-beds truncate the low angled finer-grained trough cross-beds below.

Interpretation

Trough cross-beds are commonly found in fluvial deposits (DeCelles *et al.* 1983) and they are formed by the migration of ripples or dunes downstream, or by the infilling of an eroded channel. The erosional and depositional processes have previously been discussed in A12, and are not repeated here.

Although two sets of cross-stratification were recognized in Facies G based on the angle of dip, McKee (1957) suggested that cross-strata may dip at a steeper angle if the sediment becomes coarser, or the velocity of the transporting water decreases. However, the two sets of trough cross-stratification may represent two different channels with the upper set eroding down into the lower set. DeCelles *et al.* (1983) suggested that based on the asymmetry of the basal contact and the angle at which the trough cross-beds are truncated, the palaeocurrent direction can be inferred. The steeper trough cross-beds in Facies G may therefore represent a different cut through a second channel that was flowing in a different direction to the channel below.

The trough cross-stratification identified in Facies G is interpreted to have been formed by subaqueous channels that were subsequently filled by suspension settling from pulses of sediment that were introduced into the system. The steeper cross-strata within the facies are slightly coarser indicating that coarser sediment was being transported to the site perhaps by an increase in the velocity of the transporting water, or by a secondary channel that eroded into the primary channel and had a slightly different palaeocurrent direction.

Faults

Throughout the exposure there are numerous small-scale reverse faults that are found within all types of sediments, and are not restricted to a specific unit. The fault planes dip towards the north and south and dip at angles of 11°, 24°, 35° and 50°. Between 5 and 30 beds are displaced and the amount of displacement varies from 0.5 to 17 cm (Plate 6.6).

Small-scale reverse faults are commonly associated with coarse grained glacial sediments and therefore it is unusual to find reverse faults in such fine-grained sediments. McKee and Goldberg (1969) examined the contorted structures that were formed in mud. They suggested that thrust faults may form beneath sediment that was being deposited at an angle rather than normal to the surface of the mud. This created a strong lateral, as well as compressive, force causing the underlying sediments to be faulted. However, the faults within Section A10b dip towards the north and south, and therefore they cannot have been formed by lateral stresses created by sediment deposition on a slope with only one flow direction.

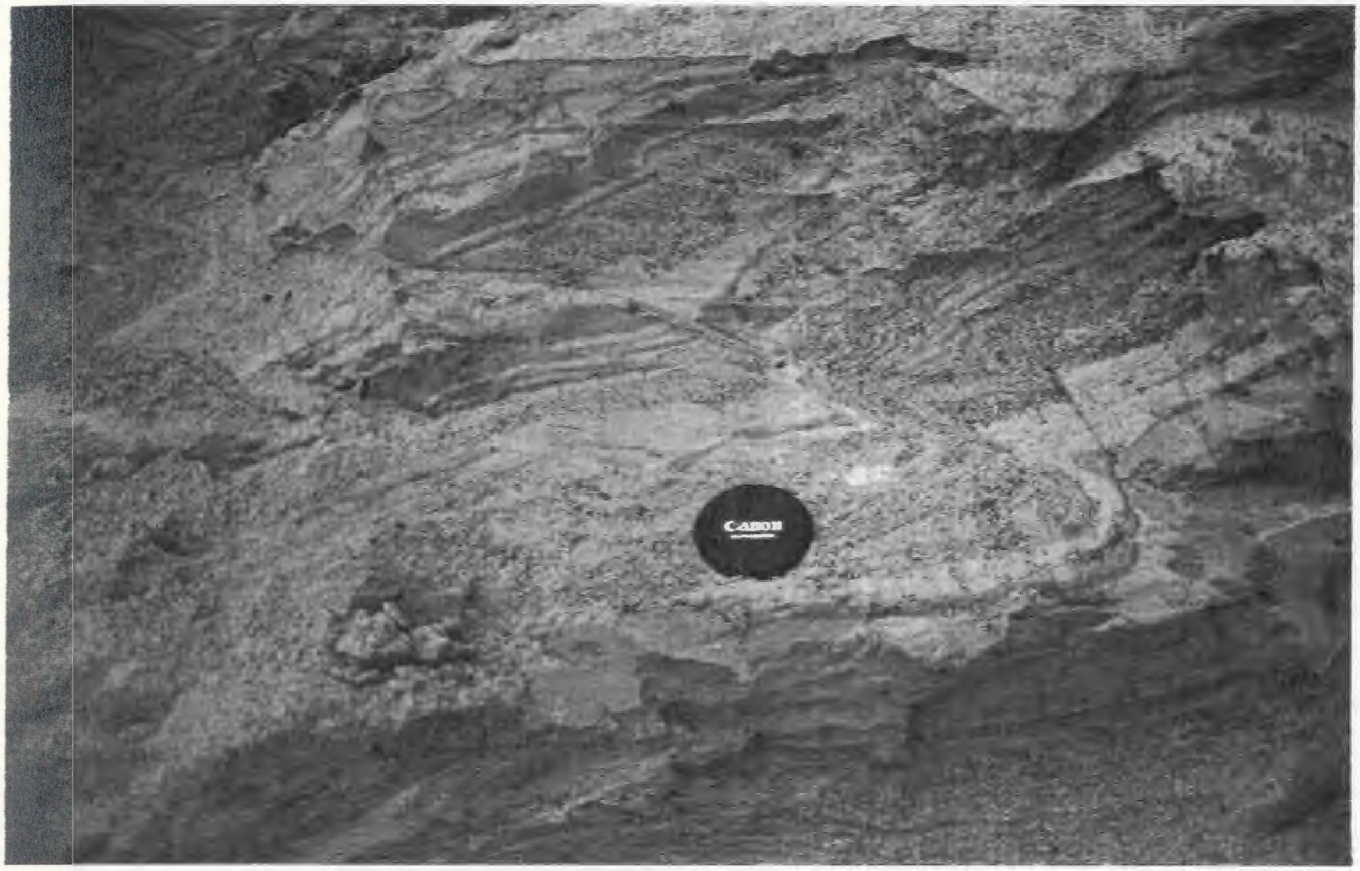

Plate 6.6 - One of the reverse faults in Section A10b. The fault is 40 cm long and within Facies A1/A2.

McDonald and Shilts (1975) discussed the formation of faults in glaciofluvial sediments. They assigned low angle reverse faults with dips less than 45° to be a result of horizontal compressive stresses. These stresses may have formed due to sediment sliding down a slope under the influence of gravity, or by an overriding glacier. The section at A10b is surrounded by glaciofluvial sediments and there is no evidence to suggest that a glacier overrode the sediments with subsequent deformation. McDonald and Shilts (175) and Reineck and Singh (1980) suggest that melting of ice underneath the sediment may produce small-scale faulting as the sediment settles under the force of gravity. Normal faults are usually associated with this process, however Sanford (1959) suggested that if there was a vertical displacement of a lower boundary due to the melting of buried ice, small-scale reverse faults would form along the fractures (Figure 6.10). Sanford (1959) recognized that a series of reverse faults would form rather than a single fault, and that the faults would have progressively lower dips towards the surface. The reverse faults within the section at A10b have progressively lower dips farther up the section and are therefore interpreted to have formed by the melting out of buried ice that caused a large block of sediment to be displaced downwards.

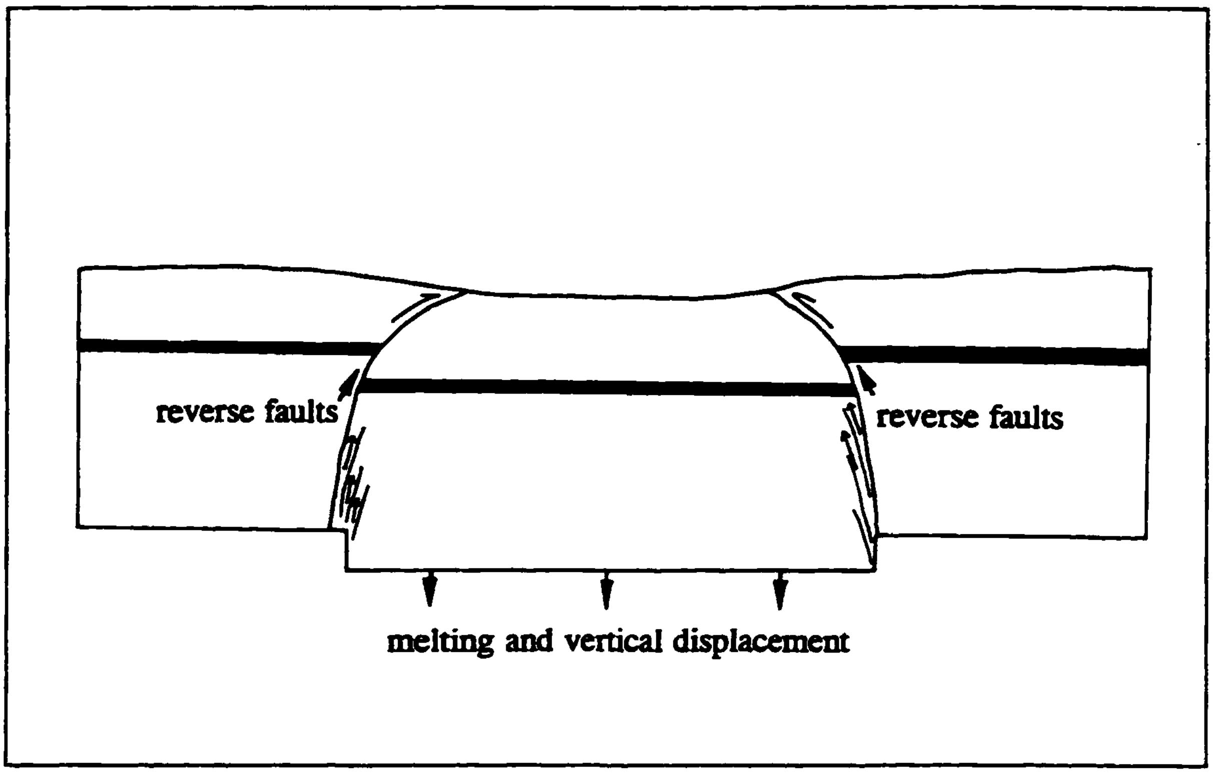

Figure 6.10 - Vertical displacement of lower boundary due to melting ice producing reverse faulting (after Sanford 1959).

Discussion

The sediment in Section A10b is very fine in contrast to the surrounding sediment which is composed of sand and gravel, as previously described in Section A10a. The sediments in Section A10b are similar to the lake sediments described by Shaw (1975, 1977), Gilbert and Shaw (1981), Catto (1987) and Liverman (1991). I suggest that the sediments in Section A10b represent a small lake adjacent to the main meltwater channel that was periodically influenced by overflowing water throughout the meltwater season.

Almost half the units in Section A10b are dominated by structureless sediments (Facies A1 and A2). These have been interpreted to have been deposited by overflows or interflows that formed from high concentrations of suspended sediment within the inflowing water. The plumes may have formed at the beginning or throughout the melt season, however the finer units are thought to represent suspension settling throughout the winter.

Facies B, C, and E are represented by normal grading, climbing ripples and reverse grading, respectively. They are all interpreted to have been deposited by turbidity currents that formed due to slumping at the delta front. Slumping may have occurred as a result of oversteepening from rapid sedimentation during the main meltwater season, or as a result of lake levels falling throughout the winter and reducing the internal strength of the deltaic sediments (Shaw 1977; Shaw and Archer 1978).

The homogenous sediments described in Facies D are similar to those in Facies A, however there are small lenses throughout the units that have been interpreted as frozen balls of sand that were dropped into the homogenous silty-clay. The lenses may have been deposited by underflows or turbidity currents, however they are located in a relatively thick bed of fine silty-clay which takes a considerable amount of time to settle out of suspension. I therefore conclude that Facies D was deposited during the winter when there was very little activity within the lake.

In conclusion, the sediments at A10b likely represent a small lake that was situated close to the main meltwater stream that flowed across the surface of the sandur towards the coast. Sedimentation in the lake was dominated by overflows or interflows, and by turbidity currents. The trough cross-beds in Facies G indicate that subaqueous channels were formed, possibly by strong turbidity currents, that eroded a substantial amount of sediment. Small normal faults within Facies F signifies deposition over a small block of ice that was rapidly buried by the overlying sediment. Larger reverse faults throughout the section indicate that the sediment was deposited in a lake that formed on top of a relatively large block of ice. The ice subsequently melted and caused part of the section to collapse into the resultant void below (cf. Sanford 1959).

6.2 - Summary of Glaciofluvial Deposits in Terra Nova National Park and Vicinity

The glaciofluvial deposits throughout Terra Nova National Park and vicinity reflect deposition in a variety of glaciofluvial environments. The most complex system is located along Big Brook valley adjacent to Route 301. Close to Terra Nova Lake there are two large sand deposits (A12 and A15) that are interpreted to have been deposited in a sandy braided fluvial environment in front of a proglacial lake. I propose that the proglacial lake (Terra Nova Lake) acted as a sink for the coarse sediment that was transported away from the ice front by meltwater streams.

Downstream in Big Brook valley, the sediment is coarser (A11b) and represents deposition in longitudinal bars in a braided river. The coarse sediments at A11b were deposited prior to the formation of the proglacial lake. As the ice retreated over the drainage divide between Big Brook valley and Terra Nova River valley, meltwater ceased to flow down Big Brook valley restricting the supply of glaciofluvial material to the valley. Meltwater was concentrated into Terra Nova River valley. The trough cross-bedding on top of the Traytown delta (A10a) and the ponded sediments (A10b) may have formed at this time. Further retreat of the ice created the proglacial lake into which the coarse grained sediment was deposited.

Coarse sediments representing deposition in a braided river are also located adjacent to the Terra Nova River at A9. Although these deposits are less than 10 m above the present river, they were deposited when sea level was higher. A subsequent fall in sea level resulted in the incision of these sediments.

The braided stream deposits at Sandringham (A2) were also deposited when sea level was higher than present. There is a lack of modern fluvial deposition, and this suggests that the drainage patterns were different during deposition of these sediments than they are today.

