LATE QUATERNARY MARINE, DELTAIC AND FLUVIAL DEPOSITS, KANAIRIKTOK VALLEY, COASTAL CENTRAL LABRADOR

SHERIF ABDEL MONEM AWADALLAH

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BY

© Sherif Abdel Monem Awadallah, B.Sc.

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ABSTRACT

The Kanairiktok River Valley, central coastal Labrador, extends more than 100 km inland at about 54°45'N latitude. Three of the five sedimentary units in the bluffs of the modern river were deposited in a marine embayment that occupied the isostatically depressed area during the Late Quaternary. The lateral relationships between the units are poorly known due to limited chronological control. The oldest units (units A and B) were deposited from surface sediment plumes originating from several points of fresh water input into the marine fjord, and from a variety of density currents. Unit B contains rare shells that have been dated at 7950 ± 95 yr B.P. (Beta-28885). Evidence for marine conditions comes from the local occurrence of (a) a periglacial shell community including *Portlandia arctica* and *Mya arenaria*, (b) benthic foraminifera including *Elphidium excavatum* f. *clavata*, *Cassidulina reniforme* and *Islandiella helenae*, and (c) marine dinocysts (*Brigantedinium*). Density currents became progressively more dominant as fjord-head and side-entry deltas prograded. Due to rapid rates of emergence, marine conditions existed for less than 2000 years in the thesis area (from about 8-6 ka), and perhaps for only a few hundred years in the western part of the area nearest the ice margin. Proximal deltaic sediments were deposited from fjord-head and side-entry deltas as the middle and eastern parts of the study area were emerging (upper part of Unit B and parts of Unit C). The marine and deltaic sediments are overlain by either local aeolian sands of Unit C or fluvial deposits of Unit D. The fluvial system truncated and reworked part of the underlying deltaic and marine sediments. A glacial re-advance may have occurred after the western part of the area became emergent, depositing ice-contact glaciofluvial sediments of units C and D and possibly till (Unit E) over marine muds. Alternatively, some Unit E deposits may have been deposited by debris flows of glaciogenic sediments.

The main contributions of the thesis are: (1) description and interpretation of the post-glacial sedimentary record of the Kanairiktok Valley area, (2) a local chronology of deglaciation and sea level, (3) the first sedimentological, micropalaeontological and process link between marine studies (Labrador Shelf and modern bays) and the land record of Quaternary events in coastal Labrador.

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CHAPTER 1

INTRODUCTION

The Kanairiktok Valley area is located in east central Labrador and extends approximately 100 km inland. Unconsolidated sediments exposed in bluffs along the banks of the modern river were deposited during the Late Quaternary. Initially, sediment-laden glacial meltwater deposited fine grained sediments in a marine fjord-embayment when the area was isostatically depressed. Subsequent land emergence due to isostatic rebound caused the withdrawal of the sea, so that fluvial-deltaic sediments were deposited over the marine and glaciomarine fjord sediments.

The Late Quaternary sediments of the Kanairiktok Valley were briefly described by Batterson *et al.* (1988), who interpreted their formation to be a result of deposition during marine inundation of the area following ice retreat followed by fluvial deposition as the area isostatically emerged. These sediments are an important indicator of the post-glacial history of the area, ice-margin positions, and sea level history. This thesis provides the first detailed sedimentological study of Quaternary deposits in central Labrador. Results from this study, together with data from other studies elsewhere in Labrador, will lead to a better understanding of the deglaciation history of central Labrador.

1.1. THESIS OBJECTIVES

The objectives of this study are as follows: (1) to describe the post-glacial sedimentary units exposed along the Kanairiktok Valley in order to deduce sedimentary environments; (2) to use the fossil flora and fauna together with numerical dating techniques to provide insight into the age of the sediments as a guide to the history of deglaciation and sea level; (3) to evaluate the different models that have been proposed for the deglaciation of the area in the light of field and laboratory studies.

1.2. LOCATION

The part of the Kanairiktok River that was studied in this thesis encompasses the area covered by the National Topographic System map sheets 13K/15 (60°30'- 61°00' west longitude, and 54°45' - 55°00' north latitude), 13K/10 (60°30'- 61°00' west longitude, and 54°30' - 55°45' north latitude) and 13K/16W (60°15'- 60°30' west longitude and 54°45' - 55°00' north latitude). Three communities are located near the study area (Fig. 1.1): Makkovik, Hopedale and Postville.

1.3. PHYSIOGRAPHY

The physiography of the Canadian Shield in which the thesis area lies is a product of Precambrian peneplanation and Pleistocene glaciation (Shilts *et al.*, 1987). The Kanairiktok Valley area lies within the George Plateau Physiographic subregion of the Davis Physiographic Region (Bostock, 1970). This subregion is characterized by an undulating plateau that slopes towards Ungava Bay and the Whale Lowland (Bostock, 1970). The part of the Kanairiktok River Valley examined for this thesis is characterized by two contrasting physiographies: low lying areas and rugged mountainous areas (Lopoukhine *et al.*, 1977). The low lying areas are located near the course of the modern river and lakes; these areas are underlain mainly by deltaic deposits and glaciofluvial outwash terraces. There are small dune fields in the northeastern part of the area, small level bogs, and lowlands underlain by marine deposits in small restricted areas (Lopoukhine *et al.*, 1977). The size of this low lying area varies along the course of the Kanairiktok River: in the upstream part it is 1 - 1.5 km wide with local relief of 0 - 20 m (elevation ranges from 100 - 120 m); in the middle part it is 4 - 12 km wide with local relief of 0 - 40 m (elevation ranges from 60 - 100 m) except where isolated hills composed of igneous and metamorphic rocks reach elevations of approximately 200 m; in the downstream part it is 3 - 9 km wide with local relief of 20 - 80 m (elevation ranges from 40 - 120 m).

The rugged mountainous areas are composed of igneous and metamorphic rocks that form ridges. These ridges trend in a northeast-southwest direction and border the low lying area in the upstream and middle part of the study area. The ridges mainly range in elevation from 280 m to less than 440 m; the maximum elevation of 452 m is in the area east of Joe Pond in the upstream part of the study area.

In the downstream part of the study area, the low-lying ground on the eastern side of the river is narrow (1 km) and is bordered by rounded mountains (elevation 200 - 300 m with gentle

Fig. 1.1. Location map showing the study area, communities and inlets

slopes) underlain by bedrock and veneers of till and colluvium; glaciofluvial deposits are restricted here to cirque-like features at higher elevations (Lopoukhine *et al.*, 1977).

1.4. THE MODERN RIVER VALLEY

The Kanairiktok River originates northeast of the Smallwood Reservoir where small tributaries connect to form two main river branches flowing in an easterly direction. A branch flows into Shipiskan Lake and then both flow into Snegamook Lake. From Snegamook Lake the Kanairiktok River flows in an easterly-northeasterly direction for approximately 100 km until it reaches Kanairiktok Bay on the Atlantic Coast (Fig. 1.2). The modern river has a sinuous channel pattern. Its general northeast trend is probably controlled by bedrock structure (Klassen and Thompson, 1993).

A longitudinal profile of that part of the river with elevation less than 130 m is shown in Figure 1.3. The profile was constructed from topographic maps and field estimation of abrupt river level changes in areas of rapids and falls. The only delta observed in the area is at an elevation between 80 - 100 m above sea level (Fig. 1.3), about 45 km downstream from the eastern end of Snegamook Lake. Downstream from the delta, for a distance of approximately 50 km, the general gradient of the river is 125 cm/km.

Two major groups of waterfalls and rapids are found along the river. The first is approximately 13 km downstream from the delta, where the river drops approximately 15 m over a distance of 8 km. The area upstream from these rapids to the delta has an average gradient of 95 cm/km. The second falls and associated rapids are at approximately 50 km from the delta (they are beyond the limits of the study area). Here the river drops about 20 m in a distance of 5 km. The average gradient of the river between the two sets of rapids is 68 cm/km.

During its northeasterly flow, the river varies in width, depth and bank materials. The variation in the width and depth of the river is mainly a reflection of differences in the underlying bedrock. The river narrows in areas where bedrock crops out and widens away from these outcrops. The river also widens where its banks are composed of unconsolidated alluvial or marine deposits.

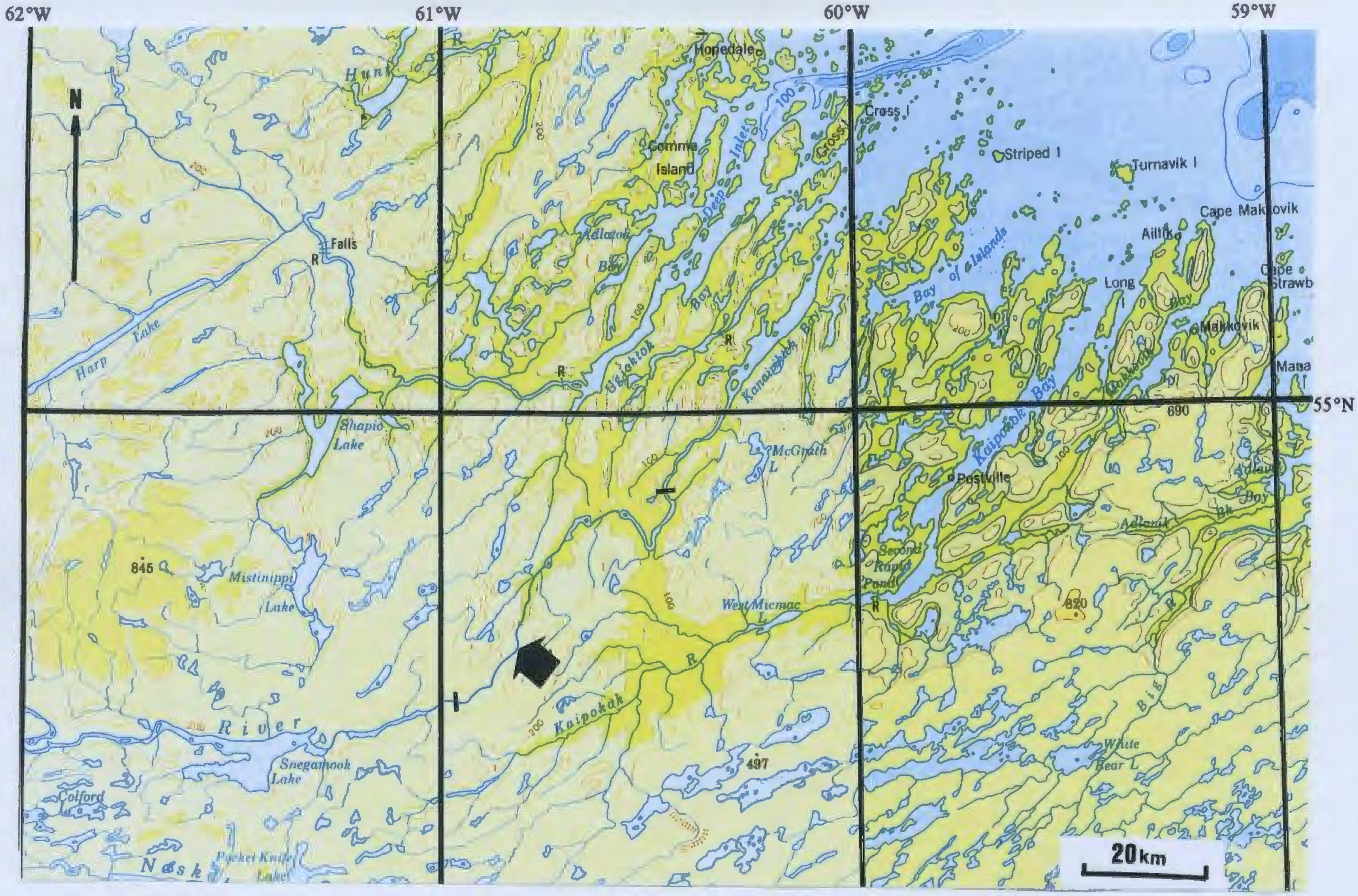


Fig. 1.2. Topographic map on which are noted the Kanairiktok River (large arrow), limits of the study area (between black bars on the modern river), Kanairiktok Bay and Lakes. Pale yellow > 100 m, deep yellow < 100 m. Water is blue. Reproduced from map sheets NN-20 and NN-21, copyright dates 1970 and 1973, respectively. Copyright granted by Her Majesty the Queen in Right of Canada with permission of Energy, Mines and Resources Canada.

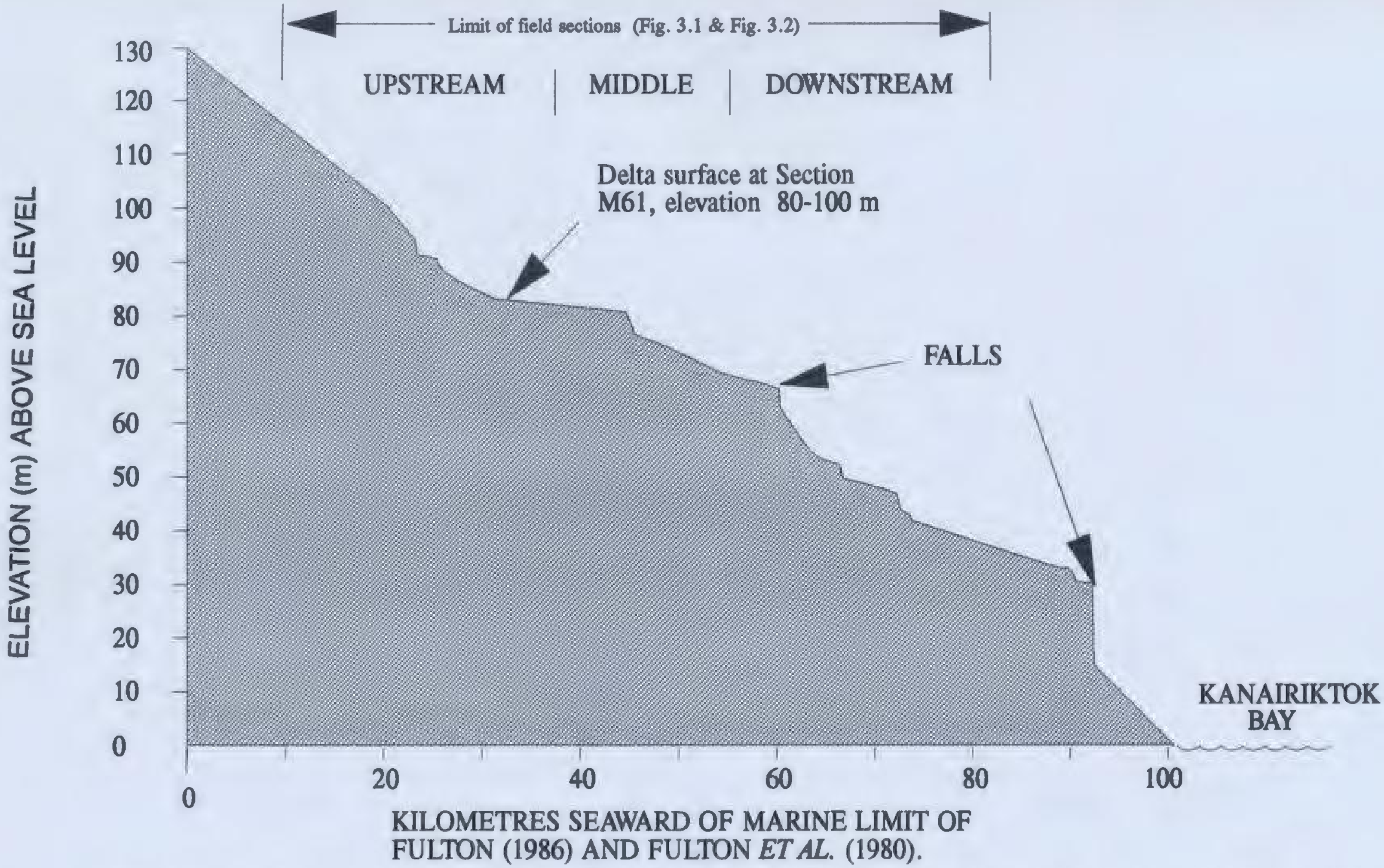


Fig. 1.3. Longitudinal profile of the Kanairiktok River. Also shown are locations of waterfalls and a delta surface at an elevation 80 - 100 m . Upstream, middle and downstream parts of the study area are marked. Vertical exaggeration is ca. 500 x.

The banks in the upstream area are mainly composed of sandy and gravelly alluvial and glaciofluvial deposits. Along the remainder of the river, the banks are composed of sandy, silty and muddy material of marine or glaciomarine origin, capped by gravelly and sandy units of variable thickness.

In this thesis, the area studied along the modern river is divided into three parts: the upstream, middle and downstream parts. The upstream part extends from approximately 20 km east of Snegamook Lake to Section U54 (approximately 29 km along the course of the river). The middle part extends between sections U54 and M88 (approximately 18 km along the course of the river). The downstream part extends from the area of Section M88 to Section D103 (approximately 27 km along the course of the river). Areas downstream of Section D103 were not studied.

1.5. VEGETATION

The study area lies entirely within the Northeastern Transition and Forest Tundra subregions of the Boreal Forest Region of Rowe (1972). Vegetation in most of the area is dominated by spruce and lichens, with white spruce (*Picea glacua*) and black spruce (*P. mariana*) being the most common vegetation on well drained glaciofluvial sands, tills and sand dunes (Lopoukhine *et al.*, 1977; Environment Canada, 1989). Poorly drained areas are dominated by black spruce, Labrador tea (*Ledum groenlandicum* and *L. palustre*), bog rosemary and cloudberry.

In the eastern part of the area, white spruce and moss are the most common flora; lichen is common in drier areas (Environment Canada, 1989). Exposed bedrock is covered by lichen and moss (Environment Canada, 1989).

1.6. CLIMATE

Most of the area falls within the Interior Labrador Climatic Zone of Banfield (1981). The eastern area near the coast is warmer due to the moderating effect of the Labrador Sea (Lopoukhine *et al.*, 1977). Mean annual temperature for inland areas is -2.5°C (Lopoukhine *et* *al.*, 1977), with four to five months of cool summers and very cold and snowy winters (Environment Canada, 1989). Areas near the coast have a mean annual temperature of -2.5°C to 0°C (Lopoukhine *et al.*, 1977), with warm summers that are approximately six months long.

Annual precipitation of 70 to 80 cm occurs mostly during the summer in inland areas; in coastal areas precipitation is distributed throughout the year (Lopoukhine *et al.*, 1977; Environment Canada, 1989). Annual snow accumulation inland is 300 to 400 cm. Ice-free periods in lakes range from 180 days near the coast to 150 days inland. Permafrost occurs in scattered areas over the whole region (Lopoukhine *et al.*, 1977).

1.7. MODERN TIDE AND WAVE REGIME ON ADJACENT COASTAL AREAS

Tidal measurements in Makkovik and Hopedale show that the tidal regime is semidiurnal to mixed (Rosen, 1979) and microtidal to mesotidal (*sensu* Davis, 1964), with mean and spring tidal ranges of 1.5 m and 2.23 m for Makkovik and 1.7 m and 2.8 m for Hopedale (Canadian Hydrographic Service, 1992).

Wave height data for the Makkovik Bay area (Rosen; 1979; Barrie, 1980) show that the maximum significant wave height seaward of Makkovik is 7.3 m with a 16 second period. The wave energy levels along the Makkovik Bay shoreline decrease from more than 6 m in the outer part of the bay, to 2 - 6 m in the middle part and less than 2 m in the inner part of the bay (Rosen, 1979).

1.8. BEDROCK GEOLOGY

The study area lies mainly in the Nain Structural Province (Taylor, 1971). The boundary between the Makkovik Subprovince and the Nain Structural Province bisects Kanairiktok Bay and trends northeast-southwest through the entire area. Basement rocks are of Archean and Aphebian age; younger basement rocks are present in the Kaipokok Valley area (Ryan, 1984).

Archean rocks in the area have been studied by Taylor (1971), Ermanovics and Raudsepp (1979), Ermanovics (1980) and Ryan (1984). They have been informally assigned to the Kanairiktok Valley Complex (Ryan, 1984) and consist of gneisses, schists, metavolcanics and granites cut by several diabase dikes. Adjacent to the Kanairiktok Valley in the Kaipokok Valley, similar gneisses were transformed by retrograde metamorphism into muscovite-rich schists of the Kaipokok Valley Complex (Ryan, 1984).

Aphebian supracrustal rocks rest unconformably on Archean rocks. These were deposited during a time interval close to the Kenoran deformation (ca. 2500 Ma [2500 million years before present]; Ryan, 1984). Other Aphebian rocks in the Kanairiktok River area include metamorphic, plutonic, metavolcanic and metasedimentary rocks that formed in an interval close to the Hudsonian deformation (ca. 1750 Ma; Ryan, 1984).

1.9. SURFICIAL SEDIMENTS

Fulton (1986) and Fulton *et al.* (1980) mapped the regional surficial geology, based mainly on interpretation of aerial photographs and a limited number of ground observations in the Kanairiktok Valley area. The surficial deposits in the western part of the study area were mapped in greater detail by Batterson (1991), again based mainly on interpretation of aerial photographs.

Fulton (1986) shows that Late Wisconsinan to Holocene moraines, proglacial and non-glacial surficial deposits overlie basement rocks over most of the area, except in the Kanairiktok Bay area where bedrock exposures dominate (Fig. 1.4). In detail, morainal deposits (tills, sands and gravels) of variable thickness (1 - 5 m) cover a wide area northwest and southeast of the river valley. These deposits form gently rolling surfaces or a thin till that mimics the underlying topography. Organic deposits cover some of these morainal deposits.

Glaciofluvial deposits are present as ridges, hummocks, terraces and plains on both sides of the modern river (Fulton, 1986; Batterson, 1991). These deposits overlie glaciomarine deposits of variable thickness. Small abandoned river channels cutting through the glaciofluvial deposits are present locally (Fulton, 1986). Alluvial deposits, up to 1 - 15 m thick and consisting of sand and gravel, are present along the middle and downstream segments of the modern river (Fulton, 1986).

According to Fulton (1986), marine, sublittoral, laminated silts and clays with minor sand

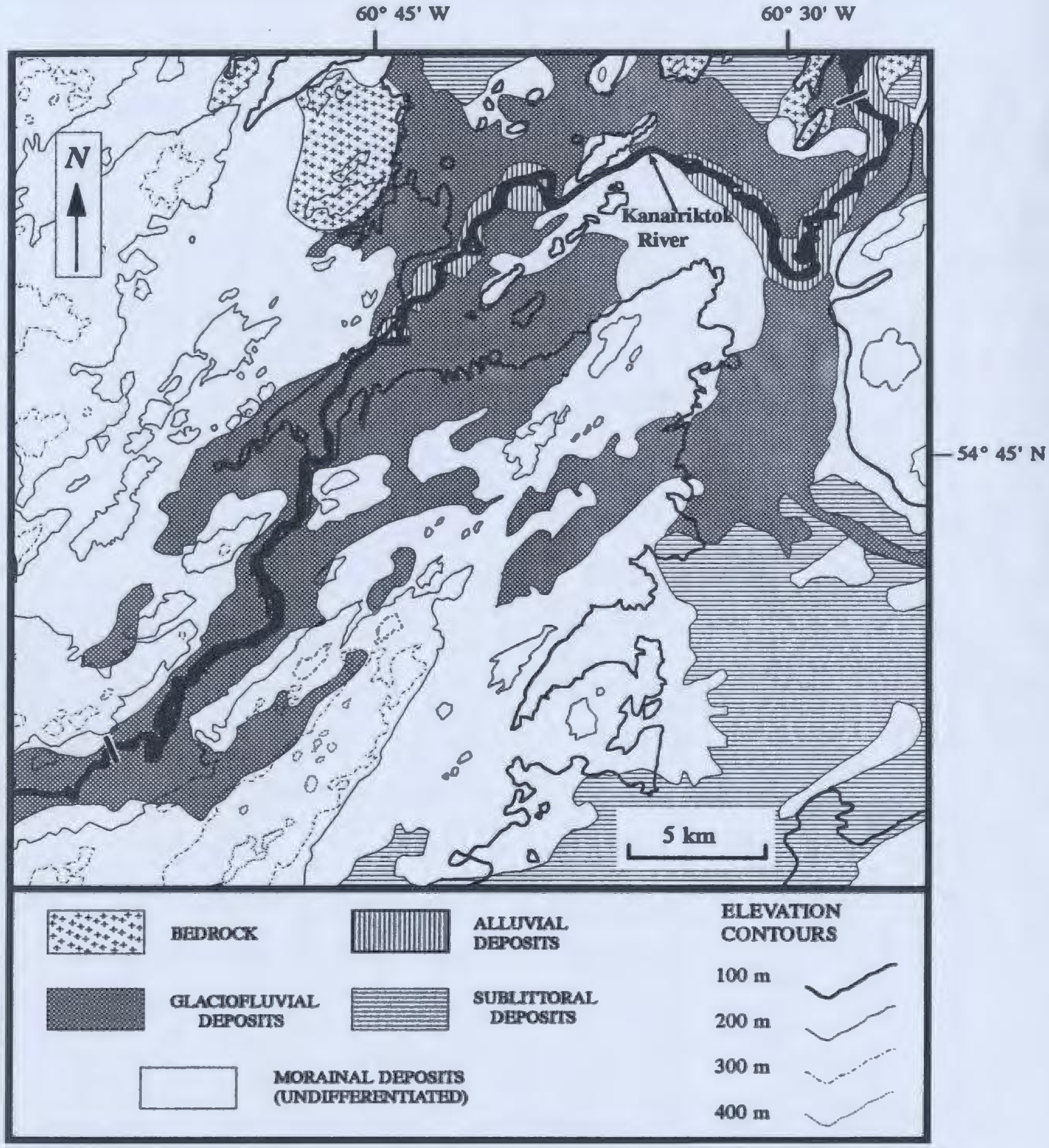


Fig. 1.4. Surficial geology of the Kanairiktok Valley and adjacent areas . The limits of the thesis area are shown by short black bars on the modern river. The river is shown in solid black. Simplified from Fulton (1986).

are present only in a small area along the downstream part of the river. There, these sediments are overlain by a layer of peat and bog deposits.

In the upstream part of the area, detailed mapping by Batterson (1991) shows that the narrow belt of low-lying areas bordering the modern river and small lakes is covered by glaciofluvial deposits that are generally greater than 1 m thick. These deposits mainly form terraces and plains, some of which have been eroded and dissected by meltwater or ice marginal channels. Bedrock dominates the upstream part of the area and borders the narrow low-lying sediments (Batterson, 1991). Thin (<2 m) till veneers are locally present over bedrock. These till veneers occur at an elevation of more than 100 m and are locally eroded.

1.10 FIELD LOGISTICS

The field work was carried out with the aid of field assistant Mr. Donald Stevens in the period June 12 to July 24, 1988. Access was by floatplane from Otter Creek in Goose Bay, 100 km north to a point approximately 10 km downstream from Snegamook Lake on the Kanairiktok River. Transport along the river was by canoe, and by portage around rapids and water falls.

Supervision in the field was provided by Mr. Martin J. Batterson and Dr David Proudfoot, Newfoundland Department of Mines and Energy, who visited during the period June 28-29, 1988. Most of the sections visited were along the river banks. The few traverses attempted away from the river were of little value because of thick vegetation.

Newfoundland Department of Mines and Energy provided some field equipment from their base camp in Goose Bay and transported the field equipment back to Goose Bay at the end of the field season. Helicopter transport for the field party from the Kanairiktok River to Postville at the end of the field season was provided by Sealand Helicopters as part of their scholarship program in support of Memorial University student field projects.

A marine seismic survey and shallow coring were planned for Kanairiktok Bay during the period July 30 to August 1, 1989, using the CSS Baffin. Unfortunately, this survey was cancelled due to a lack of detailed bathymetry in the bay. A second short field trip was planned for the summer of 1989 to complete the survey of the downstream part of the area to the head of Kanairiktok Bay, and to recheck some field relationships. This second field season was cancelled due to a lack of the funding required to reach this remote area.

1.11. THESIS OUTLINE

Chapter 2 consists of two parts. The first part is a literature review of the glaciation history of the area and the different models suggested for the relative sea level history of the area. The second part is a critical evaluation of the sea level history of the area; conclusions from this evaluation are used in the other parts of the thesis.

Chapter 3 is the largest chapter of the thesis. The different sedimentary units observed in the area are described from representative sections in different parts of the area. The sedimentary history where these sections occur is based on the interpretation of these units and their vertical relationships.

Chapter 4 consists of a description of the palaeontological data collected in the area. These data, although very limited, are used in Chapter 5 with the results of Chapter 3 to interpret the Late Quaternary history of the area. In addition, Chapter 5 summarises the results of the thesis and discusses the relationships between this study and other studies carried out in adjacent areas.

CHAPTER 2

HISTORY OF DEGLACIATION AND RELATIVE SEA LEVEL

Relative sea level variation is a major control on the sedimentary history of the area and must be critically evaluated in order to provide a realistic framework for subsequent discussion of the development of sedimentary successions in different parts of the thesis area (Chapter 3). Sea level is intimately tied to glacial history and the regional chronology provided, in part, by palaeontological studies. First, previous work on these topics is reviewed. Then, the relative sea level history of the study area is critically examined. This latter discussion is intentionally separated from the documentation of previous work because personal conclusions of the author are presented and some published ideas are seriously questioned.

2.1. PREVIOUS STUDIES

Over the last century, several models have been proposed for the Late Wisconsinan glacial history of the Laurentide Ice Sheet, but none of these models has gained general acceptance (Dyke *et al.*, 1989). The parameters that have been debated are the configuration, limits, thickness, and deglaciation chronology of the ice sheet. The Quaternary history of the southeastern part of the Canadian Shield, in which the study area lies, has been recently reviewed by Vincent (1989). The Quaternary history and remaining controversies are reviewed below.

2.1.1. LATE WISCONSINAN GLACIATION AND ICE MARGIN POSITION

The inferred configuration of the Late Wisconsinan Laurentide ice sheet is complex, consisting of several domes (Dyke *et al.*, 1982, 1989; Andrews, 1987a; Dyke and Prest, 1987a, b). The Labrador Sector of the Laurentide Ice Sheet covered most of Labrador and extended eastward onto the modern continental shelf, northward to Hudson Strait and the Hudson Bay area, westward beyond Ontario, and southward beyond the Great Lakes and into the northern part of the United States (Fulton, 1989). It is believed that most of central and southeastern Labrador, except the summits of the Mealy Mountains, was covered by ice during the last (Wisconsinan) glaciation (Vincent, 1989).

There are conflicting views on the exact position of the limits of the ice sheet in the southeastern part of Labrador (Vincent, 1989). Based on the different nature of glacial deposits east and west of the Paradise Moraine, Fulton and Hodgson (1979) placed the Late Wisconsinan (ca. 18 ka [18,000 years before present]) ice limit at this moraine and inferred that the area east of the moraine was ice-free. This view has been supported by others (e.g., Rogerson, 1981, 1982). Evidence for an ice-free area in southeastern Labrador was provided by radiocarbon dates for organic matter in marine cores from the Cartwright Saddle (Vilks and Mudie, 1978); pollen and spores in the dated Late Wisconsinan interval suggest an adjacent ice-free land. The reliability of these radiocarbon dates has been questioned by King (1985), who argued that the small amount of dated organic material could have been contaminated by older, reworked organic material. Instead, King (1985) placed the Late Wisconsinan ice limit at the Brador Moraine approximately 100 km southeast of the Paradise Moraine. This ice limit is based on a set of consistent radiocarbon dates from the area east of the Paradise Moraine (King, 1985). Vincent (1989) placed the ice margin at Belles Amours Moraine at 12 ka, whereas King (1985) suggested that the ice margin had not retreated to that position until 11 ka.

The extent of the Late Wisconsinan Ice Sheet offshore of the present coastline of central Labrador is controversial. Andrews (1987a) and Dyke and Prest (1987a,b) speculated that a floating ice shelf extended well offshore, whereas Josenhans *et al.* (1986) suggested that, although ice extended well offshore of the present coastline, it was grounded on the inner shelf. Grounded ice was present in Kaipokok Bay (Kontopoulos and Piper, 1982) and Makkovik Bay (Barrie and Piper, 1982) during the Late Wisconsinan glacial maximum.

Rogerson (1981, 1982) placed the glacial ice limit offshore of the modern coastline, and Vincent (1989) concurred but believed that the ice limit was even further offshore than advocated by Rogerson (1981, 1982) by 12 ka. These authors concluded that most of Labrador was ice covered except for the southeastern part, which may have been ice free (Vincent, 1989).

2.1.2. DEGLACIATION

The retreat of the ice sheet from its maximum limit was not synchronous along the ice margin. Retreat of the western and southern margins started as early as 18 ka (Andrews, 1987a; Dyke and Prest, 1987a), whereas the eastern margin remained stable until about 12 ka (Fulton and Hodgson, 1979; Prest, 1984; King, 1985; Andrews, 1987a). The large inferred ice shelf off Labrador probably disintegrated by 14 ka, so that only grounded ice remained beyond the coast on the inner shelf (Dyke and Prest, 1987a).

The chronology of the ice retreat on land and the age of the main moraine systems in southeastern and central Labrador are still subject to debate (Fulton and Hodgson, 1979; Rogerson, 1981, 1982; King, 1985; Dyke and Prest, 1987b; Vincent, 1989) (Fig. 2.1). At 12 ka, only the area east of the Belles Amours Moraine was deglaciated (King, 1985). The area seaward of the modern coastline to the east of the study area was still ice covered (King, 1985; Vincent, 1989). At 11 ka, the area east of Hamilton Inlet became ice free (although Rogerson, 1981 suggested that Sandwich Bay was deglaciated as late as 8 ka). The 11 ka ice margin was at the Paradise Moraine in southern Labrador according to Dyke and Prest (1987b) and Vincent (1989) but was only at the Belles Amours Moraine according to King (1985). King (1985) placed the ice margin at the Paradise Moraine at 10 ka.

At 10.5 ka, the eastern parts of Lake Melville and Makkovik Bay were ice free (Fulton and Hodgson, 1979; Barrie and Piper, 1982; King, 1985). At 10 ka, the ice margin was at the Sebaskachu and Little Drunken moraine segments of the Quebec North Shore Moraine (Dyke and Prest, 1987b; Vincent, 1989).

King (1985) suggested that the Little Drunken moraine segment was formed at 9.2-7.5 ka as a result of change in the ice flow pattern from a southeasterly flow to a northeasterly flow. This change in flow affected only the area south of Lake Melville and occurred because of extensive ice drawdown in Lake Melville (King, 1985).

At 9 ka, the ice margin had retreated to the western part of Lake Melville and to the west of Snegamook Lake in the Kanairiktok Valley area (Dyke and Prest, 1987b; Vincent, 1989). This ice margin position differs from that shown by King (1985), who placed the 9.0 ka ice margin at the middle of Lake Melville and extended it northward to a position near the head of Kanairiktok and Kaipokok bays. King (1985) concluded that the Sebaskachu moraine segment was

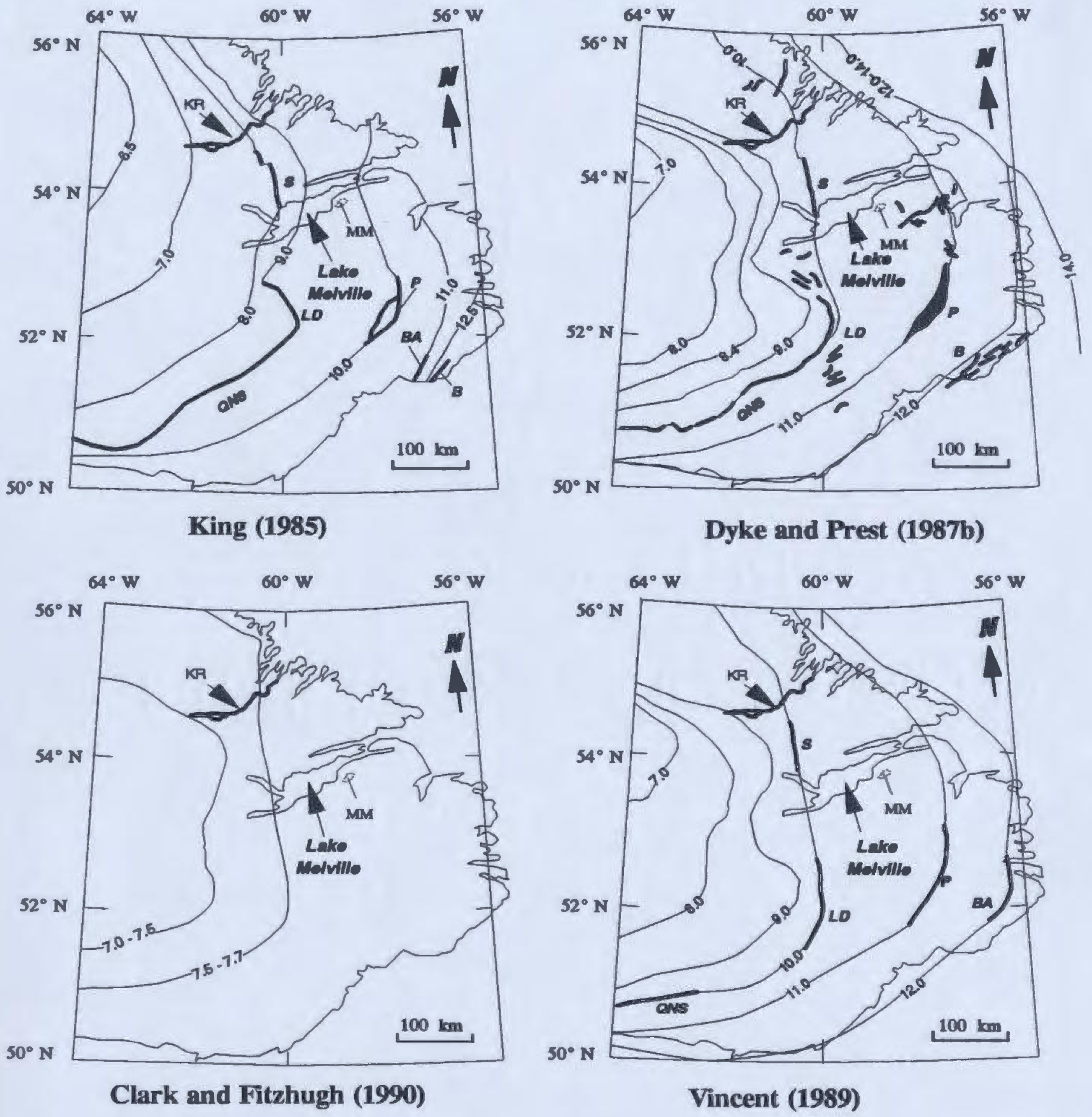


Fig. 2.1. Palaeogeographic maps of southeastern Labrador showing the ice cover at different time periods as suggested by several authors. The locations of major end moraines are shown as they were placed in the original publications (B= Brador, BA= Belles Amours; QNS=Quebec North Shore; LD= Little Drunken; S=Sebaskachu). The Kanairiktok River (KR) and Mealy Mountains (MM) are also shown.

formed at about 7.6 ka. He placed the 8.0 ka ice margin at this moraine and argued that the Kanairiktok Valley area east of Snegamook Lake became ice free after 7 ka.

Recently, Clark and Fitzhugh (1990) claimed that at 7.5-7.7 ka (a) the ice margin was situated at the head Lake Melville, and (b) the area of the Kanairiktok Valley was still ice covered. They inferred that the Kanairiktok Valley area was deglaciated at 7.0-7.5 ka, by which time the ice margin was west of the Kanairiktok and Goose rivers (Clark and Fitzhugh, 1990). Dyke and Prest (1987b) placed the ice margin at the same positions suggested by Clark and Fitzhugh (1990) much earlier, at 8.4 ka.