Chapter 7

Emerged Marine Landforms and Sediments

7.0 - Raised Marine Features

In Atlantic Canada, postglacial sea-level changes mostly occur as a result of climate-influenced changes in sea level and isostatic recovery from glacial unloading. Throughout Terra Nova National Park and vicinity there are several raised features that are evidence of higher sea-levels (Figure 7.1). Individual raised landforms, measured in July 1995, are listed in Table 7.1.

Table 7.1 - Raised landforms and elevations within Terra Nova National Park and vicinity.

Delta/Terrace	Elevation asl
Traytown	39 m
Eastport	16 m
Sandy Cove	25 m
Big Brook	31 m
Port Blandford	16 m
Culls Harbour	16 m
Charlottetown	33, 27, 21, 5 m

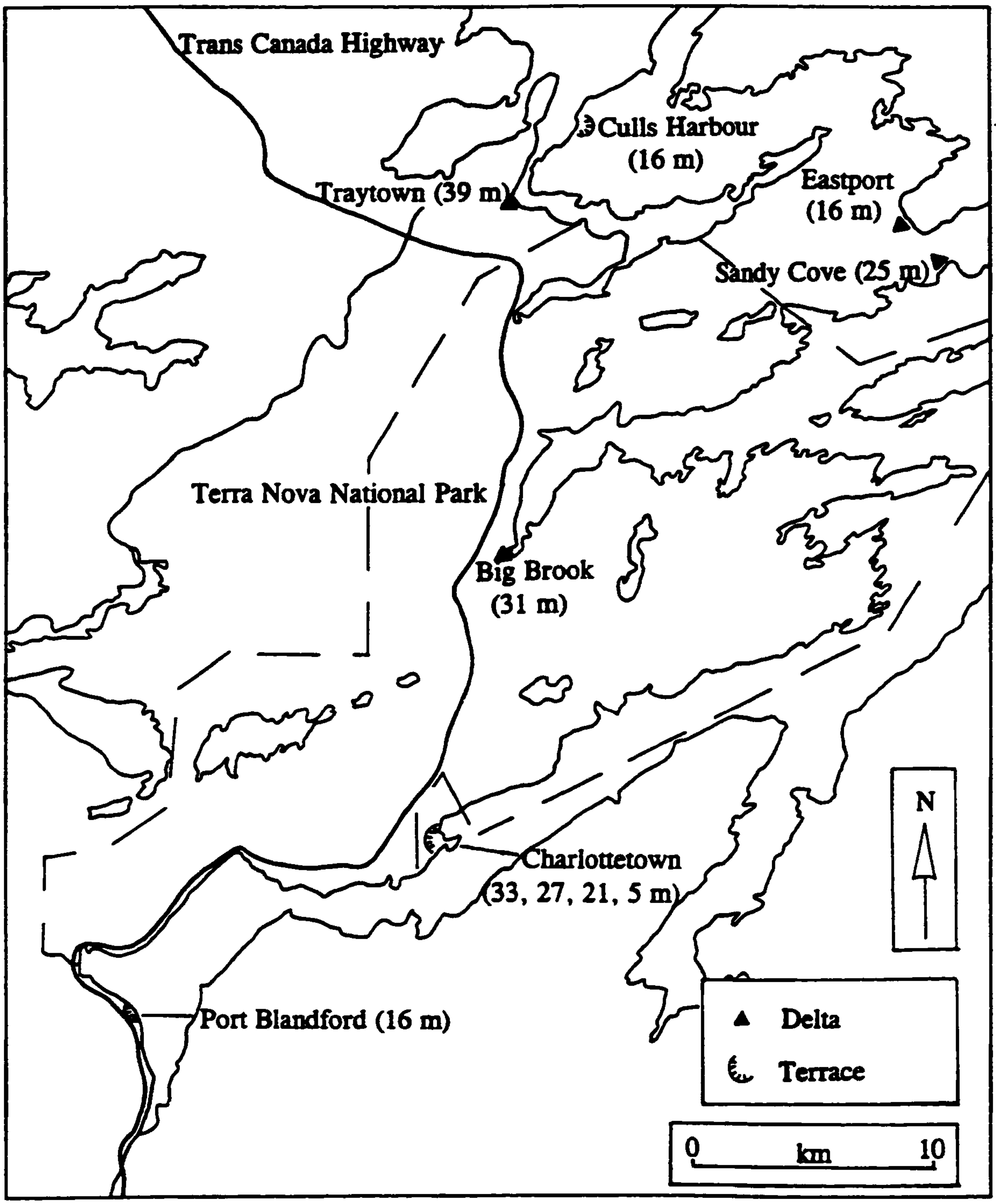

Figure 7.1 - Map showing the location of raised marine features measured during July 1995.

Large expanses of sand and gravel crop out at Traytown, Eastport, and Sandy Cove. They are all well-developed, form prominent bluffs close to the present shoreline, and are interpreted to be the remnants of raised marine deltas. The surfaces of these raised features are dominated by agriculture reflecting the very sandy and well-drained soils in the area. Depressions and hills have been colonized by black spruce (*Picea mariana*), balsam fir (*Abies balsamea*), trembling aspen (*Populus tremuloides*), and white birch (*Betula papyrifera*). The lower elevation marine terraces ($\leq$16 m asl) are fairly well developed, and tend to be covered by boreal forest.

7.1 - Raised Marine Sediments

The elevations of raised marine features were measured during July 1995. Few exposures of raised marine sediments were recognized in the study area. The exposed faces of the raised deltas at Sandy Cove, Eastport, and Traytown are covered by colluvium and therefore detailed analysis of the sediment is impossible. The observations and interpretations made by Jenness (1960) and Dyke (1972) have been used in the overall reconstruction of post-glacial sea-level history.

Two temporary sections (A6) near the top of the Traytown delta (39 m asl) were exposed in July 1995 due to housing construction. The sections are approximately 3 m thick, 20 m wide, and are composed of coarse cobble gravel and trough cross-bedded sand and fine gravel. The cobble gravel is clast-supported and clasts vary in size from 4-16 cm. The basal contact with underlying sand is trough shaped, and there is a lag of large clasts at the base of the unit.

Sand beds overlie and underlie the pebble gravel, and form trough shaped units that are commonly truncated by other trough structures. Individual beds within the units are defined by a thin layer of fine gravel and the beds conform to the trough shaped basal contacts. DeCelles *et al.* (1983) discussed the determination of palaeocurrent directions from trough cross-stratification and concluded that the axes are generally aligned parallel with the local flow direction. The trough cross-beds at A6 suggest that they were formed by flow towards the east or northeast.

The coarse gravel and sand units are interpreted to have been formed by river channels that wandered across the top of the delta eroding and depositing sediment. Although the sections are poorly exposed the trough cross-beds in the sections indicate flow directions towards the east or northeast. The deposits are downstream of the coarser trough cross-beds described in Section 6.1.4.1 (Section A10a) and are thought to be part of the same braided stream system.

Two other exposures were located northwest of Port Blandford (A18 and A19) and each provides evidence of the existence of higher sea levels during deglaciation. Both sites are characterized by clay, silt and fine-grained sand overlain by medium-coarse grained sand.

7.1.1 - Clay, Silt and Sand

Description

Only one coastal exposure interpreted to be deposited as marine clay, silt and sand was found in the study area. This exposure is located approximately 2.5 km northwest of Port Blandford (A18) in a small cove close to the main road that runs through Port Blandford. The exposure is at least 2 m thick, as the clay and silt extend to below sea level (Plate 7.1). The exposed section was restricted to approximately 10 m in width, due to colluviation and vegetation growth.

The lower 1.3 m of the unit is characterized by beds of clay, silt and sand. The beds are 0.5-3 cm thick and all are well sorted. There is some normal grading within the beds and most fine upwards from sand to clay. Contacts between the beds are generally sharp.

The lower unit has been deformed and convoluted bedding extends along the length of the exposure (Plate 7.1). Antiforms and synforms found throughout the unit vary in height and width. The convolutions in the centre of the exposure show the best development, with broad synforms separating relatively narrow antiforms and diapirs which widen towards the top of the structures. The diapirs are approximately 0.7 m wide and 0.85 m high. Towards the west the convolutions are less pronounced, although the deformation of a single bed can be followed throughout the sequence. The antiforms and synforms are approximately 0.4 m and 0.7 m wide respectively, and the structures are approximately 0.9 m high. As the structures

Plate 7.1 - Clay, silt and sand deposit at Port Blandford (A18). The clay and silt continue below sea level and the exposed section is restricted by colluviation and vegetation growth. The section is approximately 10 m wide and 5 m high. The lower 1.3 m of the section has been deformed and convoluted bedding extends along the exposure. Above this there is 0.7 m of horizontally bedded clay, silt and sand. the remainder of the section is composed of sand and is discussed in A19 (Section 7.1.2).

extend below sea level, the exact height of these structures is not known. The convolutions show horizontal displacement, and the fold axes indicate that deformation was produced by flow towards the east-northeast. Throughout the deformed unit there are pockets and lenses of coarse sand that are approximately 1-1.5 cm thick and 1 m long. The lenses follow the convolutions indicating that the sand was deposited before the sediment was deformed.

The upper 0.7 m of the unit is composed of approximately 20 horizontal beds of clay, silt and sand. Some of the lower beds deform over the convolutions below. The unit coarsens upwards from clay and the successive sand beds increase in thickness. The thicknesses of the clay beds gradually decrease up to 3 m asl. Above this level, clay beds are not observed within the exposure. All contacts within the unit are sharp.

At the top of the unit there are lenses of medium sand within a bed of coarser sand. The lenses are 1.1-5.7 cm long and 0.7-1.5 cm thick. They have an irregular shape and the contacts with the surrounding sediment are sharp (Plate 7.2). The texture of the lenses is similar to the texture of the bed below.

Within the basal deformed unit there is a granitic clast that is roughly 30 cm long and is aligned vertically within the surrounding sediment. Beds within the sediment have been deformed by the clast and can be traced on either side of, and below, the clast.

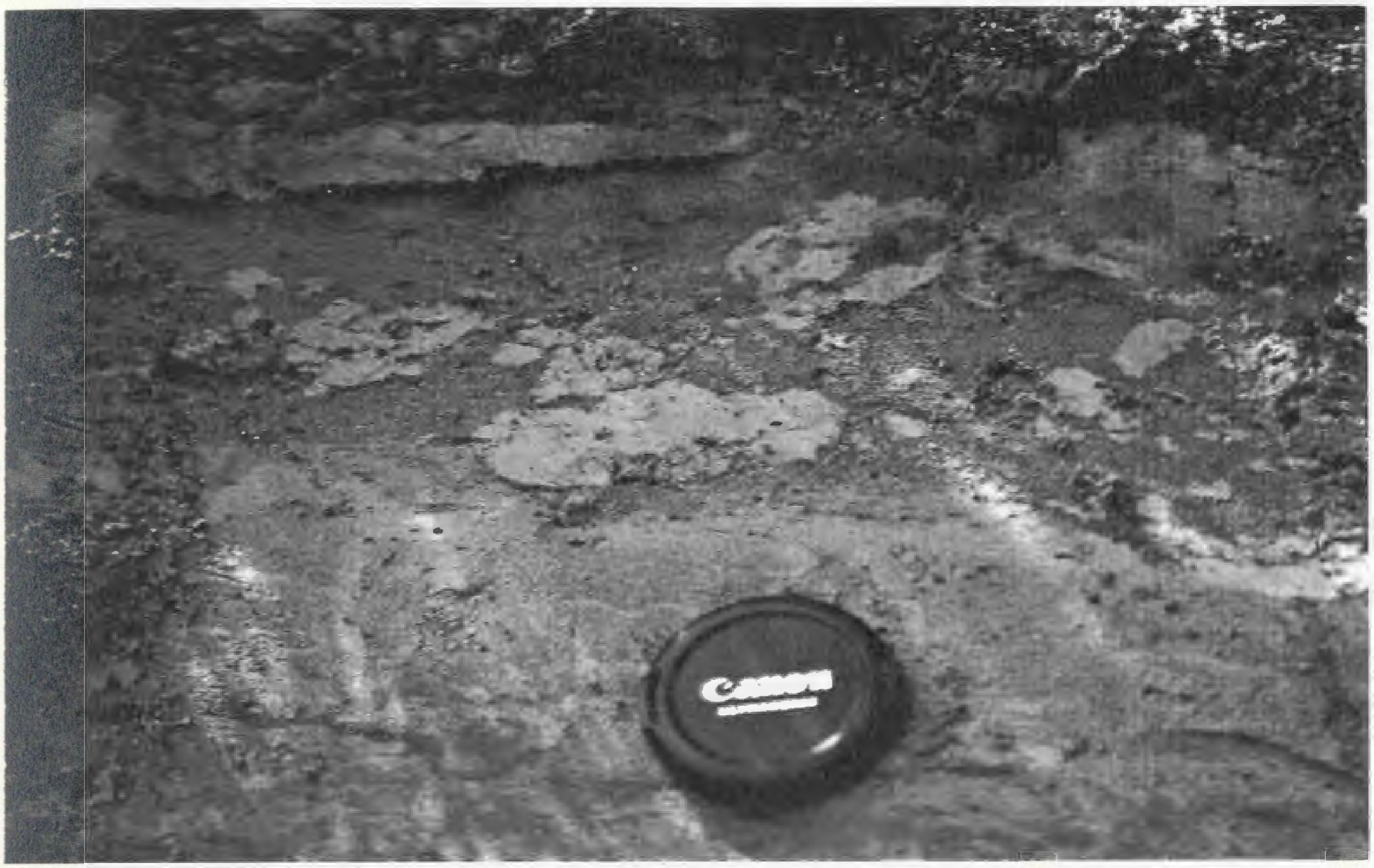

Plate 7.2 - Lenses of medium sand at the top of the clay, silt and sand unit (A18). The lenses are 1.1-0.7 cm wide and 0.7-1.5 cm thick. All the lenses have an irregular shape and sharp contacts with the surrounding coarse sand.