The final disintegration of the Laurentide Ice Sheet was near Schefferville in western Labrador (King, 1985; Dyke and Prest, 1987a,b; Vincent, 1989). The time of final disappearance of glacial ice from that area is not well known (King, 1985) but was probably complete by 6.5 ka (Vincent, 1989).

The ice flow near the study area during deglaciation was controlled by the Mealy Mountains and Lake Melville (Vincent, 1989). The flow was in a northeast direction in central Labrador and in an eastward direction in the area of Lake Melville (Vincent, 1989). In the study area, based on the distribution of glacial erratics and orientation of striae, Klassen (1983, 1984) and Thompson and Klassen (1986) concluded that the ice came from the west and northwest and flowed in a northeasterly direction parallel to the long axes of the major ridges in the area (Fig. 1.2). The ice descended into low lying areas now occupied by the Kanairiktok and Kaipokok rivers, and then flowed northeastward to the sea. Batterson *et al.* (1988) confirmed this northeasterly direction of flow.

2.1.3. POST-GLACIAL EMERGENCE AND RELATIVE SEA LEVEL CHANGES

Terminology used in this thesis is as follows. Eustatic sea level changes are defined by Mitchum (1977) as "worldwide changes of sea level that affect all the oceans (Gary *et al.*, 1972); or a relative change of sea level on a global scale, produced either by a change in the volume of sea water, or a change in the surface area of the ocean basins, or both (Fairbridge, 1961)". Relative sea level change is defined by Mitchum (1977) as "an apparent rise and fall of sea level with respect to the land surface. Either sea level itself, or the land surface, or both in combination may rise or fall during a relative change. Generally speaking, a relative change may be operative on a local, regional, or global scale". Marine limit is defined by Andrews (1974) as "the maximum elevation which the late- or post-glacial sea reached at a coastal site. The elevation is measured with respect to present sea level". An isobase is defined by Andrews (1974) as "a line joining points of equal postglacial uplift (glacial lake shorelines), where these elevations are the result of postglacial emergence or uplift operating over the same length of time. By convention, time is measured from the present". A relative sea level curve is defined as " a curve that reflects the combined interaction of glacial rebound and the global rise in sea level associated with melting ice sheets" (Andrews, 1989; p. 553)).

The field area lies in what has been called Zone I of relative sea level change (Clark *et al.*, 1978; Andrews, 1987b; Devoy, 1987), which is characterized by progressive land emergence with the maximum emergence in areas nearest the former ice load. The manner in which the crust of the Canadian Shield responded to glacial unloading in Zone I had several important consequences for the sea level changes in that zone: (1) deglaciated areas rebounded faster than areas still under ice cover, so that the land surface became tilted toward the ice centres (Clark *et al.*, 1978), (2) the rate of emergence decreased with time, such that the isostatic rebound in the first ca. 1000 years probably accounted for up to 30% of the total rebound (Clark *et al.*, 1978), and (3) the total amount of rebound has been greater than the amount of global sea level rise due to melting of the ice sheets (Clark *et al.*, 1978).

Tilt of originally horizontal features (e.g., glacial lake shorelines) in central Labrador is now towards the east-northeast (Barnett and Peterson, 1964; Andrews, 1989). This suggests that the ice centre was near Schefferville (Barnett and Peterson, 1964). Fitzhugh (1973) suggested that the deflection of the crust was about 0.4 - 0.6 m/km towards the interior of Labrador.

During the glacial maximum, sea level was at its eustatic minimum (Dyke and Prest, 1987). Relative sea level along the Labrador coast at that time is not known because ice probably extended beyond the present coastline (Dyke and Prest, 1987a; Andrews, 1987a). Quinlan (1981) showed that the zero line for relative sea level was east of the ice front, well seaward of the present coastline.

As the ice retreated, isostatically depressed areas were inundated with marine waters (Fig. 2.2). The estimates for the marine limit in western Lake Melville (all elevations are relative to the modern sea level datum) range from 37 m at 5.3 ka (Fitzhugh, 1973; Quinlan, 1981) to 61 m (Andrews, 1989) to 135 - 147 m (Vincent, 1989; Clark and Fitzhugh, 1991). The estimates for the marine limit for the outer coast of Hamilton Inlet range from 18 - 15 m at 5.3 ka (Fitzhugh, 1973; Quinlan, 1981) to 85 - 76 m (Vincent, 1989) to 75 m at 9.7 ka at Goose Bay (Clark and Fitzhugh, 1991). The difference in the amount of emergence between the Hamilton Inlet area and Lake Melville has been attributed to the persistence of a stagnant ice sheet to the west (Andrews and Tyler, 1977; Quinlan, 1981; Peltier and Andrews, 1983).

North of the thesis field area, at Hopedale, and east, at Makkovik, the marine limit has been estimated to be 119 m and 107 m, respectively (Clark and Fitzhugh, 1991). These marine limits are dated at 7.6 ka and 9.0 ka, respectively. These dates were estimated by extrapolation of ^{14}C-constrained emergence curves (Clark and Fitzhugh, 1991).

In the thesis field area, the marine limit for the entire area mapped by Fulton (1986) and Fulton *et al.* (1980) ranges from 80 - 135 m. Batterson *et al.* (1988) recognized two delta levels in the Kaipokok Valley at 125 and 105 m above present sea level. A radiocarbon date of 7690 ± 60 yr B.P. (TO-1123) was obtained from the delta bottomset deposits in the same area. This may be the age of the higher marine limit (M.J. Batterson personal communication, 1990).

Of the relative sea level curves that have been published, those for the Lake Melville-Goose Bay area (Quinlan, 1981; Andrews, 1987a; 1989; Clark and Fitzhugh, 1991), Makkovik (Clark and Fitzhugh, 1991) and Hopedale (Clark and Fitzhugh, 1990, 1991) are important for this study. These curves, with or without corroborating field data, have previously been used to construct isobase maps showing relative changes in sea level in Labrador for the period 8 - 7 ka to the present (Andrews and Tyler, 1977; Andrews, 1989; Clark and Fitzhugh, 1991).

The maps of Andrews (1989) and Andrews and Tyler (1977) show isobases parallel to the Labrador coastline; their 60 m isobase for emergence since 8 ka (Andrews and Tyler, 1977) or 7 ka (Andrews, 1989) is in the middle of the area. These isobases and emergence models

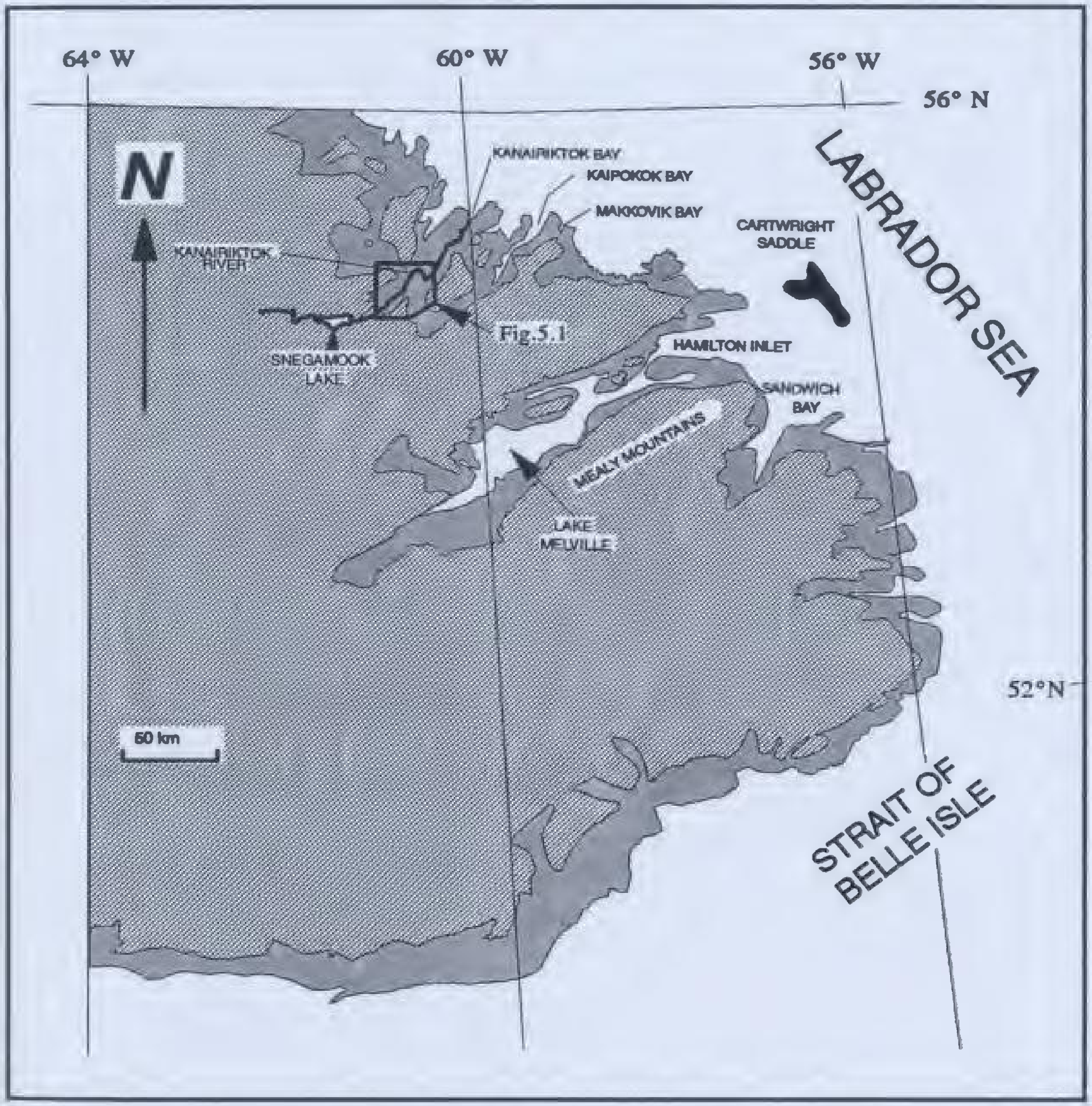


Fig. 2.2. Map of part of the southeastern Canadian Shield showing the areas inundated by marine waters during the last deglaciation (dark stippled area). Simplified from Vincent (1989). Cartwright Saddle on the Labrador shelf is shown in black. The marine inundation of the area in the small box is shown in more detail in Chapter 5.

suggest a rate of emergence of 5-10 m/1000 yr near the present coastline (Andrews and Tyler, 1977; Andrews, 1989) to 20 m/1000 yr inland (Andrews and Tyler, 1977) from 8 - 6 ka. The rate of emergence decreased to 5 m/1000 yr in coastal areas and 15 - 20 m/1000 yr in inland areas west of the study area for the period 4 - 3 ka. The rate of emergence decreased to 1 - 5 m/1000 yr for the period 1 ka to present.

The isobase maps of Clark and Fitzhugh (1991) also show isobases roughly parallel to the Labrador coast, but the maps suggest that the area of Hopedale has experienced a greater amount of emergence than areas to either the south or north (the difference in emergence between Hopedale and Makkovik since 7.0 ka is estimated at 50 m). The greater amount of emergence in the Hopedale area was attributed by Clark and Fitzhugh (1991) to the inferred presence there of thicker ice (Clark and Fitzhugh, 1991; P.U. Clark, personal communication, 1993).

According to Clark and Fitzhugh (1991), the rate of emergence was 50 m/1000 yr in the middle part of the thesis study area and 20 m/1000 yr near the coast from 7 - 6 ka. The inferred rate of emergence decreased to 6 m/1000 yr in the middle part of the study area and 3 m/1000 yr near the coast for the period from 4 - 3 ka.

2.1.4. ESTIMATION OF PALAEOCLIMATIC CONDITIONS

Palaeoclimatic conditions in the area are based on palaeontological studies. The palaeontological studies relevant to the thesis are based on microfossils collected from cores from the Labrador Shelf and Lake Melville (Vilks and Mudie, 1978, 1983; Mudie, 1982; Vilks *et al.*, 1984, 1987) and palynological studies on nearby lake bottom sediments (Jordan, 1975; Lamb, 1980, 1985; Macpherson, 1981; Fitzhugh and Lamb, 1985). Both the data from land and marine cores reflect the change in climate and water circulation following deglaciation. The palaeontological data have been recently summarised by Piper *et al.* (1990). The data from Lake Melville and nearby lakes are most relevant to this study.

Pollen and spore data reflect the change in vegetation because of climatic change after deglaciation. Four palynozones are recognized by Piper *et al.* (1990) in southeastern Labrador. The oldest zone (CS1; ca. 20 ka(?) to ca. 9 ka) is dominated by *Sphagnum* and *Betula* species reflecting an arctic tundra vegetation that was established over the recently deglaciated area. This tundra vegetation reflects a low arctic climate (Piper *et al.*, 1990). It is followed by a shrub-tundra vegetation zone (CS2; ca. 9 ka to ca. 7 ka) dominated by *Betula* and *Alnus* species reflecting a cold climate (Vilks and Mudie, 1983). The third zone (CS3; ca. 7 ka to ca. 2.6 ka) is dominated by *Picea* species reflecting a spruce forest vegetation indicating cool and wet climate conditions (Vilks and Mudie, 1983; Piper *et al.*, 1990). The last zone (CS4; ca. 2.6 ka to the present) is characterized by *Picea, Betula, Alnus* and *Abies* species indicating a boreal forest that was established because of the cooling of the climate relative to zone (CS3) (Vilks and Mudie, 1983; Piper *et al.*, 1990). The age of each zone decreases from coastal areas to inland areas reflecting the earlier colonization of vegetation in coastal areas (Jordan, 1975).

Foraminifera and dinoflagellate data are good indicators of depositional environments (Piper *et al.*, 1990). In Lake Melville, Vilks *et al.* (1987) observed four benthic foraminifera zones, the lower zone (Zone D) is barren of foraminifera. The overlying zone (Zone C) is dominated by the species *Elphidium excavatum* f. *clavata*, overlain by a middle zone (Zone B) dominated by the species *Islandiella helenae*, *Buccella frigida* and *Protelphidium orbiculare*. The upper zone (Zone A) is dominated by agglutinated species. Vilks *et al.* (1987) concluded that the transitions between these three zones are a function of changes in palaeosalinities, which are in turn due to fluctuations in fluvial discharge in coastal subarctic inlets. For areas where these four zones have been recognized, the lower zone has an early Holocene age, and occurred where meltwater was entering and filling the basin. Zone C corresponds to early deglaciation with glacial meltwater diluting the marine waters to an estimated salinity range of 31‰ to 32‰. Continuous reductions in the amount of fresh water input as glaciers retreated at the end of deglaciation resulted in establishment of Zone B, with an estimated salinity range of 32‰ to 34‰. Zone A corresponds to present oceanographic conditions in the Labrador Sea (Vilks *et al.*, 1987).

Four dinoflagellate zones have been recognized in Lake Melville (Vilks and Mudie, 1983). In the youngest Zone 1, fresh water and estuarine dinocysts (*Peridinium* species) are dominant with a mixture of arctic and temperate marine species reflecting a fluvial dominated surface water layer (Vilks and Mudie, 1983). In Zone 2, arctic marine dinocyst species, dominated by *Multispinula minuta*, are dominant. Some arctic and subarctic species of low diversity reflect the influx of low salinity arctic shelf water into Lake Melville from ca. 4.0-6.5 ka (Vilks and Mudie, 1983). Zone 3 has less estuarine freshwater dinocysts that the overlying Zone 2, reflecting a decrease in the influx of fresh water into Lake Melville (Vilks and Mudie, 1983). The oldest Zone 4 is characterized by a high dinocyst concentration dominated by *Operculodinium centrocarpum* and *Brigantedinium* species indicating an influx of warm deep water from the Labrador Shelf to Lake Melville during the early Holocene. This assemblage is similar to modern dinoflagellate assemblages found in modern sediments in eastern Baffin Bay below the West Greenland Current (Vilks and Mudie, 1983). These four dinoflagellate zones were reclassified into three zones by Piper *et al.* (1990). The age relationships between the different microfossils zones are shown by Piper *et al.*(1990; their Table 10.4).

2.2. CRITICAL EVALUATION OF RELATIVE SEA LEVEL HISTORY

Relative sea level history of the area is not well known due to a lack of local detailed studies (Andrews, 1989), and scarcity or absence of dateable geological material in raised marine deposits (Clark and Fitzhugh, 1991; p. 198). Therefore, the relative sea level history of the area must be based on (a) physical and mathematical models, with assumptions, for relative sea level changes in deglaciated areas (Clark *et al.*, 1978; Clark, 1980; Quinlan, 1981; Peltier and Andrews, 1983; Peltier, 1989; W.R. Peltier, personal communications, 1993), (b) regional studies (Andrews, 1989), and (c) studies in areas adjacent to the thesis area (Fitzhugh, 1972, 1973; Clark and Fitzhugh, 1990, 1991).

In glaciated areas such as Labrador, changes in relative sea level are attributed to the net effect, through time, of isostatic depression under ice loads, isostatic rebound after ice melting, and changes in the total volume of sea water. Ocean water volume in the Quaternary was controlled mainly by the volume of polar ice caps (Clark, 1980).

Changes in relative sea level are examined from three perspectives: (1) the amount of crustal deflection due to ice loading; (2) emergence curves based on dated marine material, marine limit estimates and isobase maps constructed by others (Fitzhugh, 1972, 1973; Andrews, 1989, Clark and Fitzhugh, 1990,1991; W.R. Peltier, personal communications, 1993); and (3) the amount of postglacial emergence and its variation across the area.

2.2.1. CRUSTAL DEFLECTION

As a first approximation, the Earth can be viewed as consisting of a thin elastic crust overlying a viscous mantle (Andrews, 1974); in reality, the structure of the Earth is more complex. This approximation is used to explain the effect of glacial loading and unloading. An elastic substance will deform instantaneously when a load is applied and will return to its original form and shape when the load is removed. A viscous material will flow when stress is applied; the rate of flow depends on the amount of stress applied, and when the stress is removed the material does not return to its original form (Andrews, 1974). The viscous mantle of the Earth shows a delayed response to crustal unloading (e.g., England, 1983; Syvitski and Prage, 1989). The nature and duration of this delay is complex and depends on many factors such as the thickness of the ice in adjacent areas, viscosity of the mantle, rate of ice melting and thinning, nature and type of crust beneath the load and many other factors (J.T. Andrews, personal communications, 1993; G. Quinlan, personal communications, 1993).

According to Walcott (1970) the effect of ice loading on deflection of the crust can be expressed as four general rules: (1) the land is depressed most near the ice centre and less away from it; (2) beyond the ice margin the depressed region rises to an equilibrium point (i.e., point of zero deflection) at about 100 - 300 km from the ice margin; (3) beyond the equilibrium point there is a peripheral bulge; (4) changing large ice and water loads may affect the planet's shape (i.e., the geoid) and result in crustal deflection far from the ice loads. Fitzhugh (1973) suggested that the difference in the crustal depression between coastal Labrador and the Labrador interior was approximately 120 metres.

Tilt of dated glacial lake shorelines can be used to determine the position and relative thickness of ice centres. Andrews and Barnett (1972) and Andrews (1989) concluded that the area north of the thesis area was deflected towards 238° at a gradient of 0.44 m/km for inland areas at 8.8 ka (area 12 in Fig. 8.5 and Table 8.2 of Andrews, 1989), and deflected towards 235° at a gradient of 1.2 m/km for coastal areas at 10 ka (area 9, op. cit.). The shoreline tilt values are summarised in Figure 2.3. It is assumed here that the amount of 8-9 ka deflection in the study area was between these two values. For the purpose of discussion, the 9 ka southwestward deflection in the study area is estimated at 0.4-0.6 m/km, slightly higher than the essentially contemporaneous 8.8 ka deflection at area 12 of Andrews (1989) and substantially less than the older 10 ka coastal deflection of 1.2 m/km. The inferred deflection of 0.4-0.6 m/km is similar to the deflection suggested for the Lake Melville-Hamilton Inlet area by Fitzhugh (1973). The greater tilt of the 10 ka features is probably due to the greater deflection of the crust at 10 ka which is probably a function of greater ice thickness in the vicinity of those features at 10 ka (Walcott, 1970; equation on page 723). The ice sheet may have been thinner at 9 ka leading to less deflection near the 9 ka ice margin.

The thesis area extends approximately 50 km in the direction of the deflection, so that the upstream part of the area may have been depressed a few tens of metres (~20-30 m) more than the downstream part of the area (based on the suggested 0.4-0.6 m/km deflection). Thus, if the marine limit in the upstream region is suggested to be 130 m (an average of the maximum marine limits suggested by Fulton (1986) and Fulton *et al.* (1980), and observed by Batterson *et al.* (1988) in the Kaipokok Valley area) then in the downstream part of the area the contemporaneous sea level would have been at the modern 100-110 elevation contour.

2.2.2. MARINE LIMITS AND EMERGENCE CURVES

Relative sea level history and emergence curves have been based on estimates of the age and elevation of marine limits observed in the study area and adjacent areas (Fitzhugh, 1972,1973; Andrews, 1989; Clark and Fitzhugh, 1990, 1991) and glacial isostatic modelling (Peltier and Andrews, 1983; Peltier, 1989; W.R. Peltier, personal communication, 1993). Marine limit estimates are based on the elevation of dated raised shorelines and deltas (Andrews, 1989). Maps showing marine limit isolines for Labrador are shown in Andr[illegible] (1989). Isolines are approximately parallel to the modern Labrador coastline, suggesting that [illegible] e centre was to the

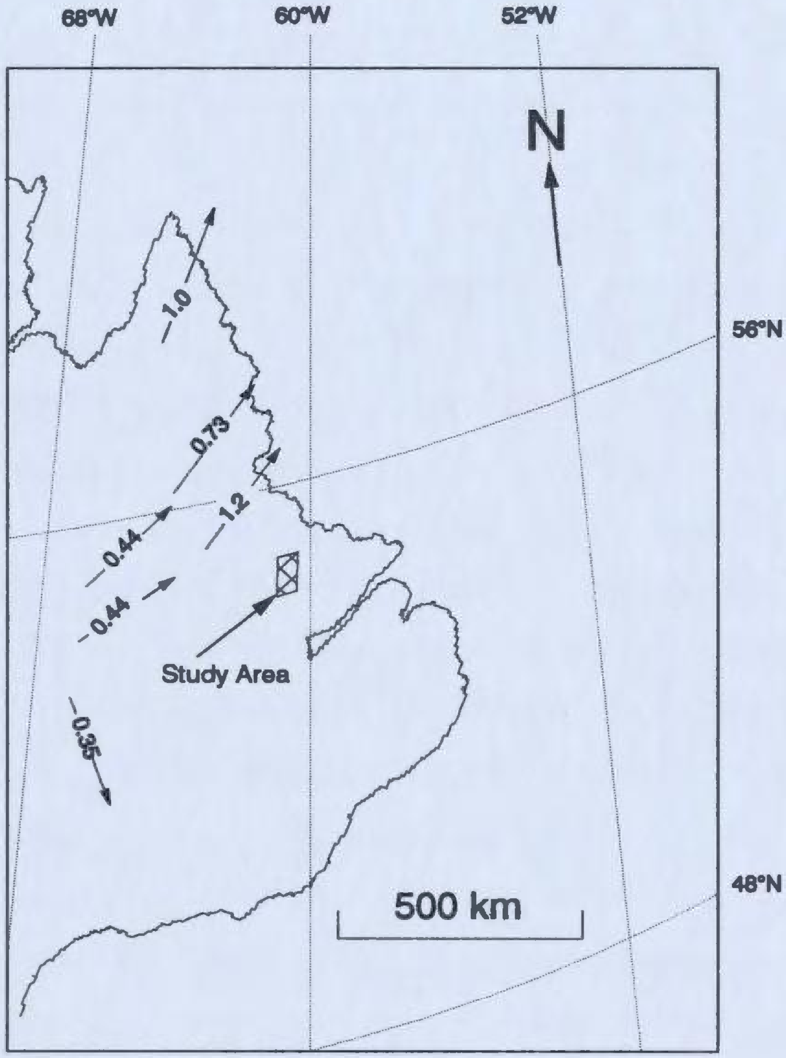


Fig. 2.3. Map showing the direction and amount of tilting of early Holocene glacial lake and raised marine shorelines (Modified from Andrews, 1989). Units are metres per kilometre. The study area is also shown.

west in the interior of Labrador.

Batterson *et al.* (1988) observed two delta levels east of Moran Lake (located approximately 11 km south of the upstream part of the study area) at 125 m and 105 m, and suggested that the higher delta elevation represents the maximum marine limit in the area and the lower level represents a still-stand related to deglaciation. A third delta, at 115 m elevation, was observed between the Kanairiktok Valley and the Kaipokok Valley at a site 2 km to the south of the upstream end of the middle part of the study area. Fulton (1986) and Fulton *et al.* (1980) suggested a maximum marine limit of 135 m for the Kanairiktok Valley area. In the Makkovik Bay area, Barrie (1980) estimated the 10 ka marine limit to be at an elevation of approximately 50 m, based on delta surfaces west of the head of Makkovik Bay.

Clark and Fitzhugh (1990, 1991) suggested a relative sea level history for coastal Labrador that differs in important ways from the ideas of other authors. Of great relevance to this thesis are their marine limit estimates for Makkovik and Hopedale. Clark and Fitzhugh (1990) estimated the marine limit in Hopedale to be 119 m, based on beach features observed by Daly (1902) and personal observations of washing limits by W.W. Fitzhugh (Clark and Fitzhugh, 1990, page 301). Clark and Fitzhugh (1990, their Fig. 3a.) constructed an emergence curve for the area of Hopedale based on dates from charcoal samples that range in age from 425 $\pm$65 yr B.P. (SI-5829) to 5129$\pm$95 yr B.P. (SI-1796). The charcoal samples were obtained from altitudes ranging from 5 to 27.6 m and suggest an exponential decrease in relative sea level. The curve, derived from an exponential model for postglacial emergence, was extrapolated to the otherwise undated 119 m marine limit estimated for Hopedale. Clark and Fitzhugh (1991; p. 203) also estimated the marine limit at Makkovik to be 107 m at 9.7 ka. This estimate is 57 m greater than the estimate of Barrie (1980).

Based on the emergence curve from Hopedale and a curve from Nain, Clark and Fitzhugh (1990) produced two palaeogeographic maps (their Fig. 4, and Fig. 2.1 in this thesis). The first map shows the ice margin east of the thesis area at 7.5-7.7 ka (implying that the study area was under glacial ice). The second map shows the ice margin west of the thesis area at 7.0-7.5 ka, but with a different shape and age than suggested by others (e.g., Fig. 2.1). These maps are incompatible with radiocarbon dates obtained from marine shells during the course of thesis research and with dates from shells collected in the Kaipokok Valley area (Awadallah and Batterson, 1990). These newly dated shells were collected from ca. 8.0 ka marine muds that lack any ice rafted component, suggesting that glaciers terminated landward of the coastline by 8.0 ka. Therefore, the palaeogeographic maps of Clark and Fitzhugh (1990) must be incorrect in the Kanairiktok/Kaipokok area.

Clark and Fitzhugh (1991) subsequently produced a series of isobase maps that show the change in relative sea level from 7 ka to 3 ka. These are different from isobase maps of others (e.g., Andrews, 1989). The maps of Clark and Fitzhugh (1991) show the isobases converging over the central part of the Labrador coast (area of Hopedale) and diverging to the north and south. This reflects their view that emergence was greatest at Hopedale (based on the 119 m marine limit). This they attribute to a later deglaciation and locally thicker ice near Hopedale.

The Clark and Fitzhugh (1990, 1991) estimates of marine limits for the area of Hopedale and Makkovik are considered too high. They place great weight on the work of Daly (1902), but some marine limit estimates of Daly (1902) are known to be incorrect; e.g., Daly (1902) estimated the marine limit at St. John's to be 508 feet above the present, whereas Rogerson (1982) and Andrews (1989) conclude that the marine limit at St. John's is near or at the present sea level. Daly (1934) revised his earlier work and suggested that the amount of emergence on the central Labrador coast was approximately 100 m (he also corrected his 1902 estimate for the marine limit at St. John's; his 1934 Fig. 64). This revision by Daly (1934) is not discussed by Clark and Fitzhugh (1991).

Clark and Fitzhugh (1991) used the following equation to calculate the amount of emergence

$$Y = e^{-kx}$$

where Y is the amount of emergence, in metres, achieved in x years, and k is the decay constant that describes the rate of emergence. Using this equation for a marine limit of 130 m in the thesis area and deglaciation ages of either 8 ka or 9 ka leads to estimates of the exponential decay constant of 0.61 and 0.54, respectively. The 8 ka and 9 ka alternatives are based on the minimum and maximum deglaciation ages for the upstream area suggested by most previous workers (King, 1985; Vilks *et al.*, 1987; Andrews, 1989; Vincent, 1989). These decay constants are reasonable in light of the decay constants for Goose Bay and Groswater Bay suggested by Clark and Fitzhugh (1991). Emergence curves based on these values are plotted in Figure 2.4, with the curve for the downstream area constrained by deflection considerations to have a maximum value 25 m less than the 130 m upstream suggested marine limit (§.2.2.1).

The isobase maps of Andrews (1989), based on smoothed regional data, suggest that 100-80 metres of emergence have occurred in the study area since 7 ka. Using these values, another emergence envelope for the thesis field area can be constructed (Fig. 2.5). Extrapolation of these curves to the 8 ka or 9 ka possible ages for deglaciation suggests that 198 m or 298 m of emergence has occurred since deglaciation. These emergence estimates are inconsistent with the observed 125 m marine limit (Batterson *et al.*, 1988) and the maximum marine limit suggested by Fulton (1986) and Fulton *et al.* (1980).

Mathematical modelling of the behaviour of the mantle because of glacial loading and unloading has been used to construct emergence curves (Peltier, 1989; Peltier and Andrews, 1983). Emergence curves based on the most recent ice sheet modelling (W.R. Peltier, personal communication, 1993) (Fig. 2.6) suggest less than 60 m of emergence in the Kanairiktok Valley since 8 ka. This curve cannot be correct, because marine shells dated at 7950 ± 95 yr B.P. (Beta-28885) were collected at ca. 75 m elevation from silts and muds interpreted to represent either prodelta or fjord-bottom sediments. These sediments and shells suggest water depths of at least a few tens of metres and therefore require a minimum emergence of ca. 100 m at the sample site.

From the above discussion it can be seen that the relative sea level history of central Labrador is still not well known and many of the models that are based on wide geographic areas or theory are inappropriate for the thesis area. Nevertheless, some conclusions are clear. First, the emergence curves of Clark and Fitzhugh (1990, 1991) are highly questionable because they are not supported by any other study, including this one, and their suggestions contradict other theories and models for the deglaciation of the area. Also, Clark and Fitzhugh (1990, 1991; P.U.

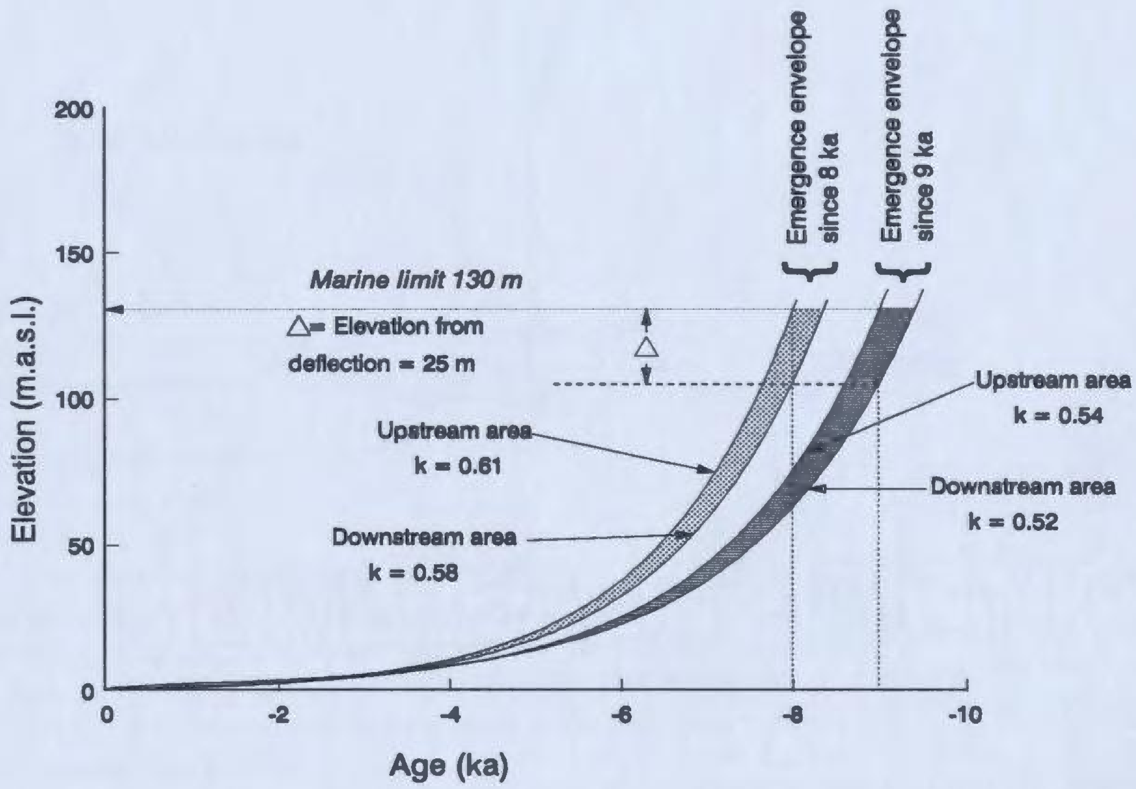


Fig. 2.4. Two emergence envelopes for the study area depending on whether deglaciation occurred at 8 ka or 9 ka. Both envelopes assume a 130 m marine limit, and an initial westward crustal deflection of ca. 0.5 m/km (the difference in the marine limits of the upstream and downstream parts of the area (Δ) is inferred to be about 25 m).

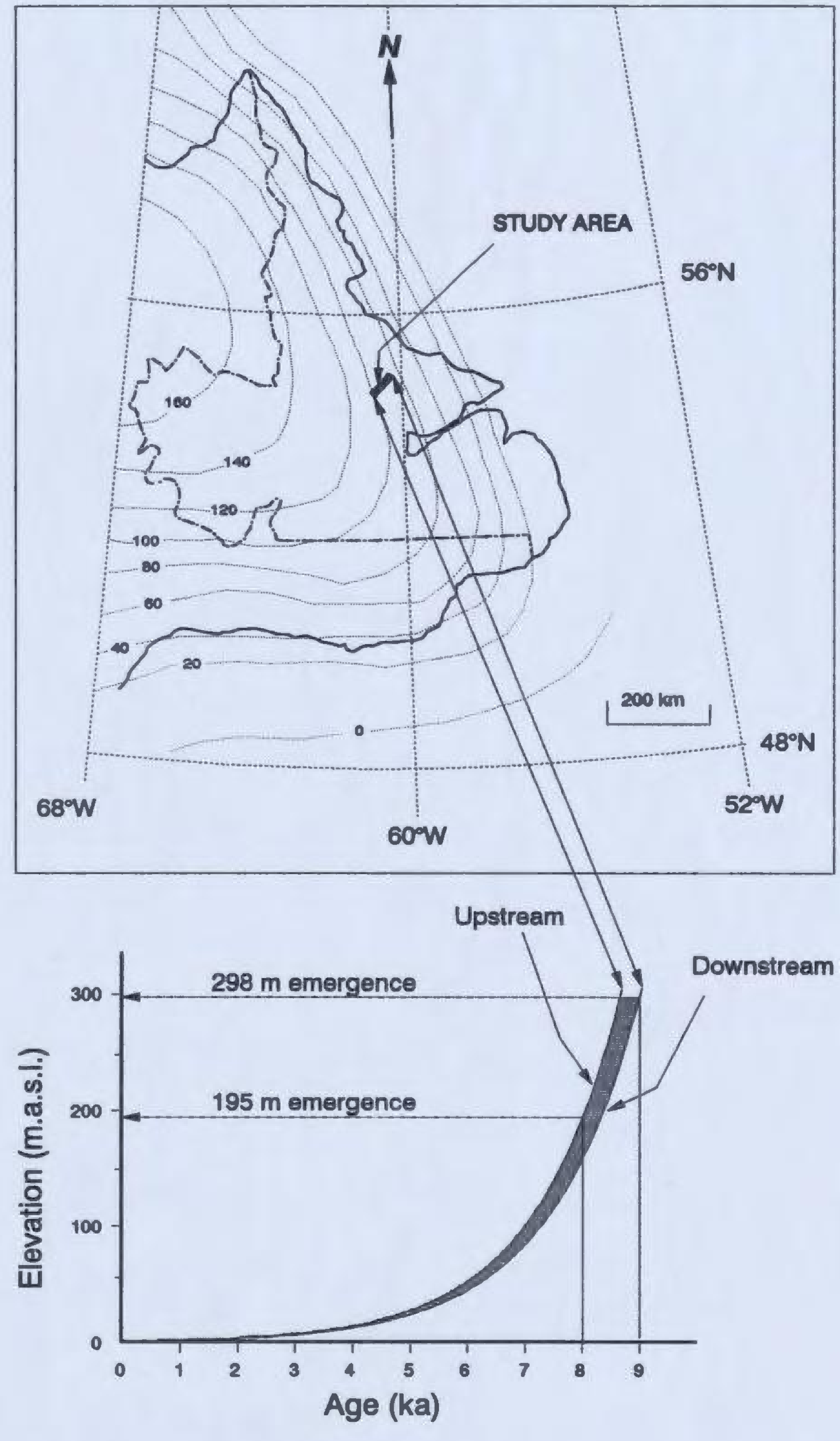


Fig. 2.5. Emergence envelopes for the study area, calculated from the upper isobase map for Labrador at 7 ka (the upper figure is modified from Andrews, 1989).