Interpretation

The horizontal bedding of clay, silt and sand throughout the unit may have been formed by underflows, interflows or overflows, from a meltwater stream entering into the sea. Within the glaciolacustrine environment coarse-grained rhythmites are quite common and are produced by underflows that flow along the lake bottom (Reineck and Singh 1980). Sedimentary structures associated with these underflows include coarse-grained graded laminations or rippled sand and silt (Reineck and Singh 1980). In the marine environment, however, underflows are relatively rare due to the greater density of sea water (Powell 1983; Domack 1984; Syvitski 1989). As a stream enters into the sea the discharge rises up as a turbid plume and forms an interflow or overflow (Mackiewicz *et al.* 1984; Cowan and Powell 1990; Lemmen 1990). The sediment is deposited by suspension settling, resulting in normal graded beds. Mackiewicz *et al.* (1984) discussed the formation of cyclically interlaminated mud within the marine environment. They suggested that each new turbid plume brings a separate supply of sediment to the water column. The coarse grained sediment is deposited out of suspension faster than the finer sediment, and this produces normally graded beds. Due to the lack of current features within the sediment at A18, the horizontal beds are interpreted to have been deposited by suspension settling from interflows and/or overflows.

The basal part of the unit exhibits soft sediment deformation and is dominated by convoluted bedding. Convoluted bedding is characterized by laminated silt and fine sand that has been internally contorted, and has planar upper and lower contacts (Mills 1983). There are several hypotheses for the formation of convoluted bedding, but all require a liquefied bed for the deformation to take place (Reineck and Singh 1980; Mills 1983; Lindholm 1987; Owen 1996). Liquefaction occurs when loosely packed grains in under-consolidated sediment are shaken apart, causing the sediment particles to move downwards and the pore fluid to move upwards (Mills 1983). The process is associated with rapid deposition (Lindholm 1987).

Allen (1982) suggested that vertical folds within convoluted bedding are the result of vertical forces acting on the liquified sediment. Cheel and Rust (1986) proposed that some vertical convoluted bedding or diapirs may be the direct result of rapid dewatering. They suggested that rapid deposition of silty-sand creates an excess pore water pressure that causes the sediment to be ejected up through more cohesive sediments when a critical value is reached.

Folds within convoluted bedding are often oriented in a preferred direction, and may be formed by gravitational slumping due to oversteepening of a slope, rapid sedimentation, or undercutting by water or turbidity currents (Allen 1982; Mills 1983; Scott *et al.* 1991). Mills (1983) described two types of slumps, both of which involved the movement of under-consolidated sediments downslope as a result of gravitational forces. Coherent slumps were characterized by very little mixing of the sediments, and bedding within the unit was generally preserved. Incoherent slumps were composed of broken bedding and balls of mixed sands, silts and muds. Slump structures are often confused with tectonically induced deformational features (Allen 1982; Mills 1983). Gravitational slumps are distinguished by undisturbed bedding above and below the deformed unit, whereas tectonically induced deformation affects the entire unit (Allen 1982).

The convoluted bedding at A18 has a preferred orientation towards the coast, and therefore the sediments are interpreted to have been deformed by slumping after their initial deposition. Individual beds can be traced throughout the unit indicating that there was little mixing of the sediment during deformation. Bedding directly above the deformed unit is undisturbed, suggesting that the slumping was induced by oversteepening or undercutting.

The lenses of medium sand at the top of the unit are interpreted as rip-up clasts that were formed by the action of turbidity currents or underflows. Turbidites are characterized by massive or graded beds that are overlain by upper flow regime planar beds, ripples, and lastly finer sediments that are deposited from suspension (Walker 1984). Due to the lack of current features in the overlying beds I suggest that the rip-up clasts were formed by underflows that incorporated lenses of sediment from the underlying bed into the flow. Rapid deposition resulted in the rip-up clasts of finer sediment being deposited within the coarser grained sand with no specific orientation or shape. Similar deposits of rhythmically bedded silts and clays with some convoluted bedding were recognized in raised marine sediment in the Springdale-Hall's Bay area by Scott *et al.* (1991) and in the Botwood area by Mackenzie and Catto (1993).

The granitic clast within the basal deformed unit is interpreted as a dropstone that was deposited from an iceberg. Liverman (1991) and Scott *et al.* (1991) described similar rhythmically bedded sediments that were deformed by clasts in a glaciolacustrine and glaciomarine environment respectively, and interpreted the clasts to be ice-rafted dropstones. Thomas and Connell (1985) discussed iceberg drop, dump and grounding structures within glaciolacustrine sediments. The majority of the clasts they analyzed were deposited by icebergs, and approximately 95% of the clasts were inclined or sub-vertical. The lower contact is often deformed and the amount of deformation depends on the size and shape of the clast, and the type of sediment into which it falls (Thomas and Connell 1985). The clast at A18 is relatively large, has a bladed shape and was deposited into underconsolidated marine clay, silt and sand. These conditions and the vertical alignment of the *ab* plane of the clast caused the underlying sediments to undergo a large degree of deformation.

7.1.2 - Alternating Silty-Sand and Sand

The marine clay is overlain by approximately 3.5 m of alternating silty-sand and sand beds. The sediment directly above the clay was covered by colluvial debris, and therefore, an exposure within a gravel pit 300 m northwest of the clay section was analyzed in detail.

The section at A19 is characterized by horizontal beds of silty-sand and fine, medium, and coarse sand. Approximately 60% of the beds are dominated by sand, 30% by silty-sand and the remaining 10% is silt. Eighteen units were recognized within the section, and these can be assigned to six facies (Figure 7.2).

Facies A

Facies A is represented by reverse grading of beds from silt to fine, medium, or coarse sand. Approximately 80 beds throughout the section show reverse grading, as found in Units 1, 7, 9, 11 and 12. Unit 1 is typical and is 30 cm thick. There are approximately 40 beds within the unit which are 0.5-1.5 cm thick. Each bed coarsens upwards from silt to medium or coarse sand, and the upper contact of each bed is abrupt and sharp. The proportion of sand in successive beds progressively increases towards the top of the unit.

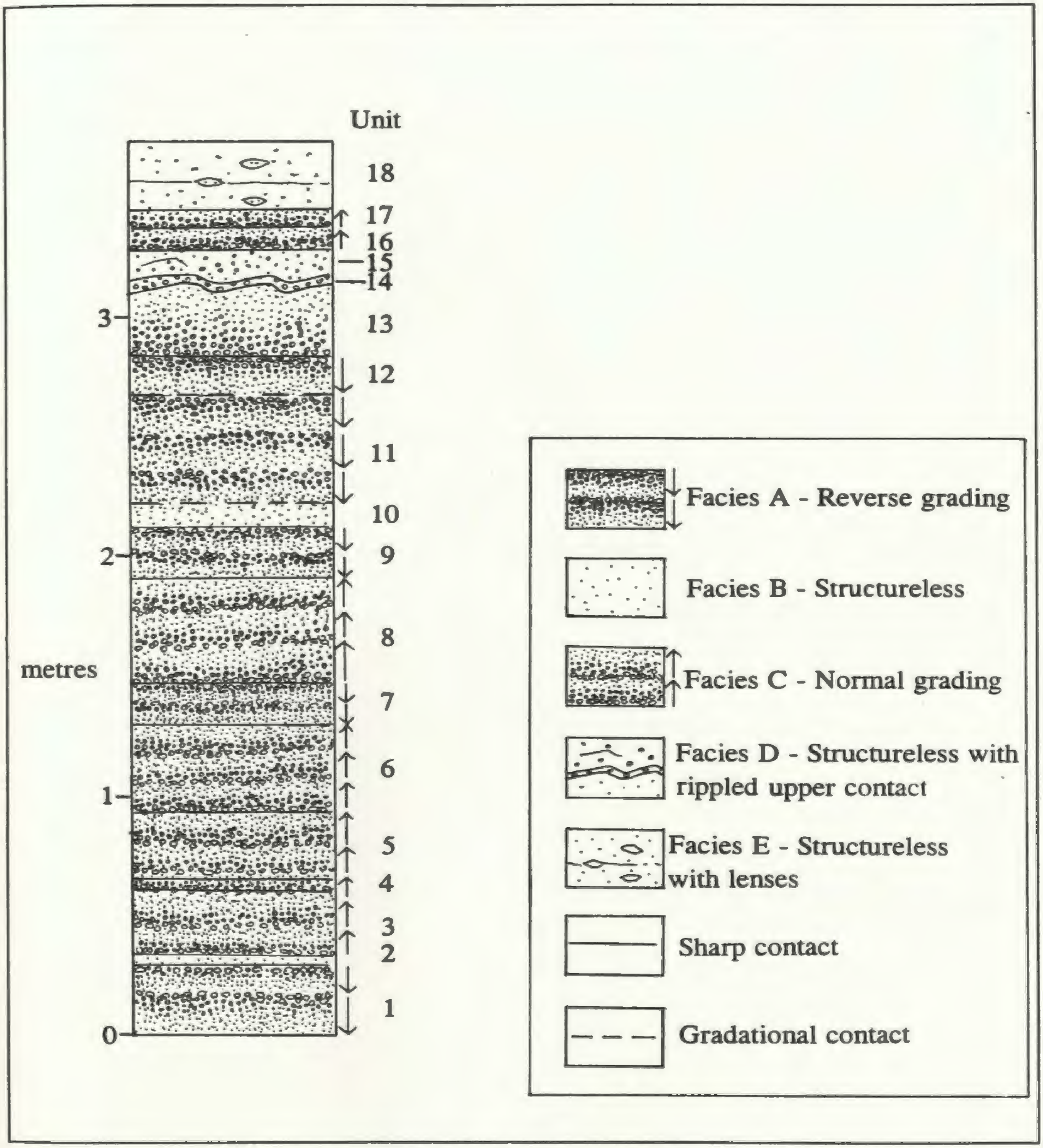

Figure 7.2 - Stratigraphic log of Section A19 at Port Blandford.

Interpretation

The reverse graded beds of Facies A are interpreted to have been deposited by subaqueous grain flows. In a well-sorted sand-dominated environment coarser particles are concentrated near the bed forming a traction carpet that is maintained by grain collisions creating a dispersive pressure (Bagnold 1954). Coarser particles rise above the traction carpet towards areas of lower shear stress, and as the sediment continues to fall out of suspension the frictional forces within the traction carpet increase until the layer freezes almost instantaneously. The process starts again above this layer and preserves the reversely graded bed below. Another method for the formation of reverse grading in grain flows was discussed by Middleton (1970). He proposed a kinetic sieve mechanism whereby the smaller grains fall through the pore spaces between the larger grains, and due to frictional forces the layer freezes preserving the reverse graded nature of the bed.

Facies B

Facies B is composed of silt beds with no internal structure. Units 2 and 10 are typical of Facies B and are 5 and 10 cm thick respectively. The upper and lower contacts of Unit 2 are abrupt and sharp. Unit 10 has a sharp lower contact, and the upper contact is gradational.

Interpretation

The structureless sediments of Facies B are interpreted to have been deposited by suspension sedimentation, grain flows or high-density turbidity currents.

Structureless beds commonly form when there is a large influx of sediment in a short period of time (Reineck and Singh 1980). The sediment is deposited very quickly from suspension and due to continual high sedimentation rates, bedforms are unable to form. Lowe (1982) suggested that although reverse grading is normally associated with grain flows, some structureless beds may form from grain flows when there are low dispersive pressures between the grains. It was also proposed by Lowe (1982) that structureless beds can be formed in sandy high-density turbidity currents by suspension sedimentation, as described previously. On compaction a structureless bed may form (Middleton and Hampton 1976), however dish and pillar structures are characteristic of these beds due to dewatering (Lowe 1982). The beds within Facies B do not display any dish or pillar structures and therefore were not deposited by high-density turbidity currents. Grain flows usually display some kind of reverse grading and therefore the structureless beds within Facies B are interpreted to have been deposited by rapid suspension sedimentation.

Facies C

Beds within Facies C are normally graded. Approximately 50 beds throughout the exposure show normal grading, as found in Units 3, 4, 5, 6, 8, 16 and 17. Unit 3 is 28 cm thick and consists of 9 beds which each fine upwards from medium sand to silt. The beds are 2-8 cm thick. The contacts between the beds are generally sharp, and small loading structures were observed along one of the contacts. The loading structures are formed in silt with sand from the upper bed protruding into the silt below. The structures are approximately 1-5 cm wide and 1.5-3 cm deep, and have a convexo-planar or wedge shape. Units 4, 16 and 17 have several small clasts scattered throughout the units. Unit 16 is 10 cm thick and consists of medium sand that fines upwards to fine sand. Within the fine sand, there are 3 clasts 2-3 cm in length. One of the clasts is aligned vertically, and deforms the gradational contact between the coarse and fine sand. The upper contact is also deformed and drapes over the vertical clast. The other two clasts are inclined 45°, and the contacts above and below the clasts are similarly deformed.

Interpretation

Facies C is characterized by normal graded beds, and these are interpreted to have been deposited by underflows, interflows or overflows. Fining-upwards sequences are generally associated with suspension settling from overflows and interflows (Reineck and Singh 1980; Collinson 1986), however Mackiewicz *et al.* (1984) described fining-upwards sequences in ice-proximal underflow deposits in Alaska. The deposits were thick, dominated by fine to coarse sand and although they exhibited some reverse grading, once the maximum particle size had been reached there was a change to normal grading (Mackiewicz *et al.* 1984). Underflows are relatively rare in the marine environment due to the density differences between fresh and saline water, however Mackiewicz *et al.* (1984) suggested that they can form close to the source of the incoming water. The large number of units with normal grading representing Facies C, and the grain flows and suspension settling of Facies

A and B respectively, suggest that underflows, close to the incoming source, were responsible for the deposition of all the sediments.

Small load structures along one of the contacts were formed by intrusion of the overlying sand into the underlying silt. The deformation occurred due to overloading of the denser sand into the less dense and possibly liquified silt layer below (Reineck and Singh 1980; Lindholm 1987). The formation of the load structures indicates that the upper sand bed was deposited quickly, preventing the water within the silt from escaping. This caused the silt to become liquefied and deformation to occur as the water was expelled (Lindholm 1987).

Small clasts within Units 4, 16, and 17 appear to have been dropped from above, and they deform the upper and lower contacts of the units. Similar structures were described by Thomas and Connell (1985), and therefore the clasts are interpreted as dropstones that were released from icebergs.

Facies D

Three units constitute Facies D, characterized by generally structureless sand with ripple structures along its upper contact. Unit 13 is 31 cm thick and fines upwards from gravelly sand to structureless fine sand at the top of the unit. The lower and upper contacts are sharp. The upper contact is marked by ripple structures that are continuous along the bed (Plate 7.3). The ripples have a wavelength of 17 cm and an amplitude of 3 cm. Within the ripples there is some internal structure, with lee side preservation dominating the bedforms. The ripples are slightly coarser

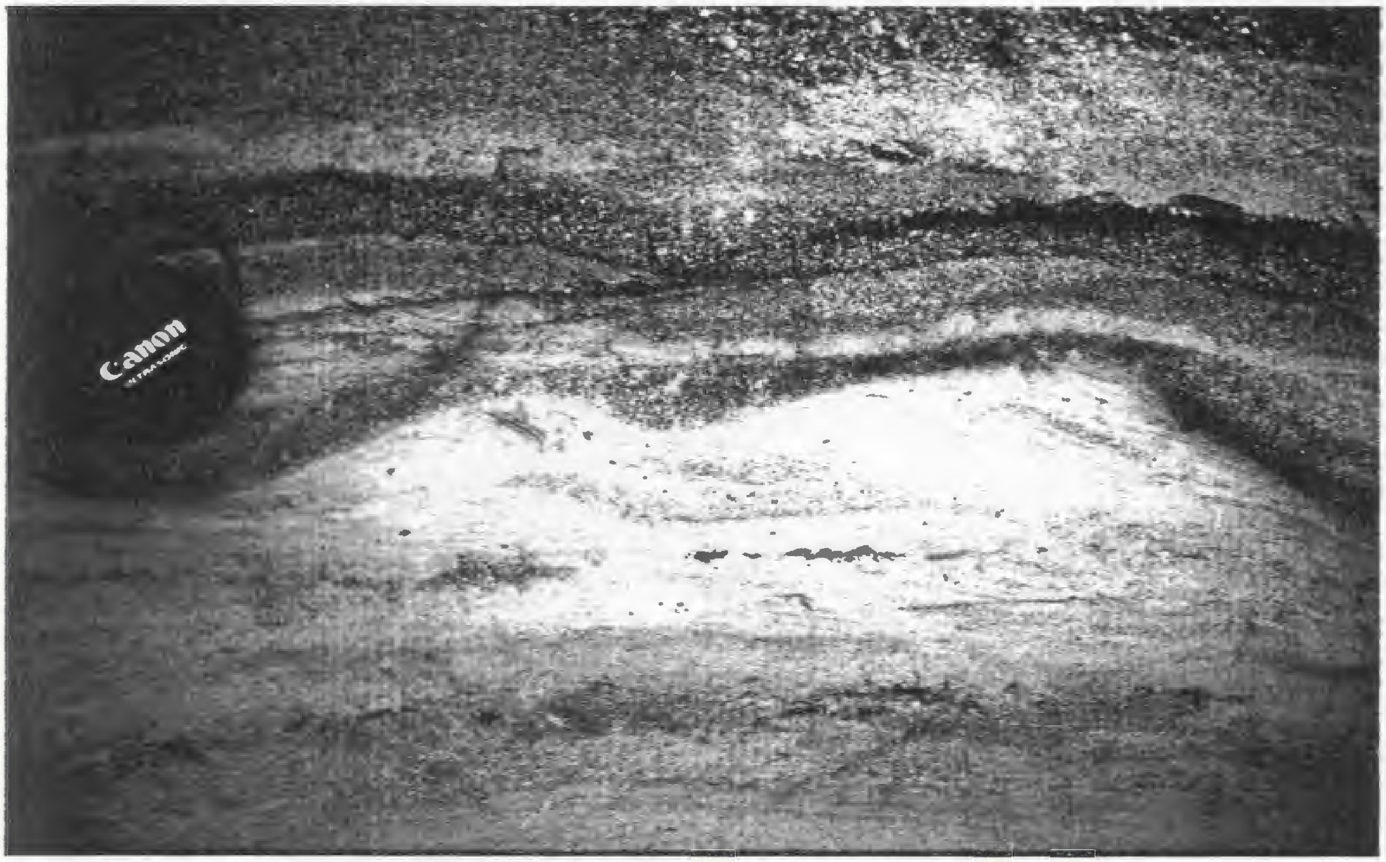

Plate 7.3 - Ripple structures along the upper contact of Unit 13 (A19). The ripples have a wavelength of 17 cm and an amplitude of 3 cm. There is some lee side preservation within the ripples and the ripples were formed by water flowing towards the east (left to right).

on the lee side. Unit 14 directly above the ripples mirrors the underlying structures. It is composed of coarse sand and is 4 cm thick. Unit 15 is 10 cm thick and is composed of beds of medium sand that drape the underlying bedforms. A ripple structure was observed in Unit 15 and is composed of fine sand overlying the coarse bed below. The ripple has a wavelength of 17 cm and an amplitude of 3 cm and the bed thins laterally eastwards, draping the underlying bed. Contacts between each unit are generally sharp.

Interpretation

Facies D is the only facies within the section that has ripple structures along its upper contact. Mackiewicz *et al.* (1984) recognized underflows within ice-proximal sediments that were related to turbidity currents. The sediments were composed of normally graded beds that became progressively finer towards the top of the unit. The upper contact was commonly characterized by a wavy structure thought to be ripples. Walker (1984) discussed the sedimentary structures associated with turbidity currents (Figure 7.3). The structures were initially described by Bouma (1962). At the base of the turbidite Bouma identified a structureless or normally graded unit that passed into planar beds and then into rippled or convoluted sediments. Above the rippled sediments there were additional planar beds and the top of the turbidite was characterized by mud. The type of sedimentary structures produced depends on the amount of sediment available and the flow velocity of the turbidity current. The sequence described by Walker (1984) represents a continual

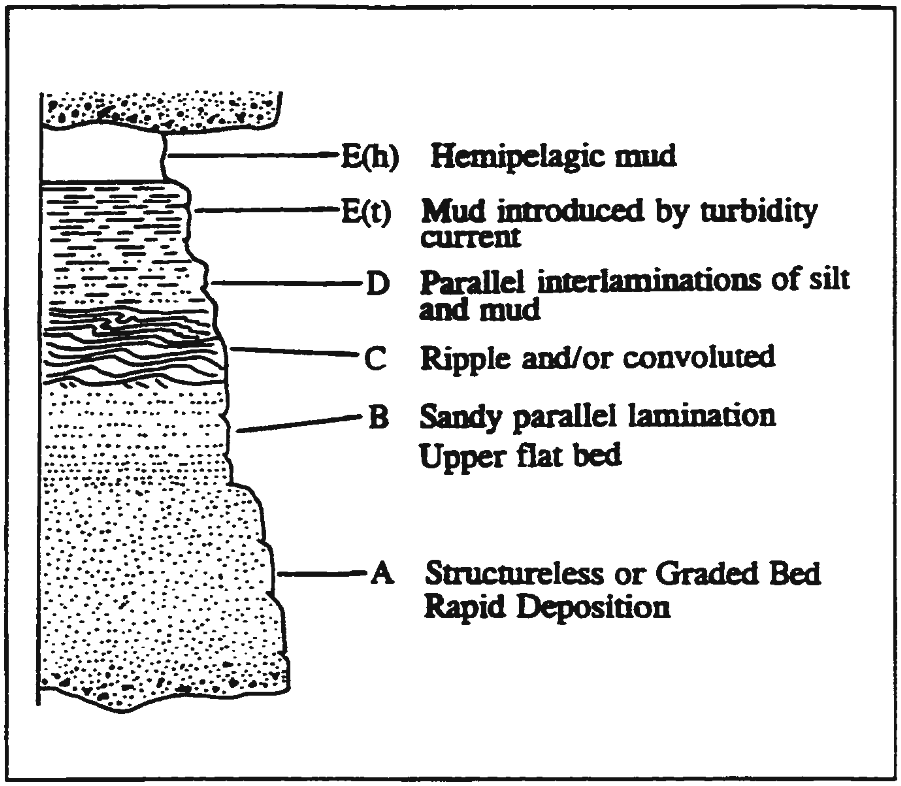

Figure 7.3 - The sedimentary structures associated with the Bouma sequence (after Walker 1984).

decrease in flow energy with the normally graded unit deposited by very rapid deposition, followed by upper flow regime plane beds, and then lower flow regime ripples. As the flow velocity continues to decrease lower flow regime planar beds are deposited and finally the turbidite mud is deposited by suspension settling. The complete sequence is rarely observed, and one or more units are usually missing from the sequence (Walker and Mutti 1973). The units within Facies D fine upwards from gravelly sand to fine sand and there are ripple structures at the top of the unit. I propose that these sediments represent turbidite deposition from an underflow. Finer sediments representing suspension settling from the underflow were not observed above the ripples. This may be due to current reworking or the finer sediments may have been carried farther out to sea.

Facies E

Only one unit represents Facies E (Unit 18) and it is composed of structureless sand that contains lenses of finer sediment. Unit 18 is 30 cm thick and consists of silty sand with an iron rich layer approximately 4 cm above the base of the unit. The lower 20 cm of the unit has 4 lenses of fine sand which are 1-5.5 cm long and 1-3 cm thick. The lenses have a variety of forms, however there is no internal structure within the lenses. Two of the lenses have a biconvex shape and they are surrounded by structureless sand. There is an irregular shaped lens 14.5 cm above the base of the unit. A bed of silt continues on either side of the lens and remains at the same height. The contacts with the lens are sharp and horizontal. The final lens has a plano-convex shape and it too is characterized by a bed of silt which continues on either side of the lens. The bedding on the east side of the lens is displaced upwards, however the contacts remain sharp and horizontal. There is no indication of the beds continuing below or above either of the lenses, and the lenses do not deform the beds in any other way.

Interpretation

The structureless sand of Facies E may have been deposited by suspension settling, a grain flow, or a high-density turbidity current, and these processes were described previously in Facies B. The lenses of finer sand within the structureless sand are interpreted to have been deposited as frozen blocks of sediment that were deposited contemporaneously with the surrounding sediment or by the infilling of cavities in a turbidity flow (cf. Catto *et al.* 1989). The frozen blocks may have come from the collapse of a frozen bank, close to the site of deposition, and the clumps of sand were deposited within the structureless sand, where they were covered by sediment and subsequently melted. Deformation of the unit by the lenses may have occurred, however due to the structureless nature of the surrounding sediment, very little deformation was observed (cf. Gilbert 1990). Lenses that are formed in association with turbidity currents may have a planar or convex basal contact and the lenses commonly show fining upward sequences. Overlying sediments tend to drape the underlying lenses, and the surrounding sediment is commonly deformed. The lenses within Facies E are structureless and there is a lack of deformation and draping in the surrounding sediment. They are, therefore, interpreted to have been deposited by the melting of frozen blocks of sediment.

The structureless sediment of Facies E is interpreted to have been deposited by a turbidity current. I suggest that the frozen blocks of sediment were eroded and incorporated into the turbidity flow, deposited by rapid sedimentation and melted with little or no post-depositional movement or deformation.

7.1.3 - Discussion

Interpretation of the facies recognized within the two exposures described above suggests the following sequence of events. The marine clay, silt and mud were deposited by suspension settling in relatively deep water. Slumping, initiated landwards of the exposure, caused the underconsolidated sediments to be deformed resulting in the formation of convoluted bedding. After the gravitational slumping event there was a return to suspension settling with increasing amounts of sand being brought into the system. Eventually clay and silt were no longer deposited at this site and medium to coarse sand was deposited by grain flows, turbidity currents and suspension settling. All of these processes can occur as a result of underflows, however for underflows to develop in a marine environment the incoming water must have a high concentration of suspended sediment or the sediment must be deposited close to the incoming source (Mackiewicz *et al.* 1984).

The progressive increase in sand within the exposure may represent a sandy delta that prograded over glaciomarine muds, or it may represent a marine regressional sequence that formed as sea level fell during the last deglaciation. The sand beds are horizontal and therefore would represent the bottomsets within a delta sequence that overlie glaciomarine clay and silt. No deltaic foresets were recognized within the gravel pit at A19, although much of the sediment has been removed. McCabe *et al.* (1994) discussed a shallow marine emergent sequence in Northern Ireland. They suggested that an emergent sequence should show a progression from glaciomarine, to shallow marine and wave-influenced units, followed by upper shoreface sands and gravels. Due to erosion or periods of little or no deposition some of these units would not be preserved.

The sediments at Port Blandford (A18/19) may represent such a sequence with glaciomarine clays and silt being progressively overlain by shallow marine sediments and upper shore face sands. I thus suggest that the sediments at A18/19 represent a marine regressional sequence and are not be part of a prograding deltaic sequence in the absence of deltaic foresets. The marine clay, silt and sand were deposited in fairly deep calm water by suspension settling and were interrupted only once by slumping that was initiated closer to the shore. As sea level fell, from isostatic rebound, the sandy nearshore environment migrated seawards depositing increasing amounts of sand over the marine clay, silt and sand, by grain flows, turbidity currents and suspension settling. The marine clay, silt and sand sediments also migrated seawards and are now being deposited below sea level.

7.2 - Dating of Raised Marine Features

Raised deltas and beaches are usually dated by ^{14}C dating of shells within the marine sediments. This aids in reconstructing the sea-level history of the area. There is, however, a lack of shells within the raised features in Terra Nova National Park and vicinity. One site that contained *Hiatella arctica* is reported from St. Chad's, 4 km north of Eastport. The shells were located at 14 m asl and provided a date of 12,400±110 years BP (GSC-5413) (Liverman 1994). *Hiatella arctica* are found at depths between 1 and 75 m (Peacock 1993), although Dyke *et al.* (1996) indicated that they are commonly found in water depths between 2 and 45 m. It is thought that the shells at St. Chad's were deposited at depths between 15 and 25 m, and that both the shells and the 30 m raised deltas at Sandy Cove, Traytown and Big Brook were deposited contemporaneously. This implies a minimum age of 12,400 years BP for the formation of the deltas and the marine limit of 39 m asl. A maximum date (12,800 years BP) for the formation of the deltas was suggested by Cumming *et al.* (1992) who dated an intact mollusc shell in a core from Bonavista Bay to 12,790±115 years BP (Beta 27227). Given the lack of ice-rafted debris in this part of the core, Cumming *et al.* (1992) concluded that at the time of deposition ice had retreated onto the land and that there may have been a sparse vegetation cover at some coastal locations.