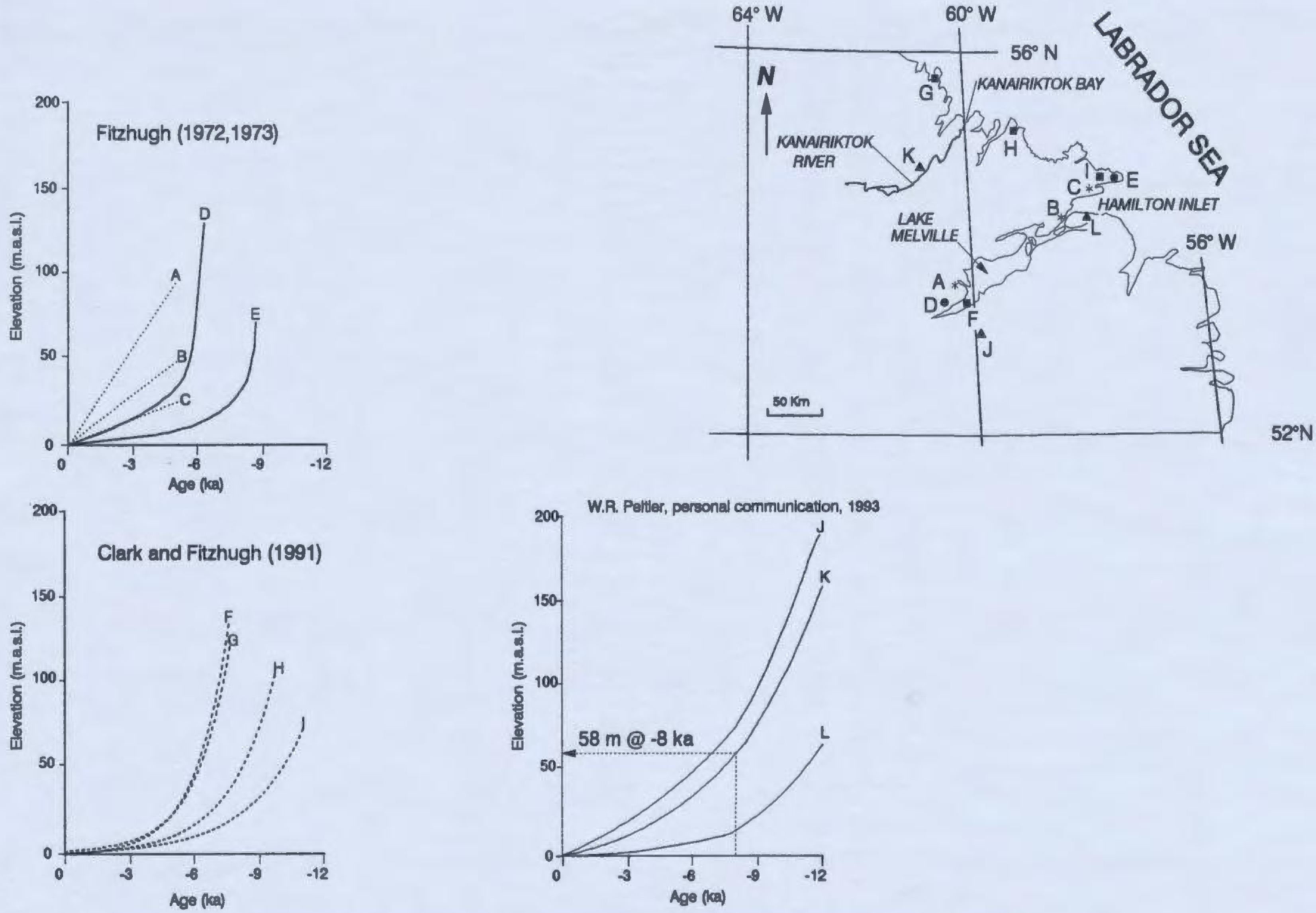


Fig. 2.6. Emergence curves for different parts of southeastern Labrador as suggested by other workers.

Clark, personal communication, 1993) based their 119 m marine limit at Hopedale solely on acceptance and checking of one of the observations of Daly (1902), without critically evaluating the other washing limits reported by Daly (1902). Second, an emergence curve based on modelling of the response of the mantle to ice sheet loading (W.R. Peltier, personal communication, 1993) is inappropriate for the Kanairiktok Valley area; this may reflect the wide area on which the model is based and the paucity of control points near the thesis area (W.R. Peltier, personal communication, 1993). Third, the isobases shown by Andrews (1989) suggest an unreasonable amount of emergence in the area when extrapolated to 8 ka or 9 ka deglaciation. The 7 ka isobase can only be consistent with the suggested 130 m marine limit if early emergence (between 9 ka or 8 ka and 7 ka) was retarded by the presence of ice in nearby areas. Decreased or no emergence due to the presence of nearby ice loads just after local "deglaciation" has been suggested for other areas (e.g. England, 1983).

For this study, the emergence curves for a marine limit of 130 m and an age of deglaciation of either 8 ka or 9 ka are considered to best reflect the sea level history for the area because they take into account the elevation of the marine limit estimated by others (Fulton *et al.*, 1980; Fulton, 1986; Batterson *et al.*, 1988) and the minimum and maximum age of deglaciation of the upstream area suggested by most studies (King, 1985; Vilks *et al.*, 1987; Andrews, 1989; Vincent, 1989). The shape of these emergence curves does not account for the effect of nearby ice. If the load of nearby ice was broadly distributed, there may have been an initial period of slow emergence (gentler slope than what is shown in Fig. 2.5) followed by a period of rapid emergence (steeper slope than what is shown in Fig. 2.5) which decreases exponentially to the present (*cf.* England, 1983). The hypothetical effect of delayed rebound on the shape of the emergence curves is shown in Figure 2.7, constrained by deflection considerations and the suggested marine limit. Note that the downstream part of the area may have experienced a slow rate of emergence while the upstream area was still depressed by ice load, and the period of accelerated emergence of the downstream part may have started while the upstream part of the area was still only slowly emerging.

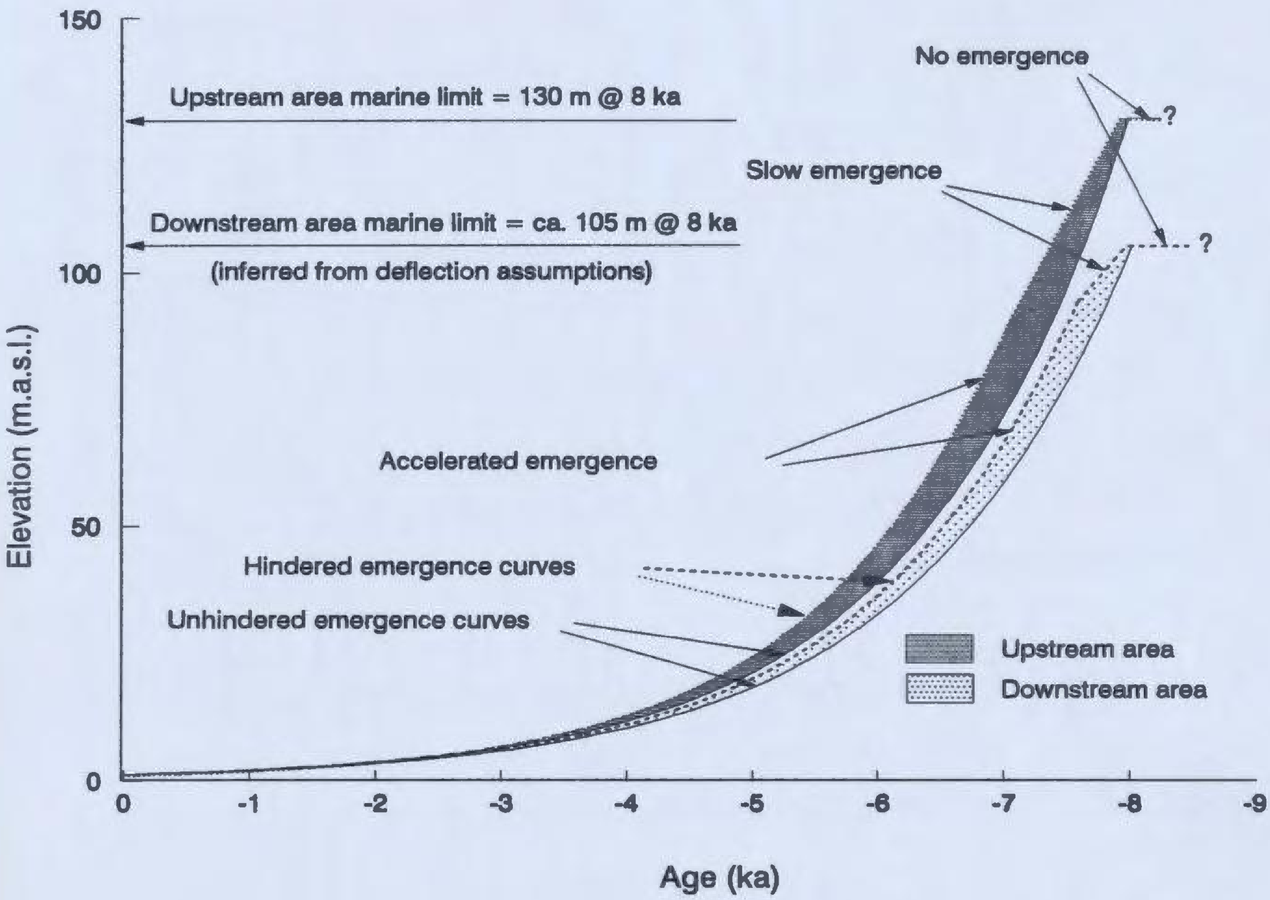


Fig. 2.7. Hypothetical emergence curves for the upstream and downstream parts of the study area showing the possible effect of nearby ice on the shape of the emergence curves.

2.2.3. AMOUNT AND RATES OF POSTGLACIAL EMERGENCE

Labrador lies in Zone I of Clark *et al.* (1978); that is, near an ice centre and exhibiting a large amount of land emergence following deglaciation. The nature and rate of land emergence depend on many factors such as the thickness of the ice in adjacent areas, rate of ice melting and thinning, nature and type of crust beneath the ice load (including flexural rigidity), viscosity of the mantle, the rate at which the mantle material returns to the area that was isostatically depressed, and many other factors (Peltier, 1989; G. Quinlan, personal communications, 1993).

In general, the amount and rate of emergence decrease with time and distance from the ice centre, so that the upstream part of the study area should have experienced a faster rate than the downstream part of the area (Fig. 2.7). Table 2.1 shows the amounts of emergence during 1000-year intervals for both the upstream and downstream part of the area extracted from the curves in Figure 2.4. As before, two possible deglaciation dates (8 ka and 9 ka) are considered. The tabulated values may be incorrect if residual nearby ice loads kept the land surface depressed (e.g., Fig. 2.7).

Based on the high initial 1000-year increments of emergence, and estimates of water depth during the deposition of muds and silts of Units A and B (Chapter 3), marine deposition lasted less than 2000 years (less in the upstream part of the thesis area). After 1000-2000 years of rapid emergence, the former seabed would have risen above sea level.

Table 2.1. Suggested amounts of emergence at 1000-year intervals for the upstream and downstream part of the area calculated from emergence curves (Fig. 2.4) and a marine limit of 130 m.

Time Interval	Deglaciation at 9 ka		Deglaciation at 8 ka	
	Upstream emergence	Downstream emergence	Upstream emergence	Downstream emergence
9 - 8 ka	54 m	42 m	------	------
8 - 7 ka	31 m	24 m	59 m	47 m
7 - 6 ka	19 m	16 m	32 m	26 m
6 - 5 ka	10 m	8 m	18 m	14 m
5 - 4 ka	7 m	6 m	10 m	8 m
4 - 3 ka	3 m	2 m	4 m	3 m

CHAPTER 3

SEDIMENTOLOGY

Sedimentology will be discussed in the context of lithologic units present in representative sections in different parts of the study area (downstream, middle and upstream). Sediments observed in the different parts of the area are first described and then interpreted. From these interpretations and the sea level history (Chapter 2), sedimentary environments are deduced.

Sections are identified by unique numbers preceded by an upper case letter to indicate whether the section is in the upstream (U), middle (M) or downstream (D) part of the area (e.g., U50 indicates Section 50 in the upstream part of the area). The sections are described from the bluffs of the modern river. The description covers all the materials exposed in these bluffs down to the water level in the modern river. Often, the lower parts of the bluffs could not be trenched and the sediments in an interval from the base of the trench to the water level of the modern river are unknown; these intervals are recorded as covered intervals in the description and all the illustrations. In some sections (e.g., Section U50) there are intervals that could not be described either because of the impossibility of digging or the frequent collapse of sediments into the trenches before they could be described. These parts of the sections are also recorded as covered intervals in the text and the illustrations.

Based on examination of more than thirty vertical sections throughout the study area (Fig. 3.1), five main lithologic units are recognized (Fig. 3.2). Seven key sections are sufficient to reveal the essential characteristics of the five units. These sections are D98, D90, M85, M61, U50, U32 and U25. Some units are only present in part of the study area. It should be noted here that the assignment of sediment to a unit is based solely on its texture and sedimentary structures, and not on its stratigraphic position or lateral extent.

Each key section is described below in detail and interpreted. The sedimentary history of that section is then discussed, based on the vertical sequences exposed at that section. The spatial relationships between the various units, and the sedimentary history of the upstream, middle and downstream parts of the area are discussed after the description and discussion of the sections in that area. The sedimentary history of the whole area in relation to other studies is

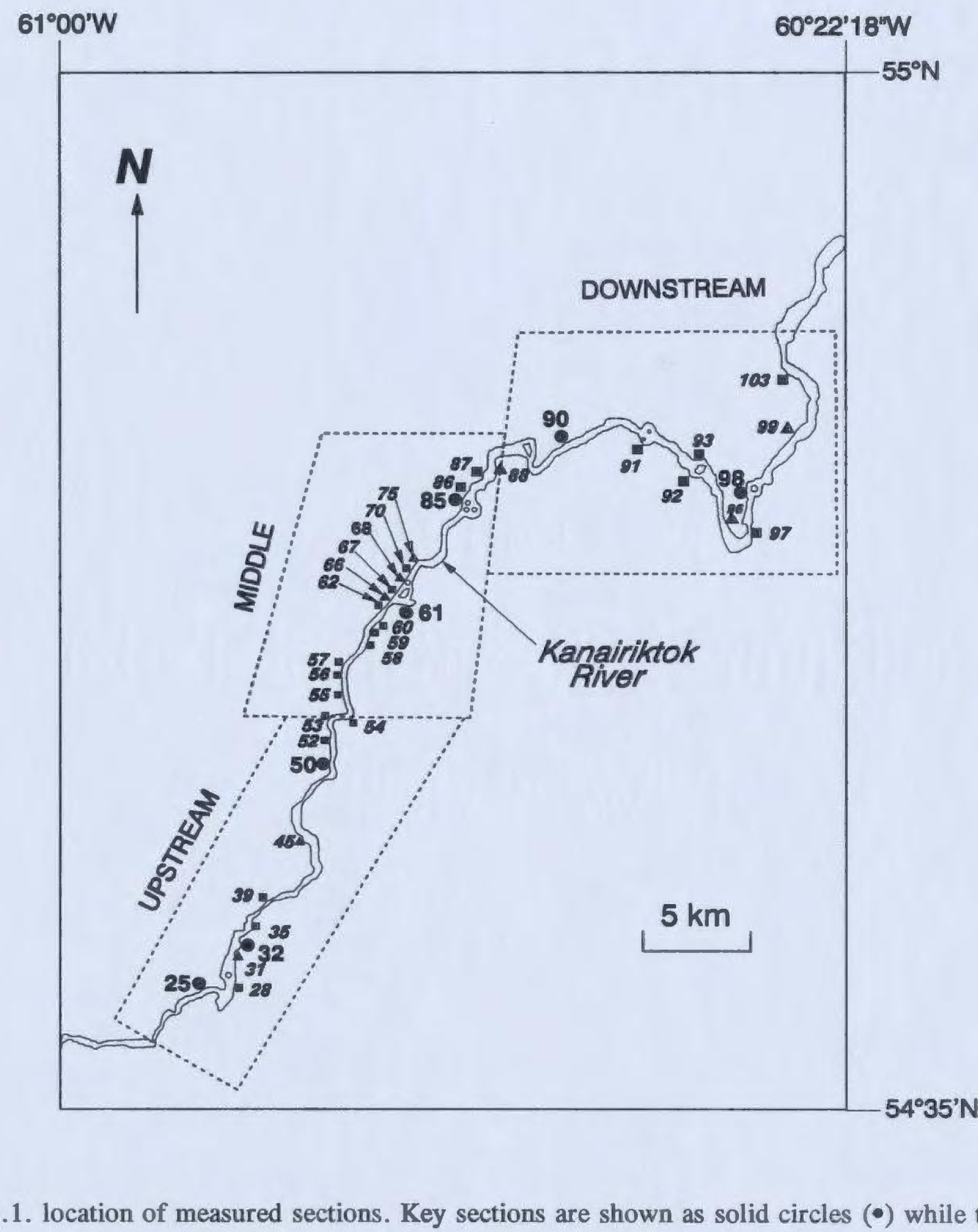


Fig. 3.1. location of measured sections. Key sections are shown as solid circles (•) while their associated sections are shown as solid triangles (▲). Other sections are indicated by solid square (■).

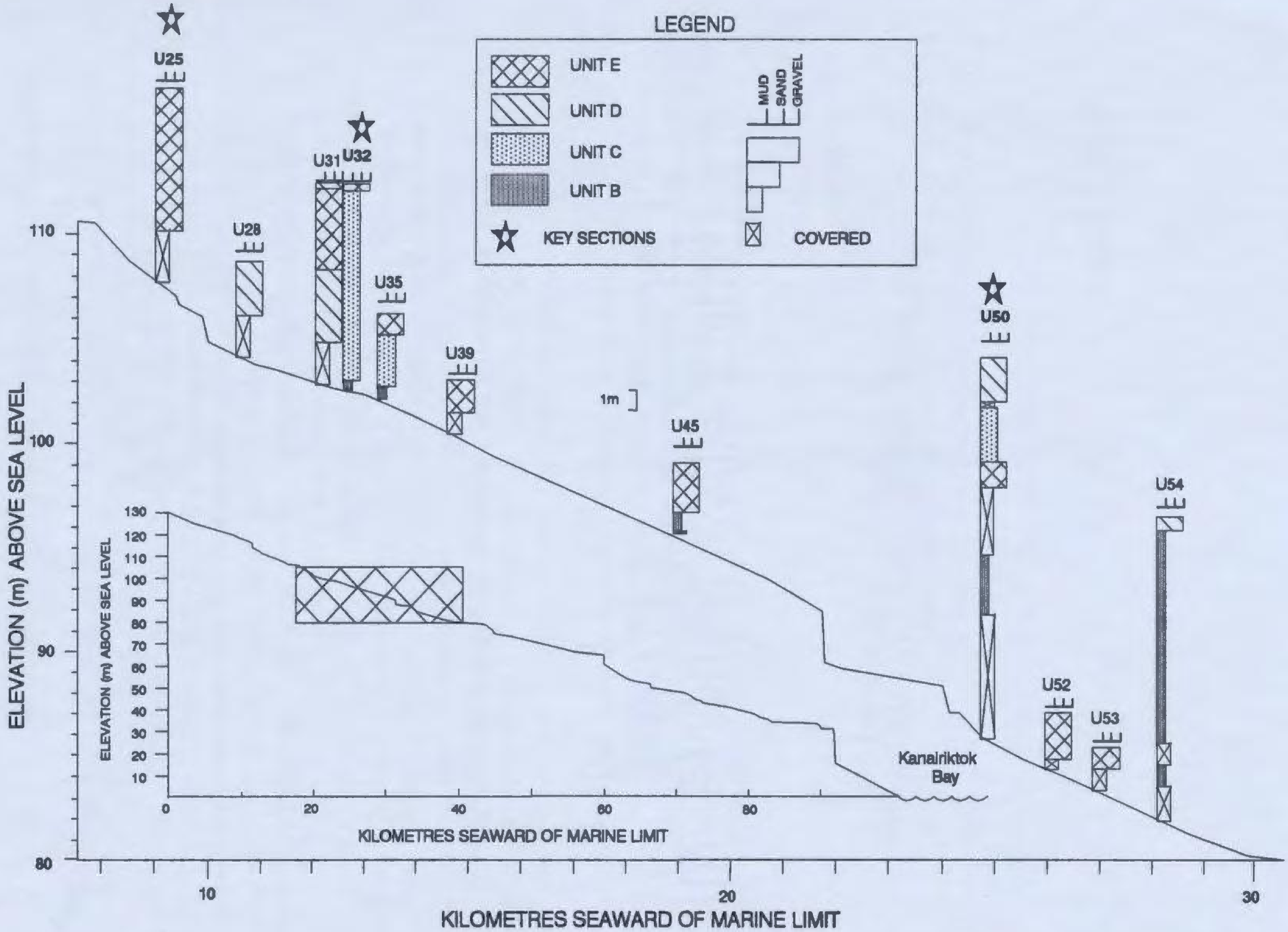


Fig. 3.2a. Distribution of the different sedimentary units in the sections of the upstream part of the area.

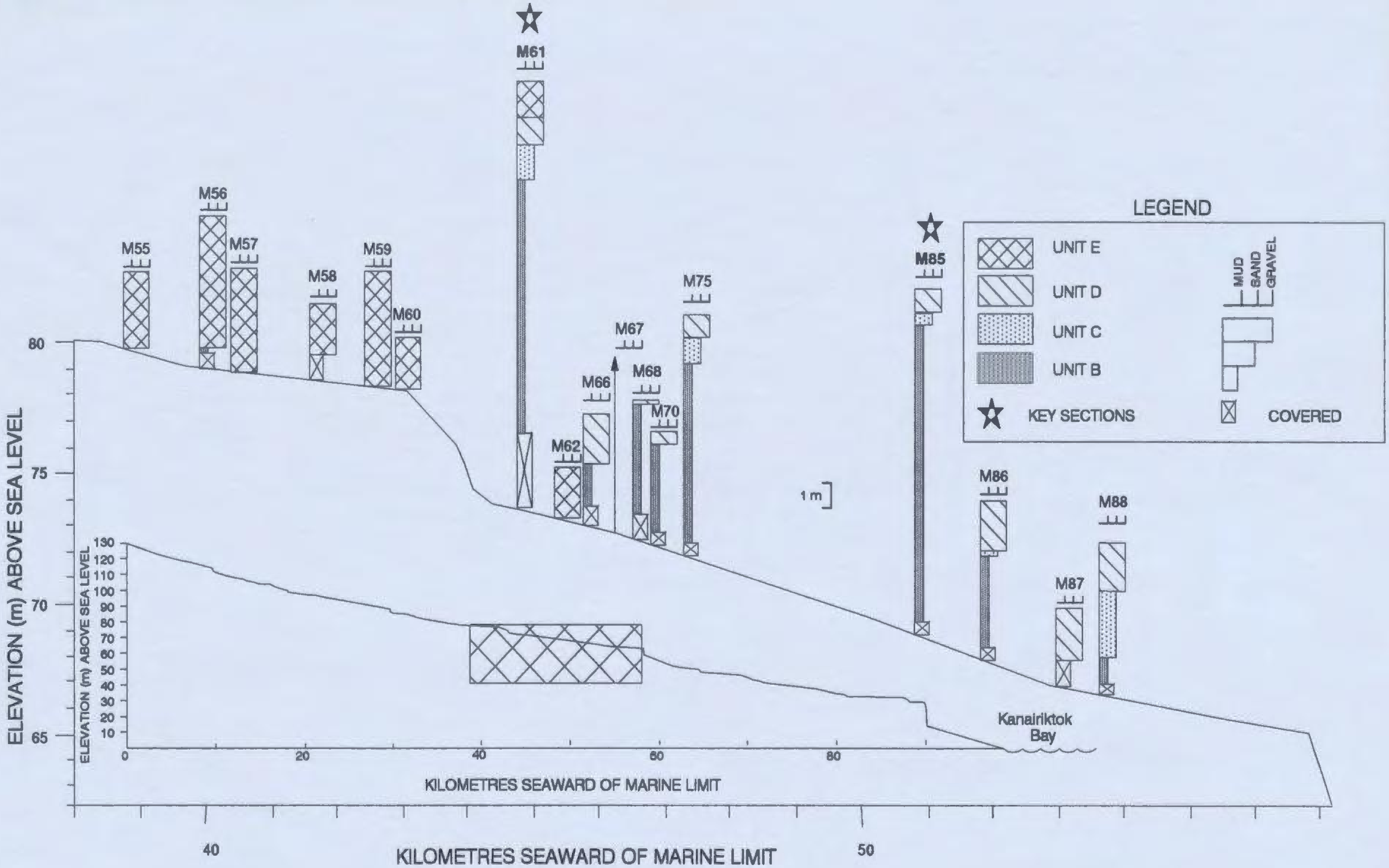


Fig. 3.2b. Distribution of the different sedimentary units in the sections of the middle part of the area.

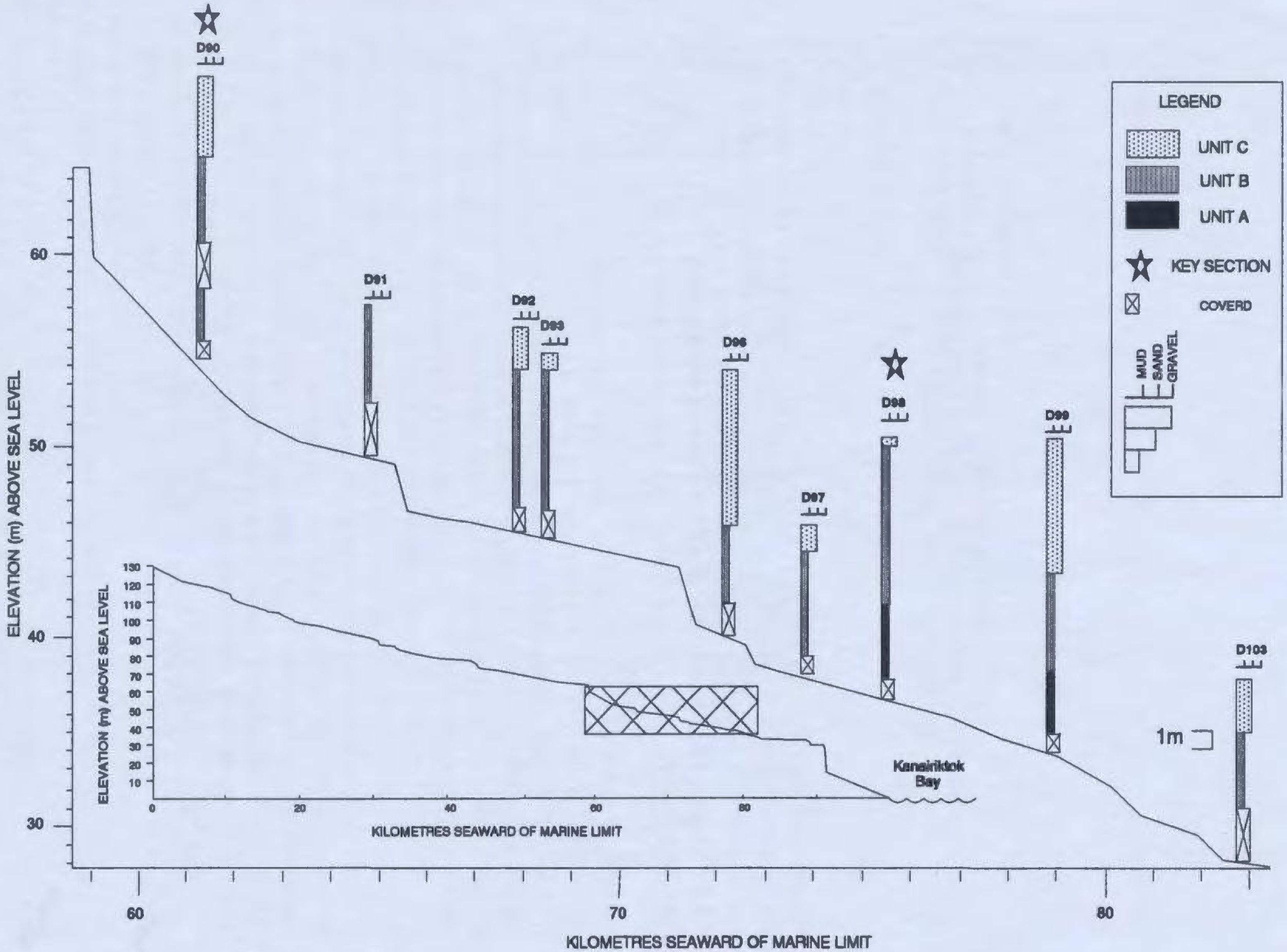


Fig. 3.2c. Distribution of the different sedimentary units in the sections of the downstream part of the area.

discussed in Chapter 5.

The five lithologic units have the following composition. Unit A is massive mud, Unit B is interstratified sand and mud, Unit C is sand, Unit D is gravel, and Unit E is diamicton. Selected representative photographs of the different units are shown in Figure 3.3. These units tend to be superimposed in such a way as to form upward coarsening sections with units A or B at the base and units D or E at the top. The sequence of units in most sections (except in Section U50 and U31) is A→ B→ C→ D→ E, although all five units are never exposed at one section. In Section U50, Unit E underlies Unit C while Unit D forms the top of the section. In Section U31, Unit D sediments overlie and underlie Unit E. These units are not stratigraphic units with relative chronological significance and their recognition is based solely on the texture of the sediments that form the unit. Unit A is present in two sections in the downstream part of the area where it forms the lowermost parts of exposed sediments in those sections. Unit B is the most common unit in that it is present in almost all the sections in all parts of the area and forms a large thickness of the sediments exposed in those sections. It forms the lower part of most of the sections (except the two sections in the downstream part where it gradationally overlies Unit A). Unit B may contain thick sandy intervals (e.g., Section M61); nevertheless, these are included in Unit B rather than separated as a separate unit (e.g., Unit C) because the overall nature of the enclosing sediments is similar to Unit B and the sandy beds have a texture similar to the thinner sand strata within Unit B. Unit C is present in different parts of the area, but it usually sharply overlies Unit B. Unit D is present in the upstream and middle part of the area where it usually forms the upper parts of the exposed sections, sharply overlying Unit B or Unit C and underlying a thin recent vegetation cover. Unit E is present in some sections in the upstream and middle parts of the study area where it usually forms the upper part of the sections (except Section U50). It sharply overlies different units and underlies a recent vegetation cover.

Many of the units are divided into subunits that are differentiated on the basis of variations in texture or sedimentary structures in the units. Subunits at one section (e.g., b1, b2 at Section D98) are distinct from subunits with the same designation at another section (e.g., b1 to b6 at Section D90).

Fig. 3.3. Selected examples of the different sedimentary units in the study area. All scale divisions except in H are equal to 10 cm. In Fig. H, the bar is equal to 1 m.

A= Unit A in Section D98. Note the massive nature of the unit.

B= Middle part of Unit B in Section D98. Note the gradual upward increase in frequency of sand layers, each 1-3 cm thick, from the lower part of the photograph to the upper part. One such layer is ~20 cm from the base of the photograph.

C= Interstratified sand and mud in the upper part of Unit B in Section D98. Note the cross set of sand of Unit C that sharply overlies this Unit.

D= Unit B in Section M85. Note the apparently random nature of alternation of laminae of different thicknesses.

E= Unit C in Section U50, interpreted as aeolian sands. Note the well developed stratification. The sand is well sorted.

F= Cross bedded sand of Unit C in Section D99.

G= Diamicton of Unit E in Section U50 (lower part of the photo) overlain sharply by sands of Unit C. Note the poor sorting in the diamicton and the roundness of the clasts. Note also the bullet-rod shape of the clast near the middle of the bar scale.

H= Unit D gravel in Section M75 (at the top) overlying dipping sand beds of Unit C (5°-8° to the right). Interstratified sand and mud of Unit B is at the base.

(A)

(B)

(C)

(D)

Fig. 3.3.

(E)

(F)

(G)

(H)

Fig. 3.3. (Continued).

Essential lithological characteristics of the units are summarized in Table 3.1. The chronological relationships between the five lithologic units across the area are not known, so that (a) units with the same lithological character may not have equivalent age, and (b) some units of different lithological character may be of equivalent age.

The seven key sections are described below, starting with sections in the downstream part of the area because the downstream sections generally expose the lowest units A and B. The benefit of describing single key sections in their entirety is that the vertical relationships between units are emphasized, at the same time that characteristics of typical occurrences of each unit are described. Units B and C, present in several of the key sections (Fig. 3.2), show sufficient spatial variability to warrant multiple description and interpretation.

Microscopic examination of more than 100 thin sections made from Unit B shows that four main types of laminae and thin beds are present in this unit (Fig. 3.4). The basic characteristics of these laminae are summarized in Table 3.2; they are not described at length for each occurrence of Unit B to avoid repetition.

In this chapter, terms used to describe lithology, texture and statistical parameters of grain size analysis (e.g., sorting and skewness) follow Folk (1974; see Appendix § A). The sediment colour was determined in the field using the rock colour chart published by the Geological Society of America (1984).

3.1. SECTIONS IN THE DOWNSTREAM PART OF THE AREA

Two key sections will be discussed in detail (D98, D90) while associated sections (D99, D96, M88) located upstream and downstream of the key sections will be discussed briefly. Some associated sections are not located directly next to the key sections (e.g., D96) but warrant consideration either because they have thicker sediments or distinctly different sediments. When any associated section has sediments distinctly different from those in the key section these sediments will be discussed in detail. The sedimentary environment of the downstream part of the area will be discussed after all sections have been described and interpreted.

Table. 3.1. Summary of the character of lithologic units A-E

Unit	Lithology	Short Description	Distribution[1]	Cumulative Thickness in single sections
E	Muddy gravel (diamicton)	Massive and inversely graded. Matrix to clast supported. Sharp upper and lower contacts.	U, M	0.3 m to ~ 8 m
D	Sandy gravel	Massive, tabular beds. Sharp upper and lower contacts. Matrix to clasts supported. Normally graded beds (rare). Inclined bedding.	U, M	0.2 m to 3.5 m
C	Sand	Massive with sharp upper and lower contacts. Normally graded with sharp lower and graded upper contacts. Cross stratified. Parallel laminated. Inclined bedding. Inversely graded with sharp upper and lower contacts.	U, M, D	0.1 m to ~ 10 m
B	Interstratified sand and mud	Massive sand strata with sharp upper and lower contacts. Normally graded sand/silt with sharp lower contacts and gradational upper contacts. Inversely graded sand/silt with sharp upper and lower contacts. Parallel stratified sand with upper and lower contacts. Sand strata with diffuse upper and lower contacts. Mud may be laminated or massive.	U, M, D	0.15 m to ~13 m
A	Massive mud	Mud without laminations. Upper contact gradational.	D	3.6 m to 4 m.

[1] U = Upstream area, M = Middle area, D = Downstream area. Underlined = main occurrence. See Fig. 3.2 for the stratigraphic position of each unit.

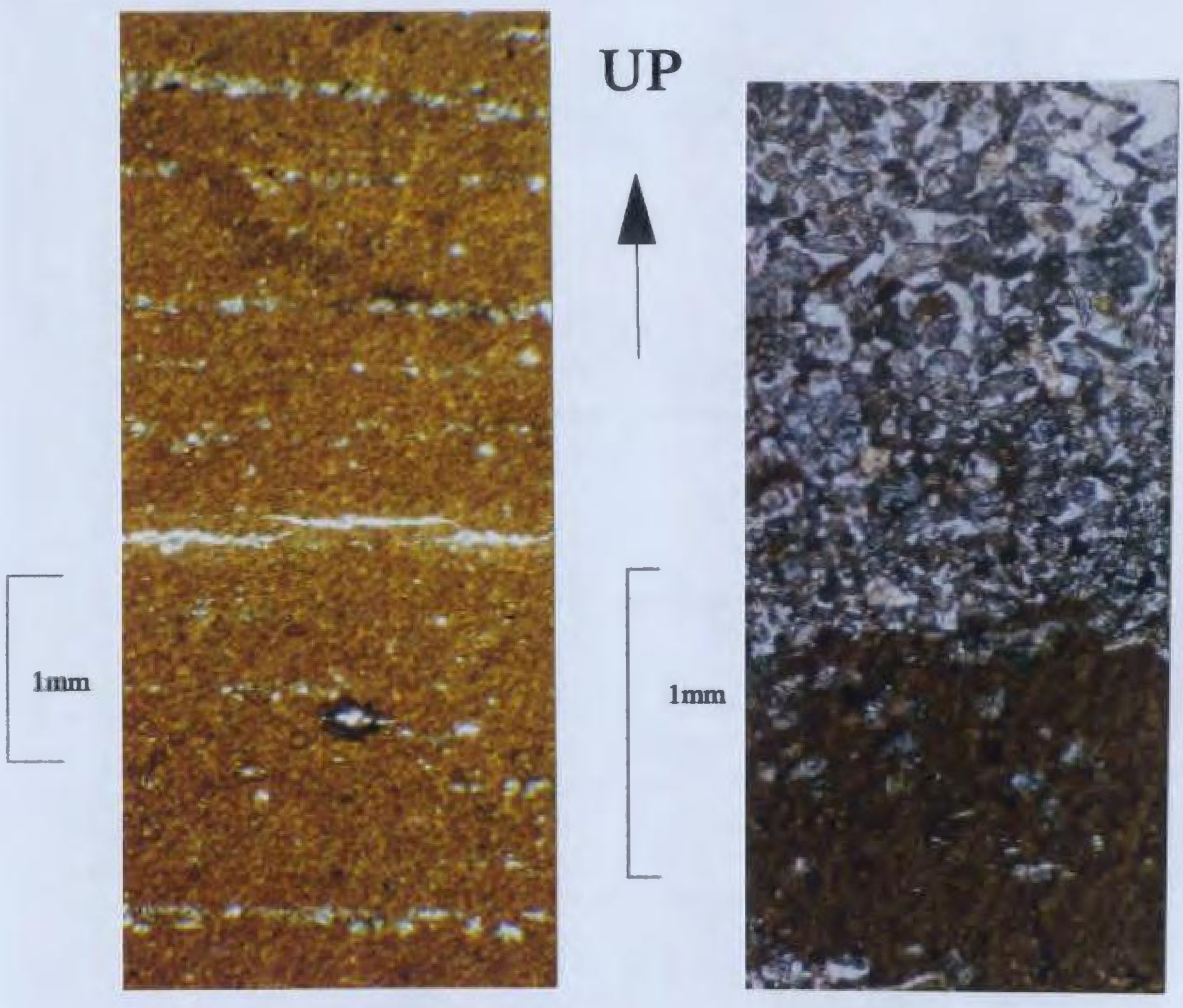
Fig. 3.4. The different types of laminae observed in Unit B (Table. 3.2).

A= Laminae type L_{I}. Note that the laminae are one to a few grains thick and the diffuse contacts. The thin section was made from a slab tray taken from the lower part of Unit B in Section D98.

B= The lower part of an example of laminae type L_{II} showing inverse grading. The thin section was made from a sample taken from the upper part of Unit B in section D98.