7.3 - Sea-Level History

A possible marine limit of 39 m asl has been suggested for Terra Nova National Park and vicinity based on the recorded elevations of the delta at Traytown. This elevation is similar to that proposed by Dyke (1972) who measured a marine limit of 38.5 m asl for the Eastport delta system. The marine limit is tentatively dated to between 12,800 years BP (Cumming *et al.* 1992) and 12,400 years BP (Liverman 1994). Although there are no ^{14}C dates for Terra Nova National Park and vicinity, shells from marine deposits 14 m asl at St. Chad's were dated to 12,400±110 years BP (GSC-5413). These shells are typically found at depths up to 30 m and I suggest that the shells and raised deltas were deposited simultaneously.

Based on raised delta and raised beach elevations throughout the area there were at least five major stands of sea level between the onset of deglaciation and the present. These were at 39 m, 31 m, 21 m, 16 m and 5 m asl. The raised marine features in Terra Nova National Park and vicinity follow the gradual decrease in marine limit from the Northern Peninsula towards the Avalon Peninsula as discussed by Liverman (1994). The marine limit progressively decreases from 75 m at Springdale (Scott *et al.* 1991), 58 m at Botwood (Mackenzie and Catto 1993) and

52 m at Carmanville (Munro and Catto 1993), to 39 m at Traytown. These features are all thought to have formed approximately 12,800-12,400 years ago.

The sea level dropped below present levels between 12,000 and 10,000 years BP and reached a possible minimum of -17 m prior to 8,600 years BP (Shaw and Edwardson 1994). Sea levels rose to -4 m at Eastport by approximately 5,500 years BP, and by 3,000 years BP sea level was -0.7 m below present (Shaw and Forbes 1990). Shaw and Forbes (1990) suggest that present sea levels were reached approximately 2,000 years BP and that there has been very little change since this time.

Chapter 8

Permafrost and Periglacial Activity

8.0 - Introduction

Identification of fossil structures associated with permafrost within Newfoundland is rare, and there are relatively few documented examples of periglacial activity. During the 1970's, Quaternary research in Newfoundland by Brookes (1971), Eyles (1977), and Tucker (1979) identified structures interpreted as ice-wedge casts.

Brookes (1971) described ice-wedge casts from two locations on the west coast of Newfoundland. One cast was found at St. David's, and several others were described at York Harbour. The casts were formed in well-sorted, bedded, sandy and pebbly gravel associated with coastal marine deposition, and were filled with a mixture of the surrounding and overlying sediment. Some bedding was apparent in the ice-wedge casts at York Harbour, as the gravel at this site was more permeable. Casts at both locations were typically 40 to 60 cm wide, and 1.8 to 2.5 m deep. Brookes (1971) concluded that for ice-wedges to form in coarse, permeable material at these sites the sediment would have to be frozen to prevent water movement.

Eyles (1977) identified several ice wedge casts from Birchy Bay, northeast of Lewisporte. The wedges developed in fine-grained sediments that underwent postdepositional deformation and faulting before the formation of the ice-wedge casts. The main ice-wedge cast had a maximum width of 2.5 m and was 2 m deep. Infill and small normal faults around the margins of the wedge structure record the degradation of the ice-wedge. The wedge formed after the main period of faulting, due to its transection of the surrounding fault planes. Eyles (1977) attributed the large size of ice-wedge casts in the Birchy Bay area, compared to the smaller casts found on the west coast by Brookes (1971), to the fine grained sediments which are more suitable for ice-wedge development, given their "low permeability and enhanced water retention" (Eyles, 1977, p2804).

Tucker (1979) described ice-wedge casts from two sites on the Burin Peninsula, southeast Newfoundland. Near Swift Current, an ice-wedge cast 2.7 m wide and 2.4 m deep was identified within trough cross-bedded and ripple laminated glaciofluvial sand and gravel. The infill of the wedge structure was similar to that of the host material and there was some stratification. The bedding surrounding the cast was upturned, one of the characteristics of true ice-wedge casts (Black 1976).

The second cast was located 2 km southeast of Dunn's Pond and was situated in "contorted and slumped cross and parallel-bedded, sand and gravel" (Tucker, 1979, p198). The wedge structure was 2.6 m deep and 70 cm wide, and the perimeter of the cast was characterized by a cemented Fe-Mn shell. The top 80 cm of the infill was chaotic and slumped, and the lower 1.8 m had sub-horizontal sand beds, similar in grain size to that of the surrounding material.

More recent work by Liverman and St. Croix (1989), Liverman *et al.* (1991), Scott (1993), Batterson (*In preparation*), and Liverman *et al.* (*In preparation*) has identified ice-wedge casts in other parts of Newfoundland. Liverman and St. Croix (1989) identified three wedge structures in deltaic sand deposits near Baie Verte Junction. The casts were 1.5 m deep, up to 2 m across, and filled with structureless pebble gravel. Many of the clasts within the wedge structures were aligned with near-vertical long axes. Sand beds sloping at high angles parallel to the side of the casts were contorted due to deformation as the ice-wedges were formed.

Liverman *et al.* (1991) studied the Quaternary geology of the Springdale area and identified several wedge structures 80 to 100 cm deep and a maximum of 40 cm wide, tapering to 1 cm at the base, on the northern side of Indian Brook valley. The ice-wedge casts were formed in planar bedded, well-sorted, medium sand. The infill of the casts was moderately to poorly sorted gravel, similar to that of the overlying material, and showed some internal stratification parallel to the sides of the wedge structures.

Scott (1993) analysed the glaciofluvial and raised marine deposits adjacent to Route 340, near Lewisporte. An ice-wedge cast, 3 m deep and 5 to 40 cm wide, was identified within interbedded sand and gravel. Beds on either side of the cast were vertically deformed. Pebbles within the upper part of the wedge were randomly aligned, whereas those near the base of the wedge were aligned with the long axes approximately vertical.

Several ice-wedge casts have been recognized by Batterson (*In preparation*) south of Pasadena in small gravel pits, at an elevation of 135 m above sea level. A well-defined wedge structure, 20 cm wide and 1.5 m deep, was formed in planar bedded, poorly sorted sand and gravel. Clasts along the margins of the wedge were vertically aligned, and the centre of the wedge was filled with structureless sand and gravel. Smaller ice-wedge casts in South Brook valley were 30 to 40 cm wide, 50 to 60 cm deep, and were filled with poorly sorted sand and gravel. These wedges truncated interbedded moderately sorted sand and gravel. Along the margins of the wedges, the beds were deformed and dipped parallel to the wedge sides.

Six ice-wedge casts found in an active gravel pit south of Botwood, were described by N. Catto (Department of Geography, Memorial University of Newfoundland, pers. comm.) and Liverman *et al.* (*In preparation*). The wedges have a maximum width of 1.5 m and are 1.5 to 2.5 m deep. The surrounding material was horizontally bedded, well sorted coarse to fine sand, while the wedge was infilled with moderately sorted granule to pebble gravel, similar to the overlying sediment. Internal stratification within the wedge was parallel to the wedge sides, and the long axis of many of the clasts were approximately vertical. The surrounding strata were deformed and dipped gently towards the cast.

8.1 - Ice-wedge Casts in the Terra Nova National Park area

Two gravel pits in the vicinity of Terra Nova National Park contain structures interpreted to have formed in a periglacial environment. The first was found approximately 3 km north of Port Blandford (A19).

A well-defined wedge structure (Plate 8.1) was formed in planar interbedded medium to coarse sand and silt. Most of the individual beds coarsen upwards from silt to sand, or from medium sand to coarse sand with scattered small pebbles. The wedge is a maximum of 0.7 m wide and 1.35 m deep, and the west side of the wedge is clearly differentiated from the host material by vertically aligned clasts along the margin (Plate 8.1). The infill of the wedge is coarser than the surrounding material, and was apparently derived from the overlying sediment, although the sediment has been disturbed. There is some internal stratification parallel to the sides of the wedge. Where beds of the host material make contact with the wedge structure, they are deformed and dip parallel to the margins of the wedge. The east side of the wedge appears to contain coarser material than the western side. The wedge is light brownish red in contrast to the host material, and it is suggested that due to the movement of water through the more permeable infill material, the sediment within the wedge has become stained with iron (cf. Svensson 1988; Walters 1994).

Plate 8.1 - Well-defined wedge structure in sand and gravel at Port Blandford (A19). The wedge has a maximum width of 0.7 m and a depth of 1.35 m.

A second wedge structure was identified at Southwest River, approximately 5 km south of Port Blandford. The wedge is found within a silty sand diamicton with a variety of clast sizes, and it protrudes at a high angle from the exposure face. It is 65 cm long and 60 cm wide. The sediment within the wedge is very similar to the host material, although there are more clasts present in the infill, aiding in the identification of the wedge structure. Many of the clasts in the wedge are vertically aligned. This suggests that the infill was derived from overlying sediment, although there has been some soil development above the wedge. The host and wedge material are very similar and therefore it is not possible to identify conclusively any deformation of the host material. The base of the wedge extends into sandy sediment below, and again there appears to be no definite deformation of this host material.

8.2 - Discussion

The wedge structures described above are thought to be casts of former ice-wedges. The morphology and sedimentology of these wedges correspond to the criteria established by Black (1969, 1976, 1983). With the melting of the ice within a wedge, material lying above the ice-wedge should fall into the space created by the melting ice. The ice-wedge would seldom melt completely before sediment was deposited within the wedge, and therefore as overlying material falls into the wedge it becomes foliated. The foliation is parallel to the sides of the wedge, and may extend across the base of the wedge (Black 1969; Walters 1994).

A second diagnostic feature of ice-wedge casts is the deformation of sediment surrounding the wedge. As the ice-wedge grows it may create pressure on the surrounding sediment, causing this sediment to be deformed. Beds of the surrounding sediment may be pulled down as the ice-wedge grows, and frequently become parallel with the margins of the wedge (Black 1983). This may also occur if the frozen material surrounding the wedge thaws and sinks downwards before the wedge can be infilled with secondary material (Walters 1994). The enclosing sediment may also exhibit upturning at the margins of the wedge. Drifting sediment collects in the wedge as it grows, and this creates a space problem during the summer as the permafrost melts and expands. This forces excess material around the wedge to the surface (Black 1976, 1983; Watson 1981; Walters 1994). Although deformation of these marginal beds can be obscured, small pebbles within the sediment may be realigned and serve to indicate the deformation.

8.3 - Palaeoclimatic Implications

Ice-wedge casts are one of the most diagnostic features of permafrost and have been used in many studies to infer past climatic regimes (Burn and Smith 1983; Péwé 1966). Many of these areas now experience a temperate climate and are no longer influenced by permafrost, however palaeoclimates can be inferred using data collected in areas currently experiencing permafrost conditions. Péwé (1966, 1973) and Hamilton *et al.* (1983) studied present permafrost conditions to determine the mean annual air temperature (MAAT) that is required for ice-wedges to form. Péwé (1973) conducted most of his research in central Alaska and suggested that ice-wedges would form only in areas of continuous permafrost where "the ground must cool at a certain rate for a certain period of time in a winter cold snap" (p18). He proposed that the MAAT must be at least -6 °C to -8 °C, and at these temperatures the ground will thermally crack and ice-wedges will begin to form (Péwé 1966, 1973). Ice-wedges can form in discontinuous permafrost zones, but they are usually inactive and were probably formed when the climate was cooler and the area was part of the continuous permafrost zone (Péwé 1973). Péwé concluded that the identification of ice-wedge casts in temperate areas provided evidence for the existence of permafrost conditions around the margins of the Wisconsinan ice sheet, and that the MAAT at that time must have been at least -6 °C to -8 °C.

Hamilton *et al.* (1983) studied active ice-wedges in Alaska and proposed that the MAAT does not need to be as low as that proposed by Péwé (1973) for ice-wedge formation and growth, and that the location and type of sediment in which the ice-wedges grow may be influencing factors. The MAAT at the sites was -3.5 °C, 3 °C warmer than that suggested by Péwé (1973). The wedges formed and grew in peat bogs which are usually cooler than the surrounding land, and their location at the base of a slope encouraged the drainage of colder air into the depression (Hamilton *et al.* 1983). Hamilton *et al.* (1983) concluded by proposing that active ice-wedges can form and exist in areas where the MAAT is 3 °C warmer than that proposed by Péwé

(1973), but that due to the topography in which the ice-wedges were found, care must be taken when inferring a regional climate.

Harry and Gozdzik (1988) discussed the formation of ice-wedges, their transformation as they thaw, and their palaeoenvironmental significance. They recognized that permafrost conditions are required for the development of ice-wedges, but that there are several problems when trying to reconstruct the environment in which they were formed. The first is the relationship between the mean annual air temperature and active ice-wedge growth. Péwé (1966, 1973) proposed a MAAT of -6 °C to -8 °C, whereas Washburn (1973) suggested -5 °C, Hamilton *et al.* (1983) proposed -3 °C, and Mackay (1975) thought that infrequent cracking may occur when the MAAT was slightly below 0 °C. The second is that other site-specific factors may be involved including the rate at which the temperature drops, microclimatic effects and the depth of snow cover over the ground surface (Harry and Gozdzik 1988).

Mackay (1992) discussed the frequency of ice-wedge cracking on Garry Island, western Arctic coast, between 1967 and 1987. His results from various parts of the island showed that the frequency of ice-wedge cracking varied from site to site and that factors other than the air temperature, and the rate at which the air temperature fell, must be involved. Mackay (1992) concluded that snow cover was an important factor and that if the snow was deep enough, the permafrost would be insulated and the frequency of ice-wedge cracking would be reduced.