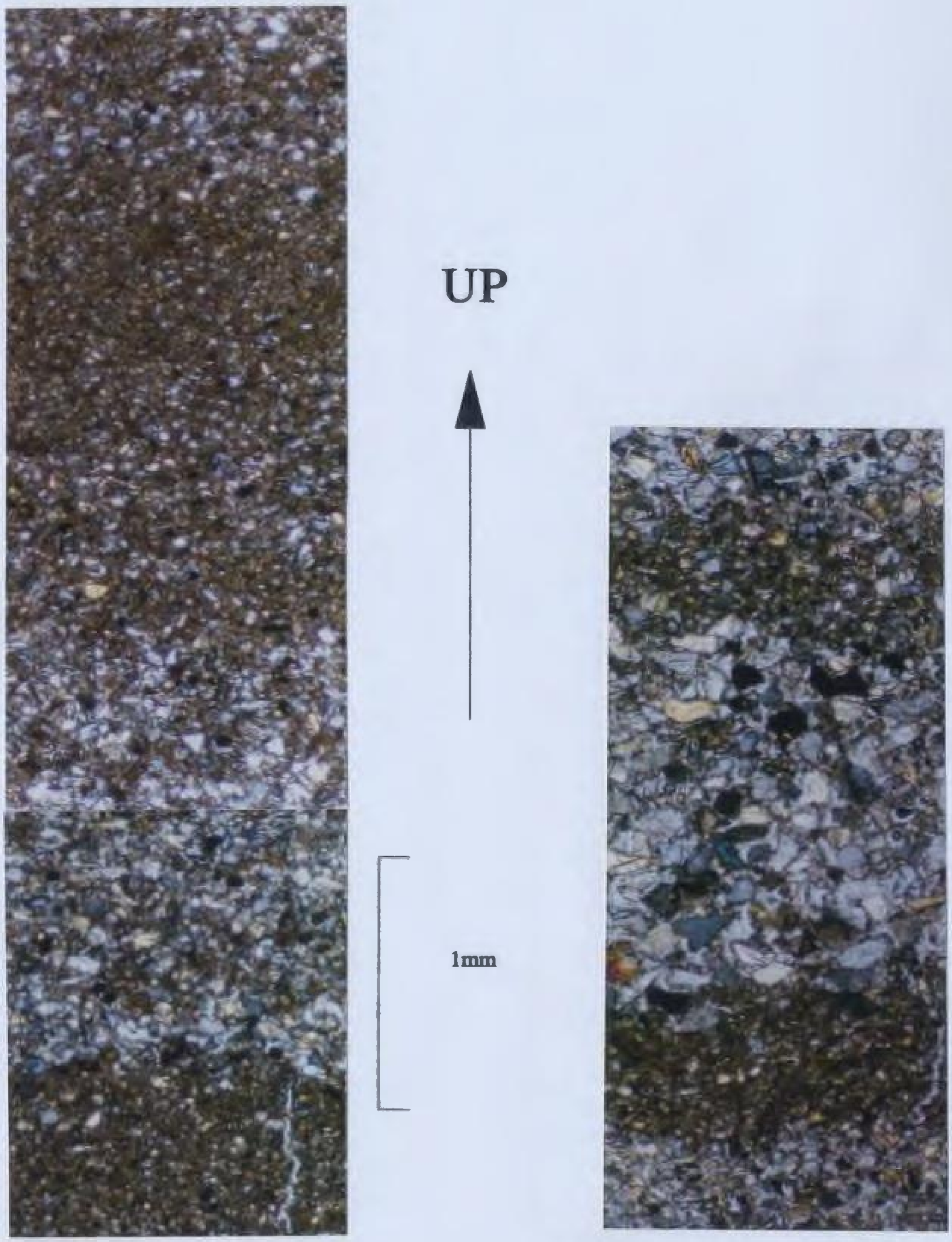
C= Normally graded lamina of type L_{III}. Note the gradual upward increase in the amount of mud. The sample was taken from a tray sample taken from the lower part of Unit B in Section U50.

D= An example of Lamina type L_{IV} consisting of massive sand with sharp upper and lower contacts. The thin section was made from a slab tray taken from the more muddy part of Unit B in section U50.



(A) L_I (B) L_{II}

Fig. 3.4.



(C) L_{III} (D) L_{IV}

Fig. 3.4. (Continued)

Table 3.2. Summary of the characteristics of the different laminae/thin beds observed in Unit B in this study (F = fine; M = medium; C = coarse; g = gradational; d = diffuse; s = sharp; e = erosive; v = variable; R = reverse; N = normal; MS = massive, GRD = graded).

Lamina Type	Grain Size	Sand Content	Sorting	Thickness	Contacts	Structure	Key Section(s)
L_I	F. Sand, Silt	Low	Poor	<1 mm - 4 mm	d Lower d Upper	MS, N GRD	D98, D90, M85
L_{II}	M. Sand, F. Sand	High	Improves Up., Moderate	1 - 12 mm	s Lower s/d Upper	R to N GRD, N GRD	D98, M85
L_{III}	F. Sand to Mud	Decreases up., v	Decreases up., Poor	0.4 mm -10 mm	s Lower g Upper	N GRD	D90, M85, U50
L_{IV}	F. Sand, Silt	High	Moderate - Poor	~1 mm - 12 mm	s Lower s Upper	N GRD, MS	U50

3.1.1. KEY SECTION D98

This section is located approximately 25 km upriver from the head of Kanairiktok Bay. Sediments exposed in this section are 13.5 m thick. The base of Unit A is not exposed. Unit A forms the lowermost part of the section and underlies Unit B with a gradational contact. Unit C overlies Unit B with a sharp planar contact and underlies 30 cm of recent vegetation (Fig. 3.5).

3.1.1.1. UNIT A

Unit A, 4.0 m thick, is dominated by clay-sized particles (67-71%) mixed with 26%-32% silt and 0.9-2.3% sand (Sedigraph analyses of three representative samples: SN2, S7, and S8 in Appendix B). This unit lacks sedimentary structures. Colour on fresh wet surfaces ranges from pale reddish brown (10 R 5/4) to greyish red (10 R 4/2).

Thin sections made from slab trays taken from the upper part of the unit (T98B) consist of mud that contains scattered 0.3-1.2 mm sand grains that are concentrated in distinct 1-7 mm-thick bands. The sand grains do not form continuous layers, but are separated and surrounded by finer mud. Silt is common in the sandy intervals. No particles coarser than 1.2 mm were found in Unit A in any part of the study area. The samples have polymodal to bimodal grain size distribution so that the calculation of the statistical parameters of grain size are unreliable. Sample SN2 contains more than 16% of material finer than 13 ϕ ($\phi = -\log_2 d$ where d = grain diameter in millimetres [Krumbein, 1934]). These samples do not plot in any of the fields of the C-M diagram of Passega (1964; also used to evaluate textural data from fine grained sediments -- see Mackiewicz *et al.*, 1984), but instead plot to the left of the field for settling from suspension (Fig. 3.6).

Samples from this unit were not analyzed for pollen, spores, dinoflagellates nor algae. Two poorly preserved foraminifera tests (*Elphidium excavatum* f. *clavata* were recovered from samples taken from this unit.

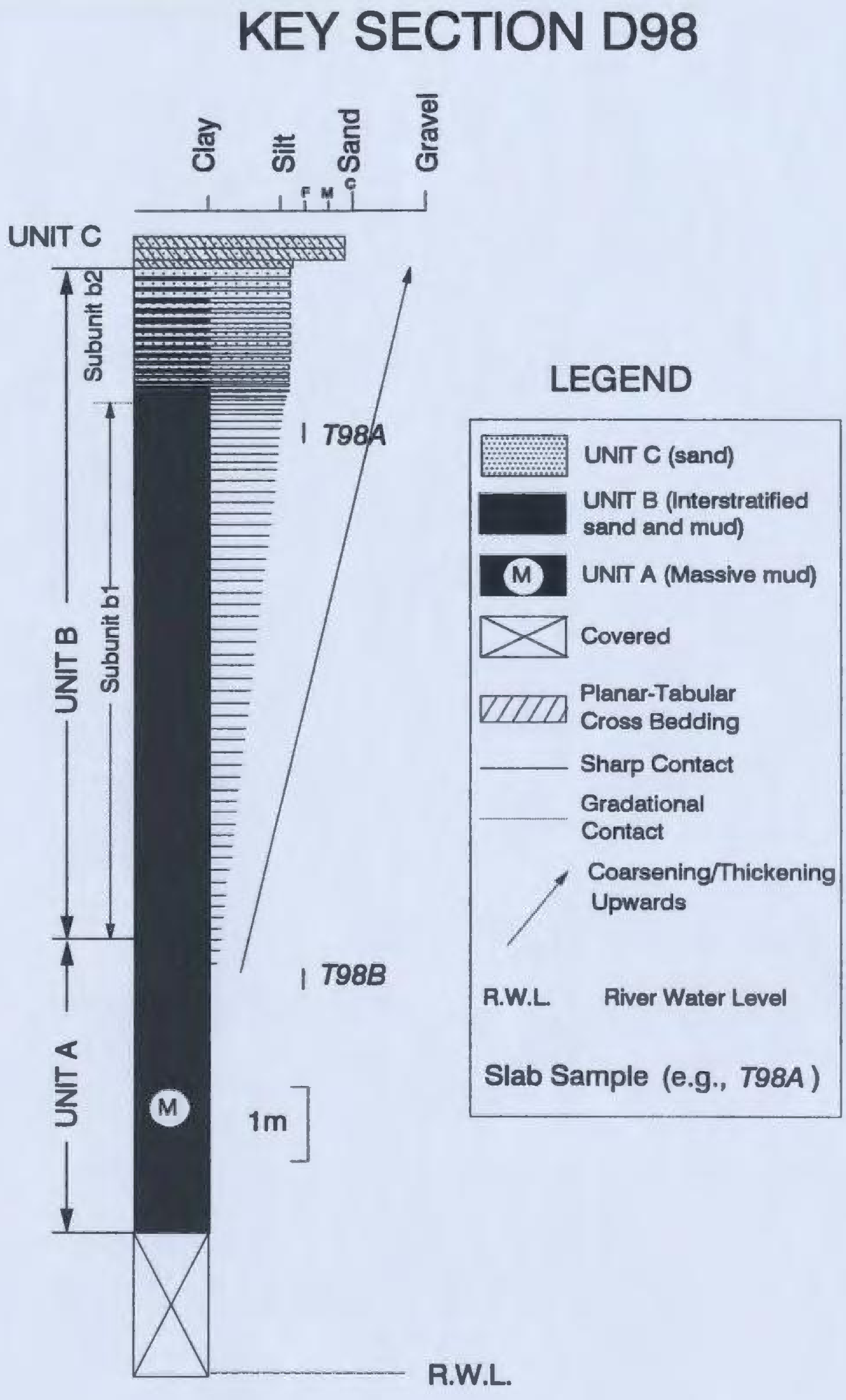


Fig. 3.5. Key Section D98 stratigraphic column. See text for details.

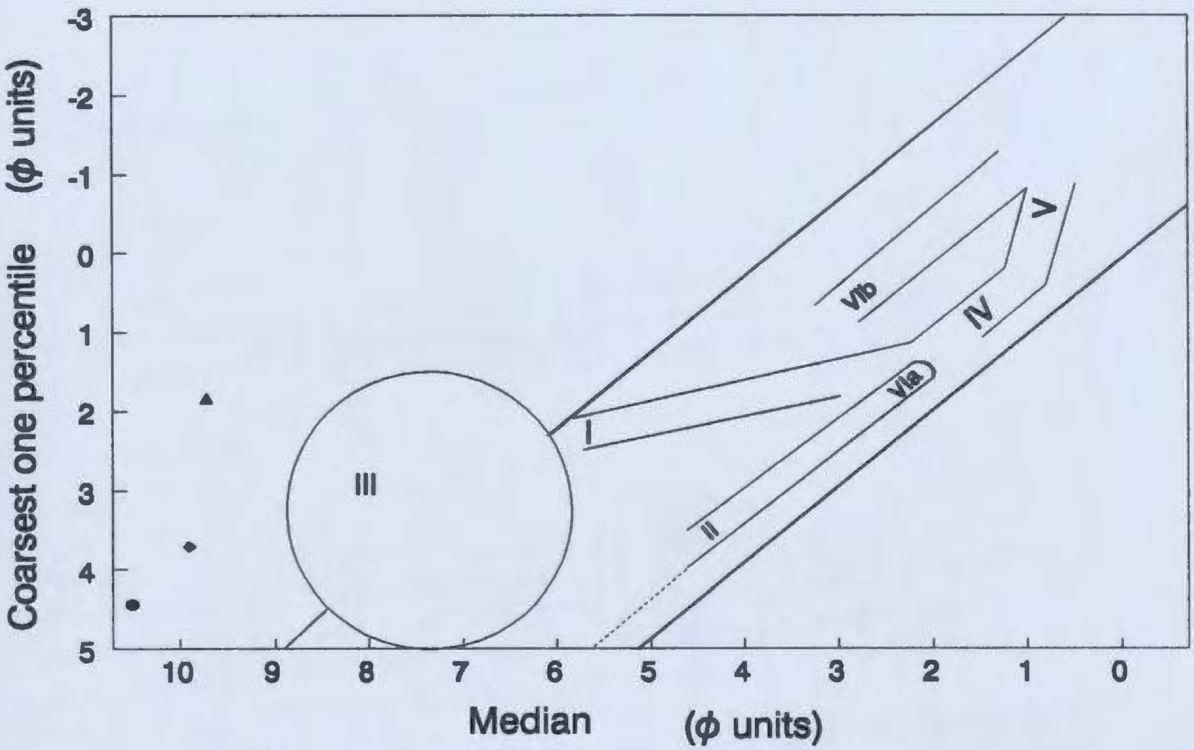


Fig. 3.6. Plot of three samples from Unit A, Section D98, on the diagram of Passega (1964), as modified by Mackiewicz *et al.* (1984). All the samples fall to the left of the suspension settling field (III), probably because the original particles were deposited as flocs. The flocs must have been disaggregated during sample preparation for analysis. I = low-intensity traction currents; II = low density turbidity currents; III = settling from suspension; IV = moderate-intensity traction currents; V = high-intensity traction currents; VIa,b = moderate- to high-density turbidity currents.

3.1.1.2. INTERPRETATION OF UNIT A

The fine grain size and general lack of sedimentary structures in this unit suggest low energy depositional conditions. The presence of dispersed sand at certain levels in an otherwise fine grained mud, with no sharp contacts, suggests two independent mechanisms of deposition for the mud and sand modes (Gilbert, 1983; Mackiewicz *et al.*, 1984; Gilbert *et al.*, 1990).

The dominance of clay-sized particles suggests that Unit A was deposited from suspension by settling of flocculated clays and fine silt (Syvitski *et al.*, 1987). Flocculation is known to take place where mud settles from surface freshwater sediment plumes through saline to brackish water of salinity as low as 3 ‰ (Syvitski *et al.*, 1987). The absence of any material coarser than fine silt (except for the scattered coarse sand particles present at certain levels) suggests that the source of the sediment plumes was distant from this section (Cowan and Powell, 1990); surface sediment plumes in arctic fjords have been observed to carry sand-sized particles 2-3 km away from river mouths (Gilbert, 1980). Density currents, if active, were either very weak or the section was far away from the source of such currents (*cf.* Liverman, 1989).

Alternatively, Unit A may have been deposited when the area was under permanent or semipermanent winter sea-ice cover. Many present-day arctic fjords have winter ice cover for nine months or more (Syvitski *et al.*, 1987). The duration of winter ice cover is affected by many factors such as temperature, waves, tides, wind, snow fall and pack-ice conditions (Gilbert, 1983; Taylor and McCann, 1983; Syvitski *et al.*, 1987). During the winter, formation of sea ice may lead to the development of a homogeneous surface layer under the ice by the process of salt rejection (Gilbert, 1983; Syvitski *et al.*, 1987). In northern fjords, fresh water may enter the fjord as a glacial meltstream that sinks to the pycnocline. Sediments carried in the meltstream may spread down the fjord and settle across the homogenous surface layer as flocs (Syvitski *et al.*, 1987).

The reason that samples do not plot in any of the fields of the C-M diagram of Passega (1964) is probably that the primary grain size of the flocs was modified by sample disaggregation before analysis. Floc disaggregation during analysis of clay grade sediments has been reported by others (Hoskin and Burrell, 1972; Gilbert, 1982). If the effective fall diameter of flocculated mud is taken to be approximately 6.5 ϕ (Gilbert, 1983), these samples would plot in the suspension settling field of the diagram.

The coarse sand particles that are present at certain intervals within this unit have several possible origins. The particles may have been mixed from originally well defined layers by bioturbation, but the apparent lack of burrows in the slabs and X-radiographs does not favour this interpretation. The coarse particles may have been transported by aeolian action, either directly to the surface of an open body of water, or to the surface of sea ice. Reineck and Singh (1980) claim that the maximum particle size carried in suspension by wind is about 0.05 mm, and that wind blown sands are well sorted. Both the coarser sediment size and fairly poor sorting of the coarse mode in Unit A do not support direct aeolian transport for these sand particles.

The coarse grains may have been rafted by plants, icebergs, or sea ice. Although plant-rafted dropstones are rare in muddy sediments of non-glacial origin (Anderson, 1983), rafting by algae is a potentially important transport mechanism (Gilbert, 1990). Algae may transport particles as large as pebbles (Gilbert, 1990). The algae may lift particles to which they are attached or the algae may be incorporated in sea ice or anchor ice on the sea bottom in the littoral zone. The lack of algal debris does not support an algal rafting origin for the coarse sand grains.

The lack of particles coarser than 1.2 mm is difficult to explain if transport was by ice rafting, because ice-rafted materials, particularly from bergs, range in size from clay to gravel (Anderson, 1983; Gilbert, 1990). Sediment-laden sea ice, however, may lack gravel if the ice obtained its sediments by freezing to a sea bed composed of only sand or mud (e.g. Anderson, 1983); preferential incorporation of fine sediments by this process has been observed in Makkovik Bay (Rosen, 1979). Alternatively, if the coarse particles were transported to the surface of winter sea ice by wind, as saltation load, then the anomalously coarse maximum size and poor sorting can be explained by accumulation over a period of weeks or months as the wind fluctuated.

In summary, sediments of Unit A are interpreted to have been deposited from suspension fall-out of fine silt and clay particles from surface sediment plumes in areas far away from river outlets. Some of Unit A may have been deposited below winter sea ice from subglacial meltstreams. The primary particles flocculated as they settled through saline or brackish waters.

Sand particles present at certain intervals were probably transported by algal rafting or by wind transport to local winter sea ice that released the sand when the ice melted.

3.1.1.3. UNIT B

Unit B in Section D98 is divided into two subunits, described separately below. Subunit b1 gradationally overlies Unit A and gradationally underlies subunit b2.

3.1.1.3.1. Description of subunit b1

Subunit b1 consists of 7.25 m of interstratified sand and mud. Its contact with Unit A is marked by the gradual appearance of thin (1-3 mm thick) silt to fine sand laminae. Colour on fresh wet surfaces is moderate reddish brown (10 R 4/6) to moderate red (5 R 5/4).

Thin sections made from a slab tray sample (T98A) show scattered coarse particles as in Unit A; some of these coarse particles deform the faint laminae and are surrounded by finer mud. The laminae in subunit b1 are of two types. The first consists of thin faint laminae of very fine sand and silt (type L_I in Table 3.2). The second type of lamina is sharp based, and 1-5 mm thick (type L_{II} in Table 3.2). The two types of laminae occur together but the second type becomes progressively more common in the upper part of this subunit.

The mud between both types of laminae is massive and does not show any indication of bioturbation or grading. There is a sharp break in grain size between the fine sand or silt and the mud. Below some thick laminae (type L_{II}) there are scattered sand grains similar in size to the sand in the overlying lamina.

Grain size distributions of bulk samples (i.e., including mud and sand/silt laminae) from different parts of this unit consist of 12%-16% sand, 28%-29% silt and 57%-59% clay (samples S71 and S72 in Appendix B). These bulk samples have a bimodal grain-size distribution, with a coarse mode (3.5 to 4 ϕ) and a fine mode (9.5 ϕ). These modes correspond to the sand/silt laminae and the massive mud interlayers, respectively.

3.1.1.3.2. Interpretation of subunit b1

The dominance of fine grained material suggests mainly low energy depositional conditions. Laminae of silt and sand in mud can be formed by different mechanisms in different environments (Reineck and Singh, 1980). These mechanisms include a variety of density/buoyancy currents (turbidity currents, grain flows, overflows, interflows, and underflows; Gilbert, 1983; Mackiewicz *et al.*, 1984; Stow and Piper, 1984) and reworking of bottom sediments by marine currents (Barrie, 1980; Barrie and Piper, 1982; Piper *et al.*, 1983; Johnson and Baldwin, 1986).

In this thesis the term "overflow" is used for both overflows and interflows because it is difficult or impossible to differentiate between their deposits (Mackiewicz *et al.*, 1984; Smith and Ashley, 1985). These surface or mid-water flows rain sediment to the sea floor as they decelerate and mix with sea water. Couplets of mud and sand/silt laminae formed in this way have been termed cyclopels by Mackiewicz *et al.* (1984, p. 114).

Mackiewicz *et al.* (1984) described deposits formed by traction transport beneath continuously flowing density currents fed from subglacial tunnels. The fluid phase of these currents is fresh water. To distinguish these currents from surge-type turbidity currents, Mackiewicz *et al.* (1984) called these special currents "underflows". Genetically similar density currents also form where sediment-charged, fresh glacial meltwater has sufficient density to flow beneath a body of marine water (Powell, 1981; Syvitski *et al.*, 1987). These are also called underflows. Clearly, underflows (*sensu* Mackiewicz *et al.*, 1984) can only form in marine environments if the suspension formed of sediment and cold fresh water has a higher density than that of sea water. This usually only occurs during periods of peak discharge when suspended sediment concentration is very high (Powell, 1981; Mackiewicz *et al.*, 1984; Syvitski *et al.*, 1987). The restricted definition of the term "underflow" provided by Mackiewicz *et al.* (1984) is used throughout this thesis. Underflows can be transformed into turbidity currents if sufficient ambient sea water is entrained by the current. Many researchers (e.g. Sanders, 1965; Gilbert *et al.*, 1990; R. Gilbert, personal communications, 1993) have used the term "underflow" to describe density currents generated by the collapse of sediments on steep slopes. These density currents should be termed turbidity currents rather than underflows according to Mackiewicz *et al.* (1984) and Smith and Ashley (1985). Syvitski (1989) suggested that the textural features such as the upward increase in sorting and sand content used by Mackiewicz *et al.* (1984) to distinguish underflow deposits from other surge-type gravity flow deposits have to be verified in the modern environment because these features can also be formed in slump generated gravity currents. Smith and Ashley (1985) claimed that the lower sandy and silty parts of couplets deposited from underflows in lakes may be normally graded or massive rather than solely inversely graded. Mackiewicz *et al.* (1984) also suggested that underflows formed in a marine basin may be massive or normally graded.

Although fine grained turbidites are common in a variety of environments (Stow and Piper, 1984), they usually exhibit diagnostic features like a sharp basal contact, a preferred sequence of sedimentary structures and normal grading (Stow and Shanmugam, 1980; Stow and Piper, 1984). The lack of both sharp basal contacts and normal grading in the faint laminae of subunit b1 suggests that these are not turbidites.

The faint laminae may have been formed as a result of settling of fine sand and silt particles from overflows (Mackiewicz *et al.*, 1984). Laminae formed by deposition from surface sediment plumes even in areas distant from the source of the fresh water usually have sharp basal contacts and show normal grading (Mackiewicz *et al.*, 1984). The lack of such features may be due to either very low concentration of sediment in the surface sediment plumes (Mackiewicz *et al.*, 1984), or postdepositional reworking by turbidity or tidal currents (Mackiewicz *et al.*, 1984). Reworking breaks the soft clay flocs and resuspends them, forming turbid layers that tend to sort the clay and silt during deposition, generating deposits with diffuse contacts and variably developed lamination (Mackiewicz *et al.*, 1984). Such features are exhibited by the faint laminae of subunit b1 (type L_I).

The second thicker type of laminae in this subunit (type L_{II}) was probably not deposited by suspension settling of particles from surface sediment plumes because this process cannot account for inverse grading. Wave reworking of bottom sediments may produce thin storm beds that are characterized by sharp bases, rare grading or cross lamination, and internal flat laminations (Johnson and Baldwin, 1986). In modern nearby inlets such as Makkovik Bay, thin beds of similar appearance have been interpreted as thin storm beds by Piper *et al.* (1983). Water depth cannot be reconstructed from the deposits of Unit B, so that it is possible that wave height and fetch need not have been great to produce some storm reworking.

L_u laminae may be underflow deposits. Near the point of meltwater entry into the basin (within about 0.5 km according to Mackiewicz *et al.* (1984)), underflows transport sand by traction and form sharp based laminae that may show reverse, normal or reverse-to-normal grading. According to Powell (1981), underflows may persist far from the fresh water source.

Grain flows may also produce laminae similar to L_u, with sharp upper and lower contacts, upward increase in sorting and sand content, and reverse to normal grading (Gilbert, 1983). Grain flows are sediment gravity flows (*sensu* Middleton and Hampton, 1976) supported by dispersive pressure and fluid escape, which are usually associated with sediment failures from steep slopes (Gilbert, 1983). If L_u laminae in subunit b1 consist of grain flow deposits, then the flows probably originated from a nearby source characterized by steep slopes, because dispersive pressure in thin grain flows is unlikely to be maintained for long distances unless sufficient slopes are present (Gilbert, 1983).

Considering the spectrum of possible origins for the laminae in subunit b1, it is concluded that the well defined, thicker laminae were probably deposited either (a) beneath underflows (*sensu* Mackiewicz *et al.*, 1984) emanating from a nearby source (probably a side entry-river), or (b) by grain flows generated by the collapse of sediment in a nearby area on steep slopes (probably nearby steep fjord walls), or (c) by periodic wave reworking of bottom sediments. All these processes may have played a role. The faint laminae were probably formed by settling of fine sand and silt from surface or mid-water sediment plumes. The plumes were characterized by very low silt and fine sand concentration, probably because Section D98 was far from the main fresh-water source to the W-SW. The faint laminae are not interpreted as the deposits of distal underflows because they are so thin, suggesting a dilute suspension. Underflows in a marine setting, in contrast, could not be dilute without losing their positive density contrast with sea water (Mackiewicz *et al.*, 1984).

The upward increase in the proportion of thicker, well defined laminae in subunit b1 suggests a temporal increase in the capacity and/or duration of the depositing currents. This may reflect (a) an increase in fresh water discharge, or (b) an increase in the volume of sediment failures due to the greater amount and rate of sediment deposition, or (c) increased proximity of river outlets that contributed underflows.

The scattered coarse particles that are distributed throughout subunit b1 are a product of either ice or algal rafting (see interpretation of similar particles in Unit A). This interpretation is based on the observation that these coarse particles disturb underlying faint laminae. This type of disturbance is characteristic of rafted debris (Edwards, 1986; Gilbert, 1990).

The muddy material between both types of laminae was probably deposited from surface sediment plumes, as flocs, during periods of low discharge (see massive mud of Unit A). The lack of bioturbation in this mud may suggest rapid rates of sedimentation or simply unfavourable living conditions for burrowing organisms. The lack of a significant number of microfossils (compared with subunit b1 in Section D99, Chapter 4) suggests unfavourable environmental conditions such as reduced salinity of bottom waters or high turbidity (Scott *et al.*, 1984).

3.1.1.3.3. Description of subunit b2

Subunit b2 consists of 1.7 m of interbedded sand and mud that gradationally overlies the lower interstratified sand and mud (sand : mud ratio increases from 1 : 1 to 3 : 1). The thickness of the sand layers increases from less than 1 cm in the lower part to 10-15 cm in the upper part. The thickness of the mud layers ranges from 1-5 cm. These mud layers may be massive of laminated. Colour on fresh wet surfaces is medium light grey (N6) for sand and pale reddish brown (10 R 5/4) for mud. The sand layers consist of fine to very fine, well sorted sand. The sand layers have sharp upper and lower contacts. Most sand beds are massive but some have faint parallel laminae, < 1 cm thick, defined by a concentration of dark heavy minerals.

3.1.1.3.4. Interpretation of subunit b2

Interbedded sand and mud suggests two modes of sedimentation. The sharp contact between each sand bed and the overlying mud suggests that processes that deposited the sand and overlying mud were independent.

Parallel laminated, well sorted sand beds interbedded with mud can form as proximal storm beds (Anderton, 1976). In this case, the sand beds are deposited under upper flow regime plane bed conditions while the mud is deposited from suspension after the storm subsides (Anderton, 1976). The parallel laminated sand in storm deposits is generally overlain by ripple lamination that forms as the storm abates (Blatt *et al.*, 1973; Harms *et al.*, 1982). Ripples do not characterize the deposits of subunit b2, and heavy-mineral lamination is not a typical feature of storm sands.

Parallel laminated sands characterized by dark heavy mineral concentrations are common on beaches where they are formed by wave swash and backwash (Clifton *et al.*, 1971); each lamina is characterized by inverse grading. The absence of the inverse grading in the laminae and the presence of mud interbeds rule out a beach setting for these beds.

Scott *et al.* (1991) interpreted parallel laminated sand in glaciomarine deltaic sand as the result of deposition from pulsating currents. Unlike the laminae observed in this study, individual laminae of Scott *et al.* (1991) were normally graded. Ungraded laminae characterized by the alternation of fine grained dark heavy minerals and light minerals have been interpreted to form by deposition from turbulent suspension currents characterized by fluctuating velocity (Sanders, 1965). However, the exact mechanism for the segregation of heavy and light minerals during deposition from turbulent suspensions is unknown (Slingerland and Smith, 1986).

The massive (structureless), well sorted sand beds may be interpreted as sediment gravity flow deposits, with the texture being inherited. Suitable sediment gravity flows are cohesionless debris flows (*sensu* Postma, 1986), which derive their particle support from grain collision and pore-fluid escape, and liquefied flows (*sensu* Lowe, 1976). The lack of normal or reverse grading and sedimentary structures typical of turbidites (i.e., turbidite divisions of Bouma, 1962) rules out turbidity currents.

The high volume concentration of cohesionless debris flows/liquefied flows would have hindered traction transport, while a narrow grain size range inherited from the source would have limited grain segregation during flow, so that no grading nor lamination was developed (Middleton and Hampton, 1976; Middleton and Southard, 1984). Cohesionless debris flows/liquefied flows probably originated from the collapse of an originally well sorted sand from a nearby source because these flows are expected, from theory and experiments, to have a limited distance of travel (Lowe, 1976).

Alternatively, the structureless sand beds may be underflow deposits formed during high sediment discharge from a nearby source. Although these deposits lack the upward increase in sorting and sand content considered to be characteristic of underflows (Mackiewicz *et al.*, 1984), the sharp contact between each sand bed and the overlying mud suggests that these may be underflows (Smith and Ashley, 1985). Lack of grading at the top of the sand beds should be expected in a marine environment because any muddy material present in the flow would probably separate and move upwards as an interflow or an overflow (Mackiewicz *et al.*, 1984; Smith and Ashley, 1985). Increase in sand bed thickness requires that the underflows increased in discharge with time, due to their source becoming closer to the location of the section. This may have taken place during the progradation of deltas (either side entry of fjord-head). The mud layers between the sand beds were probably deposited from suspension or low concentration density currents.

3.1.1.4. UNIT C

Uppermost Unit C at Section D98 consists of three sets of planar tabular cross bedded sand, each about 15 cm thick. The lower set consists of very fine grained, moderately well sorted sand while the upper two sets consist of coarse to very coarse, moderately well sorted sand. The contacts between the sets are flat and sharp. Foreset strata within the lower set are less than 1 cm thick; in the upper two sets they are 1-2 cm thick. Foresets in each set are straight and dip at an angle of 20°-25° towards the E-NE (three measurements: N28E, N37E, N41E). Colour on fresh wet surfaces is moderate to light grey (N5 to N6). Unit C overlies Unit B with a sharp planar contact and underlies a thin (20 cm) recent vegetation cover.

3.1.1.5. INTERPRETATION OF UNIT C

Planar tabular cross beds with planar foresets dipping in the same direction are usually interpreted to have been deposited by migration of 2D dunes (*sensu* Ashley *et al.*, 1990), beneath unidirectional currents. The thin sets suggest that these were formed beneath shallow flows, or that each set was partly eroded before the deposition of the overlying set.

3.1.1.6. DISCUSSION OF KEY SECTION D98

The sedimentary units exposed at this section suggest a general increase in the energy of the system from the lower massive mud (Unit A) to the interstratified sand and mud (Unit B). Only minor rafting of sand by local winter ice or algae can be advocated. Some sand was probably transported by wind to the surface of winter sea ice, from which it was released when this ice melted. Wind-blown grains are very common in fjord sediments (Syvitski and Hein, 1991).

A deep embayment or a fjord is the most probable depositional environment for Unit A. The source of the fresh-water plume was far away from the area or it was characterized by low discharge or the area was covered by sea ice for long periods. Relative sea level may have been high at the time of deposition of Unit A, following the retreat of glacial ice from the area and marine inundation of the isostatically depressed area.

The lack of dropstones suggests one or a combination of the following conditions (McCabe and Eyles, 1988; Liverman, 1989; Ashley *et al.*, 1991): (a) glacier ice had retreated landwards of the head of the fjord, (b) sea ice existed for long periods restricting the movement of ice bergs, or (c) winds kept any ice bergs away from the area of this section.

During the deposition of the subunit b1, accelerated melting of the inland ice sheet caused an increase in fresh-water supply. Due to isostatic recovery, the cross section of the fjord probably became narrower, so that the velocity of the fresh-water outflow would have increased (i.e., focused flow). Hence, coarse sediments could be transported further seaward. Rebound also would have initiated the migration of the fjord-head delta, and migration of the fjord-side deltas towards the fjord centreline. Suspension settling of clays and silts was initially still the dominant sedimentation process but the surface sediment plumes sometimes carried some fine sand and deposited faint fine sand laminae. Deposition from underflows and grain flows, probably generated from a local source, also occurred during periods of high discharge.

Tidal and wave action may have affected sedimentation of subunit b1 through the resuspension and deposition of some laminae, but since the present tidal range is small (microtidal to low mesotidal and is assumed to have not changed considerably) and the exact basin shape is not known (thus fetch length, wind energy and tidal amplifications are poorly constrained), the effect of wave and tidal action cannot be evaluated. However, waves and tidal currents do influence sedimentation in nearby fjords (Barrie and Piper, 1982; Piper *et al.*, 1983). More prolonged open marine conditions probably existed during the deposition of subunit b1, as suggested by the presence of microfossils in an adjacent section (D99) at approximately the same stratigraphic level.

The increase in the frequency of thin, sharp based sand beds from the lower part to the upper part of subunit b1 suggests an increase in the amount of fresh water discharge and the probable gradual advance of deltaic deposits over fjord-bottom sediments as the area rebounded. Specifically, the river outlets discharging into the fjord became nearer to the area of the section and strata deposited from underflows and other gravity flows become thicker and more distinct. Reduction in the salinity of the basin may have allowed the underflows to form more easily (Syvitski *et al.*, 1987). This trend continues in subunit b2.

The sediments of Unit A and subunit b1 may represent basin muds and delta bottomsets, the upper part of subunit b1 may represent the delta bottomsets-lower delta foresets (Smith and Ashley, 1985). The progradation of a fluvial-deltaic system into the area would then be responsible for the deposition of more sand in the upper part of Unit B (*cf.* Syvitski and Farrow, 1989). Sand may have been initially deposited as mouth bars that prograded onto delta foreset beds that eventually failed when the slope increased above a certain angle. Alternatively, most of the sand beds may have been emplaced by underflows generated at river outlets. The lack of an erosional contact between subunits b1 and b2 (unlike what is observed in nearby sections D96 and D99) suggests that the progradation was gradual in the area of this section.

The cross sets in the upper part of the section were probably deposited in a shallow fluvial channel after the area became emergent. This channel was probably separate from the main fluvial channel that was the site for deposition of thick cross bedded units in associated sections (Unit C in sections D96 and D99, below). Alternatively, these sediments may represent distributary channel deposits of the delta plain that were preserved from later erosion by the presence of nearby bedrock outcrops, immediately upstream and downstream of Section D98.

3.1.2. ASSOCIATED SECTION D99

This section is located approximately 3.5 km downstream of Section D98 (Fig. 3.2c). It is the furthest downstream section described in this chapter. Units A, B and C (Fig. 3.7) are exposed in an approximately 16.5 m high terrace.

3.1.2.1. DESCRIPTION OF SEDIMENTS IN SECTION D99

Units A and B are similar to those at Section D98, except that the contact between subunit b1 and b2 is sharp and planar to slightly irregular (across a 1 m-wide trench); 2-5 cm-thick sand beds appear suddenly above the laminated sand and mud of subunit b1. Sand beds are either massive with sharp upper and lower contacts or normally graded from sand to mud with sharp lower contacts. In the upper part of subunit b2, the sand beds contain 1-2 cm soft irregular mud clasts consisting of material similar to the mud interbeds between the sand beds.

A slab tray (T99B) was taken from the upper part of subunit b1. Fresh water algae *Pediastrum*, rhizopoda and *Botryococcus*/Diterma are present in this sample whereas marine and fresh water/estuarine dinocysts are rare (Chapter 4). A significant number of benthic foraminifera tests were separated from a subsample from the upper part of tray T99B (Chapter 4).

Unit C is 7.4 m thick and sharply overlies Unit B with an erosional contact marked by irregular scours (across a 1 m-wide trench). This unit is divided into two subunits. Subunit c1 is 4.4 m thick and consists of cross beded, fine to very coarse sand with sets ranging in thickness from 5-50 cm (Fig. 3.8). A sample from one of the sand beds in the lower part of this subunit (S99X in Appendix B) shows that the beds consist of fine to medium sand (mean = 1.91 ϕ), well

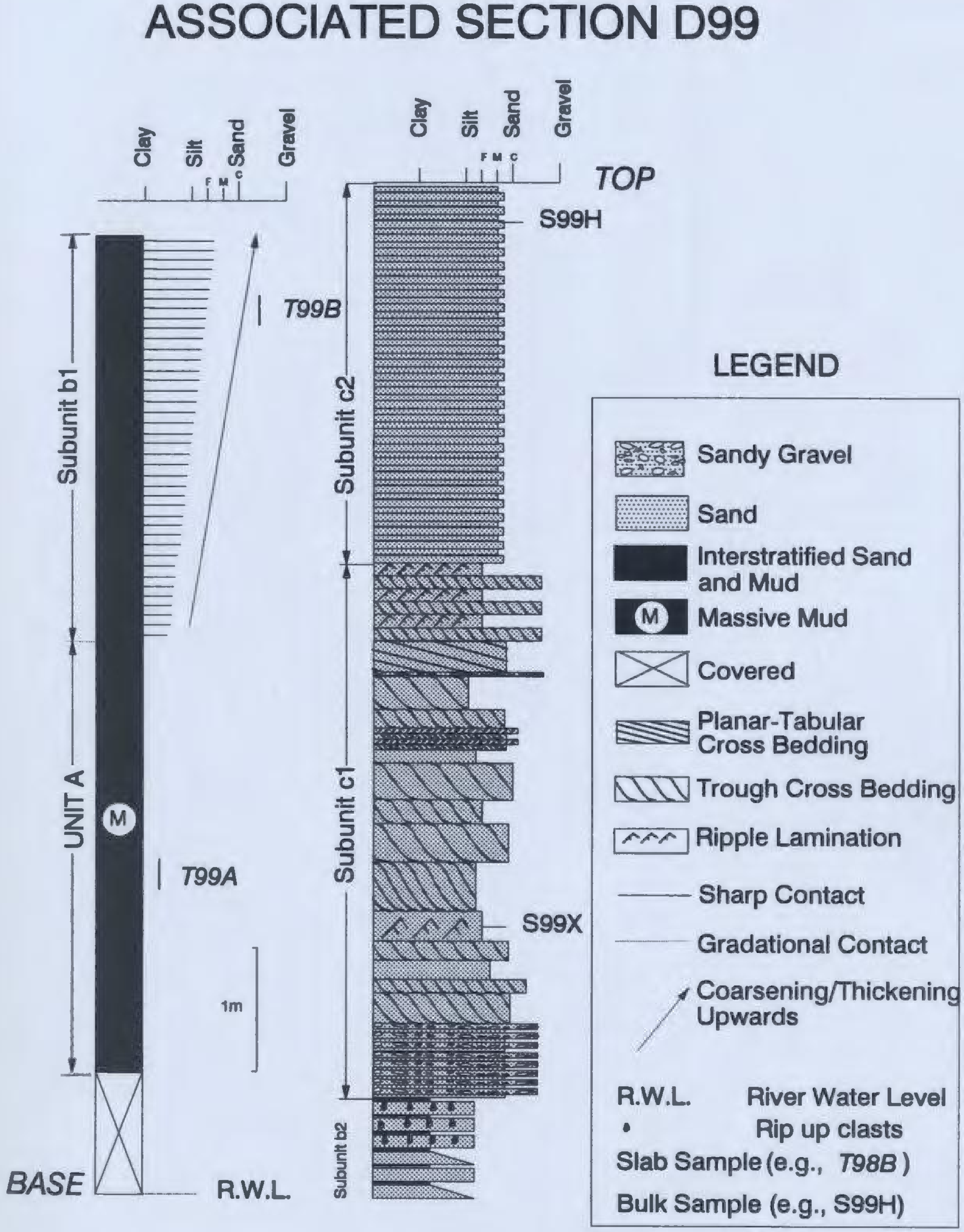


Fig. 3.7. Associated Section D99 stratigraphic column. See text for details.