Several authors have tried to use the size of ice-wedge casts to infer the duration of the climatic event that formed them (Eyles, 1977; Johnson 1990; Burn and Smith 1993). Based on his work in the N.W.T (Mackay 1986) and Alaska, Mackay (1992) concluded that smaller wedges that crack infrequently may be much older than larger wedges that crack frequently. The result is that ice-wedge casts can be unreliable indicators of time and that the duration of activity cannot be determined from the wedge size. Milder winters or increased snow cover may prevent cracking of ice-wedges, and therefore once the ice melts, the cast that forms may be smaller than that expected for the length of time that the ground was frozen.

The ice-wedge casts identified in Newfoundland indicate that periglacial conditions did exist at some time after glacial retreat. The present MAAT at the southern boundary of the discontinuous permafrost zone in southern Labrador is -1 °C, and isolated patches of discontinuous permafrost currently exist in the highest areas of the Long Range Mountains (Heginbottom *et al.* 1995). Current MAAT temperatures throughout Newfoundland range between 2 °C and 5 °C, and therefore it would not take a significant drop in temperature for frost cracking to occur (Liverman *et al. In preparation*). The suggested MAAT for ice-wedge formation and growth ranges from -6 °C (Péwé 1973) to -1 °C (Mackay 1992), which indicates that ice-wedges may form in areas of discontinuous permafrost which are characterized by occasional winters with abnormally low temperatures, and limited snow cover. Once the ice-wedges have formed, they do not require low temperatures to persist (Mackay 1992). This suggests that temperatures may not have had to fall as much as 10°C lower than the present, as Eyles (1977) suggested, for ice-wedges to form in Newfoundland.

The amount of snow cover is a critical factor in the development of ice-wedge cracks in the modern environment. Mackay (1992) suggests that areas experiencing less than 50 cm of snow throughout the year may be suitable for the growth of ice-wedges if the temperatures are low enough. Under present conditions the volume and persistence of snow, and the temperatures experienced in Newfoundland are enough to prevent the formation of ice-wedges. At Terra Nova National Park the daily mean temperature in February is -6.6°C, and an average of 53 cm of snow accumulates per month between December and April (AES 1993). The presence of ice-wedge casts in Newfoundland indicates that the winter climate was drier and colder in the past than it is today.

Ice-wedge casts throughout Newfoundland are developed in similar sediments (Liverman *et al. In preparation*) and at similar stratigraphic positions suggesting that they were all developed within a single broad period of climatic cooling that affected the exposed coastline of the island. At the end of the last glaciation, sea levels dropped and newly exposed unvegetated terrain was subjected to colder, and drier winters (Liverman *et al. In preparation*). Most of the ice-wedge casts in the vicinity of Terra Nova National Park are found at coastal locations. The areas are relatively flat and low-lying, and their exposure to the coast ensures that they are currently not affected by a prolonged snow cover due to the windier conditions experienced in these areas. There are, however, many similar sites which show no evidence of periglacial processes. Permafrost in this area is thought to have been discontinuous, governed by topography and microclimatic conditions.

Chapter 9

Holocene

9.0 - Introduction

Deglaciation of the Island of Newfoundland was completed between 10,000 and 8,000 years ago (Rogerson 1982; Anderson and Macpherson 1994), and since this time it has experienced climatic change that has influenced vegetation growth, and isostatic recovery that has affected coastal and fluvial processes throughout the Island. Many of the present features and vegetation have already been discussed in previous sections (e.g. physiography, soils and vegetation). However, it is important to realise that the climate and physiography continue to change even after the direct influence of the ice caps has disappeared.

9.1 - Climate Change

Macpherson (1981, 1995), Davis (1984, 1985, 1993) and Irwin (1993) have reconstructed Holocene climatic changes by pollen analysis and ^{14}C dating of lake and peat sediments. The development of peat did not begin immediately after deglaciation, and therefore these records are commonly truncated at the base (Macpherson 1981). Early Holocene climatic changes are consequently reconstructed from lake sediments.

Macpherson (1981, 1995) analysed and dated pollen grains from lake cores located throughout Newfoundland. One lake core was analysed by Wolfe and Butler (1994) from Pine Hill Pond in Terra Nova National Park. The latter study concentrated on the late-glacial environment rather than the Holocene, and only briefly mentioned that by 9,000 years ago mixed forest communities had become established. Given the lack of data from the Terra Nova National Park area, lake sediments from Central Newfoundland are assumed to be comparable with lake sediment deposition within western areas of the Park (J. Macpherson, Department of Geography, Memorial University of Newfoundland, pers. comm. 1996). The earliest tree pollen recognized within central Newfoundland were white spruce (*Picea glauca*) and balsam fir (*Abies balsamea*), and these replaced shrub birch (*Betula* spp.) heath at approximately 9,500 years BP (Macpherson 1995). Temperatures and the length of the growing season continued to increase after deglaciation, and by 8,500 years BP boreal forest, including red pine (*Pinus resinosa*), had rapidly expanded over much of the interior (Macpherson, pers. comm. 1996). Coastal areas continued to be dominated by shrub birch due to the influence of low ocean temperatures, however, by 5,500 years BP ocean surface temperatures had risen and black ash (*Fraxinus nigra*) had expanded into northeast Newfoundland (Macpherson 1995). Approximately 4,500 years ago the climate started to cool and deteriorate, and there was a reduction in pine (*Pinus* spp.) adjacent to coastal sites as ocean temperatures decreased. Since 4,000 years BP summer temperatures and the growing season have continued to decrease, and moisture has steadily increased. This is reflected in the decrease of red pine and black ash, both of which prefer a slightly drier climate, and an increase in speckled alder (*Alnus rugosa*), sweetgale (*Myrica gale*), grasses and sedges, and sphagnum, all of which prefer wetter and cooler habitats (Macpherson, pers. comm. 1996). Vegetation today in Terra Nova National Park is dominated by black spruce (*Picea mariana*), with lesser amounts of balsam fir, white spruce, white birch (*Betula papyrifera*) and trembling aspen (*Populus tremuloides*). Red pine has been extensively cut throughout the past 150 years and therefore no longer grows in central Newfoundland, however it has recently been replanted in north-central Newfoundland.

Davis (1984, 1985, 1993) and Irwin (1993) discussed the development of peatlands in Newfoundland, and how their formation was related to climatic change. Peatland formation requires a surplus amount of water that occurs with high precipitation, relatively low temperatures, and high relative humidities (Wells and Pollett 1983; Davis 1984, 1993). Pollen analysis from bogs and lakes indicates that prior to 3,000 years BP, Newfoundland was mainly covered in trees, and that peatlands had a limited extent (Davis 1984, 1993). After this time there was a rapid change in the regional vegetation due to a climatic deterioration. Peatlands quickly developed and expanded by paludification over mineral substrates that had previously supported boreal forest (Davis 1984, 1985, 1993).

Lake and bog stratigraphy throughout Newfoundland indicates that Newfoundland experienced a mid-Holocene hypsithermal between 6,000 and 4,000 years BP, approximately 2,000 years after central North America. Macpherson (1981) and Davis (1984) suggested that the lag in the climatic optimum was due to ice remaining over Labrador until 6,000 years BP and the cold Labrador Current. During the Hypsithermal most of Newfoundland was covered by boreal forest, although coastal locations may have been characterized by shrub birch. Since 4,000 years BP the climate has steadily deteriorated, becoming colder and wetter and providing the ideal conditions for peatlands to develop. Large areas within Terra Nova National Park, notably Gros Bog and Saltons Marsh, are typical of these peatlands. Macpherson (1995) suggested that the lower temperatures and increased moisture were initiated by a decrease in ocean surface temperatures.

9.2 - Coastal Development

Most of the coastline of Terra Nova National Park is characterized by bedrock which forms cliffs that rise directly out of the sea. Beaches are restricted to small bays where sediment is trapped (e.g. Dumpling Cove, Bread Cove and Platter Cove), or where there is an ample supply of sediment, as at the base of raised marine features (e.g. Sandy Cove [Plate 9.1], Eastport Beach and Buckley Cove). Deposition and erosion of the raised marine features and the subsequent beach formation is a result of sea level changes throughout the Holocene. Shaw and Forbes (1990, 1995)

Plate 9.1 - Sandy beach at the base of the raised marine delta at Sandy Cove. The beach is 10 m wide and the bluff is 25 m high. The upper surface of the raised feature has an elevation of 30 m asl.

discussed the sea level history of northeast Newfoundland. They suggested that the raised marine features were formed when sea level was higher than present at the end of deglaciation approximately 12,000 years BP. Due to isostatic recovery sea level dropped below present levels to -17 m by 8,600 years BP (Shaw and Edwardson 1994), and by 5,490 years BP sea level had risen to -4 m below sea level.Shaw and Forbes (1990) indicate that by approximately 2,000 years BP sea level had reached its present position.

Sediment spits are presently being formed at the mouths of Terra Nova River and Cobblers Brook. Their formation suggests that large amounts of sand are being transported down the rivers. Due to longshore drift, the sediment is moved along the shore in a southwesterly and northeasterly direction, respectively, forming spits rather than deltas.

9.3 - River Incision

Rivers pass through large volumes of glaciofluvial sediments within Terra Nova National Park and vicinity. As sea level has risen and fallen due to the melting of the ice caps and isostatic recovery, the rivers have had to adjust to new base levels. Several rivers have cut through the glaciofluvial deposits and are now flowing within valleys with banks 30 m high on either side (e.g. Northwest Brook, Terra Nova Brook and Big Brook). The fine-grained sand is carried downstream and is deposited in estuaries (Forbes 1984) where rivers, such as Southwest Brook and Big

Brook, debouch into the sea. Due to the large quantities of glaciofluvial material in the area, rivers within Terra Nova National Park and vicinity tend to be braided, and continue to incise into the glaciofluvial material.

Chapter 10

Conclusions

10.0 - Quaternary History

Reconstruction of the Late Quaternary history of Terra Nova National Park and vicinity is important for the overall understanding of the Quaternary history of the Island. By reconstructing the glacial history of small study areas, like Terra Nova National Park, local ice-flow movements are recognized and these may differ from the regional ice-flow directions.

There is no firm evidence to suggest that any of the landforms or deposits throughout Terra Nova National Park and vicinity are older than the Late Wisconsinan. The striations are relatively fresh, especially where they are covered by a thin layer of sediment, and erratics within the valleys and on the hill summits are unweathered. Brookes (1989) identified a small unglaciated plateau at the tip of the Bonavista Peninsula that was covered by felsenmeer. The summits within the study area are free of felsenmeer and this indicates that during the Late Wisconsinan glacial ice covered the entire study area leaving no nunataks protruding above the ice surface.

Terra Nova National Park and vicinity were influenced by one main ice-flow event throughout the Late Wisconsinan. At the maximum of the last glaciation, approximately 20,000 years BP (Table 10.1), glacial ice extended out onto the continental shelf as a grounded ice sheet (Cumming *et al.* 1992). Glacial landforms, such as flyggbergs, roches moutonnées, crag and tails, and striations, all indicate a regional ice-flow direction towards the northeast.

Glacial deposits throughout the study area also reflect the northeasterly movement of glacial ice. Erratics are found within almost all the deposits. Terra Nova Granite is the most distinctive erratic, and its source is west/southwest of Terra Nova National Park, near the village of Terra Nova. The presence of these erratics within the sediments and on the surface of the land indicates that glacial ice moved across the bedrock source and continued in a northeasterly direction before depositing its load. Four sites studied in detail deviate from the general northeast transport pattern. They indicate that there are local ice-flow directions within the overall pattern. These deposits are preserved only locally due to reworking, the shortage of sediment, and erosion. Although the tills at A11a, A21 and A14 were deposited by local ice-flow directions, the high percentage of Terra Nova Granite and other erratics within these deposits indicates that the main ice flow direction was towards the northeast.

Time (BP)	Event	Evidence/Source
0	Cool climate. Rising sea levels.	Macpherson (1995)
~5,000	Climatic optimum followed by climate deterioration	Macpherson (1995)
5,490	Sea level stand at -4 m	Shaw and Forbes (1990)
8,600	Sea level minimum at -17 m	Shaw and Edwardson (1994)
~10,000		
	Ice wedge formation (Younger Dryas)	Ice wedges at Port Blandford and Southwest River.
11,300	Deposition of inorganic bed in organic lake sediments	Wolfe and Butler (1994) Macpherson (1995)
	Sea level stands 27, 21 and 15 m asl	Terraces at Charlottetown and raised marine sediments at Port Blandford
12,400	Deposition of shells at St. Chad's. Marine limit of 38.5 m asl	Liverman (1994) Deltas at Traytown and Sandy Cove. Dyke (1972)
12,790	Ice retreats to inland position. "Normal" marine conditions outer Bonavista Bay.	Cumming *et al.* (1992)
13,000	Ice retreats to present coastline. Underlying topography increasing effect on ice flow directions. Sparse vegetation on land.	Cumming *et al.* (1992) Striations and deposition of basal tills (A11a, A21, A14)
	Ice shelf over Bonavista Bay. Onset of deglaciation.	Cumming *et al.* (1992)
20,000	Late Wisconsinan maximum. Regional ice flow direction towards the NE. Grounded ice sheet in Bonavista Bay.	Grant (1989)

Table 10.1 - Chronology of Terra Nova National Park and vicinity, northeast Newfoundland

The northeasterly trend of landforms and glacial erratics throughout Terra Nova National Park conforms with the glacial history of the Bonavista Peninsula (Brookes 1989) and the Gander area (Vanderveer and Taylor 1987). The three areas indicate radial flow away from an ice centre located over Middle Ridge (Rogerson 1982; St. Croix and Taylor 1991).