Fig. 3.8. Subunit c1 in Section D99. Note the different geometries of the cross bedded sand beds and the variable direction of dip of the foreset laminations. Divisions on the bar scale= 10 cm.

sorted (σ = 0.49 ϕ) and near symmetrically skewed (Sk = 0.07). The cross beds in this subunit show variable palaeoflow direction (varying by as much as 100°) but most dip towards the NE. Some 5-7 cm-thick cross bed sets dip towards the SW. Foresets have either angular or tangential contacts with the lower bounding surface of the set. Set contacts are planar to curved.

Subunit c2 is approximately 3 m thick and consists of parallel stratified medium to coarse sand. Each bed is 2-5 cm thick and is differentiated from surrounding beds by a variation in grain size or by the proportion of heavy minerals. A sample from one of the sand beds in the upper part of this subunit (S99H in Appendix B) shows that the beds consist of coarse to medium sand (mean = 0.89 ϕ), moderate to well sorted (σ = 0.71 ϕ) and near symmetrically skewed (Sk = 0.03). Contacts between the beds are sharp and planar.

3.1.2.2. INTERPRETATION OF SEDIMENTS IN SECTION D99

The massive mud of Unit A is interpreted in the same way as Unit A in Section D98; i.e., the sediment was deposited from suspension fall out of fine silt and clay particles from surface sediment plumes either below winter sea ice or at areas distal from the river outlets. The coarse sand particles were probably transported by algal rafting or by wind transport to local winter sea ice and subsequent release by melting.

For subunit b1, the different types of laminae are interpreted to have been deposited by underflows, grain flows, wave reworking, and suspension settling from surface plumes (see § 3.1.1.3.2). Microfossils present in the upper part of this subunit suggest a surface layer of fresh or brackish water (algae and dinoflagellates) and normal marine bottom waters (benthic foraminifera). The occurrence of microfossils indicates conditions of reduced water turbidity and, probably, reduced sediment supply, either from surface plumes or from bottom currents (Scott *et al.*, 1984).

Sediments of subunit b2 are interpreted to have been deposited by turbidity currents, liquefied flows, grain flows and underflows. Beds that are sharp based and normally graded are interpreted as the deposits of (a) turbidity currents that originated by collapse of sediments on steep slopes or (b) underflows generated at river outlets. Mud caps at the top of the graded beds are interpreted to have been deposited from the tails of the turbidity currents or from suspension. Massive sand beds are interpreted to have been deposited from liquefied flows/cohesionless debris flows (*sensu* Postma, 1986) or underflows. Mud clasts near the base of some beds were eroded from semi-consolidated mud by the density currents. Laminated mud interbeds were probably deposited from suspension settling from surface plumes or low concentration turbidity currents. The sharp contact between subunit b2 and subunit b2 suggests that part of subunit b1 may have been eroded or that sedimentation rates were highly variable. The thinness of subunit b2 in this section compared to Section D98 and its sharp erosional contact with Unit C suggest that part of subunit b2 also may have been eroded.

The lower part of Unit C (subunit c1) suggests deposition under conditions of variable velocity and direction (as indicated by the variable grain size and foreset dip directions). Deposition was probably from migrating 2D and 3D dunes. The size of the dunes is unknown because the bedforms were probably truncated by erosion. Opposed cross bed sets suggest flow reversals such as those found in herringbone cross bedding formed by tidal currents. However, the lack of three-dimensional exposure does not permit the unambiguous recognition of herringbone cross bedding. Thus, cross beds in this subunit suggests deposition dominated by variable flow speeds and directions. These may have been in multiple shallow channels oriented in different directions. Alternatively, sets showing dip in opposite directions may have been deposited by tidal currents that were enhanced in areas of channels or basin constriction (Nichols and Biggs, 1985). The lack of any information on the tidal regime in the area, the lack of 3-D outcrop, and the poorly known configuration of the basin (for enhancement of tidal currents) makes a tidal interpretation speculative.

The parallel bedded sand of subunit c2 may have been deposited by the swash and backwash of waves on a beach (Clifton *et al.*, 1971). Alternatively, these may represent low angle aeolian deposits formed by wind ripple migration (Glennie, 1970). Determination of which interpretation is correct is difficult, although both processes may operate on sandy beaches. Marine sediments are not present at this stratigraphic level in nearby sections, whereas an aeolian interpretation is consistent with the mapping of Fulton (1986).

3.1.2.3. DISCUSSION OF ASSOCIATED SECTION D99

The environmental significance of units A and B is the same as at Section D98. Subunit c1 may have been deposited by a river system that was established in this area after emergence. This river system was characterized by variable current speed and channel switching; the channel pattern may have been braided or "wandering" (Brierley, 1989). This river system may have eroded whatever upper delta foresets and delta plain deposits that were present at the top of Unit B. Alternatively, the sediments of subunit c1 may represent distributary channel deposits of the delta topsets. It is unknown if the cross-bed sets that dip up-valley were deposited by reverse currents or if they formed by infilling of large scour troughs. Opposite dipping cross bedding have been reported point bars and fluvial deposits outside sharp river bends (Taylor *et al.* 1971). Tidal currents may have been responsible for the deposition of parts of subunit c1, with the cross bed sets that dip up-valley being deposited by flood currents. This tidal hypothesis is not supported by any other field data, and therefore remains speculative.

Subunit c2 may represent a beach deposit formed by wave swash and backwash, but the 3 m thickness is believed to exceed what would be expected for beach growth during rapid regression suggested by the sea level history of the area (Chapter 2). It is more likely that subunit c2 represents low angle aeolian sheet deposits formed by wind ripple migration. Winds may have been present for long periods and mobile surficial sand was probably readily available.

3.1.3. ASSOCIATED SECTION D96

This section is located approximately 3.0 km upriver from Section D98. Exposed sediments are approximately 13.1 m thick. Only units B and C are present (Fig. 3.9).

3.1.3.1. DESCRIPTION OF SEDIMENTS IN SECTION D96

Unit B is approximately 4.25 m thick. It is lithologically very similar to Unit B in Section D98. Sixty centimetres from the top of the unit there is an abrupt change from sand beds approximately 1 cm thick to sand beds 3-5 cm thick. Mud laminae between the sand beds are 0.5-1 cm thick throughout the unit.

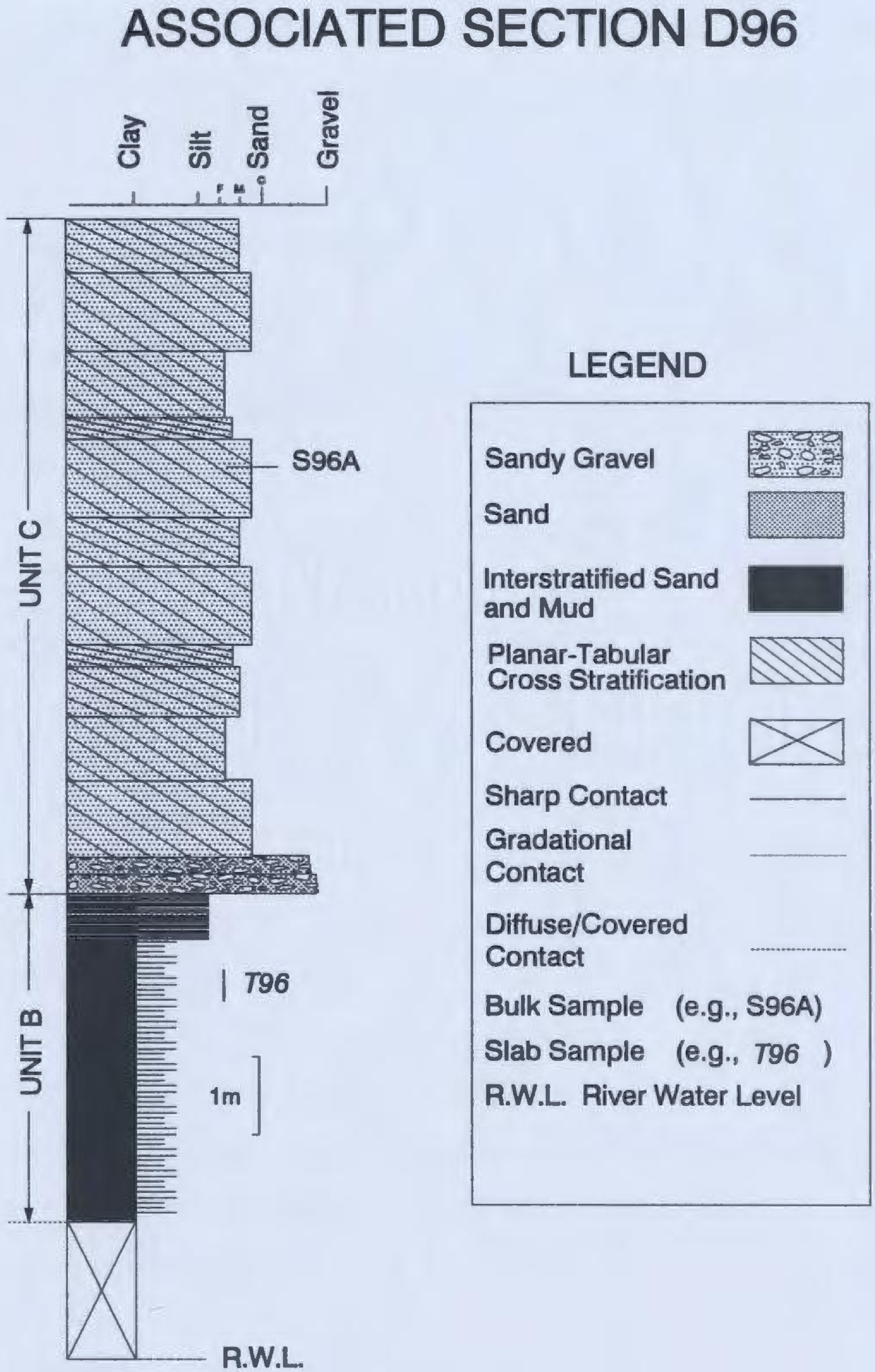


Fig. 3.9. Associated Section D96 stratigraphic column. See text for details.

Unit C is approximately 8.7 m thick. It is different than Unit C in Section D98 in that it consists mainly of stacked sets of planar tabular cross bedded sand that sharply overlie Unit B. At the base of Unit C, there are two 25 cm-thick pebbly gravel layers (clasts 2-5 cm) that are present across the width of the outcrop (1 m wide trench). The matrix of these gravel beds is coarse sand. The gravel beds are clast supported to matrix supported. The rest of Unit C consists mainly of medium to very coarse sand (sample S96A in Appendix B was taken from one of the cross bed sets). Sets vary in thickness from 30-100 cm; foreset strata vary in thickness from 1-5 cm. Foresets have an angular contact with the lower bounding surface of the set. The sets have a dip of 20°-30° towards the SE. Foreset dips are all within ± 10° of the mean palaeoflow direction.

Within the stacked sets of cross bedded sand there are 10-20 cm intervals of cross laminated sand with preserved ripple forms. The ripples are 10-15 cm long and 3-5 cm high. The foreset laminations in these ripples dip in a direction opposite to the dip of the foresets in the thick cross bed sets (Fig. 3.10). Some of these ripples are similar to type A ripple-drift cross lamination of Jopling and Walker (1968) in which only stoss side laminae are preserved.

3.1.3.2. INTERPRETATION OF SEDIMENTS IN SECTION D96

The sediments of Unit B are interpreted to have been deposited by a combination of suspension settling from sediment plumes, turbidity currents, cohesionless debris flows/liquefied flows and underflows (see interpretation for Section D98 for details). The sharp upward increase in grain size and thickness of sand beds suggests that the density currents (probably underflows generated at the river outlets) either became stronger or more focused at this site. This suggests that the stream building the subaqueous delta may have changed direction. Fluctuation and channel switching are very common in proglacial and recently deglaciated areas (Ashley, 1975).

The planar tabular cross beds of Unit C are interpreted to have been deposited by the migration of two dimensional dunes (*sensu* Ashley *et al.*, 1990). The original size of the dunes is not known, since the thickness of preserved sets only provides a minimum estimate of the dune height (Harms *et al.*, 1982). Thick sets (100 cm thick) were certainly deposited by the migration

Fig. 3.10. Photograph of back-flow ripples (migrating to the left at mid-point of the scale) at the base of a cross bedded sand set of Unit C in Section D96. Large sets dip to the right (base of the trench). Divisions on the scale = 10 cm.

of large dunes (*sensu* Ashley *et al.*, 1990). The ripples that migrated in a direction opposite to the larger dunes probably represent back flow ripples that formed in the zone of flow separation downstream from dune crests (Blatt *et al.*, 1973; Allen 1980, 1982). Type A ripple cross lamination indicates that deposition from suspension was minimal (Jopling and Walker, 1968).

The lower two gravel beds probably represent channel bottom lag deposits. Their sand matrix may have been deposited later as a result of infiltration of sand particles into the gravel bed as the flow waned (Smith, 1985).

3.1.3.3. DISCUSSION OF SECTION D96

The sequence described in this section records two distinctly different depositional conditions. The first is the deposition of mud and laminated sand and mud from (1) surface sediment plumes, interrupted by (2) turbidity currents, liquefied flows/cohesionless debris flows and/or underflows. As the basin shallowed due to emergence, the progradation of a fluvial deltaic system transported more and coarser sediment to the area. The cross bedded sands forming Unit C probably represent channel fill deposits of a river system that may have eroded part of the underlying deposits in the area of Section D96. This river may have been the same river that fed the delta that formed in the area, so that these cross bedded sands may represent delta plain deposits. Alternatively, these cross bedded sands may represent channel fill deposits that formed later after the area emerged. The greater thickness of these deposits compared to Section D98 suggests that a larger or deeper channel formed here.

3.1.4. KEY SECTION D90

This section is located approximately 11 km upstream of Section D96. It was measured in an approximately 15 m-high terrace. Unit B is partly exposed in the lower 10.4 m of this terrace while Unit C, 4.65 m thick, forms the upper part (Figs. 3.2c, 3.11). Unit C is differentiated from Unit B by the lack of mud interbeds.

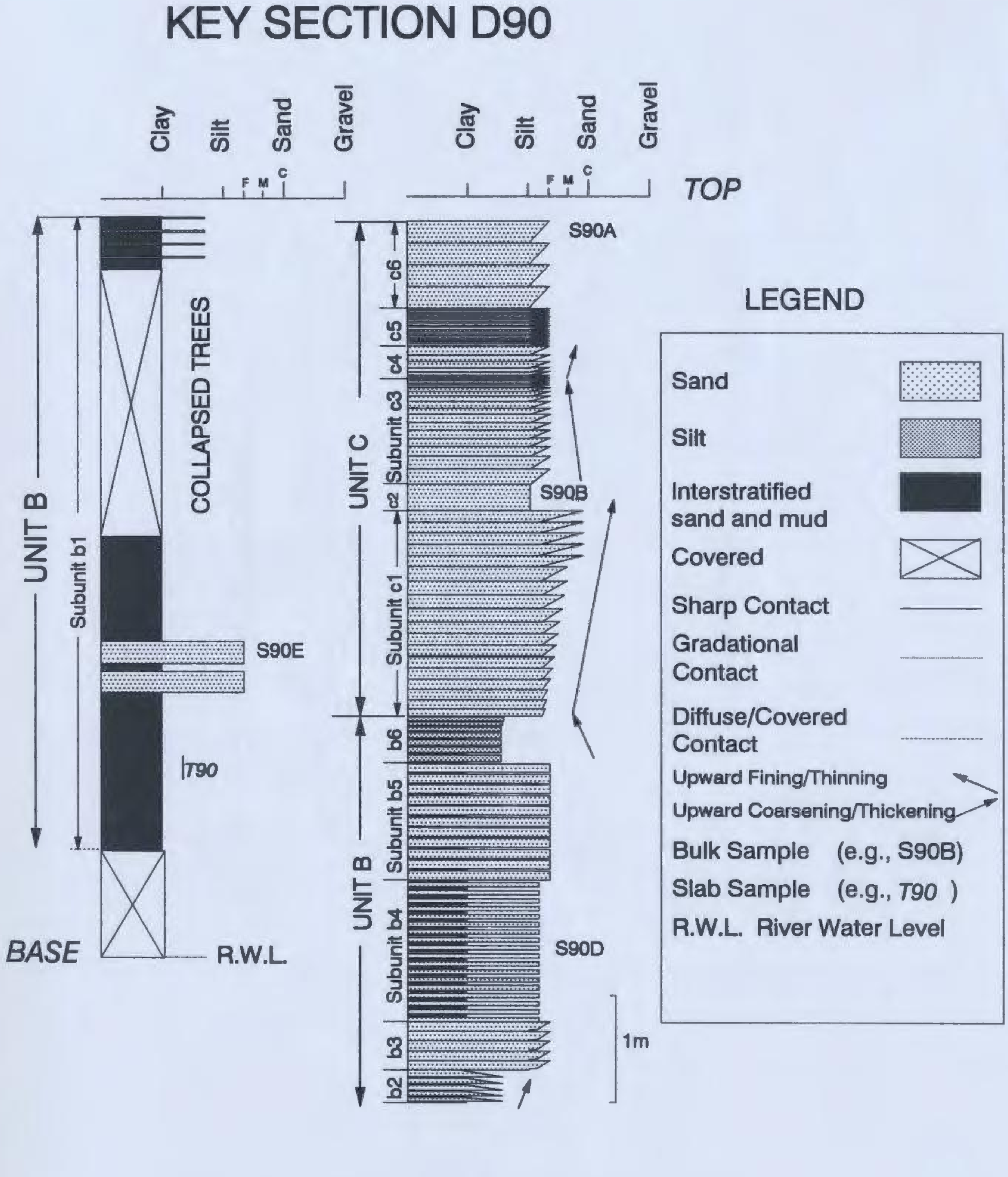


Fig. 3.11. Key Section D90 stratigraphic column. See text for details.

3.1.4.1. UNIT B

Unit B in Section D90 is divided into two main parts: a lower part consisting mainly of interstratified sand and mud (subunit b1), and an upper part consisting of thin to medium bedded sand interbedded with mud (subunits b2 to b6).

3.1.4.1.1. Description of subunit b1

This subunit is approximately 6 m of mainly interlaminated sand and mud, partly covered. The lowermost exposed part of this subunit lies approximately 1 m above the modern river level. There is an approximately 2.5 m interval with its base 3.0 m below the top of this subunit which could not be trenched due to the presence of collapsed trees. Thin sections and X-radiographs show a variety of laminae similar to the laminae in other sections through Unit B, but two main types of laminae dominate. The first type is assigned to L_I (Table 3.2). The mud that surrounds the fine sand and silt in these laminae is structureless and ungraded. Laminae of the second type are normally graded and assigned to L_{III}; this type is more common in the upper part of subunit b1 where it includes 1-1.5 cm-thick silt-fine sand strata. Sample 90F (Appendix B) contains more than 16% of material finer than 13 ϕ. This sample has a high proportion of silt (48.53%) and a very small proportion of sand (2.78%).

In the middle of subunit b1 are two beds (20 cm each) of massive sand, separated by 10 cm of laminated mud (mainly type L_{III}). Both the bases and tops of these beds are sharp. Colour of the sand on fresh wet surfaces is medium dark grey (N4). Grain size analysis of one of the sand beds (S90E in Appendix B) shows a fine sand (mean = 2.34 ϕ), moderately well sorted (σ = 0.56 ϕ) and fine skewed (Sk = 0.14).

3.1.4.1.2. Interpretation of subunit b1

The finer, thinner laminae (L_I) that are characterized by diffuse upper and lower contacts are interpreted to have formed as a result of suspension settling from fairly dilute surface sediment plumes or low density turbidity currents. The finer muddy material that surrounds the sand and silt is interpreted to have been deposited as larger flocs (Gilbert, 1983) as the fine silt and clay settled from suspension. The thicker, sharp based and normally graded laminae (L_{1n}) are interpreted to have been deposited either by turbidity currents, underflows or from moderately concentrated surface fresh-water sediment plumes. These interpretations are not mutually exclusive as the two types of laminae may have been formed as a result of suspension settling of fine sand, silt and clay from variably concentrated surface sediment plumes, periodically interrupted by turbidity currents and underflows. More than one point source may have contributed sediment.

The two sand beds in the middle of subunit b1 were probably deposited by high density turbidity currents, liquefied flows (Lowe, 1976), cohesionless debris flows (*sensu* Postma, 1986; Syvitski and Farrow, 1989) or underflows formed during times of high discharge. The lack of grading probably rules out deposition by turbidity currents (Lowe, 1976; Postma, 1986; Myrow and Hiscott, 1991).

3.1.4.1.3. Description of subunits b2-b6

These subunits total approximately 3.5 m thick and consist of interbedded sand and mud. Colour of the sand and mud on fresh wet surfaces is medium dark grey (N4) and moderate reddish brown (10 R 4/6), respectively. The nature of the sand and mud beds is variable in terms of grading and contacts. The lowest subunit b2, 30 cm thick, consists of thin, normally graded silt - mud couplets. The silt : mud ratio varies from 2.5 : 1 in the lower part to 8 : 1 in the upper part of this subunit. The lower contacts of silts are sharp and planar. Thicknesses of individual silt beds increase from 2-3 cm in the lower part to 3-5 cm in the upper part of this subunit. The mud interbeds decrease in thickness from 1 cm in the lower part to 0.5 cm in the upper part.

Subunit b3 is a 40 cm interval of approximately 10-cm thick beds, each of which coarsens up from silt - very fine sand in the lower part to fine to medium sand in the upper part. These silt-sand beds are separated by thin (1 cm thick) mud beds. The sand to mud ratio in this subunit is approximately 10 : 1. Upper and lower contacts of the silt-sand beds are sharp and planar.

Subunit b4 is a 130 cm interval with a sand : mud ratio of 4 : 1. The sand beds are mainly about 2 cm thick, massive, fine to medium grained and are separated from the silty, massive mud laminae (0.5 cm thick) by sharp, planar contacts.

Subunit b5 is 110 cm thick and is similar to subunit b4, except that the sand beds are thicker (2-5 cm) and medium grained, with a sand : mud ratio of approximately 3 : 1. The silty mud intercalations are 1-1.5 cm thick.

Subunit b6 is a 40 cm interval of interbedded silt and mud in which the silt beds decrease in thickness from 5 cm in the lower part to 2-3 cm in the upper part. Some silt beds grade into mud, but others have sharp contacts. The silt : mud ratio ranges from 3 : 1 to 5 : 1.

3.1.4.1.4. Interpretation of subunits b2-b6

For these subunits, beds that are characterized by normal grading from sand to mud are interpreted to have been deposited from turbidity currents or underflows. Sequences that show an upward increase in the thickness of the sand beds may suggest increasing flow, or local progradation, or aggradation of the basin floor (see below). Interbedded massive sand and mud with sharp contacts suggest deposition of the sand and mud by two independent processes (Ashley, 1975). The lack of internal laminations or grading and the sharp upper and lower contacts do not suggest a turbidity current origin. Instead deposition from liquefied flows (Lowe, 1976; Middleton and Hampton, 1976; Myrow and Hiscott, 1991) or underflows (Smith and Ashley, 1985) is suggested. The mud beds were probably deposited by low concentration turbidity currents or from suspension.

Finally, inversely graded sand beds with sharp contacts that alternate with mud are interpreted to have been deposited by either grain flows or underflows. Features such as an upward improvement in sorting in the sand beds, used by Mackiewicz *et al.* (1984) to support an underflow origin, were not noted in the field, so the interpretation remains equivocal.

3.1.4.2. UNIT C

This unit is approximately 4.65 m thick. It is differentiated from Unit B by the absence of mud layers. Colour of the sand on fresh wet surfaces in the subunits of Unit C is medium light grey (N6) to medium grey (N5). Subunit c1 is a lower 190 cm interval of sand beds that increase in thickness from 2-3 cm in the lower part to 10-15 cm in the upper part. Bed contacts are sharp and planar. Individual sand beds show inverse grading from very fine to fine sand in the lower part of subunit c1. In the upper part, the sand beds coarsen upward from fine to coarse sand. Subunit c2 is 25 cm of very fine sand; upper and lower contacts are sharp and planar. Grain size analysis (S90B in Appendix B) shows a very fine sand (mean = 3.48 ϕ), poorly sorted (σ = 1.04 ϕ) and fine skewed (Sk = 0.14). Subunit c3 is a 100 cm interval of sand beds that decrease in thickness from 5-10 cm in the lower part to 1-2 cm in the upper part. Contacts between the sand beds are sharp and planar. Individual sand beds coarsen upward from very fine sand to fine-medium sand. Subunit c4 is a 30 cm interval of thinly bedded sand in which sand beds increase in thickness from 1-2 cm in the lower part to 5 cm in the upper part. The sand beds are inversely graded from very fine to fine sand. Subunit c5 is a 35 cm interval of thinly bedded, upward coarsening sand beds that show no trend in bed thickness. Individual beds coarsen upward from very fine sand to fine-medium sand. Contacts are sharp and planar. Finally, subunit c6 is a 85 cm interval of medium bedded sand (20 cm beds). Beds coarsen from very fine sand in the lower part to fine to medium sand in the upper part. Grain size analysis (S90A in Appendix B) from the upper part of one of the sand beds shows a fine sand (mean = 2.2 ϕ), moderately sorted (σ = 0.74 ϕ) and fine skewed (Sk = 0.29). Contacts are sharp and planar. Subunit c6 forms the top of the section and underlies a thin recent vegetation cover.

3.1.4.3. INTERPRETATION OF UNIT C

Most sand beds within this unit consist of sharp based, inversely graded beds arranged in upward thickening and thinning cycles of different thicknesses. Inverse grading is attributed to sediment gravity flows such as grain flows and high density turbidity currents (Middleton and Hampton, 1976; Hiscott and Middleton, 1979). Inverse grading may also form as a result of deposition from underflows characterized by increasing discharge (Mackiewicz *et al.*, 1984). These alternatives are discussed in the next section of the thesis.

The unique 25 cm-thick bed of massive very fine sand (subunit c2) was deposited so rapidly from suspension that tractional structures could not develop (*cf.* Middleton and Hampton, 1976; Arnott and Hand, 1989). Alternatively, it may have been liquefied after deposition and lost all primary fabric. Sediment of this size is particulary susceptible to liquefaction (Andresen and Bjerrum, 1967; Lowe, 1976; Myrow and Hiscott, 1991).

3.1.4.4. DISCUSSION OF SECTION D90

The sedimentary units in this section are interpreted to represent deposition in a fjord that became progressively narrower as relative sea level fell due to isostatic rebound and as a fluvial-deltaic system prograded over the area. During the initial phase of sedimentation (lower part of Unit B), probably at high relative sea level, the area of the section was covered by a broad water body. The effectiveness of the inlet in funnelling the water and sediments was minimal and the sediments in the plumes were of lower concentration due to the widespread area of the plume. Sedimentation was dominated by suspension setting from surface-water sediment plumes, producing different types of laminae as a result of variation in sediment concentration in the surface plume or different sources for these sediment plumes. Turbidity currents may have also been an important method of sediment transport and deposition. Large scale sediment failures were common, depositing relatively thicker sand beds from liquefied or cohesionless debris flows (*sensu* Postma, 1986). These processes suggest unstable steep slopes in the area.

Land emergence because of the drop in relative sea level caused most of the sediment to be funnelled through a progressively more narrow fjord or channel. Turbidity currents, underflows, grain flows, liquefied flows and cohesionless debris flows became more important during the deposition of the upper part of Unit B. Progradation of a fluvial-deltaic system brought more sediments to the area. The different sediment gravity flows may have deposited their loads beyond the mouths of the distributary channels as lobes similar to chute lobes observed in modern fjords (Prior and Bornhold, 1989), or the sands may have been deposited from underflows generated by river outflow. No core data are available from modern chute lobes, but progradation of these lobes might produce thickening upward sequences while filling of small channels might produce thinning upward sequences (analogous to elements of the submarine fan model of Mutti and Ricci Lucchi, 1972). These deposits may represent delta bottomsets and lower delta foresets (Smith and Ashley, 1985).

During the deposition of Unit C, progradation and channel filling may have continued producing upward thickening and thinning cycles, respectively. The lack of mud beds between the sand beds may reflect erosion of the mud beds by the gravity flows (turbidity currents, underflows, grain flows) or no time between successive flows for mud to accumulate from suspension. These processes were greatly affected by the narrowing of the marine embayment in the area of Section D90. These deposits may represent middle delta foreset deposits (Smith and Ashley, 1985). The limited exposure did not permit the assessment of the dip of the beds.

Alternatively, the upper part of Unit B (subunits b2-b6) and Unit C may be interpreted as levees formed at the margin of a subaqueous channel, based on the presence of thinning and thickening upward trends, and analogy with fluvial channels (Collinson, 1986). Levees form as sediment-laden water overfloods a channel, depositing sand during periods of peak discharge and mud from suspension. The thin mud interbeds in subunits b2-b6 are interpreted to have been deposited from suspension in a subaqueous environment (implying that these are subaqueous levees). As the area experienced further isostatic rebound, the levees may have become subaerial, with no mud deposited from suspension (Unit C in the upper part of this section). The upward thinning and thickening trends observed in fluvial levees (Collinson, 1986) are attributed to (a) lateral encroachment of the levees as a result of channel migration or (b) fluctuation in the discharge. Levees have been observed along some modern fjord-bottom channels (Syvitski and Farrow, 1983).

Progradation, rather than levee dynamics, is favoured here because (a) levees in modern fjords are characterized by ripples and ripple lamination which are absent in the sediments of subunits b2-b6 and Unit C, and (b) the modern river in the area of Section D90 is surrounded by bedrock, so that lateral migration of a channel to produce the upward thinning and thickening sequences is unlikely to have occurred. The processes responsible for the deposition of Unit C are similar to the upper part of Unit B in this sections. The sand beds are assigned to Unit C only because they lack the mud interbeds that are present between the sand beds in the upper part of

Unit B.

Sediments interpreted as delta plain or fluvial deposits that form the top of the Quaternary sequence in sections upstream and downstream of this area were not observed at Section D90. These fluvial sediments may be present in the terrace that is present at a higher elevation than the sequence described in section D90. Unfortunately, this terrace is covered by vegetation and trenching was not possible there.

3.1.6. ASSOCIATED SECTION M88

Although this section is located in the middle part of the area it is described here because it is the nearest section upstream of key Section D90. This section is located approximately 5 km upstream of Section D90. Sediments exposed at this section are 5.5 m thick in an approximately 5.8 m high terrace. Units B, C and D are present (Fig. 3.12).

3.1.6.1. DESCRIPTION OF SEDIMENTS IN SECTION M88

The lowest exposed sediments in this section are assigned to Unit B. The base of the exposure is approximately 0.3 m above the river water level. The lower 15 cm consists of interstratified sand and mud. Sand strata are 0.5-2 cm thick and consist of sand and silt separated by 0.5 cm layers of massive mud. Some sands are sharp based and show normal grading from fine sand-silt into the overlying mud while others are massive and have sharp upper and lower contacts. No vertical trends in the thickness of the sand or mud strata were observed. This interstratified sand and mud is sharply overlain by a 90 cm interval of interbedded sand and mud with a higher sand percentage. The sand beds consist of fine to very fine sand. They have sharp bases (scoured to planar). Some show normal grading into the overlying mud while others are massive and have sharp upper and lower contacts. The sand beds are 8-10 cm thick while the mud beds are 1-2 cm thick.

Unit C consists of four sand beds with a cumulative thickness of 260 cm. The lower three beds total 130 cm thick (each 40-50 cm thick). They consist of massive, very fine to medium grained sand with sharp planar upper and lower contacts. Beds are differentiated by variations

ASSOCIATED SECTION M88

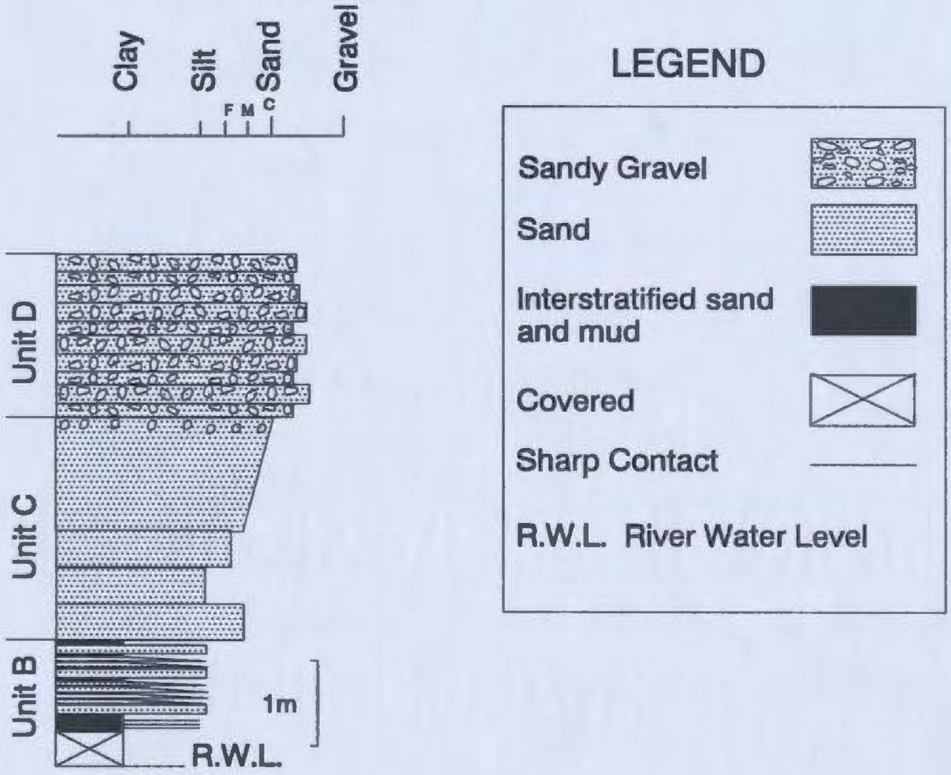


Fig. 3.12. Associated Section M88 stratigraphic column. See text for details.

in grain size (Fig. 3.12). The upper bed is 130 cm thick. It has a sharp planar base and is inversely graded from medium sand to very coarse sand to granules (0.2-0.4 cm). Granules are rounded to subrounded and constitute less that 10% of the upper 30 cm of this bed.

Unit D sharply overlies Unit C with a planar to slightly irregular contact (across 1 m wide trench). It is 185 cm thick and consists of crudely stratified gravel beds each 15-20 cm thick. Gravel beds consist of rounded pebbles (2-4 cm) in a matrix of very coarse sand. Beds are matrix supported to clast supported. Bedding planes are distinguished by variations in the amount or size of pebbles. This unit is overlain by 10 cm of recent vegetation cover.

3.1.6.2. INTERPRETATION OF SEDIMENTS IN SECTION M88

Unit B formed in the same way as Unit B at Section D90 (turbidity currents, overflows, underflows, grain flows and liquefied flows). The lower three sand beds of Unit C were probably deposited rapidly from gravity flows because they lack tractional structures. The fine grain size and lack of grading suggest that they are probably liquefied flow deposits. The lack of mud interbeds suggests that either the mud was eroded or that these beds were deposited in rapid succession. The sharp basal contact and inverse grading of the upper bed suggest that it was deposited from a grain flow that was probably generated by a failure of a sediment mixture of coarse sand and some granules.

The gravel beds of Unit D are believed to represent bed load deposits formed in shallow channels during periods of high discharge (Church and Gilbert, 1975; Hein, 1984). The clasts may have been transported individually or as diffuse gravel sheets (Hein, 1984). The variations in clast size in each bed suggest fluctuations in the flow strength. The sand matrix was probably formed as a result of sand percolation into the interstices during falling stages of flow.

3.1.6.3. DISCUSSION OF SECTION M88

The sediments exposed in this section are broadly similar to the sediments exposed in the other nearby sections and record the transition from fjord bottom-delta bottomset sediments to probably delta plain deposits as a result of a progradation of a fluvial system into the fjord area during isostatic rebound. The lower interstratified sand and mud were probably deposited by turbidity currents, and suspension settling when the source was far from this section and when the basin was probably deeper. These may represent fjord bottom sediments or delta bottomsets.

As water covering the site became shallower, thicker sand beds were deposited from liquefied flows, underflows, possibly as lower-upper delta foresets (upper Unit B, Unit C). Part of the lower interstratified interval of Unit B may have been eroded by the first of these flows.

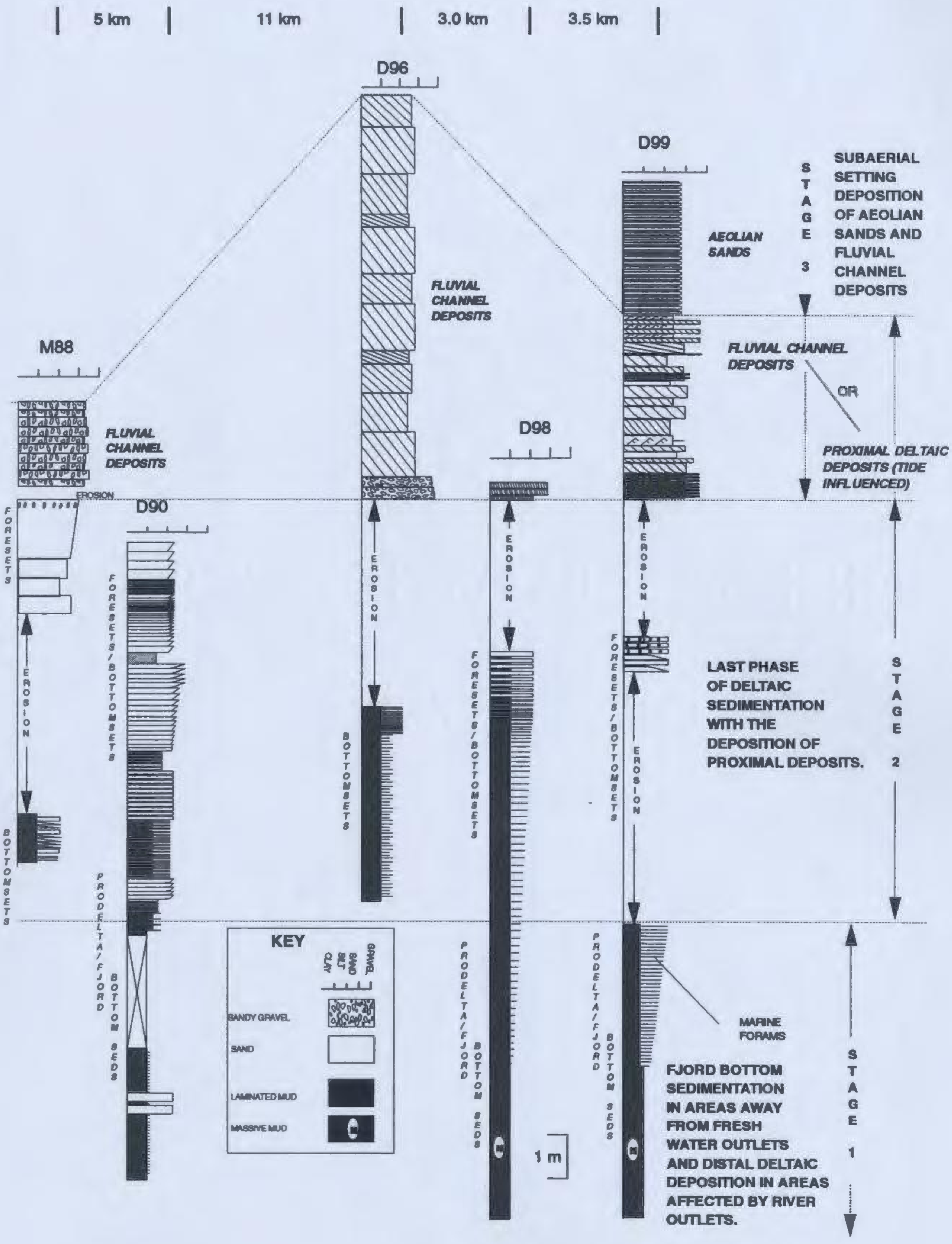
Unit D at the top of the sequence was probably deposited by a fluvial system that was established as the area became emergent. The gravelly nature of the beds suggests that this fluvial system was of high energy, and the variation in the clast size between the beds suggests fluctuating discharge. The fluvial system may have eroded part of Unit C and formed thin gravel bars in the area of this section. Alternatively, these sediments may represent fluvial channels on the delta top.