As deglaciation commenced an ice shelf rapidly developed in Bonavista Bay, and by 13,500-13,000 years BP the ice margin had retreated in a southwesterly direction to the present shoreline, creating "normal" marine conditions in outer Bonavista Bay. Jenness (1960, 1963) proposed that the ice retreated fairly rapidly and formed a stable ice front inland from the coast. Jenness (1960) identified an end moraine that stretched from Gambo, along the east end of Maccles Lake and Terra Nova Lake, south towards Gisborne Lake near the south coast of Newfoundland. The glaciofluvial deposits analysed support the deglacial history suggested by Jenness (1960). The coarse clast-supported deposits at A11b and A9 are interpreted to have been deposited in ice-proximal braided streams. They indicate that the ice retreated in a southwesterly direction. The meltwater streams first deposited coarse grained sediments west of Glovertown (A9), and as the ice retreated the coarser sediments were deposited farther inland in Big Brook valley (A11b). Glaciofluvial terraces in Big Brook valley suggest that large quantities of sediment were deposited by meltwater streams. Subsequent incision of these deposits by glaciofluvial or fluvial rivers resulted in the formation of the terraces.

Jenness (1960, 1963) identified an esker belt directly west of the end moraine. He observed that eskers commonly form close to granitic bedrock, and that their orientation was similar to that of the regional ice flow direction. Six eskers were identified in Big Brook valley and close to Terra Nova Lake. Eskers form near the ice front and provide an idea of possible terminal ice positions as the ice retreated. During deglaciation, there may have been as many as five major stillstands in the study area as the ice retreated towards the southwest.

Marine shells were found 14 m asl at St. Chad's and were dated to 12,400 years BP. Other evidence pertaining to a marine incursion are the raised deltas at Eastport/Sandy Cove and Traytown, and the marine sediments at Port Blandford (A18 and A19). The sediments at Port Blandford are characterized by horizontally bedded and deformed marine clay, silt and sand, overlain by alternating beds of silty-sand and sand. The sequence represents falling sea level. The deltas at Eastport/Sandy Cove and Traytown were fed by meltwater streams that originated at an ice margin close to Terra Nova Lake. Trough cross-bedding was identified at two sections (A6 and A10a) close to the top of the Traytown delta. The sections are approximately 2 km apart and the sediment becomes finer downstream, indicating that the deposits were part of the same braided stream system. The Traytown delta has no modern fluvial deposition on its surface and therefore the sediment was deposited when sea level was at or close to the marine limit. Ponded sediments (A10b) close to the trough cross-beds at A10a were deposited during floods when meltwater from the main river overflowed into a depression adjacent to the main river channel.

The shells found at St. Chad's are typically found in shallow cool arctic water between 2 and 45 m deep, although they can range to a depth of several hundred metres (Dyke *et al.* 1996). I suggest that the shells and the 30 m raised deltas at Eastport/Sandy Cove and Traytown were deposited contemporaneously. This implies a minimum age of 12,400 years BP for the formation of the deltas and the marine limit of 39 m asl.

The post-glacial marine limit in Terra Nova National Park and vicinity is similar to that postulated by Dyke (1972). The maximum recorded marine limit in northeast Newfoundland increases from 26.5 m on the Bonavista Peninsula (Brookes 1989) to 58 m at Botwood (Mackenzie and Catto 1993), and 75 m at Springdale (Scott *et al.* 1991). The raised marine features in the study area correspond well with a progressive increase in elevation of raised marine features towards the northwest, reflecting the influence of the Laurentide Ice Sheet.

At approximately 12,400 years BP sea level stood at the marine limit and the terminus of the ice is thought to have been directly west of the stillstand position postulated by Jenness (1960, 1963). I suggest that a large proglacial Terra Nova Lake acted as a sink for the coarse grained sediment that was transported from the ice front into the lake by meltwater streams. Beyond the lake a sandy braided river developed depositing the sediments now exposed at A12 and A15. These sand deposits are characterized by horizontal beds, trough cross-beds and climbing ripples indicating flow towards the east/northeast.

By 11,300 years BP sea level had fallen below 15 m asl and the deglacial climatic warming was interrupted by cooler conditions. With a fall in sea level the rivers throughout the study area had a lower base level to grade to causing them to incise into their beds. This initiated the terrace formation in the sandy deposits at A12 and A15. The cooler climate was recorded in lake sediments throughout Newfoundland by the interruption of organic sedimentation with an influx of inorganic sediments (Wolfe and Butler 1994; Macpherson 1995). Dating of organic deposits below and above the inorganic layer indicated that the sediment was deposited between 11,000 and 10,000 years BP (Anderson and Macpherson 1994). Ice-wedges found in the exposed marine deposits at Port Blandford (A19) are thought to have formed during this cool period. Both the inorganic lake sediments and the ice-wedge casts represent a climatic cooling event that is thought to correspond with the Younger Dryas.

Recent research throughout Atlantic Canada has provided other evidence for a Younger Dryas cool period. Pollen assemblages from peat beds and lake cores in Nova Scotia and New Brunswick record a reversion back to tundra-like flora at the expense of arboreal vegetation (Stea and Mott 1989; Mayle and Cwynar 1991). Many of the peat beds are truncated by minerogenic sedimentation, and some are overlain by glaciogenic deposits indicating that there was a readvance at this time (Stea and

Mott 1991; Mott and Stea 1993). Other ice-wedge casts have been identified recently in raised marine sediments and glaciofluvial sediments throughout Newfoundland (Liverman *et al. In preparation*). Their location and stratigraphic position indicates that they developed after the Late Wisconsinan ice had retreated and subsequent to the initial postglacial marine transgression. This suggests that the ice-wedge casts were formed between 11,200 BP and 10,400 BP (Liverman *et al. In preparation*).

The climate rapidly warmed after 10,000 years BP and final deglaciation of the study area probably took place soon after this time. Sea level continued to fall resulting in the formation of the raised deltas and beaches throughout the study area and by 8,600 years BP sea level had reached a minimum of -17 m (Shaw and Forbes 1990; Shaw and Edwardson 1994). The rivers continued to incise into their sediments forming raised terraces along the sides of the rivers such as Big Brook and Terra Nova Brook.

As the temperatures rose and the growing season lengthened the vegetation changed from a herb-dominated tundra to red pine and boreal forest in the interior and shrub birch at coastal sites (Anderson and Macpherson 1994; Macpherson 1995).

Sea level had risen from its minimum to -4 m by 5,490 years BP (Shaw and Forbes 1990) and Terra Nova National Park and vicinity were dominated by boreal forest during the climatic optimum at about 5,000 BP (Macpherson 1995).

For the past 5,000 years the climate has steadily deteriorated becoming cooler and wetter (Macpherson 1995). The rivers have experienced an increase in discharge and commonly flood the surrounding land creating low river terraces adjacent to the modern rivers. Ponds throughout the area may also have increased in size due to the cooler and wetter conditions, although an increase in beaver activity may also be responsible. Throughout Terra Nova National Park and vicinity the cooler and wetter climate has encouraged wetland development (Davis 1984, 1985, 1993), creating large wetland areas such as Gros Bog and Saltons Marsh.

10.1 - Suggestions for Future Work

Detailed regional studies throughout Newfoundland are important for the full comprehension and establishment of the glacial history and post-glacial sea level changes associated with the Island. Newfoundland was dominated by several ice centres during the Late Wisconsinan maximum and it is essential to recognize the different ice flow directions associated with each ice centre. Adjacent regions may have been influenced by separate ice centres and detailed analysis may identify more local ice flow directions. Small parts of northeast Newfoundland have recently been studied by Mackenzie and Catto (1993), Scott (1993) and Munro (1994), however the area between Carmanville and Glovertown has had relatively little attention. Detailed studies in the regions surrounding Musgrave Harbour, Wesleyville and Gambo would provide a complete and thorough understanding of the Late Quaternary history of northeast Newfoundland.

Glaciofluvial deposits are important for reconstructing the deglacial history of an area. Some parts of the study area, notably Big Brook valley and the eastern end of Terra Nova Lake, have a complex assemblage of glaciofluvial landforms and sediments. Areas directly adjacent to these features require detailed analysis. This would facilitate the establishment of ice marginal positions as the ice retreated. It would also provide a greater understanding of the meltwater systems that operated in Big Brook valley and the Terra Nova River valley.

I have proposed in this thesis that a proglacial lake existed in the area now occupied by Terra Nova Lake. Further research is required around the margin of the lake to identify flat-topped terraces that may represent deltas or outflowing river deposits. These sediments may be incised due to a fall in sea level causing the rivers to downcut into their sediments and the lake level to fall. The deltas should be coarse grained especially close to the western end of the lake where the ice margin may have been located. I suggested that the lake acted as a sink for the coarse grained sediment that was deposited directly off the ice front or by meltwater streams entering into the lake. Detailed analysis of the sediment within the lake itself should determine if the lake was a sink for the coarser sediment.

The information in this study must be combined with other regional studies throughout Newfoundland to improve the overall understanding of the Quaternary history of Newfoundland.

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Appendix 1

Location of Striation Sites and Striation Orientations

Site	Location	Orientation(°)
1	970 927 - 967 936(NTM)	294*, 316, 319, 320*, 321, 322, 323*, 326, 334, 340* 003(2), 004(2), 009(2)
2	113 585	153, 260, 296, 300
3	818 768	295
4	825 846	309
5	927 939	311, 313, 318, 324, 325, 331, 343, 335
6	947 929	338, 348, 349, 350, 358, 359, 000
7	974 944	130, 135, 140, 141, 145, 149
8	964 943	009, 026, 045, 320, 325, 336, 339, 342, 348, 350, 353, 356
9	960 950	308, 312, 320
10	958 965	305, 309, 314, 315, 317, 321, 322, 325
11	158 942	302, 308, 309, 310, 312, 314, 315, 316, 319
12	162 935	281, 302, 307, 310, 315, 332
13	828 873	284* 293, 298, 314, 315, 325

Striation measurements from all sites in Terra Nova National Park and vicinity - See Figure 4.1 in the pocket at the back of the thesis for striation locations.

Site	Location	Orientation(°)
14	159 943	230, 231, 241*, 245*, 258, 259, 260, 262
15	794 803	279, 286, 287, 293, 294, 296*, 298, 301*, 305
16	151 673	000*, 010*, 240*, 265, 303, 306, 309, 312, 319*, 320, 331, 340*, 354
17	142 674	284, 285*, 290, 299, 300, 303, 305, 306, 308
18	140 677	298*, 300*, 301, 305, 306*, 308*, 309, 310*, 311, 313, 341, 351*
19	984 907	000, 001, 002, 004*, 005, 007, 009, 012, 015, 334, 336, 342, 351, 355
20	801 917	302, 305, 306, 309, 310*, 311, 316, 318, 321, 323, 332, 340* 335(2), 343(2), 346(2)

Striation orientations from all sites in Terra Nova National Park and vicinity - See Figure 4.1 in the pocket at the back of the thesis for striation locations.

* Nailhead striations

(2) Indicates cross-cutting striations that are younger in age

Appendix 2

Textural and Lithological Data from Tills in Terra Nova National Park and vicinity

	Lithology			Shape				Fabric				
Site	TNG	HSG	M	E	R	D	B	S_1	S_3	K	Or(°)	P(°)
9	59	18	23	28	16	48	08	0.495	0.032	0.02	316.6	14.2
11a-1	39	32	29	12	16	44	28	0.609	0.051	0.31	288.9	19.7
11a-2	41	45	14	00	20	52	28	0.589	0.052	0.25	197.1	16.9
11a-3	35	62	03	11	23	51	15	0.576	0.057	0.24	292.9	21.8
11a-4	22	44	34	04	13	34	49	0.523	0.093	0.22	115.8	06.1
14	41	06	53	30	18	40	12	0.625	0.058	0.40	134.2	29.3
21	32	52	16	-	-	-	-	0.803	0.039	1.17	159.2	22.4

Clast lithology, shape and fabric determinations for the till sites analysed in Terra Nova National Park and vicinity