3.1.7. SEDIMENTARY ENVIRONMENTS IN THE DOWNSTREAM PART OF THE AREA

Hypothetical stages of sedimentation are shown in Figure 3.13. Each stage represents deposition of sediments under similar conditions. These stages are based on the vertical sequences observed at each section. Note that this figure is not a correlation chart. There is no time constraint on the different stages, although the later depositional stages probably occured earlier in the sections located nearer to the prograding fjord head delta (i.e., at Section D90). The stages have no chronological significance.

During deglaciation, marine water may have been in contact with the termini of glaciers. Marine sediments deposited in front of the glaciers would be expected to contain ice-rafted debris, allowing them to be classified as glaciomarine sediments. Such sediments were not observed in the area, although they may be present below the exposed sections. Because no ice-rafted debris is present in the marine muds in the thesis area, they are not termed as glaciomarine deposits in the thesis. Some researchers (e.g. Molnia, 1983, 1989) use the term glaciomarine more broadly to include sediments introduced to the depositional environment by fluvial transport, by ice

Fig. 3.13. Hypothetical stages during which the different sedimentary units of the downstream sections were deposited. Each stage represents deposition under relatively similar conditions in the different areas where sections occur. Positioning of the sections is an attempt to line up deposits formed in similar settings. Where erosional contacts are present, the columns have been split and the unconformity-bounded units have been independently positioned. The thickness of each erosional gap however, is intended to indicate neither duration nor depth of erosion. Stages do not have any chronological connotation because the lateral relationships between the sections are unknown. Sediments within each stage may be considered as a "sediment association" reflecting deposition in broadly similar conditions (Liverman, 1989). Some of these stages may be localized in only certain sections.


5 km
11 km
3.0 km
3.5 km
D96
D99
M88
D98
D90
FLUVIAL CHANNEL DEPOSITS
AEOLIAN SANDS
STAGE 3
SUBAERIAL SETTING DEPOSITION OF AEOLIAN SANDS AND FLUVIAL CHANNEL DEPOSITS
FLUVIAL CHANNEL DEPOSITS
OR
PROXIMAL DELTAIC DEPOSITS (TIDE INFLUENCED)
EROSION
FORESETS
FORESETS/BOTTOMSETS
BOTTOMSETS
PRODELTA/FJORD BOTTOM SEDS
LAST PHASE OF DELTAIC SEDIMENTATION WITH THE DEPOSITION OF PROXIMAL DEPOSITS.
STAGE 2
KEY
GRAVEL
SAND
SILT
CLAY
SANDY GRAVEL
SAND
LAMINATED MUD
MASSIVE MUD
MARINE FORAMS
1 m
FJORD BOTTOM SEDIMENTATION IN AREAS AWAY FROM FRESH WATER OUTLETS AND DISTAL DELTAIC DEPOSITION IN AREAS AFFECTED BY RIVER OUTLETS.
STAGE 1

rafting or aeolian transport. This thesis uses a more restricted definition of the term "glaciomarine", requiring the presence of an ice-rafted component which is characteristic of glaciomarine deposits (Flint, 1971). The different difinitions of glaciomarine sediments are discussed by many others (Andrews and Matsch, 1983; Molnia, 1983; Brodzikowski and Van Loon, 1987). Fjords receiving their sediments from glacial meltwater after the glaciers have receded onto land have been termed outwash fjords by Powell (1981, 1983, 1984).

In stage 1, sedimentation was dominated by deposition from surface sediment plumes that entered a fjord or an embayment. There are two hypotheses for the deposition of Unit A in sections D98 and D99. Either Unit A was deposited from sediment plumes below winter sea ice in a frigid fjord, or from far-travelled surface sediment plumes characterized by low sediment concentration from which clay- and silt-sized particles were deposited as flocs. Coarse grains were contributed by wind transport, perhaps to the surface of winter ice, and possibly by algal rafting.

It is possible that Unit A in sections D98 and D99 was deposited at the same time as the lower part of Unit B in the other sections in the downstream part of the area (Fig. 3.2c). The high percentage of silt in Section D90 suggests a location closer to the source of the fresh-water plumes. All these mud-rich sediments may represent deposition under the highest relative sea level recorded by the sediments in this part of the study area (although sea level may have been even higher earlier).

Figure 3.14 shows the inferred shape of the basin at 8 ka. In such a setting, all the river outlets were located well away from the locations of the sections, so that when their sediment plumes reached the area of these sections they carried only fine silt and clay. Sections D98 and D99 are to one side of the embayment, where they may have been isolated from the sediment gravity flows that distinguish Unit B from Unit A. Relative sea level may have been still rising, stationary, slowly dropping or even rapidly dropping during the deposition of Unit A.

In stage 2, sedimentation was still dominated by deposition from surface sediment plumes, at least during the initial deposition of Unit B. Sediment plumes reaching the downstream area probably had a progressively higher sediment concentration causing the transition from Unit A

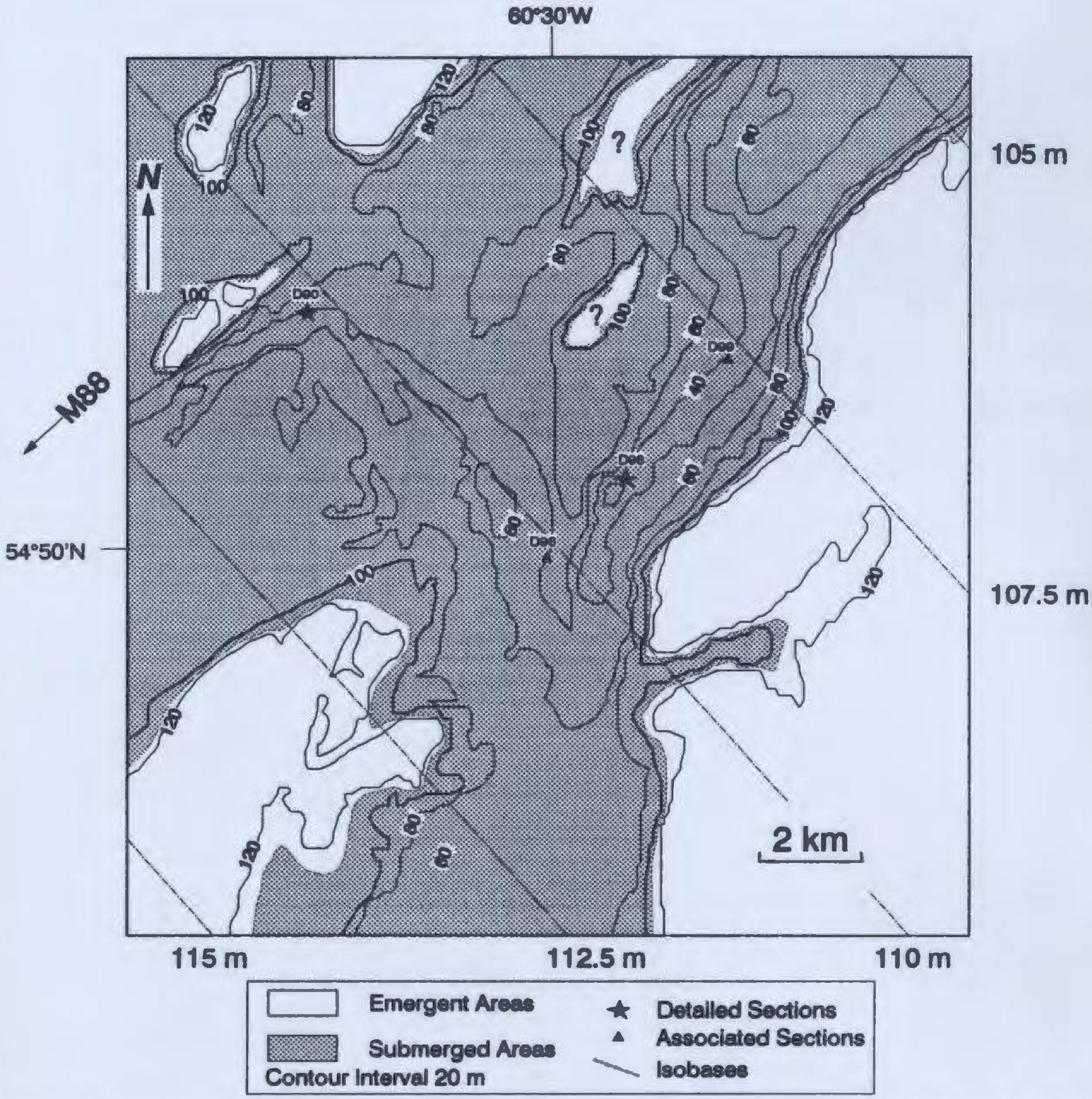


Fig. 3.14. Suggested shape of the basin or embayment in the downstream part of the study area at 8 ka. Shoreline position is based on the marine limit estimates for the entire area and the inferred crustal deflection (towards the southwest). Note that the distances between the isobases are equal, because it is assumed that the crustal deflection increases at a constant rate towards the SW. No data are available on the flexural rigidity of the crust in this area, but it probably does not vary over such as small area. Location of the key and associated sections is also shown. Note the modern contours for areas above 120 m have been omitted. Ice may have been present in emergent areas but its position is unknown.

to Unit B in sections D98 and D99. This increase in concentration may have been due either to increased melting of inland ice or progradation of the fluvial deltaic system as a result of isostatic rebound, or both. The coarsening upward trend in Unit B is attributed to the same causes.

The lower parts of Unit B in all the sections probably represent delta bottomset deposits. Several deltas may have been forming as a result of multiple entry points into the basin. Deposition from sediment gravity flows, dominated by turbidity currents, was an important factor. Grain flows and liquefied flows deposited massive and inversely graded sand strata. Underflows may have also occurred during periods of high discharge and salinity reduction. Marine microorganisms briefly inhabited the area of Section D99 during the deposition of the middle part of Unit B. This may have been due to better exchange of marine waters in the area or short periods of reduced sediment supply.

In the area of Section D90 (Fig. 3.14), sediments carried by the river flowing from the west were probably building a delta. Sediment gravity flows (grain flows, liquefied flows and turbidity currents) or underflows formed channel mouth lobes characterized by thick sand beds interbedded with mud, forming upward thickening and thinning sequences. Such lobes might be similar to the lobes in modern fjords and lakes that form at the mouths of chutes on the delta slope (Smith and Ashley, 1985; Prior and Bornhold, 1989). In the area of Section D90, the lower part of Unit B might represent fjord bottom sediments or delta bottomsets while the upper part of Unit B and Unit C might represent the transition from bottomsets to delta foresets.

During the deposition of the upper part of Unit B and Unit C in Section D90, the area of sections D96, D98 and D99 was probably in slightly deeper water. Deposition at these sections may have been affected by multiple sources causing the development of variable trends in the thicknesses of the beds. In the area of sections D96 and D99, the upward increase in the thickness of the sand strata in the upper part of Unit B is not regular, suggesting that the river system that supplied the sediments to the deltas in this area was characterized by channel directional changes and avulsion (Ashley, 1975). In Section D98, the sand beds in the upper part of Unit B show a gradual upward increase in thickness, suggesting that this phase of deposition took place without channel switching. This river system may have been different than the river system forming the deltas at the other section.

In stage 3, continued isostatic rebound and the progradation of the fluvial deltaic system over the area caused the deposition of progressively coarser material. If the assumptions on the rate of emergence (§ 2.2.3) are correct, then by 7 ka only a narrow part of the downstream area would have been submerged. The emergence would have changed the morphology of the basin considerably (compare Fig. 3.14 with Fig. 3.15), and the main river system would have delivered all its load into a narrow embayment.

During the deposition of Unit C in the area of sections D96, D98 and D99, depositional conditions were also variable. Part of Unit B in these sections was probably eroded before the deposition of Unit C. At Section D96, the cross beds forming Unit C were probably deposited in a deep fluvial channel. At Section D99, Unit C exhibits a variety of cross bedded sand beds indicating variable current speed and direction. These may have been deposited by a braided or wandering river system. The cross beds that show an up-valley current direction (although only observed in a two dimensional outcrop) may have been deposited by tidal currents in the embayment. The modern tidal range in Makkovik Bay is microtidal to low mesotidal, but enhancement of tides by changes in the size of a water body may cause temporal changes in the importance of tidal currents and their energy (e.g. Amos, 1978). Alternatively, these oppositely dipping cross beds may have been deposited by variable wind-generated marine currents generated in the different small embayments near Section D99 (Fig. 3.15).

The upper part of Unit C is characterized by planar bedded sand that may have been deposited at a beach or as aeolian sands. The beach interpretation is speculative. Fulton (1986) interpreted these sands as aeolian from aerial photographs.

In the area of Section D98, Unit C is relatively thin and Unit B is relatively thick (compared to sections D96 and D98). The thinness of Unit C suggests that only a small channel may have formed in this area, preserving more of Unit B. The presence of nearby bedrock outcrops may have prevented the establishment of a large fluvial channel near Section D98.

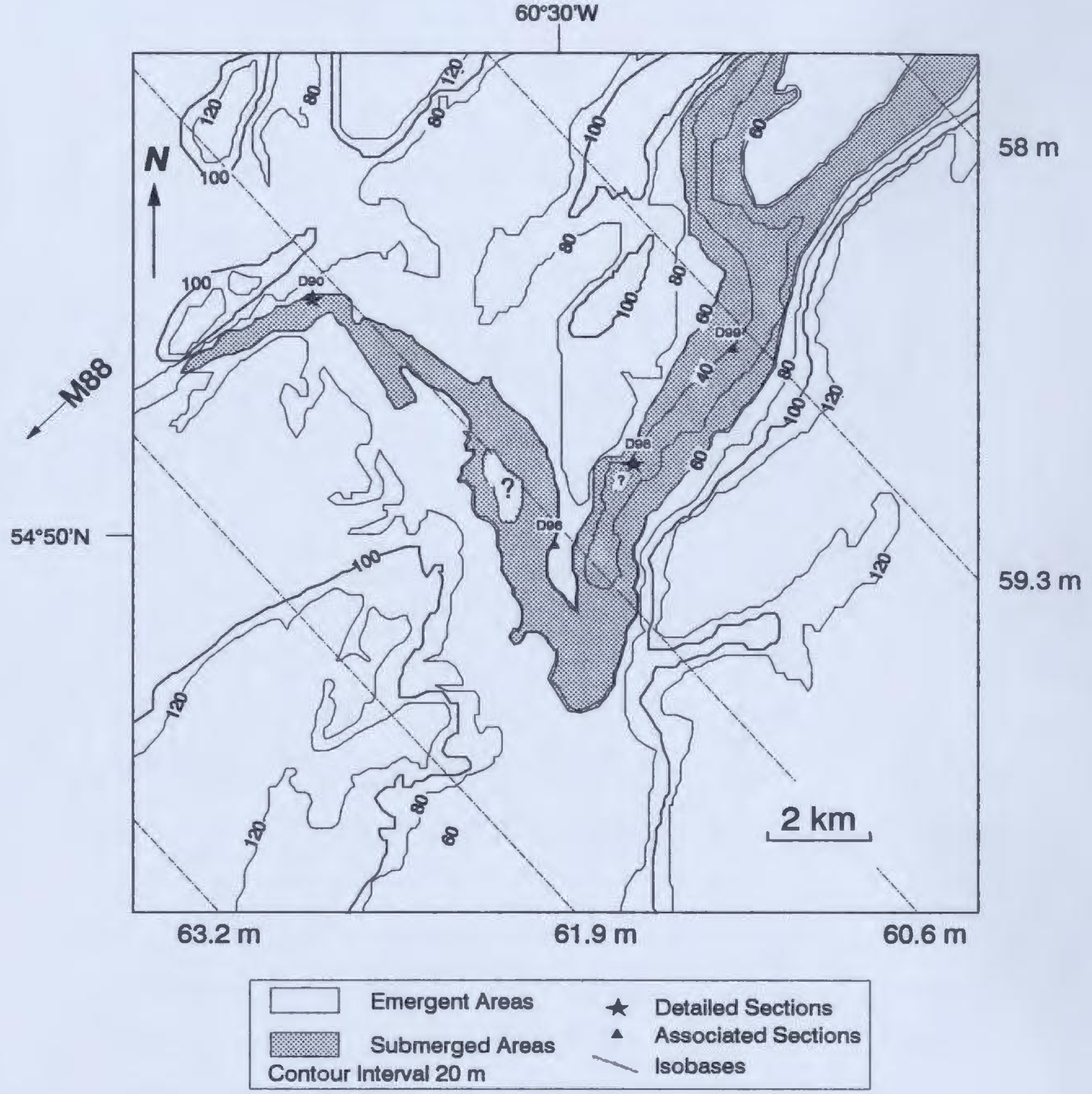


Fig. 3.15. Suggested shape of the basin or embayment in the downstream part of the study area at 7 ka. Shoreline position is based on the marine limit estimates for the entire area and the inferred crustal deflection (towards the southwest). The distances between the isobases is constant based on the assumption that the rate of crustal deflection did not vary considerably over this area. No data are available on the flexural rigidity of the area. Location of the key and associated sections is also shown. Note the modern contours for areas above 120 m have been omitted. Ice may have been present in emergent areas but its position is unknown.

3.2. SECTIONS IN THE MIDDLE PART OF THE AREA

Two sections, M61 and M85, are described in detail while features of associated sections M66, M68 and M75 are described briefly. The sedimentary environments of the middle part of the thesis area will be assessed after all sections have been described and interpreted.

3.2.1. KEY SECTION M85

This section is located 3.0 km upstream of Section M88. Only units B, C and D are present in this section. Unit B forms most of the exposed section (Fig. 3.16).

3.2.1.1. UNIT B

Interstratified sand and mud of Unit B exhibit four upward thinning cycles. Each cycle changes from interbedded sand and mud in the lower part to interlaminated sand and mud in the upper part. In the lower two cycles, the upward thinning trend is interrupted by anomalous sand beds or deformed intervals underlying sand beds. These anomalous beds are defined as beds that do not follow the upward trend in bed thickness or grain size. Each of the upward thinning cycles overlies an interval of deformed sand and mud. Colour of the sand beds/laminae on wet fresh surfaces ranges from light grey (N7) to medium light grey (N8) while the colour of the mud ranges from pale red (5 R 6/2) to moderate reddish brown (10 R 4/6).

The basal part of the section has a 90 cm interval of interbedded sand and mud that lies 50 cm above the modern river level. The sand beds are 15 - 20 cm thick. Based on one representative grain size determination (S85L; Appendix B), they consist of very fine sand (mean = 3.27 ϕ) that is poorly sorted (σ = 1.08 ϕ) and fine skewed (Sk = 0.21). Contacts between the sand beds and the overlying and underlying mud beds are sharp and irregular. The sand beds are massive and ungraded. The interbedded mud layers are silty, massive and do not grade from the sand beds below.

Overlying this basal interbedded sand and mud is a 35 cm interval of deformed sand and mud consisting of small (5-10 cm wide and 7-15 cm high) folds (isoclinal and overturned). Axial planes strike approximately NNE-SSW and dip towards the SE. Above this deformed interval is

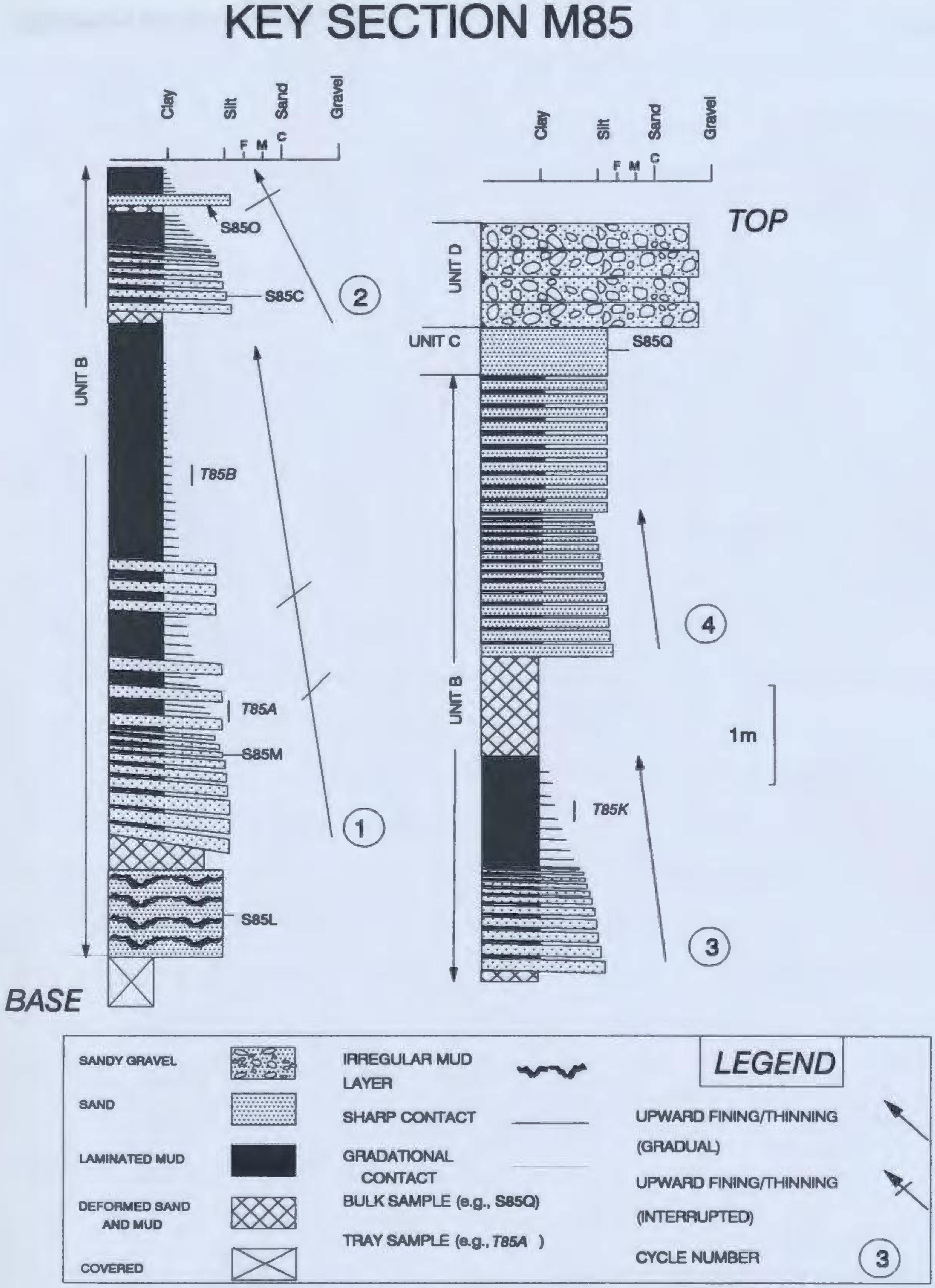


Fig. 3.16. Key Section M85 stratigraphic column. See text for details.

the first of the four upward thinning cycles (including two anomalous intervals), approximately 4.8 m thick. In this cycle, the sand beds show a general upward decrease in thickness in the lower 1.2 m from 25 - 20 cm to 1 cm. A sample from a typical sand bed in the lower part of this cycle (S85M; Appendix B) shows that the beds consist of fine to very fine sand (mean = 3.04 ϕ), well sorted (σ = 0.47 ϕ) and fine skewed (Sk = 0.14). These sand beds are massive and contacts between individual sand beds and the overlying mud beds are sharp and planar while the lower contacts are sharp and slightly irregular. These beds dip 5°-8° NNE.

Next in the stratigraphy is the first anomalous sand interval, which consists of three 15 cm-thick fine grained sand beds separated by 5 -10 cm-thick intervals of interstratified sand and mud. The base of each sand bed is sharp and slightly irregular, while the upper contact of each sand bed is sharp and planar. Thin sections made from a slab tray sample of the muddy interbeds (T85A; Fig. 3.16) show that the laminae are of types L_{II}, L_{III} and L_{IV} (Table 3.2). The mud separating these laminae varies from massive to graded to burrowed. Grain size of three bulk samples taken from one of the muddy intervals (samples S1, S2 and S3 in Appendix B) is 3.1% - 13.9% sand, 23.4% - 28.9% silt and 61.7% - 73.6% clay.

The upper 2.9 m of the cycle (including the second anomalous interval) consists mainly of interlaminated sand and mud in which the sand laminae decrease in thickness upward from approximately 10 mm to 1 - 2 mm. A slab tray (T85B; Fig. 3.16) taken from this interval shows that lamina types L_I and L_{III} (Table 3.2) are present. Two bulk samples from this tray (S13 and SM, Appendix B) have a sand : silt : clay ratios of 11.1 : 61.3 : 27.6 and 81.5 : 9.3 : 9.2. The sandier sample SM is a very fine sand (mean = 3.43 ϕ), moderately well sorted (σ = 0.58 ϕ) and fine to strong fine skewed (Sk = 0.3). Sample S13 had more than 16% material finer than 13 ϕ. Fresh water algae, *Pediastrum*, rhizopoda, fresh water dinocysts and estuarine dinocysts are present but rare in this interval (T85B).

The second anomalous interval consists of three 20 cm-thick, massive, fine grained sand beds, each separated by 10 cm of laminated mud. The upper contacts of the sand beds are sharp and planar while the lower contacts are sharp and slightly irregular.

The second upward thinning cycle is 1.5 m thick. It overlies a 20 cm interval of deformed sand and mud like that at the base of cycle 1. The sand beds of this second cycle change in thickness from 15 - 20 cm in the lower 0.75 m to interlaminated sand and mud in the upper 0.75 m. The sand beds have sharp, planar upper contacts and sharp and slightly irregular lower contacts. These beds dip in the same direction as the beds in cycle 1 but at a lower angle of 5°-3°. Grain size analysis of one of the sand beds (S85C; Appendix B) shows a very fine sand (mean = 3.29 ϕ), moderately sorted (σ = 0.85 ϕ) and strongly fine skewed (Sk = 0.33). In the upper part of the upper muddy part of this cycle is a 15 cm-thick sand bed that overlies 20 cm of deformed sand and mud similar to the other deformed units. Grain size analysis of a sample from this bed (S85O; Appendix B) shows a very fine sand (mean = 3.12 ϕ), moderately well sorted (σ = 0.67 ϕ) and fine skewed (Sk = 0.14).

The third upward fining sequence is 2.3 m thick (Fig. 3.16). The lower part consists of approximately 1 m of interbedded sand and mud in which the sand beds change in thickness upward from 20 cm to 1 cm. The remainder of the cycle is interlaminated sand and mud. Thin sections made from a slab tray taken from the interlaminated sand and mud in the upper part of this cycle (T85K; Fig. 3.16) show that three types of laminae or thin beds are present (L_I, L_{II} and L_{III}; Table 3.2); L_{III} is rare. Grain size analyses of three bulk samples (S16, S17 and S18 in Appendix B) reveal 1.6% - 26.9% sand, 25.4% - 71.2% silt and 27.2% - 47.8% clay. These beds also dip in the same direction as cycles 1 and 2 at lower angles (< 3°). Fresh water algae, *Pediastrum* and Rhizopoda are rarely present in the tray sample (T85K). Some *Brigantedinium* cysts are present while fresh water and estuarine cysts are common (T85K).

Overlying the third upward thinning cycle is a 1.0 m-thick interval of deformed sand and mud. The composition of the deformed sand and mud is similar to the sediments in the upper part of the underlying cycle. The deformation is similar to that below the first upward thinning cycle.

The fourth and last cycle consists of a 1.45 m-thick unit of interbedded sand and mud. It overlies the deformed interval with a sharp planar to irregular contact. The sand beds change in thickness from 20 cm in the lower part to 2 cm in the upper part. They are separated by approximately 2 cm-thick layers of massive silty mud. The sand beds dip in the same direction as the sand beds in the underlying cycles.

The top 1.4 m of Unit B consists of 10-15 cm-thick sand beds separated by 2-5 cm intervals of laminated sand and mud. The contact between this 1.4 m interval and cycle 4 is sharp and erosional. Sand beds in this interval are fine grained and have sharp upper and lower contacts. The laminated sand and mud consists of 0.5-1 cm-thick, sharp based fine sand laminae separated by 0.5-0.8 cm-thick laminae of mud. Contacts between the sand and mud laminae are either sharp or gradational.

3.2.1.2. INTERPRETATION OF UNIT B

Sand beds in each upward thinning cycle are massive and have sharp, slightly irregular basal contacts. These common features suggest a recurring depositional process. The massive nature of the beds suggests rapid deposition. These sand beds are interpreted to have been deposited by sediment gravity flows such as cohesionless debris flows (*sensu* Postma, 1986), liquefied flows (Lowe, 1976) or high density turbidity currents, grain flows or underflows (Gilbert, 1983; Mackiewicz *et al.*, 1984). The lack of sedimentary structures typical of turbidites (i.e. Bouma sequences) may suggest rapid deposition from immature turbidity currents (Middleton, 1967), but the fine grain size and the lack of grading suggest instead that these are liquefied flow deposits, or underflow deposits formed during periods of high discharge. Liquefied flows may have originated by the collapse of sediments on steep slopes. Moderate to good sorting may have been inherited from the failed deposits.

The deformed beds that are present at the base of upward thinning cycles are interpreted as slide deposits linked to upslope failures (Coleman and Prior, 1980). The failures may have generated gullies or scars that influenced subsequent deposition.

The laminated and thinly bedded muddy intervals that form the upper part of the upward thinning sequences were probably deposited by a combination of underflows, overflows or turbidity currents. A different origin may be considered for each of the four types of laminae observed in Unit B.

Laminae of type L_1, with their fine grain size, low sand content and diffuse upper and lower contacts, were probably deposited by settling of grains from surface sediment plumes. The low sand content and the thin nature of these laminae suggest that the sediment concentration in these plumes was low, perhaps due to the low fresh water discharge or a distal source of these sediment plumes (Powell, 1981; Mackiewicz *et al.*, 1984).

Laminae and thin beds of the second type (L_{II}) are interpreted as the deposits of underflows or grain flows. This interpretation is based on their inverse to normal grading, sharp upper and lower contacts and high sand content (Gilbert, 1983; Mackiewicz *et al.*, 1984; R. Gilbert personal communications, 1993). Laminae of the third type (L_{III}) are interpreted to represent either turbidity current deposits, probably generated by the collapse of sediments upslope, or proximal overflows deposited from surface sediment plumes generated at a nearby source or during periods of high discharge. This interpretation is based on the sharp lower and graded upper contacts of the laminae, normal grading, and the upward decrease in sorting in each lamina (Mackiewicz *et al.*, 1984). The fourth type of laminae (L_{IV}) is interpreted as the deposits of grain flows, underflows, or proximal overflows based on sharp upper and lower contacts and high sand content (Gilbert, 1983; Mackiewicz *et al.*, 1984; R. Gilbert, personal communications, 1993). The lack of normal grading suggests that these laminae were deposited from grain flows or underflows rather than proximal overflows; although grain flows are generally reverse to normally graded some may be massive (R. Gilbert, personal communication, 1993). These beds may also be interpreted as liquefied flow deposits but their thin nature would require that they were formed near the site of sediment failures so that excess pore-fluid pressures could be maintained.

The sand beds in the upper part of Unit B are interpreted as the deposits of sediment gravity flows similar to those that deposited the sand beds in the lower part of each upward thinning cycles. The dip of the beds in the lower cycles indicates deposition on originally inclined surfaces. The upward decrease in the angle of dip of the beds in the cycles suggests that each depositional surface became gentler with time. The dip of these beds and the dip of the axial planes of folds in the deformed intervals suggest palaeoslopes to the NNE.

In summary, Unit B in Section M85 contains massive sand beds probably deposited from liquefied flows, cohesionless debris flows or underflows (*cf.* Syvitski and Farrow, 1989; Syvitski and Hein, 1991). These flows were either generated by the collapse of originally well sorted sediments from a nearby source or formed during periods of high discharge. The laminated mud that dominates this unit was probably deposited by suspension settling of sediments from surface sediment plumes characterised by variable sediment concentrations interrupted by turbidity currents and other sediment gravity flows.

3.2.1.3. UNIT C

This unit consists of a single 50 cm-thick bed of very pale orange (10 YR 8/2) to yellowish grey (5 Y 7/2) fine sand (mean = 2.44 ϕ) that is well sorted (σ = 0.57 ϕ) and near symmetrically skewed (Sk = 0.088) (sample S85Q, Appendix B). The lower and upper contacts of this bed are sharp and irregular. The sand bed is structureless.

3.2.1.4. INTERPRETATION OF UNIT C

This sharp based, massive sand bed is interpreted to have been deposited rapidly from a single sedimentary gravity flow or from suspension. The lack of mud at the top of the bed suggests either that it was eroded when the overlying unit was deposited or that mud was not available for deposition. The presence of an erosional contact at the top of this bed suggests that the former interpretation is more appropriate. The flow was likely a cohesionless debris flow (*sensu* Postma, 1986) or liquefied flow that re-mobilised an originally well sorted sediment.

3.2.1.5. UNIT D

Unit D is 1.0 m thick and overlies Unit C with a sharp irregular contact (across a 1 m wide trench). This unit consists of crudely stratified gravel beds. Each bed is 20-30 cm thick and is bound by planar contacts. Beds are distinguished by the variation of the clast size of each bed. Clasts have mean diameters of 10-15 cm, and are rounded to very well rounded. Within each bed, the clast population is moderately sorted. The beds are clast supported to matrix supported. The clasts form 40 - 70% of the beds. The matrix consists of coarse sand. Clast fabric could not be determined due to the collapse of the sediments in the trenches.

3.2.1.6. INTERPRETATION OF UNIT D

Based on the clast size, crude stratification and lack of cross bedding, each gravel bed is interpreted to have been deposited as bed load in a shallow channel during a period of high discharge (Church and Gilbert, 1975; Hein, 1984). The gravel was probably transported as individual clasts or as diffuse gravel sheets during periods of high water and sediment discharge (Hein, 1984). The sand matrix in these gravel beds was probably formed when sand percolated into the interstices during falling stages of flow (Rust, 1978; Smith, 1985).

These deposits are similar to the longitudinal bar deposits formed in the proximal reaches of braided rivers and outwash plains (Gustavson, 1975; Hein, 1984; Rust, 1984; Rust and Koster, 1984; Smith, 1985). A proximal setting is suggested by the coarse clast size, diffuse bedding planes and the lack of both cross bedding and sand interbeds that are more common in the medial and distal parts of braided rivers (Hein, 1984). The lack of features such as inverse grading and greater thickness of individual beds precludes a sediment gravity flow origin for these gravel beds (Hein, 1984).

3.2.1.7. DISCUSSION OF SECTION M85

This section is unique in that it exhibits several upward thinning cycles characterised by a change from interbedded sand and mud to interlaminated sand and mud. The lower cycles dip at an angle of 3°-8° towards the NNE; the dip angle decreases in the upper cycles. These upward thinning sequences are not observed in the adjacent section (M88) which instead is characterized by interlaminated to thinly interbedded sand and mud. This suggests that the processes of deposition in the area of Section M85 were local in their influence.

Thick sand beds in the lower part of each sequence indicate frequent sediment failures on steep slopes (Lowe, 1976; Postma, 1986), or prolonged and vigorous underflow during periods of high discharge from the river outlets supplying sediments to this area. The laminated and interstratified intervals in the upper parts of the cycles were deposited by a combination of sedimentary gravity flows and suspension fallout from surface sediment plumes. Based on these features and the dip of the beds, Unit B in Section M85 is interpreted to represent delta front-prodelta deposits formed as a result of failure of fluvially derived sediments that were prograding in a NNE direction. The development only locally of upward thinning cycles suggests that the delta was formed at the terminus of a side entry river. On aerial photographs, meltwater channels are present to the northwest of the location of the section. These channels are not observed in the immediate vicinity of the section because younger alluvial sediments cover this area (Fulton, 1986). The sand in Unit B was probably sorted and deposited as river mouth bars that subsequently collapsed down the delta slope (Syvitski and Farrow, 1989). Some sands may have been emplaced by underflows during periods of high discharge.

The upward thinning sequences probably indicate either medium to long term decrease in the discharge of the river feeding the delta or migration of delta front channels (Syvitski and Hein, 1991; Hein and Syvitski, 1992) or both. During periods of low discharge, sedimentation was dominated by suspension fallout from surface sediment plumes punctuated by grain flows, turbidity currents and underflows. The presence of a local source of fresh water is supported by the presence of laminae types L_{Π} and $L_{\Pi I}$, interpreted as proximal deposits of overflows and underflows (criteria of Mackiewicz *et al.*, 1984).

Marine waters and a surface fresh water layer were present during the deposition of part of the sequence as indicated by microfossils (mainly palynological evidence; see Chapter 4). The conditions may have been unfavourable for benthic forams due to frequency of sediment failures (Scott *et al.*, 1984).

Unit C is the deposit of a single sediment gravity flow. Its thickness (50 cm) might be anomalous, or might reflect more advanced progradation of the fluvial-deltaic system.

Finally, a fluvial system developed in the emergent area. This fluvial system was characterised by high sediment discharge as indicated by the coarse grain size of Unit D (mainly coarse pebbles and cobbles). This gravel was probably deposited as bars in shallow channels characterized by successive high-discharge events. The river may have been a braided or a wandering river (Brierley, 1989). The material transported in the channels was probably the glacial and glaciofluvial material that is widespread in the area. This gravelly fluvial system may have eroded part of the underlying delta sequence (delta topsets).

3.2.2. ASSOCIATED SECTIONS M75 AND M68

Section M75 is located 3.8 km upstream of Section M85 and 0.2 km downstream of Section M68. Section M75 contains units B, C and D while Section M68 contains units B and D (Fig. 3.17). Unit B coarsens upward at both localities. Units B and D are not significantly different than in Section M85 (except that Unit B does not contain the sandy packets that thin upward) and are not described here. The main reason for including these sections here is to described the microfossils that they contain, and to point out some interesting features of Unit C in Section M75.

3.2.2.1. DESCRIPTION OF ASSOCIATED SECTIONS M75 AND M68

Unit C in Section M75 is approximately 1 m thick and consists of normally graded and massive sand beds that are inclined at an angle of 5°-8° towards the NE. Graded beds fine upward from medium to fine-very fine sand. Massive beds are medium to fine grained. Both types of sand beds have sharp planar to slightly irregular (scoured) contacts. The basal and upper contacts of Unit C are sharp and erosional.

The microfossils found in Unit B in both sections are of particular environmental importance. A small number of benthic foraminifera tests were separated from the lower part of Unit B in Section M75 (17 *Elphidium excavatum* f. *clavata* and 4 *Cassidulina reniforme*; sample T75B) while only 4 tests were present in the upper part of the unit (4 *Elphidium excavatum* f. *clavata*; sample T75A). In sample T75B, the fresh water alga *Botryococcus*/Diterma is common while *Pediastrum*, *Brigantedinium* and other marine cysts are present. In the upper part of Unit B, only fresh water algae, rhizopoda and *Pediastrum* are present.

Samples from the lower part of Unit B at Section M68 (T68B) contain a significant number of benthic foraminifera tests dominated by *Elphidium excavatum* f. *clavata* (117 tests) and *Cassidulina reniforme* (33 tests). Fresh water algae, *Pediastrum* and Rhizopoda are common.

3.2.2.2. INTERPRETATION OF ASSOCIATED SECTIONS M75 AND M68

Marine waters with a fresh water surface layer were probably present during the

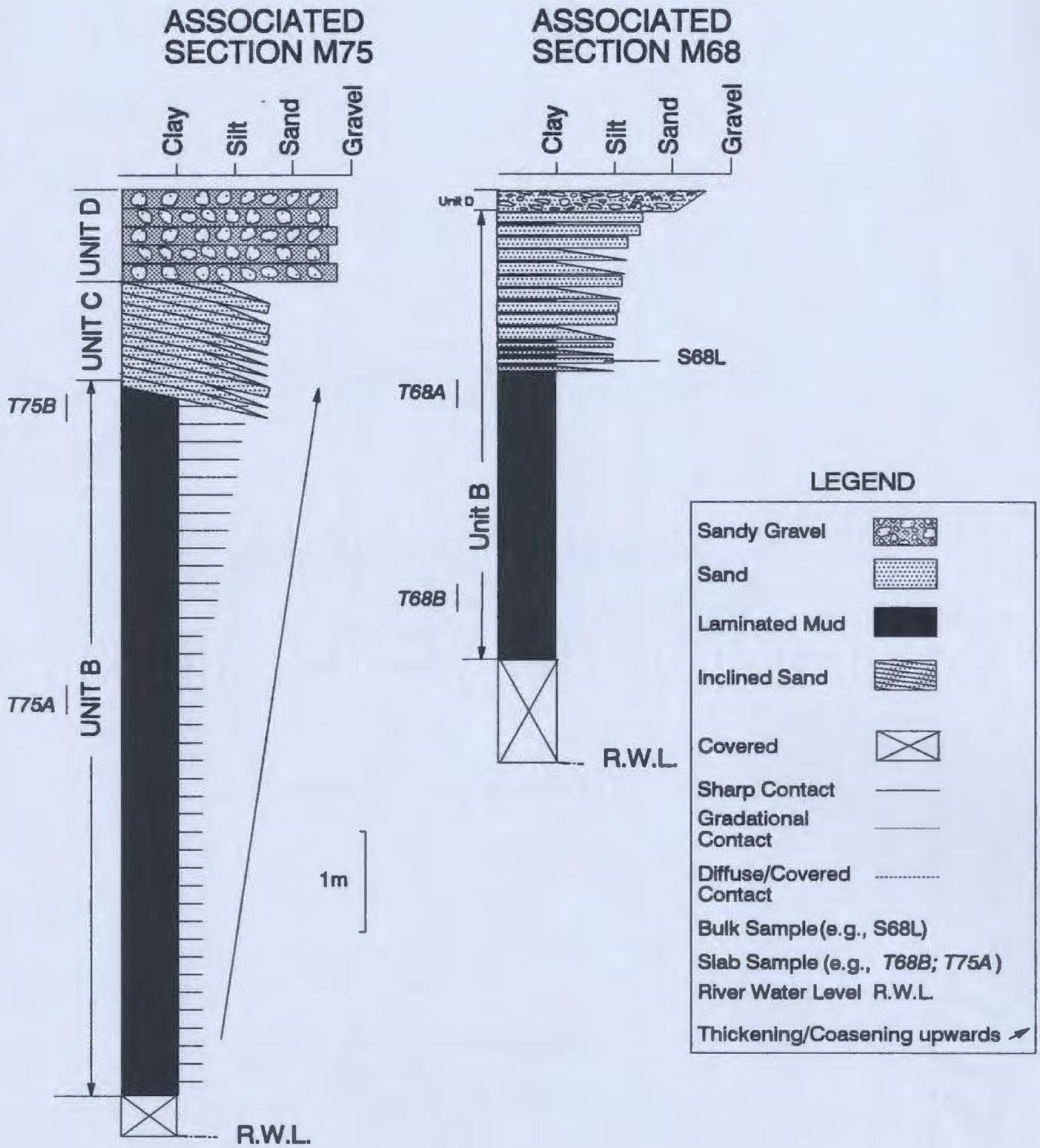


Fig. 3.17. Associated sections M75 and M68 stratigraphic columns. See text for details.

deposition of the lower part of Section M75 as indicated by the foraminifera species, dinoflagellates and algae. The salinity of the basin is believed to have been significantly reduced during the deposition of Unit B at Section M75, based on the presence of only fresh water algae and fresh water-estuarine dinoflagellates, and the disappearance of benthic foraminifera. This may have been caused by the isolation of the basin and a limited exchange of water masses due to the isostatic recovery of the downstream area and sills.

Unit C in Section M75 is believed to have been deposited by a combination of turbidity currents, liquefied flows and cohesionless debris flows due to upslope failures, and underflows generated during periods of high discharge. The inclined beds suggest deposition on an originally inclined surface. This unit may be part of the delta foreset that was prograding down the valley (i.e., fjord head delta). The sharp lower contact of this unit suggests that part of Unit B may have been eroded. Part of Unit C may have been eroded may have been eroded by the flows that deposited Unit D.

The presence of benthic foraminifera species *Elphidium excavatum* f. *clavata* and *Cassidulina reniforme* in Unit B at Section M68 suggests marine waters of reduced salinity. These foraminifera species are characteristic of recently deglaciated areas (Barrie, 1980; Vilks *et al.*, 1987). Fresh-water input is suggested by the presence of fresh-water algae.

3.2.3. KEY SECTION M61

This section is located approximately 1.5 km upstream of Section M68. Sediments exposed in a 17 m high terrace at this section are approximately 13.9 m thick. Unit B forms about 10 m of the section. It sharply underlies Unit C. Unit C is 1.4 m thick and sharply underlies Unit D (1 m thick). Unit E forms the top of the section; it is 1.4 m thick. There is a 2.8 m interval of covered sediments from the base of Unit B to the river water level (Fig. 3.18).

3.2.3.1. UNIT B

Unit B is divided into 7 subunits. Some of these subunits are similar in texture and sedimentary structures but they occur at different stratigraphic levels. Each subunit is described

KEY SECTION M61

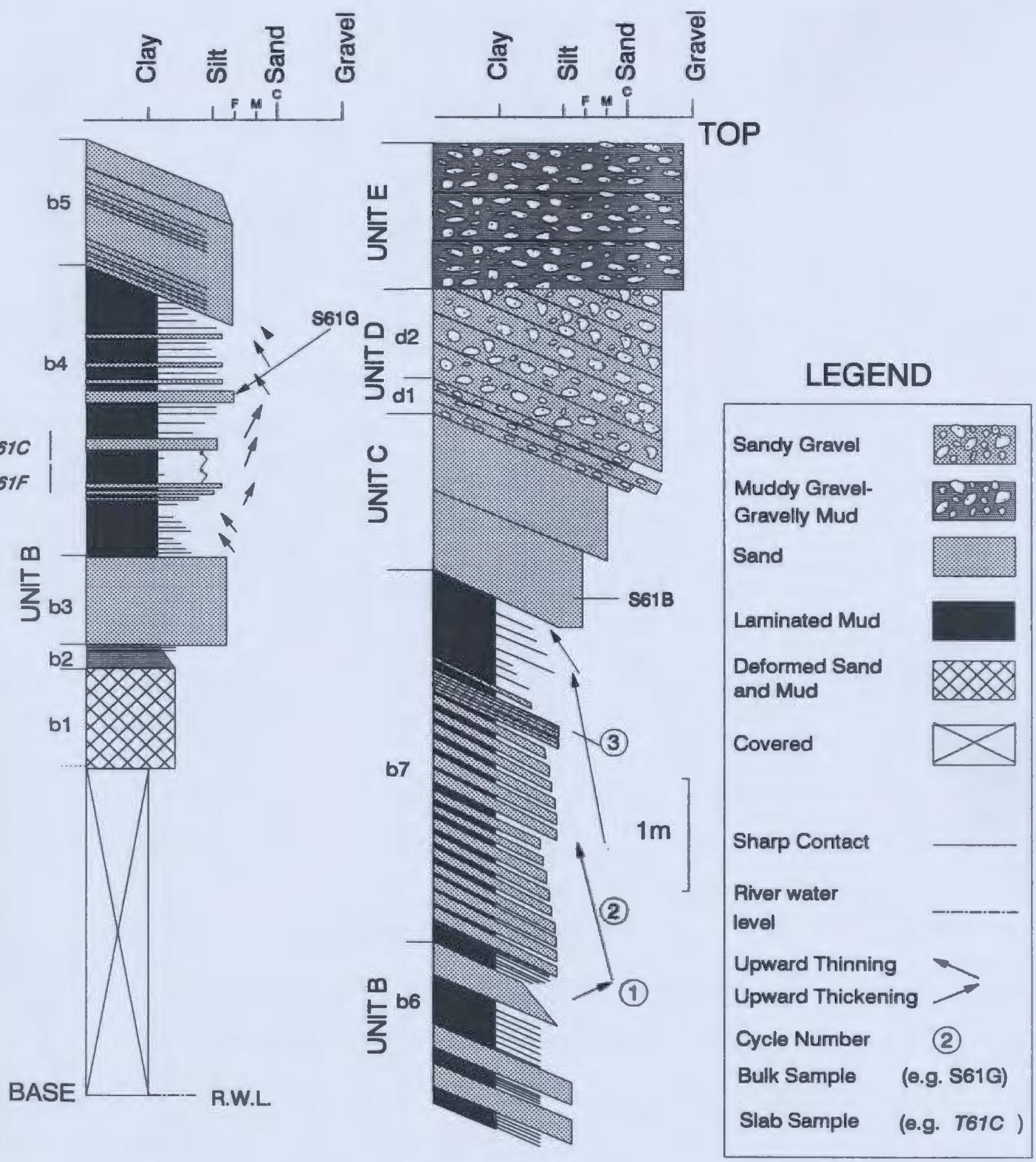


Fig. 3.18. Key Section M61 stratigraphic column. See text for details.

separately and then interpreted.

On fresh wet surfaces colour of the silty mud ranges from moderate pink (5 R 7/4) to light red (5 R 6/6) while the sand laminae within the silty mud are pinkish grey (5 YR 8/1) to greyish pink (5 R 8/2). Colour of the sand beds on fresh wet surfaces is yellowish grey (5 Y 8/1).

3.2.3.1.1. Description of subunit b1

This subunit consists of 90 cm of deformed sand and mud. The deformation is in the form of flame structures (10 cm high and 2-5 cm wide) of silty mud intruding variably thick sand layers. Within the more sandy parts, there are diffuse or irregular silty mud clasts about 10 cm across. Some bedding planes are visible and marked by load casts (10 cm wide and 7-8 cm deep). The upper contact of this subunit is wavy.

3.2.3.1.2. Interpretation of subunit b1

The presence of flame structures and fluid injection structures is interpreted to be the result of sediment loading due to rapid deposition (Anketell *et al.*, 1970; Lowe, 1975). The sand may have been emplaced rapidly by sediment gravity flows. The irregular mud clasts were probably ripped up during flow and were further modified during compaction and expulsion of the pore water associated with deformation.

3.2.3.1.3. Description of subunit b2

This subunit consists of a 20 cm-thick normally graded bed. Grain size changes from sandy-silty mud in the lower part to silty mud in the upper part. Faint 0.2-0.3 cm-thick silt laminae are present in the upper part of this bed. These laminae have diffuse contacts. This bed conformably overlies the wavy top of subunit b1.

3.2.3.1.4. Interpretation of subunit b2

This bed may have been deposited from a single turbidity current, with the lower massive part representing the A division and the upper part representing the B division of Bouma (1962). An alternative interpretation is that the lower graded part of this bed may have been deposited from suspension settling of silt from surface sediment plumes formed during periods of high sediment discharge, while the upper laminated part may have been deposited during a period of pulsations in plume discharge. The lower conformable contact with subunit b1 is taken as possible evidence against a turbidity current interpretation, because the A division of turbidites usually overlies an erosional surface.

3.2.3.1.5. Description of subunit b3

This subunit is 75 cm thick and consists of thinly bedded (1-2 cm), normally graded silty fine sand. Beds grade from fine sand at the base to silt at the top. These beds are distributed randomly within the subunit and they are irregular and contorted. Subunit b3 overlies subunit b2 with a sharp irregular contact marked by small shallow scours (1-3 cm deep and 4-10 cm wide). The upper contact of this subunit is sharp and flat.

3.2.3.1.6. Interpretation of subunit b3

This subunit may have been deposited from a series of pulsating turbidity currents or underflows that deposited a series of graded beds (Prior and Bornhold, 1988; Scott *et al.*, 1991). Alternatively, each graded bed may have been deposited from a surface sediment plume during a period of high discharge (Coleman and Prior, 1980). The lack of mud interbeds suggests proximity to the source of the flows and little time between events.

The contorted and deformed thin beds may have been produced by rapid sedimentation and pore water expulsion or by loading. The trench of the exposure was not sufficiently large to determine whether some of the irregularity of the lower contact of the subunit is due to loading.

3.2.3.1.7. Description of subunit b4

This subunit is 245 cm thick. It consists of normally graded fine sand-silt to mud couplets arranged in several upward thinning/fining and thickening/coarsening cycles. The upward thinning/fining cycles are more common than the upward thickening cycles (i.e., they are more numerous and are represented by a slightly greater cumulative thickness although individual upward thickening/coarsening cycles may be thicker). This subunit has a sharp planar lower contact marked by a sudden change in grain size and a sharp erosional upper contact where the overlying subunit truncates this subunit at an angle of 5°-10° (Fig. 3.18).

Upward thinning/fining cycles vary in thickness from 20 - 60 cm. In these cycles, the thickness of the sandy lower part of each couplet changes from 2-6 cm in the lower part to less than 1 cm in the upper part of the cycle. The thickness of the mud part of each couplet increases from 0.1-0.2 cm in the lower part of the cycle to 2-3 cm in the upper part of the cycle. The lower contact of each couplet is sharp and erosional in the lower part of the cycle, but becomes diffuse in the upper part of the cycle. Most couplets are normally graded but some sand beds that are 3-5 cm thick have sharp upper and lower contacts.

A thin section made from a slab tray sample (T61C; Fig. 3.18) taken from a muddy interval between cycles shows thinner cycles (~ 2 cm thick) in which the thickness of the sandy-silty part of the couplets increases and then decreases (Fig. 3.19a). Approximately 15 couplets are present in the 3.2 cm-high thin section. In the lower parts of these mini-cycles, the couplets have sharp basal contacts and show clear normal grading from sand to mud, while in the upper part the contacts between the sand and the underlying and overlying mud are diffuse. Another thin section from the same slab tray shows couplets with no trend in thickness (Fig. 3.19b)

The upward thickening/coarsening cycles noted in the field range in thickness from 20-50 cm. The thickness of the sandy part in each couplet increases from 1-2 mm in the lower part of the cycle to 2-3 cm in the upper part of the cycle. The lower contacts of the couplets change from diffuse in the lower part of the each cycles to sharp, irregular and erosional (with small scour marks, 1-3 mm deep). Normal grading within individual couplets is poor in the upper part of the cycles. All contacts between sand and mud are sharp.

Within this subunit, there are relatively thicker (10-20 cm), massive sand beds with sharp scoured bases and sharp planar upper contacts. Grain size analysis of a sample from one of these beds (sample S61G in Appendix B) shows that it is composed of fine sand (mean = 2.79 ϕ) that

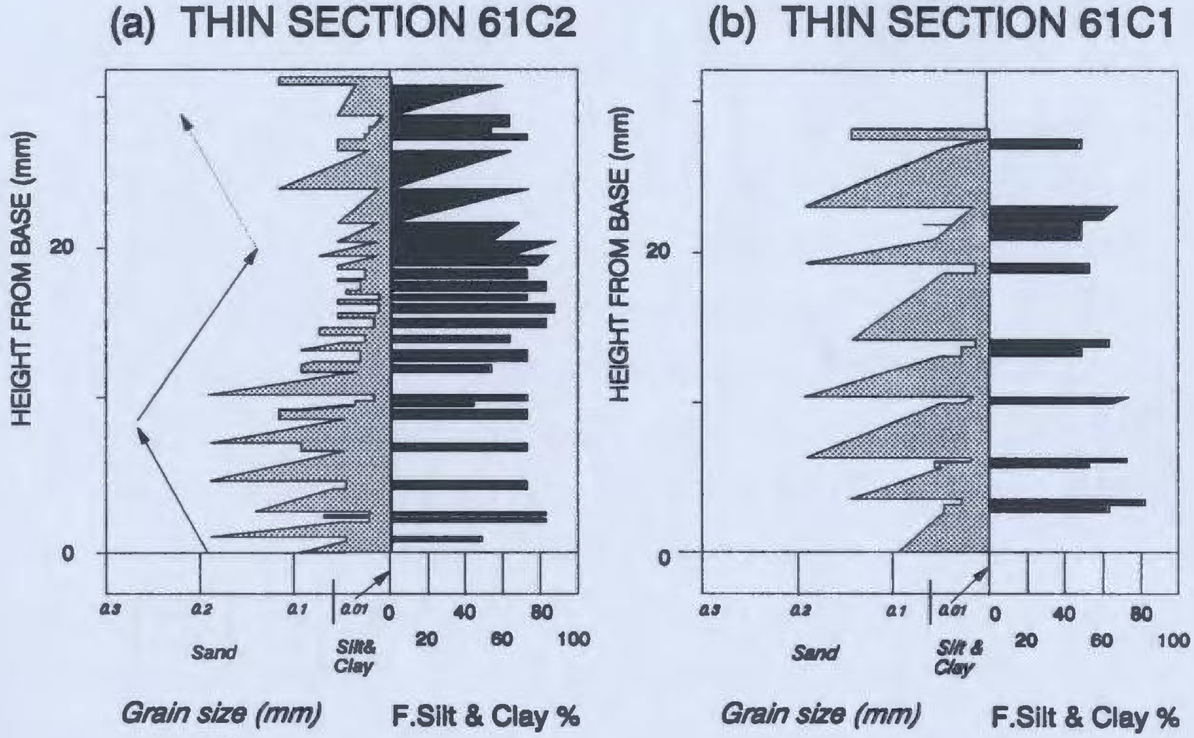


Fig. 3.19. Plot of the laminae thickness and percentage of fine silt and clay of two thin sections made from slab tray T61C. Note that one thin section shows upward thinning/fining and upward thickening/coarsening of laminae (a, 61C2) while the other (b, 61C1) does not show a trend.

is well sorted (σ = 0.43 ϕ) and fine skewed (Sk = 0.11).

3.2.3.1.8. Interpretation of subunit b4

The upward thinning/fining cycles are interpreted to have been deposited by turbidity currents, underflows and overflows (*cf.* Liverman, 1989). The high proportion of silt and fine sand in the lower couplets of each cycle suggests that turbidity currents or underflows were initially dominant. The lack of internal lamination in the sandy part of each couplet suggests deposition during a single event. The upward decrease in the thickness of the sandy part in each couplet and the increase in the proportion of mud suggests an increase in the importance of suspension settling relative to turbidity currents or underflows. The presence of several upward thinning and fining cycles (Fig. 3.18) indicates that these processes were repetitive.

Upward thickening cycles are interpreted to have been deposited from a series of turbidity currents, grain flows or underflows with the sandy part of each couplet being deposited from a single event and the muddy part being deposited from suspension. The lack of grading and the sharp contacts between the sand and mud suggest deposition by underflows or grain flows rather than by turbidity currents (Mackiewicz *et al.*, 1984; Hein *et al.*, 1990; R. Gilbert, personal communication, 1993).

Mini-cycles seen in thin section (Fig 3.19a) indicate that discharge fluctuations occurred on a variety of scales. Whether or not these mini-cycles are common is not known, because most of the unit was not studied using thin sections.

The relatively thicker sand beds (e.g., the 20-cm thick sand bed that yielded sample S61G) were probably deposited from single turbidity currents, grain flows, underflows or liquefied flows. The good sorting of the 20 cm-thick bed suggests that the sediment may have been well sorted before its incorporation into density currents or gravity flows. The basal scour and the lack of grading suggest that this bed was probably deposited by a grain flow, underflow, or liquefied flow rather than by a turbidity current (Gilbert, 1983; R. Gilbert, personal communication, 1993).

3.2.3.1.9. Description of subunit b5

This subunit is 110 cm thick. It differs from other subunits in that it lacks muddy strata. It is characterised by bedding that is inclined 5°-10° towards the SE. The bedding is marked by thin (1-7 cm) normally graded sand deposits that have dark heavy minerals grains at their top. Mica flakes (0.5 cm) are also present in these dark intervals. The sand is fine to medium grained.

Groups of normally graded beds, 20-25 cm thick, show an upward decrease in bed thickness from 3-7 cm to 1-2 cm. The uppermost upward thinning cycle is truncated by a 20 cm-thick, normally graded, fine to very fine sand bed that forms the upper part of this subunit. The lower contact of this graded bed is planar and inclined (5°-10°).

The lower contact of this subunit is sharp and slightly irregular with shallow (1-3 cm deep), broad (10-15 cm wide) scours. The upper contact is sharp and planar.

3.2.3.1.10. Interpretation of subunit b5

The graded beds showing an upward decrease in thickness are interpreted to have been deposited by a number of successively weaker turbidity currents or underflows, or by a single unsteady turbidity current or underflow characterized by fluctuating velocity (Prior and Bornhold, 1988; Scott *et al.*, 1991). The lack of mud intervals suggests rapid emplacement of this subunit. The normally graded bed that forms the top of this subunit was probably deposited from a turbidity current or an underflow, based on its sharp basal contact and the normal grading. The inclined bedding is discussed later (§ 3.2.3.7.).

3.2.3.1.11. Description of subunit b6

This subunit is 145 cm thick and is dominated by silty mud with thin (< 1 cm) silt laminae that dip at 5°-8° towards the E-SE. The contacts between the silty mud and the thin silt laminae are diffuse. In the lower part of this subunit are two 20 cm-thick, massive, sharp based, fine to medium grained sand beds. These two sand beds are separated by a 20 cm interval of silty mud with thin sand laminae. In the upper part of this subunit is a 30 cm-thick, normally graded bed. The bed fines upward from fine sand to silty mud. The bed is massive and has sharp, planar lower and upper contacts. The lower contact of this subunit is sharp and planar.

3.2.3.1.12. Interpretation of subunit b6

This subunit is interpreted to have been deposited by a combination of turbidity currents, underflows and suspension settling. The faint nature of the thin silt laminae within the silty mud suggests that overflows were dominant. The two massive beds in the lower part of this subunit are interpreted to have been deposited by grain flows or liquefied flows/cohesionless debris flows or underflows. The normally graded bed at the top of this subunit is interpreted to have been deposited from a single turbidity current or waning underflow.

3.2.3.1.13. Description of subunit b7

This subunit is 335 cm thick and consists of three intervals that show vertical trends in grain size and bed thickness (Fig. 3.18). The lower cycle 1 shows an upward thickening/coarsening while the upper two cycles show an upward thinning/fining trend (the trend in the upper cycle is not regular due to the presence of a 20 cm-thick sandy interval). The lower 20 cm-thick cycle 1 is characterised by fine sand-silt and mud couplets. In the lower part, the sandy part of each couplet is thin (0.5 cm) and fine (silt size) with diffuse upper and lower contacts. The silty mud in these couplets is about 0.5 cm thick. In the upper part of cycle 1, the sandy part of the couplets increases in thickness to 1-2 cm. These couplets are sharp based and normally graded from fine sand to silty mud. The thickness of the silty mud of each couplet is 0.5 cm. The lower contact of cycle 1 is sharp and planar.

The upper two cycles consist of fine sand-silt and mud couplets. The lower cycle 2 is 125 cm thick while the upper cycle 3 is 190 cm thick (Fig. 3.18). The basal contacts of these two cycles are sharp and erosional. In cycle 2, the sandy/silty part of each couplet decreases in thickness from 8 cm in the lower part to 4-5 cm in the upper part. In most couplets, the sand or silt is normally graded into the overlying mud. In some couplets, the sand is massive and the contact between the sand and mud is sharp. The silty mud that forms the upper part of each couplet is massive. The lower 90 cm of cycle 3 is similar to cycle 2. This is overlain by a 20 cm bed of parallel stratified fine-medium grained sand; the parallel strata are 1-2 cm thick and normally graded. This 20 cm bed has a sharp, erosional lower contact characterised by shallow scours 1-2 cm deep and 10-15 cm wide. The upper 80 cm of cycle 3 consists of two essentially identical upward thinning intervals (each ~40 cm thick). The lower interval continues the trend of decreasing sand bed thickness that starts at the base of cycle 3. The thickness of the sand/silt layers decreases to less than 1 cm in the upper part of this interval. The lower contacts of the sand strata in this interval change from sharp in the lower part to diffuse in the upper part. The upper contacts of the sand strata are sharp or gradational in the lower part while in the upper part they are diffuse.

3.2.3.1.14. Interpretation of subunit b7

This subunit is interpreted in the same way as subunit b4. Grain flows, turbidity currents or underflows were the dominant processes for the upward thickening and coarsening cycle 1; suspension settling from surface plumes and fine grained turbidity currents was followed by sedimentation from stronger turbidity currents, grain flows and underflows. For cycles 2 and 3, couplets that show normal grading from a lower sandy part to an upper muddy part were probably deposited from turbidity currents or underflows. Nongraded couplets that have a massive lower sandy part and a massive upper muddy part were probably emplaced by grain flows, cohesionless debris flows/liquefied flows or underflows (Smith and Ashley, 1985; Syvitski and Farrow, 1989; R. Gilbert, personal communications, 1993). The upward thinning and fining trends suggest that the intensity of the flows decreased with time.

The parallel stratified sand bed that interrupts the upward thinning and fining trend in cycle 3 is interpreted to have being deposited from a succession of turbidity currents or underflows, or from a single pulsating current. The lack of mud interbeds supports the hypothesis of a pulsating current. Alternatively, successive turbidity currents or underflows may have been so frequent that mud was not deposited or that mud was always eroded.

3.2.3.2. UNIT C

This unit consists of two sand beds, each 70 cm thick. The contact between the two beds and the upper and lower contacts of this unit are sharp and erosional. The beds dip 15° - 20° towards the E-NE.

The lower bed in this unit consists of massive sand that is fine grained (mean = 2.21 ϕ), poorly sorted (σ = 1.27 ϕ) and fine skewed (Sk = 0.19; S61B in Appendix B). This bed shows a very crude inverse grading from very fine sand to fine sand in the lower 10 cm of the bed, followed by normal grading to very fine sand at the top of the bed. The lower contact is marked by shallow (1-2 cm deep), broad (5- 10 cm wide) scours.

The upper bed consists of sharp based, massive, pebbly coarse sand. Small (0.5 cm) rounded pebbles are scattered throughout the bed and constitute less than 5% of the sediment. The upper and lower contacts are marked by shallow (1-2 cm deep), broad (5-10 cm wide) scours.

3.2.3.3. INTERPRETATION OF UNIT C

The lower inverse to normally graded bed is interpreted to have been deposited from a grain flow (Middleton and Hampton, 1976) or prolonged underflows (Mackiewicz *et al.*, 1984). The high primary dip would have facilitated grain flows. The massive nature of the upper bed, its pebbly texture, and lack of grading suggest deposition from a cohesionless debris flow (Postma, 1986; McCabe and Eyles, 1988) or an underflow (Smith and Ashley, 1985). The sharp erosional contacts of the two beds suggest that these flows possessed high speed and momentum (Postma, 1984).

3.2.3.4. UNIT D

Unit D is divided into two subunits (Fig. 3.18) that are separated from each other by sharp contacts marked by either an erosional surface of a sudden change in grain size. Subunits d1 and d2 dip at an angle of 20°-25° towards the E-NE. Unit E sharply overlies Unit D with a sharp flat contact (Fig. 3.20).

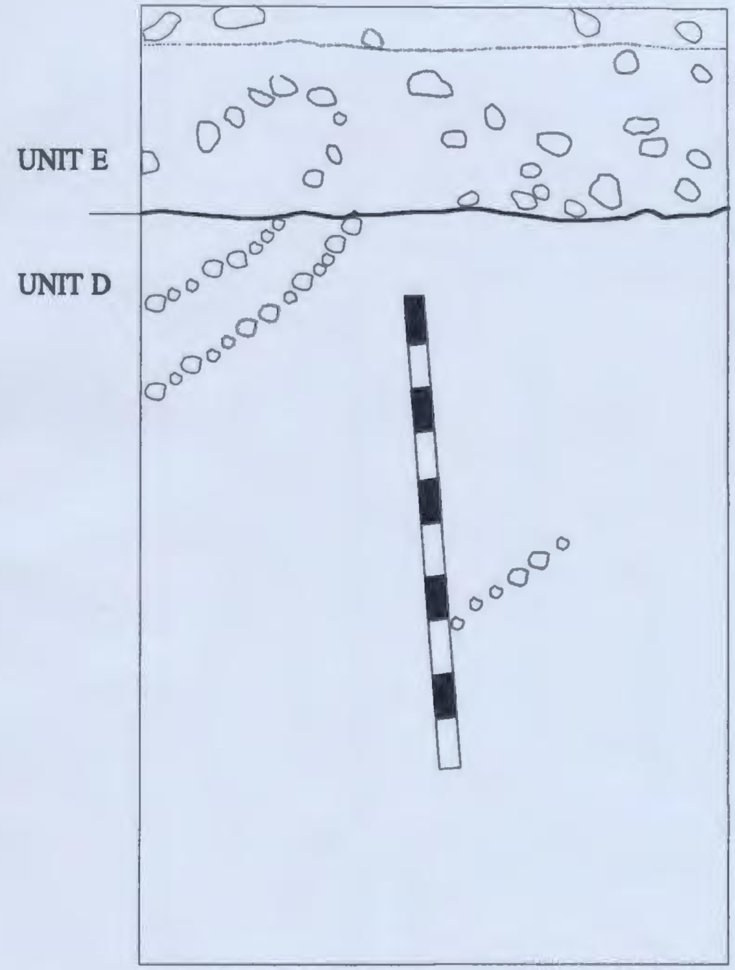


Fig. 3.20. Photograph of Unit E overlying the inclined beds of Unit D. The inclined beds below the contact dip to the left. Scale divisions = 10 cm.

3.2.3.4.1. Description of subunit d1

This subunit consists of three beds (10 cm each), one of sandy gravel and two of pebbly sand. The contacts between the beds are sharp and marked by a change in grain size and proportion of pebbles. The pebbly sand beds are inversely graded and sharp based. The sand is coarse to very coarse grained with scattered, rounded pebbles (2-3 cm). Inverse grading is recognised by an upward increase in the proportion of the pebbles within the beds. Pebbles form 20 - 30% of the beds. The sandy gravel bed is massive with sharp upper and lower contacts. Pebbles are 3-6 cm, well rounded and form 50-70% of the bed. The framework varies from clast supported to matrix supported. The matrix is coarse grained sand.

3.2.3.4.2. Interpretation of subunit d1

The alternating pebbly sand and gravel are interpreted to have been deposited by sediment gravity flows. The inverse grading in the pebbly sand beds suggests deposition from grain flows with dispersive pressure being the dominant particle support mechanism (Middleton and Hampton, 1976; Lowe 1982). Pure grain flows are deposited on slopes approaching the angle of repose of the material (Ashley *et al.*, 1991). The sandy gravel bed was probably deposited by a grain flow or as a slide deposit (Postma, 1984; Ashley *et al.*, 1991). Alternatively, both the sandy gravel and the gravelly sand beds may have been deposited from proximal underflows that formed at the river outlets during periods of high discharge (inertia flows of Prior and Bornhold, 1990).

3.2.3.4.3. Description of subunit d2

This subunit is 80 cm thick and consists of 10-15 cm-thick beds of gravel that dip 20°-25° towards the E-SE. The gravel mode consists of subrounded pebbles (3-5 cm). The beds are matrix supported to clast supported. The beds are differentiated by the amount of coarse sand matrix. Clast-supported beds have little or no sand and are loosely packed. Matrix-supported beds contain up to 30% coarse sand. Contacts between individual beds are sharp.

3.2.3.4.4. Interpretation of subunit d2

Each bed is interpreted to have been deposited from a single sediment gravity flow. These may have been debris falls (*sensu* Nemec, 1990), grain flows (Middleton and Hampton, 1976, and many other) or cohesionless debris flows (*sensu* Postma, 1986). Both clast-supported and matrix-supported beds might have been emplaced by rock falls (*sensu* Nemec, 1990) with sand infiltration providing the matrix (Postma, 1984). Alternatively, the porosity at the top of a gravel bed may have been clogged by infiltrating sand, preventing further sand infiltration into the lower parts of the gravel beds (Smith, 1985). Textural variations might also reflect changing source material.

3.2.3.5. UNIT E

Unit E is 140 cm thick and consists of three beds of massive, muddy sandy gravel to gravelly sandy mud. Each bed is 45-50 cm thick, and consists of 15-20 cm cobbles in a matrix of muddy sand. The beds show a crude coarsening upward with a subtle change in the size of the cobbles from 15 cm in the lower part to 20 cm in the upper part. The beds are matrix supported. The clasts have a wide range of shape (from very angular to rounded). The contacts between the beds are sharp and irregular. The contact between this subunit and the inclined beds of the underlying subunit is horizontal and irregular (Fig. 3.20). Although this unit is assigned to Unit E, it is not as muddy as Unit E in the upstream part of the area, but it differs from typical Unit D in that it contains fine muddy material.

3.2.3.6. INTERPRETATION OF UNIT E

This unit is characterised by a homogeneous muddy sand matrix, massive nature, weakly developed inverse grading, and stacked beds with distinct contacts. These characteristics are typically associated with debris flows (*cf.* Middleton and Hampton, 1976; Lowe 1979, 1982; McCabe and Eyles, 1988). The matrix-supported beds suggest cohesive flow while the crude inverse grading suggests complementary dispersive pressure resulting from clast interaction (Middleton and Hampton, 1976; Lowe, 1982). The wide range in clast shape and poor sorting suggest that the clasts were derived from a glacial. This texture is strikingly different to the texture of gravel beds in Unit D. This occurrence of Unit E is not interpreted as till because of its inverse grading, which is typical of sediment gravity flow deposits. This could, however, be a glaciogenic debris flow, or " flow till" (Hicock *et al.*, 1981).

3.2.3.7. DISCUSSION OF SECTION M61

Based on the inclined nature of some of the beds in this sequence (upper part of Unit B, Unit C and Unit D) and the overall upward coarsening trend, this section is interpreted as a deltaic sequence that formed as a result of progradation of a fluvial system into a fjord-type basin as relative sea level dropped. The exposed sequence may represent a Gilbert-type delta with delta bottomsets and foresets.

The lower part of the sequence (subunits b2 and b4) is dominated by fine sand-silt and mud couplets deposited mainly by low concentration turbidity currents, and suspension settling from surface plumes and the tails of turbidity currents. Salinity of the basin may have been high during the deposition of this lower part so that underflows were rare or absent. Subunits b2 and b4 probably represent delta bottomset deposits. The lack of ice rafted pebbles indicates that glaciers had receded from the head of the fjord. The cyclicity observed in some of the subunits of Unit B probably resulted from changing sediment supply and current strength and activity. Similar cycles have been observed in may fjord settings (Mackiewicz *et al.*, 1984; Powell, 1984; Syvitski and Hein, 1991; Hein and Syvitski, 1992) and in glaciolacustrine settings (Smith and Ashley, 1985; Liverman, 1989). In detail, these cycles are attributed to the changing of the discharge of the river that fed the delta, lateral migration of the fluvial channels feeding the delta, or the presence of multiple sources of fresh water discharge. These cycles may be related to seasonal and shorter variations in discharge due to variations in the melting rates of the glaciers that fed the fluvial channels.

The thin nature of the couplets in the lower part of the sequence suggests that the source of currents was not close to Section M61. Deposition probably took place during the early part of the 8 ka - 7 ka interval when relative sea level was high (see Chapter 5).

Land emergence due to isostatic rebound and the progradation of the fluvial system resulted in the deposition of more proximal deltaic sediments through time. The upper part of Unit B probably represents lower delta foresets with deposition dominated by sediment gravity flows, underflows and overflows. The thinning/fining upward trends observed in subunit b7 probably represent fluctuations in the sediment supply to the area of the delta. Migration of the fluvial channel feeding the delta may have also played a role in the development of these cycles. Episodic periods of high sediment supply or large mass failures also occurred during the development of these cycles, depositing thicker beds.

As the delta prograded, coarser sediments were being deposited by a variety of sediment gravity flows onto the middle and upper delta foresets (Unit C and Unit D). Unit E is interpreted to represent debris flow activity on the delta top; essentially identical debris flow deposits characterise nearby sections M60 and M62 (Fig. 3.2b). The sediments of Unit E were probably glacial materials that did not experience much reworking.

The delta at Section M61 may have been a side-entry delta because the dip of the beds is towards the E-SE, unlike the expected northeastward direction of progradation of the fjord head delta. This side-entry river may have descended from the NE-trending ridge that is present in the area (Fig. 1.2). The top of the section is located at an elevation of 80-100 m; thus, this delta may have formed at the same time as the lower-level delta observed by Batterson *et al.* (1988) in the Moran Lake - Kaipokok Valley area, approximately 15-25 km SE of Section M61.

3.2.4. ASSOCIATED SECTION M66

This section is located 0.3 km downstream of Section M61, and is noteworthy because marine shells (which have been dated) were collected from the lower part of the section. Only units B and D were observed in this section (Fig. 3.21). These units are exposed in an approximately 6.7 m-high terrace.