TNG - Terra Nova Granite

HSG - Hare Bay and Square Pond Gneiss

M - Musgravetown Sandstones, Conglomerates and Volcanics

E - Equants

R - Rollers

D- Discs

B - Blades

Or - Orientation

P - Plunge

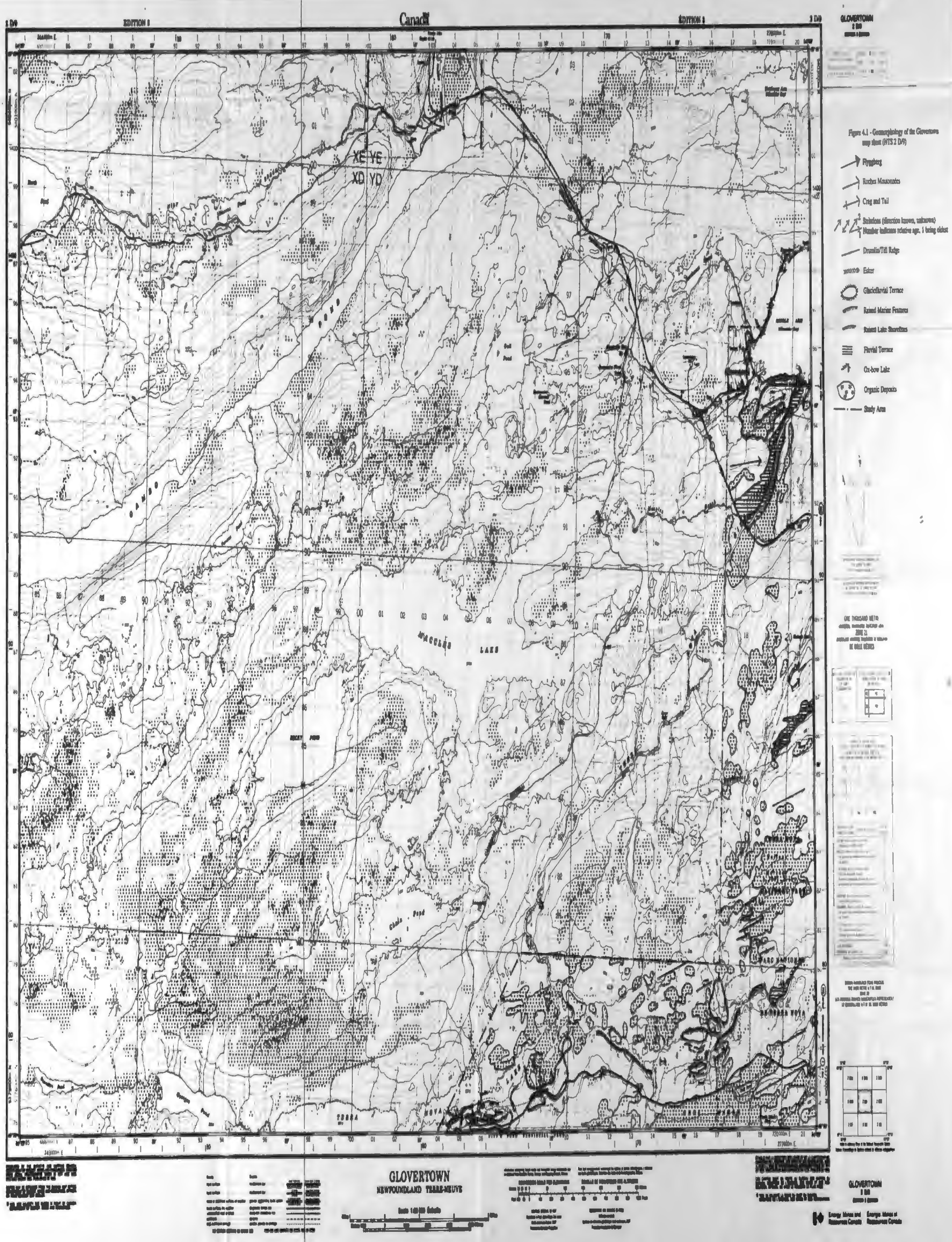
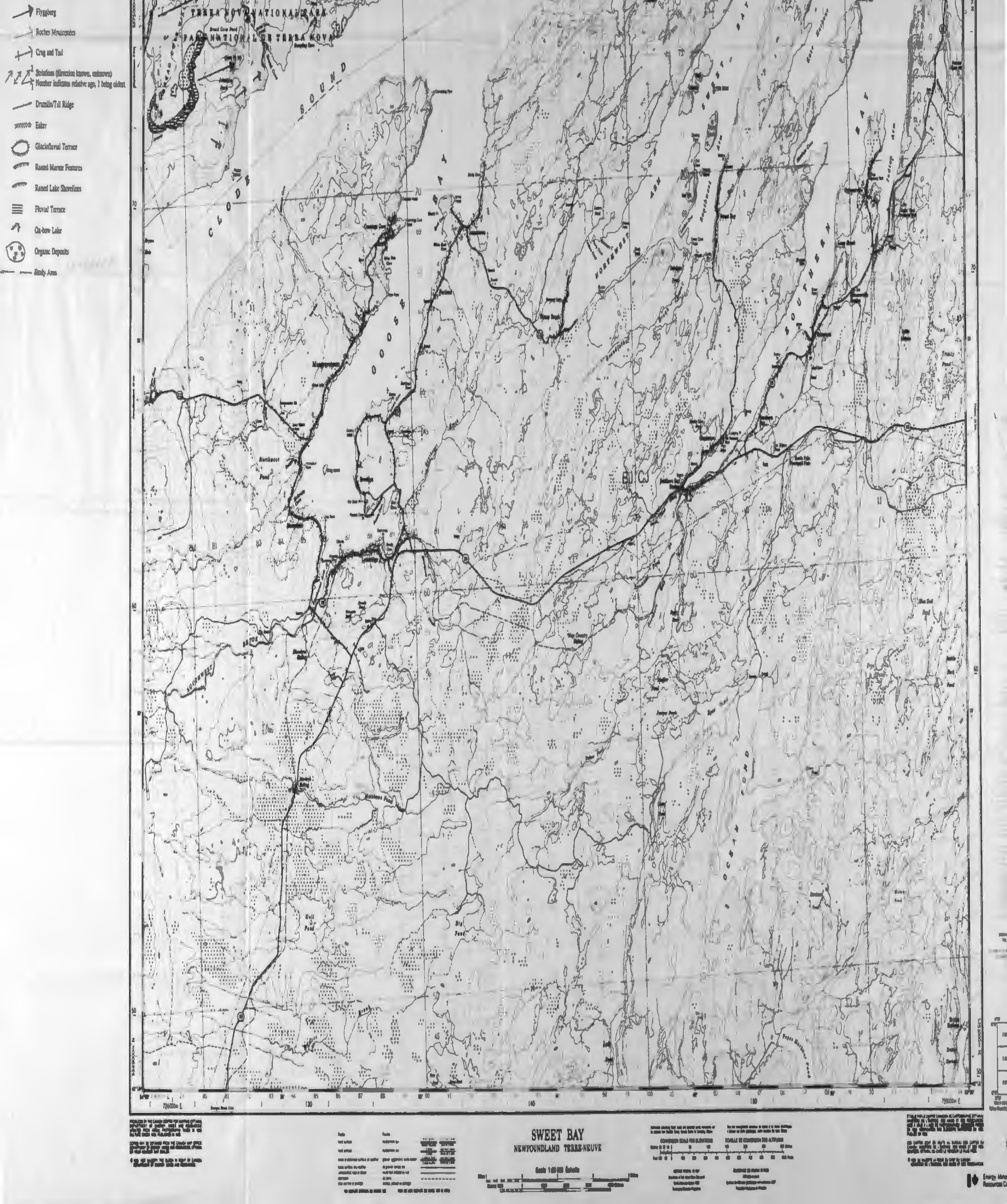


Figure 4.1 - Geomorphology of the Glovertown map sheet (NTS 2 D/9)

Figure 4.1 - Geomorphology of the Sweet Bay map sheet (NTS 2 C/5)



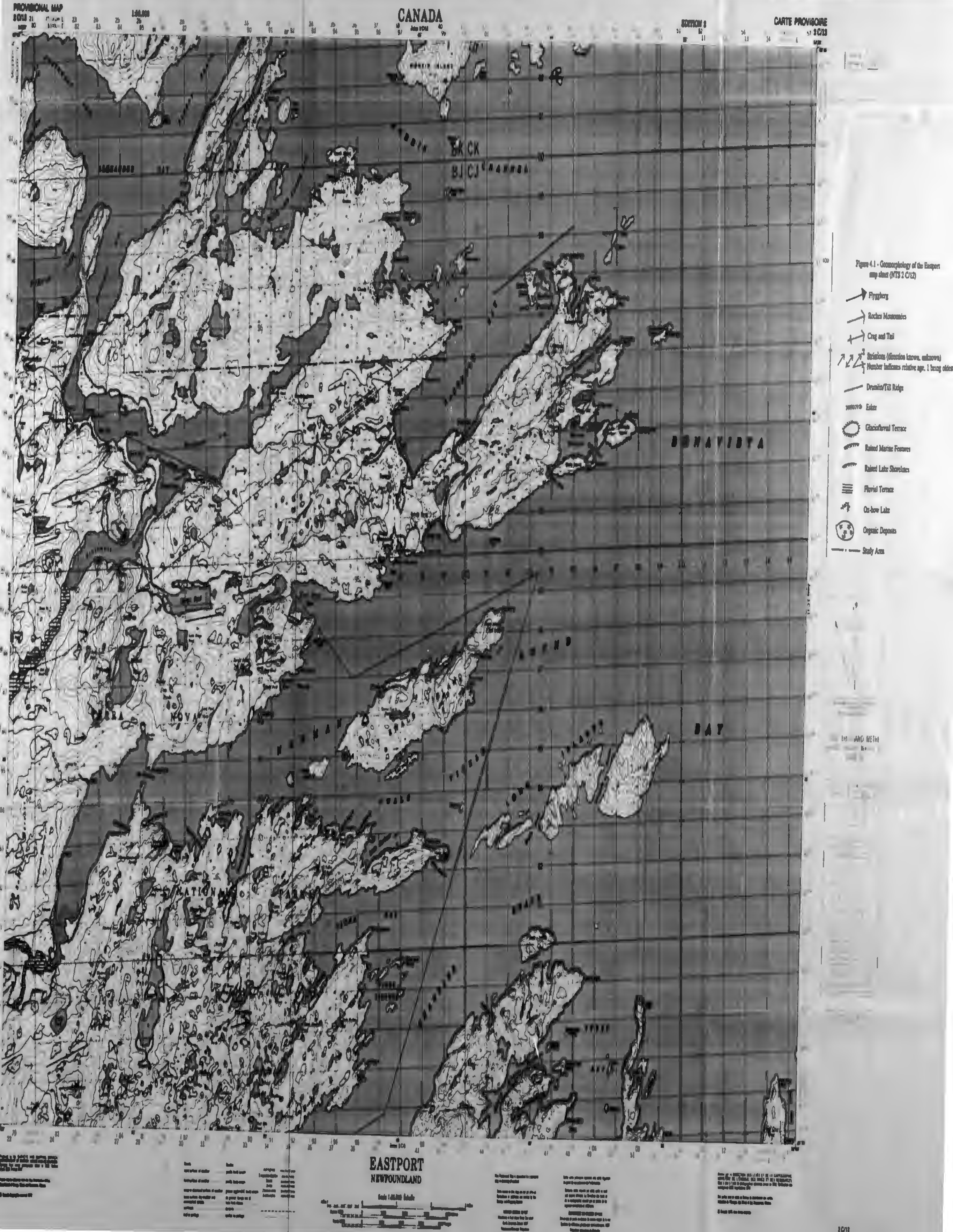
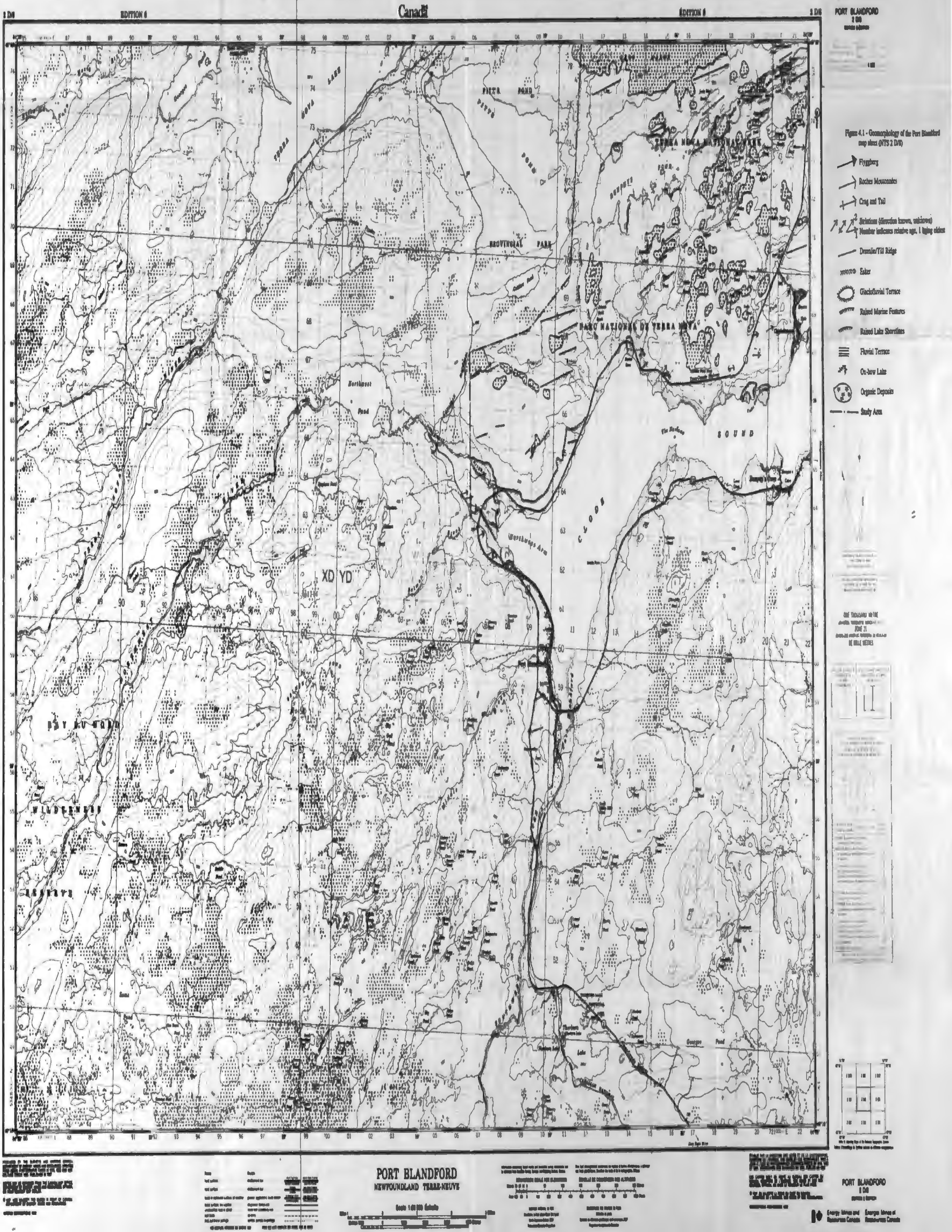


Figure 4.1 - Geomorphology of the Eastport map sheet (NTS 2 C/12)



2 D/8
EDITION 6
Canada
EDITION 6
2 D/8
PORT BLANDFORD
2 D/8
Figure 4.1 - Geomorphology of the Port Blandford map sheet (NTS 2 D/8)
Flyggberg
Roches Moutonnées
Crag and Tail
Striations (direction known, unknown)
Number indicates relative age, 1 being oldest
Drumlin/Till Ridge
Esker
Glaciofluvial Terrace
Raised Marine Features
Raised Lake Shorelines
Fluvial Terrace
Ox-bow Lake
Organic Deposits
Study Area
PITTS POND
TERRA NOVA LAKE
TERRA NOVA NATIONAL PARK
PROVINCIAL PARK
PARC NATIONAL DE TERRA NOVA
Northwest
Pond
SOUND
BAY DU NORD
WILDERNESS
RESERVE
XD
YD
Gasper
Pond
Lake
PORT BLANDFORD
NEWFOUNDLAND TERRE-NEUVE
PORT BLANDFORD
2 D/8
Energy, Mines and Resources Canada
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