3.2.4.1. DESCRIPTION OF SEDIMENTS IN SECTION M66

Unit B consists of interstratified sand and mud that is 1.5 m thick. In the lower part, this

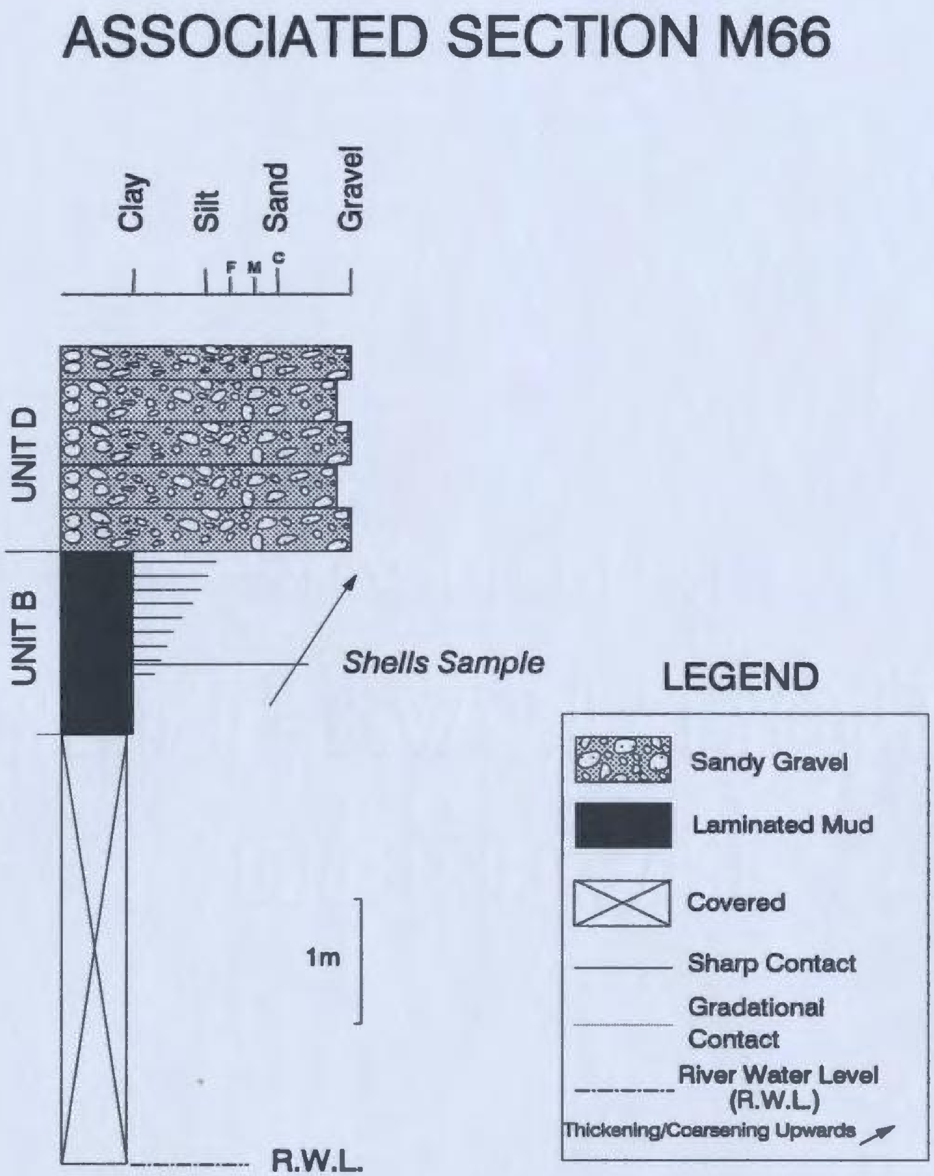


Fig. 3.21. Associated Section M66 stratigraphic columns. See text for details.

unit consists of faint silt laminae (0.1-0.3 cm-thick) (type L_1 in Table 3.2) separated by massive mud (0.5-1 cm-thick). The upper and lower contacts of the silt laminae are diffuse. Colour on fresh wet surfaces ranges from greyish red (5 R 4/2) to moderate red (5 R 4/6). Some intervals in the lower part are massive. Shells were collected from a massive 20 cm interval situated 20cm from the lowermost exposed level of this unit, 3.5 m above the river water level. The shells are described in Chapter 4. They have been dated at 7950 ± 95 yr B.P. (Beta-28885).

In the upper part of Unit B, the silt laminae become coarser, more distinct, sharp based, and normally graded from fine sand-silt at the base to silty mud at the top (L_{1II} in Table 3.2). The silty mud is massive and ranges in thickness from 0.4-0.8 cm. Laminae similar to those present in the lower part of the unit are also present. Colour of the fine sand-silt laminae on fresh wet surfaces is light grey (N7) to pinkish grey (5 YR 8/1), while the colour of the mud is light red (5 R 6/6) to moderate red (5 R 5/4). The upper contact of this subunit with the overlying Unit D is sharp and erosional with scours 20-25 cm wide and 10-15 cm deep.

Unit D is 1.6 m thick and consists of crudely stratified gravel beds. Beds are approximately 35 cm thick and are separated from one another by planar contacts. Clasts are rounded and range in size from 15-20 cm. The matrix is coarse sand. Beds vary from matrix supported to clast supported. The percentage of clasts ranges from 35-65%. Beds are differentiated by variable clast size or percentage. Clast fabric could not be determined due to collapse of the sediments.

3.2.4.2. INTERPRETATION OF SEDIMENTS IN SECTION M66

The faint laminae in the lower part of Unit B were deposited by suspension settling from surface sediment plumes or low concentration turbidity currents, or both. Concentrations in the plumes and turbidity currents increased with time. Deposition from underflows may have been common during the deposition of the upper part of Unit B.

Unit D is interpreted as gravel bars formed by bed load transport during periods of high discharge (Church and Gilbert, 1975; Hein, 1 [illegible] The planar contacts and the diffuse bedding planes of the beds suggest deposition as bars in [illegible]low channels (Hein, 1984).

3.2.4.3. DISCUSSION OF SECTION M66

Unit B is interpreted as delta bottomset deposits like those at Key Section M61, subunit b4. The upper part of the deltaic sequence at Section M66 may have been eroded by fluvial channels after the area became emergent (Unit D).

Shells collected from the lower part of this section suggest a marine environment near river outlets (see Chapter 4). Even at the highest relative sea level suggested for the area, this section would have been located near the margin of the fjord/embayment and thus any melting ice at higher elevation may have supplied fresh water to the area. The lack of coarse particles in the sediments from which the shells were collected implies that the glaciers had receded onto land by this time, so that iceberg rafting did not occur.

3.2.5. DEFORMED SEDIMENTS IN THE MIDDLE PART OF THE STUDY AREA

Deformed sediments occur in two locations in the banks of the modern river between sections M61 and M66. The first outcrop consists of deformed interbedded sand and mud (the sediments are similar to the upper part of Unit B in many sections). The second outcrop consists of deformed laminated to thinly interbedded sand and mud (like the lower part of Unit B in some sections). The outcrops are located within 100 m of each other. The exposure between the two outcrops is covered. The deformed sediments were well exposed in this area in 1987 (M.J. Batterson, personal communications, 1988).

3.2.5.1. DESCRIPTION OF OUTCROP 1

This outcrop is approximately 15 m wide and 1.8 m high (Fig. 3.22 and Fig. 3.23). There is a ca. 2.5 m covered interval from the base of the deformed sediments to the river water level. The deformed sediments are overlain and truncated by bedded gravel similar to Unit D in other sections.

The deformed sediments consist of 10-50 cm-thick beds of fine to medium sand interbedded with 5-8 cm-thick beds of silty mud. Contacts between the sand and mud are sharp. The sand beds have a concave upward form. The dip of the sand beds varies from 30°- 40°

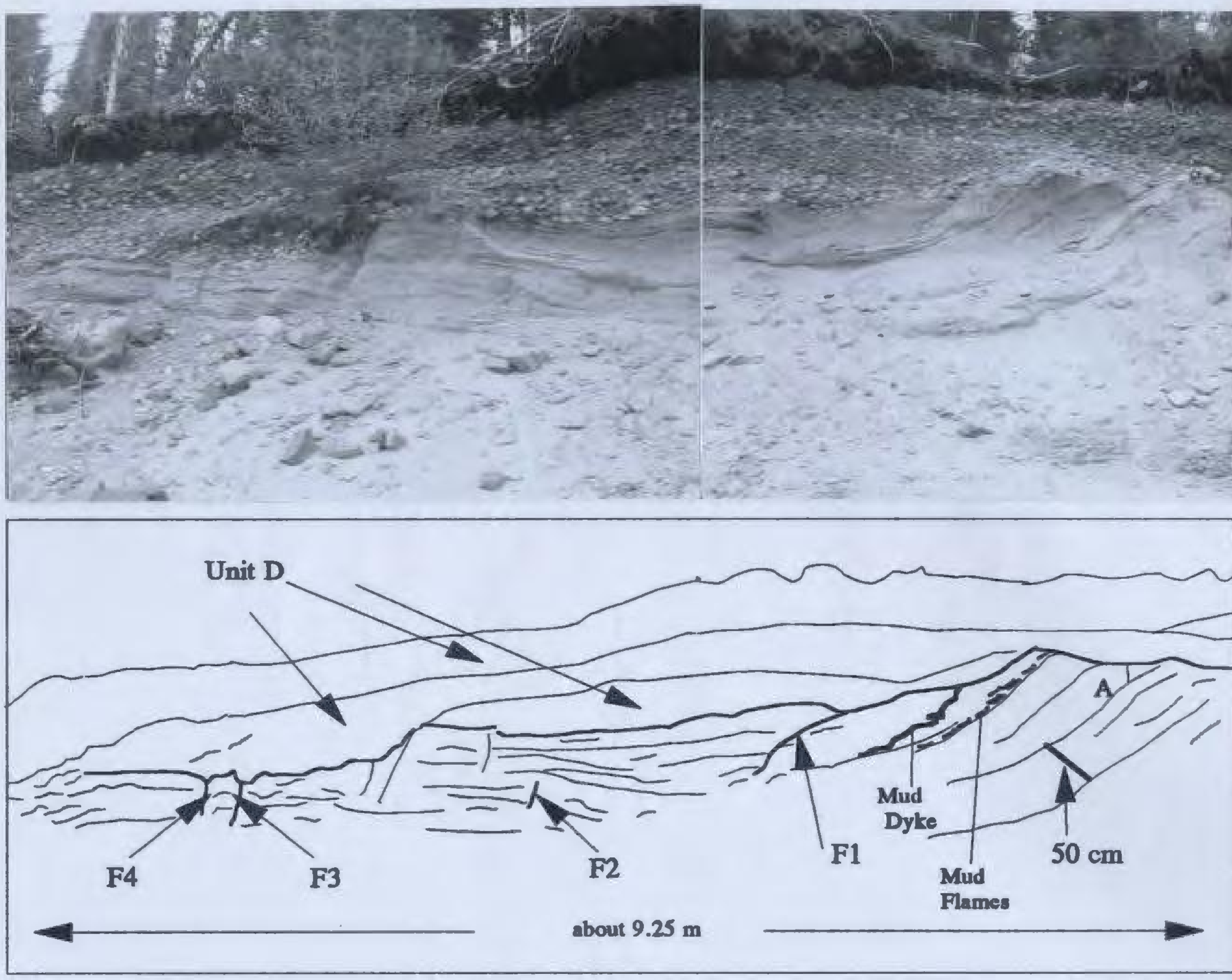


Fig. 3.22. Outcrop 1 (Upstream end). Concave upward beds are interpreted to have been formed by compression near the toe of a slump block. Note the reverse fault (F1) in the centre of the depression and the other vertical faults (F2, F3 and F4) at the upstream end of the outcrop (towards the left of the photograph). Note also the flame structures and the mud- filled joints in the sand beds. Bed A is shown in both this figure and in Figure 3.23, for orientation.

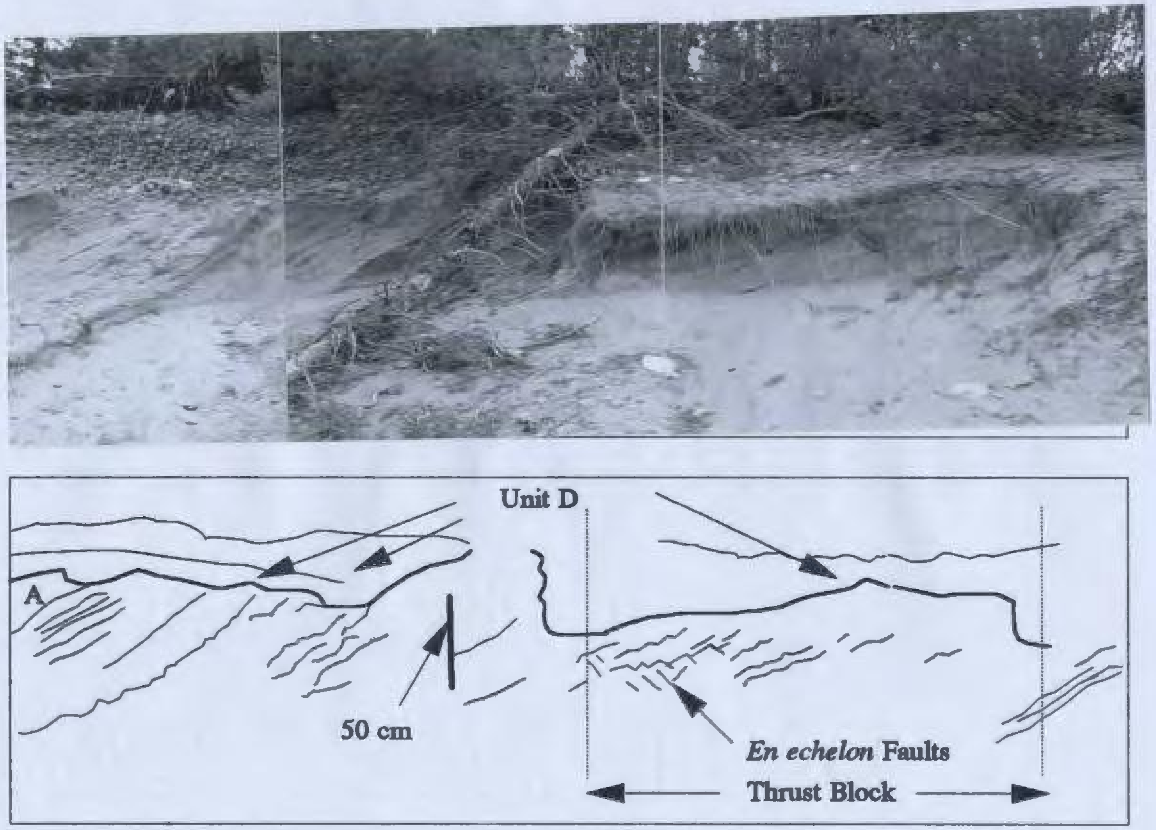


Fig. 3.23. Outcrop 1 (downstream end). The toe of the inferred slump block shows a thrust block in the extreme downstream end of the photo with *en echelon* faults. The two photographs that form this figure were taken at quite different angles from the face, and at a different angle than the photographs in Figure 3.22. The bar scale in the line drawing only applies to the beds in that part of the slump block. Bed A is the same as in Figure 3.22.

towards the SW in the downstream end of the exposure, to flat lying in the middle part, to a < 10° dip towards the NE at the upstream end. Several faults cut this outcrop. A reverse fault (F1 in Fig. 3.22) that has a strike of N76°E and a dip of 14° SW offsets beds 15-20 cm. A series of vertical faults towards the upstream end of the outcrop (F2, F3 and F4) offset beds < 10 cm. There are several closely spaced, *en echelon* faults in the extreme downriver part of this outcrop that have normal offset of < 5 cm (Fig. 3.23).

There are irregular mud lenses and flame-like structures (10-15 cm long and with variable thicknesses) along some of the bedding planes (Fig 3.22). These flame structures and lenses appear to intrude the bottom of the sand layers. Some of the mud lenses are completely isolated from nearby mud layers. Mud locally forms continuous irregular dykes (2-3 cm wide) that cut some of the sand beds diagonally.

3.2.5.2. INTERPRETATION OF OUTCROP 1

The sediments in this outcrop were originally deposited by processes similar to those that formed the upper part of Unit B in other sections. The sand beds are interpreted to have been deposited by a combination of underflows, cohesionless debris flows and liquefied flows (Mackiewicz *et al.*, 1984; Postma, 1986; Syvitski *et al.*, 1987). The sharp contacts between the sand and mud beds and the lack of turbidite divisions of Bouma (1962) suggest that the sands are not turbidites. The mud was probably deposited from suspension and low concentration turbidity currents.

The deformed outcrop is characterized by concave upward bedding planes and an inferred thrust block at the downstream end of the outcrop. Similar features have been described from rotational slump blocks in deltas, continental slopes and fjords (*cf.* Coleman and Prior, 1980; Syvitski *et al.*, 1987; Barnes and Lewis, 1991). Such rotational slumps occur in areas of high sedimentation rates, unstable slopes and where rivers enter the side of a fjord (Syvitski *et al.*, 1987). They are characterized by a compressional depression and a series of thrust faults at the toe of the slide block. Earthquakes, wave action and sediment loading have been suggested as the main causes for slumping (Coleman and Prior, 1980; Syvitski *et al.*, 1987).

The concave upward form of the beds in face 1 probably formed as a result of compression at the toe of a slide. This flexing may have caused some of the muddy material between the sand beds to inject the overlying sand beds. The mud-filled joints that cut diagonally across the sand beds have a similar origin. *En echelon* faults point to local brittle behaviour during sliding. The *en echelon* faults probably formed due to shearing during thrusting (Maltman, 1988).

This slide block may have originated near the inferred side entry delta in the area of nearby Section M61, where there would have been slopes and high sedimentation rates (*cf.* Syvitski *et al.*, 1987). The exact direction of movement of the slide block is not known due to the limited exposure.

3.2.5.3. DESCRIPTION OF OUTCROP 2

This outcrop is located approximately 150 m downriver from outcrop 1. The outcrop is approximately 6 m wide. The exposure in this outcrop is very poor; repeated washing of the outcrop only exposed the main structures.

This outcrop can be divided into four parts based on the type of sediment and intensity of the deformation (Fig. 3.24). The lower part is ca. 1.5 m thick and consists of apparently flat lying laminated sand and mud similar to the lower part of Unit B in many sections. Above this lower interval, there is a ca. 2.3 m interval of deformed laminated sand and mud. The lower 80 cm of this deformed interval conforms to the underlying flat lying laminated sand and mud in the middle of the outcrop but towards the upstream part of the outcrop the laminae dip steeply (50°-60°) towards the NE. The upper part of the deformed interval is ca. 1.5 m thick and shows large, tight, isoclinal or fan folds (both anticlines and synclines). The folds are ca. 50-70 cm across and ca. 90-120 cm high. Some sand and mud laminae wrap around to form circular cross sections that appear like the spiral folds of Allen (1982). The axial planes of all the folds trend NW (four measurements; N37°W, N45°W, N51°W and N53°W).

The deformed interval is overlain by a 40-60 cm interval of interbedded fine sand and mud. Sand beds are 5-10 cm thick while the mud beds are 2-5 cm thick. Contacts between the

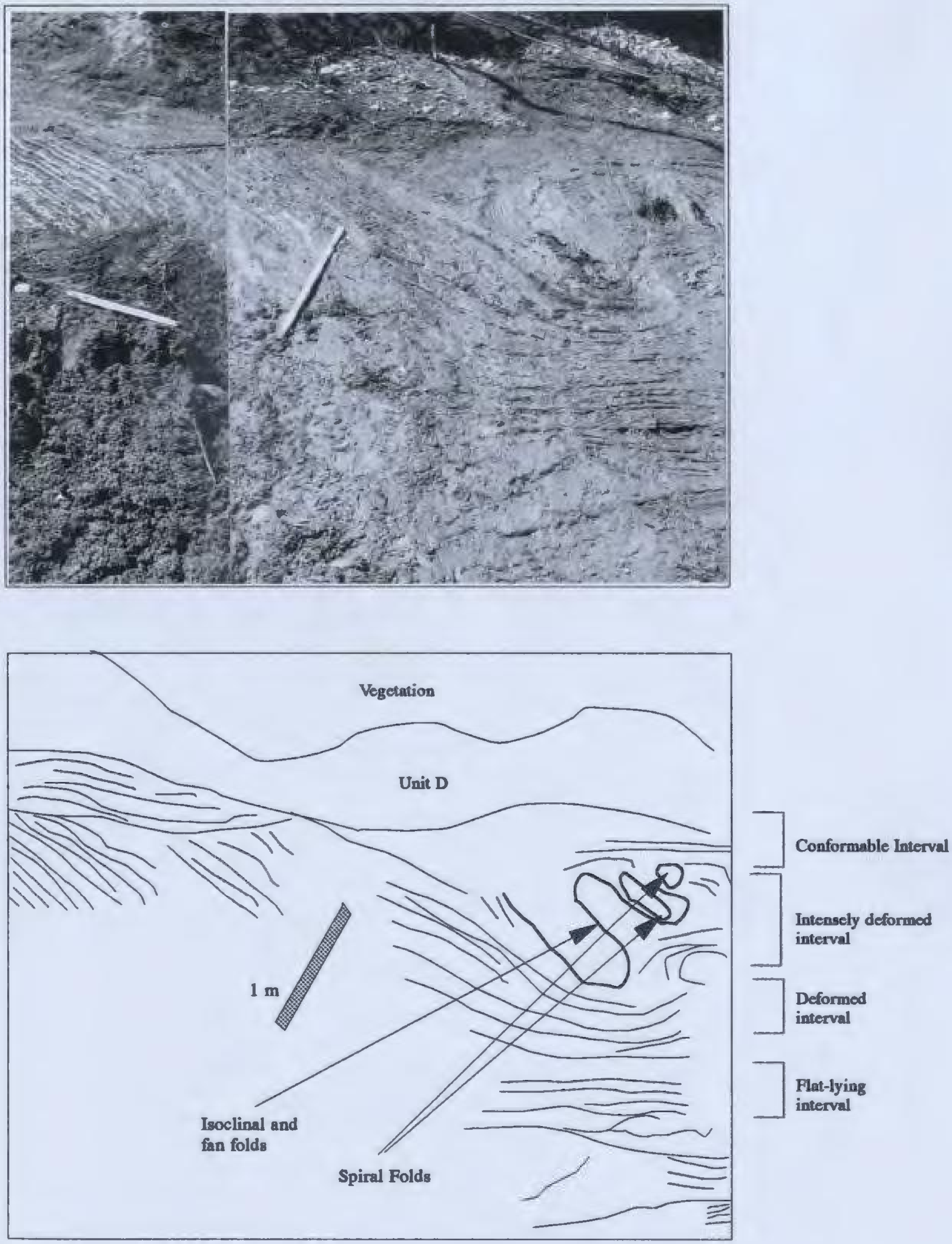


Fig. 3.24. Outcrop 2, deformed laminated mud. Note the tight folds in the centre right and the lateral change in dip of the deformed unit.

sand and mud are sharp. This interval drapes the irregularities in the underlying deformed interval. The top of the outcrop consists of gravel beds 20-40 cm thick that are similar to Unit D sediments in other sections. This gravel interval has a sharp, erosional lower contact and underlies a recent vegetation cover.

3.2.5.4. INTERPRETATION OF OUTCROP 2

The laminated sand and mud (lower two intervals of this outcrop) and the interbedded sand and mud (the third interval) are similar to Unit B ir other sections. Primary deposition was from a combination of underflows, overflows, turbidity currents and cohesionless debris flows. The gravel beds in the upper part (Unit D) are interpreted as shallow fluvial channel bars.

Folds in the middle of the section suggest sliding to the N-NW. Similar folds have been termed flow folds (Brodzikowski and Van Loon, 1985). Such structures are very common on slopes in areas of high sedimentation rate. The deformation probably formed as a result of the downslope collapse of unstable laminated sediment. The relationship of this deformation to outcrop 1 is unknown due to the limited exposure. Coleman and Prior (1980) observed folding in muddy intervals beneath rotational slumps that are similar to what was observed in outcrop 2. The spiral and isoclinal-fan folds observed in the upper part of the deformed interval indicate greater strain, perhaps due to a high initial water content (Brodzikowski and Van Loon, 1983).

The truncated contact between the lower deformed interval and the overlying interbedded sand and mud suggests that some deformed sediment was eroded before the deposition of the interbedded sand and mud interval. The deformation was probably caused by slumping rather than glaciotectonism due to ice re-advance. This interpretation is based on the inferred subaqueous origin of the draping interbedded sand and mud, and the lack of dropstones in any Unit B sediments in any part of the thesis area.

3.2.6. SEDIMENTARY ENVIRONMENTS IN THE MIDDLE PART OF THE AREA

The sections in the middle part of the area suggest a sequence of events similar to that recognized in the downstream part of the area. There was a transition from marine to deltaic to fluvial conditions that took place as a result of isostatic rebound and the progradation of a fluvial system. The sediments in the middle part of the study area show a greater lateral variability than in the downstream area, probably reflecting the influence of side entry rivers on sedimentation, especially in the areas of sections M61 and M85. Also, the transition from marine to deltaic to fluvial probably took place earlier than in the downstream region due to a faster rate of isostatic recovery of the middle part of the area.

Figure 3.25 is a hypothetical diagram that shows the suggested stages of deposition in the different sections in the middle part of the area. This figure is not a correlation chart because the age relationships between the different units in the sections are not known. The positioning of the sections within these stages is an attempt to group similar inferred sedimentary environments. Figure 3.26 is a hypothetical map of the shoreline positions at 8 ka based on emergence curves and crustal deflection (Chapter 2).

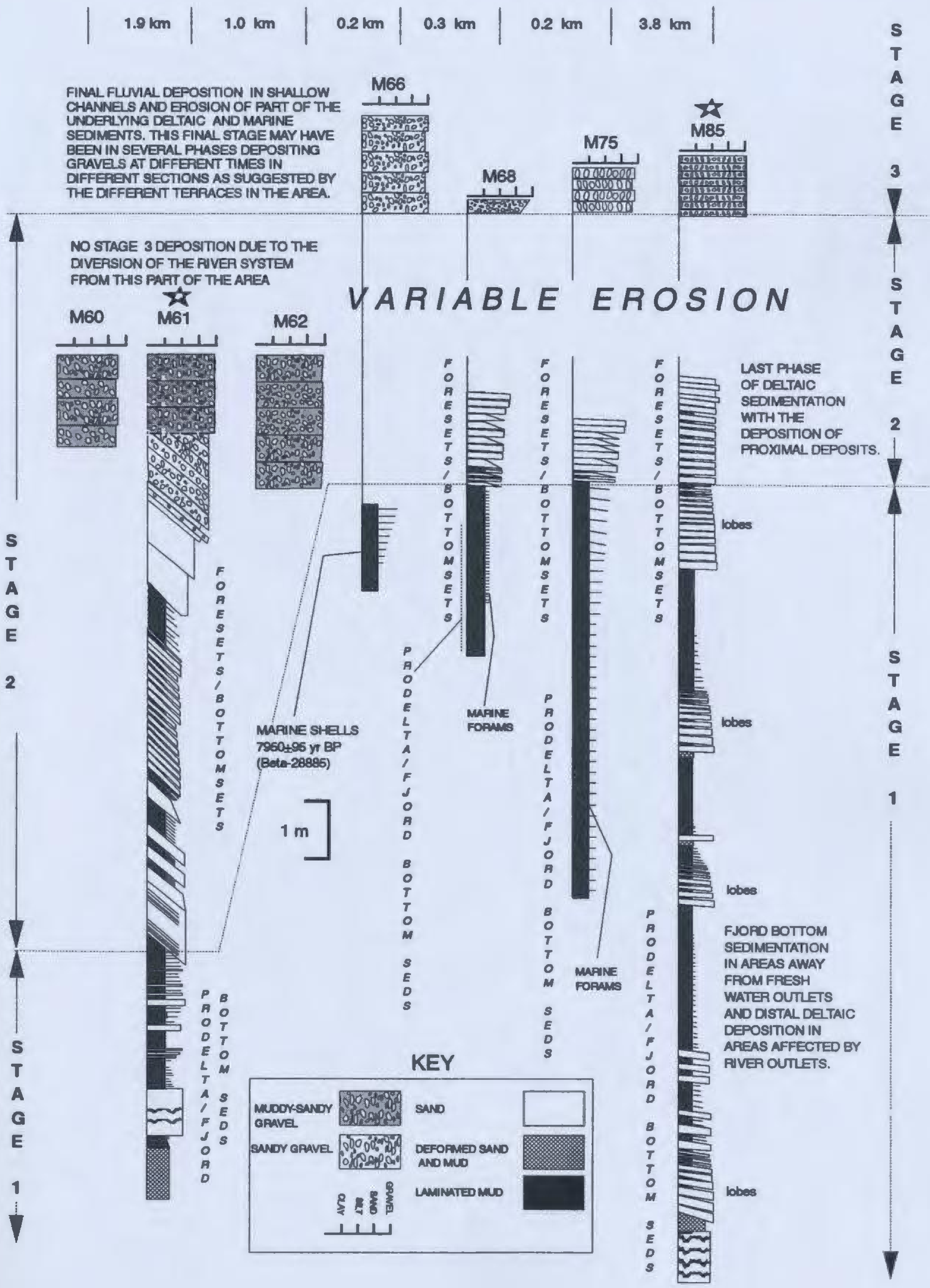
In stage 1, deposition was in a fjord-embayment setting. Marine conditions are indicated by microfossils and shells in sections M68, M66 and M75. In these sections, deposition was dominated by suspension settling of silt and clay from surface sediment plumes interrupted by turbidity currents generated at distant river outlets. Ice rafted material is absent, suggesting that the termini of the glaciers had retreated onto land before deposition of Unit B. The marine microfossils and shells suggest open-water conditions and no prolonged periods of winter ice cover.

The areas of sections M61 and M85 were more directly influenced by fresh water outlets, probably due to their proximity to these outlets. Initially, sedimentation was dominated by deposition of silt and clay from surface sediment plumes but was interrupted by sedimentary gravity flows such as grain flows, turbidity currents, liquefied flows and cohesionless debris flows (*sensu* Postma, 1986) that formed because of sediment failures on steep slopes. These different gravity flows deposited sand beds of variable thickness within the lower part of Unit B (Fig. 3.25). The different vertical trends observed groups of sand beds probably reflect both long- and short-period variations in the discharge from these outlets, and migration of river channels.

Delta foresets at sections M61 and M85 dip in a direction different to the expected cap

Fig. 3.25. Hypothetical stages during which the different sections in the middle part of the study area were deposited. Key sections are indicated by a star. This is not a correlation chart because the lateral relationships between the units are not known. The positioning of each section is based on the inferred depositional environment for each unit. Sections M60, M62, M66, M68 and M75, although not described in any detail in the text, are included to show either the extent of similar units (M60, M61 and M62 debris flow deposits), or the locations of microfossils (M68 , M75). This diagram is meant to show neither the magnitude nor the duration of erosion between stages 2 and 3.

In stage 1, deposition was dominated by suspension settling from surface-sediment plumes, except near areas of sediment input where thicker sand beds were deposited as lobes. The beds in these lobes thin upward, probably reflecting a decline in discharge or migration of the loci of deposition. In stage 2, proximal deltaic deposits were deposited as delta foresets and topsets with thicker and coarser deposits forming in areas near the river outlets feeding the deltas (sections M61 and M85). This was followed in most sections by a phase of variable erosion by fluvial channels of stage 3. In sections M61, M60 and M62, the proximal deltaic deposits may owe their preservation to diversion of the fluvial channels away from nearby highs.



1.9 km
1.0 km
0.2 km
0.3 km
0.2 km
3.8 km
FINAL FLUVIAL DEPOSITION IN SHALLOW CHANNELS AND EROSION OF PART OF THE UNDERLYING DELTAIC AND MARINE SEDIMENTS. THIS FINAL STAGE MAY HAVE BEEN IN SEVERAL PHASES DEPOSITING GRAVELS AT DIFFERENT TIMES IN DIFFERENT SECTIONS AS SUGGESTED BY THE DIFFERENT TERRACES IN THE AREA.
M66
M68
M75
M85
STAGE 3
NO STAGE 3 DEPOSITION DUE TO THE DIVERSION OF THE RIVER SYSTEM FROM THIS PART OF THE AREA
VARIABLE EROSION
M60
M61
M62
STAGE 2
FORESETS/BOTTOMSETS
LAST PHASE OF DELTAIC SEDIMENTATION WITH THE DEPOSITION OF PROXIMAL DEPOSITS.
lobes
PRODELTA/FJORD BOTTOM SEDS
MARINE FORAMS
MARINE SHELLS
7960±95 yr BP
(Beta-28885)
1 m
FJORD BOTTOM SEDIMENTATION IN AREAS AWAY FROM FRESH WATER OUTLETS AND DISTAL DELTAIC DEPOSITION IN AREAS AFFECTED BY RIVER OUTLETS.
STAGE 1
KEY
MUDDY-SANDY GRAVEL
SANDY GRAVEL
SAND
DEFORMED SAND AND MUD
LAMINATED MUD
GRAVEL
SAND
SILT
CLAY

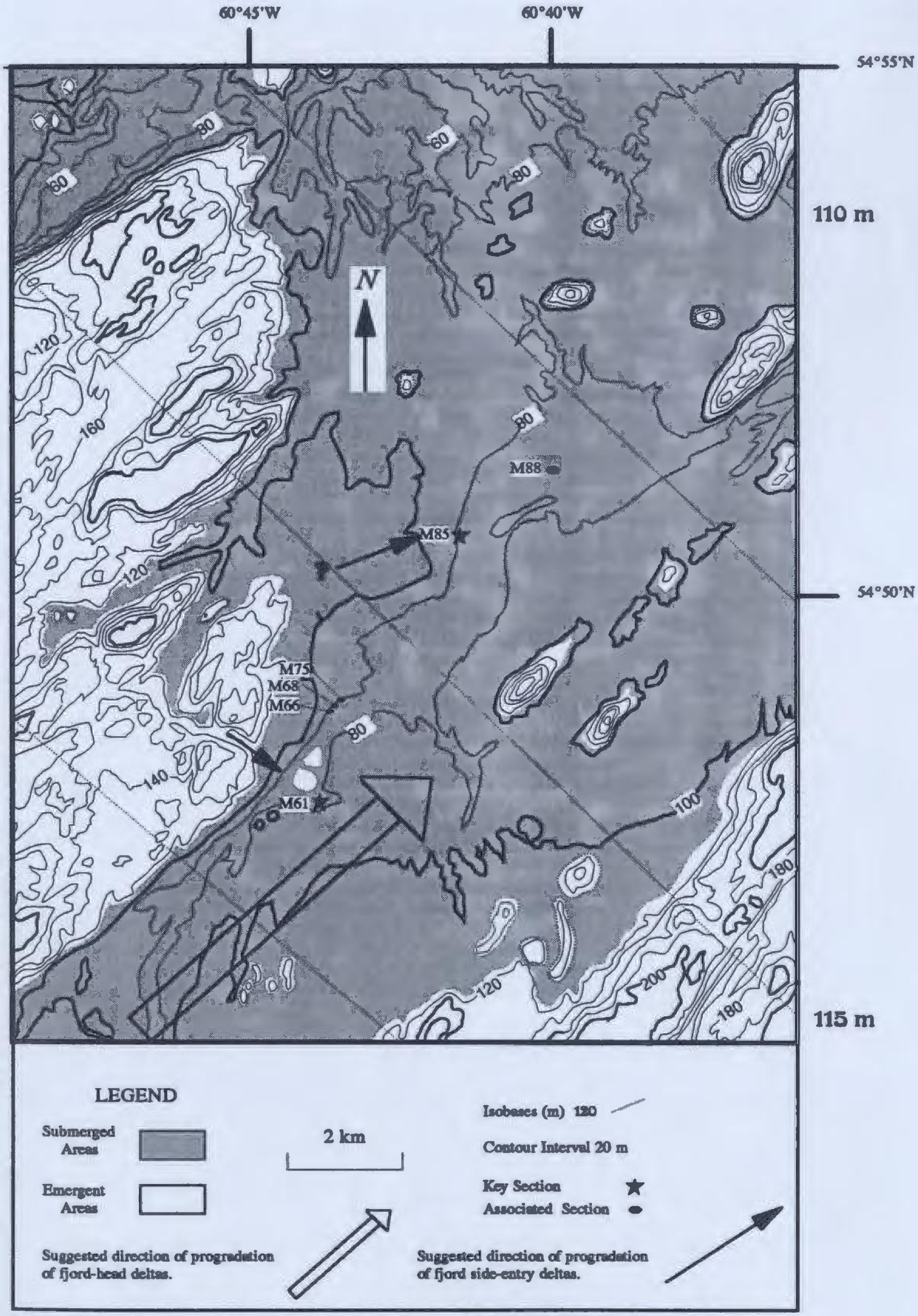


Fig. 3.26. Suggested shoreline position at 8 ka. The suggested directions of progradation of fjord-head and fjord side-entry deltas are also show. Note that the areas above 200 m elevation are not contoured. Ice may have been present in the area but its positions are unknown.

direction of progradation down the valley. These deltas are therefore interpreted as side-entry features. Deltaic deposits at other sections may be related to side-entry deltas or the main fjord-head delta.

Glacier ice is believed to have retreated from the study area by 8 ka (at least the middle part of the area). Deglaciation at 8 ka rather than 9 ka can better account for the presence of the ca. 8 ka marine shells at an elevation of 75 m (Table 2.1). Based on the emergence data in Chapter 2, most of the middle part of the area would have been emergent by 7 ka. Hence, the change from a marine to a deltaic to a fluvial setting probably took place during the 1000 yr interval from 8 ka to 7 ka.

During the deposition of the upper part of Unit B, deposition from turbidity currents and underflows became more important because of the reduced salinity of the basin waters (as suggested by the microfossil data in Chapter 4) and the gradual progradation of deltas.

In the areas of sections M61 and M85, progradation of side-entry deltas led to the deposition of progressively coarser sediments and gradually thicker sand beds from different types of gravity flows. These gravity flows probably interrupted deposition from suspension settling from surface plumes. The apparent limited areal extent of these more sandy deposits near sections M61 and M85 might indicate compartmentalization of the basin floor by sills. For example, sections M61 and M68 are 1.7 km apart but have quite different stage 1 sediments and microfossil abundances (Fig. 3.25), consistent with some isolation from each other. The presence of a small area of bedrock outcrop between these two sections implies the presence of a buried bedrock sill that may have divided the basin. This situation is very common in many modern fjords (Syvitski *et al.*, 1987; Syvitski and Hein, 1991; Hein and Syvitski, 1992; Gilbert *et al.*, 1993).

There may have been several periods of relatively high discharge that affected the area. This is suggested by the presence of several upward thinning sandy intervals within the lower part of Section M85, although channel migration is an alternative explanation for these cycles. There are several sandy units in the lower part of Section M61 (subunits b3 and b5) that may have a similar origin, but they are not so well developed as in Section M85.

In stage 2, isostatic rebound and progradation of the main fjord-head delta and the side-entry deltas resulted in the deposition of thicker sandy and gravelly deposits in the area. This probably took place in the first few hundreds of years after the area started to emerge (assuming that the rates of emergence suggested in Chapter 2 are correct and that there was no delay due to nearby ice loads). The gravelly foresets in Section M61 suggest proximity to the river mouths. This stage probably ended with the deposition of muddy debris flows of Unit E in sections M60, M61 and M62. These delta-top debris flows apparently remobilized glaciogenic material.

Delta topsets might have been deposited at many section locations during the period of still stand that was inferred by Batterson *et al.* (1988), based on a lower delta surface at 95 m in the Kaipokok Valley area. River channels of stage 3, however, are believed to have eroded the upper part of the deltaic sequence at many localities (e.g. sections M66, M68, M75, M85, M88; Fig. 3.25).

In stage 3, after the area became emergent, river systems were established. These rivers where probably characterised by high and fluctuating discharge and deposited thin gravel beds in shallow channels at several sections (Fig. 3.25). This river system may have included braided or wandering channels as in other recently deglaciated areas (Church, 1983; Brierley, 1989).

3.3. SECTIONS IN THE UPSTREAM PART OF THE AREA

Quaternary strata in this part of the area show the greatest lateral and vertical variability. Exposures are limited and very poor. Several sections are described below.

3.3.1. KEY SECTION U50

This section was described from an approximately 18.5 m-high terrace. The exposed sediments at this section are approximately 9.15 m thick (Fig. 3.27). Unit B forms the lower 2.9 m of the exposed sediments. There is a an approximately 6 m covered interval between the base of Unit B and the river water level that consists of semiconsolidated material that could not be trenched. Above Unit B is a 3.4 m covered interval that could not be described due to the continuous collapse of the sediments from the sides of the trench. Unit E overlies this unstable

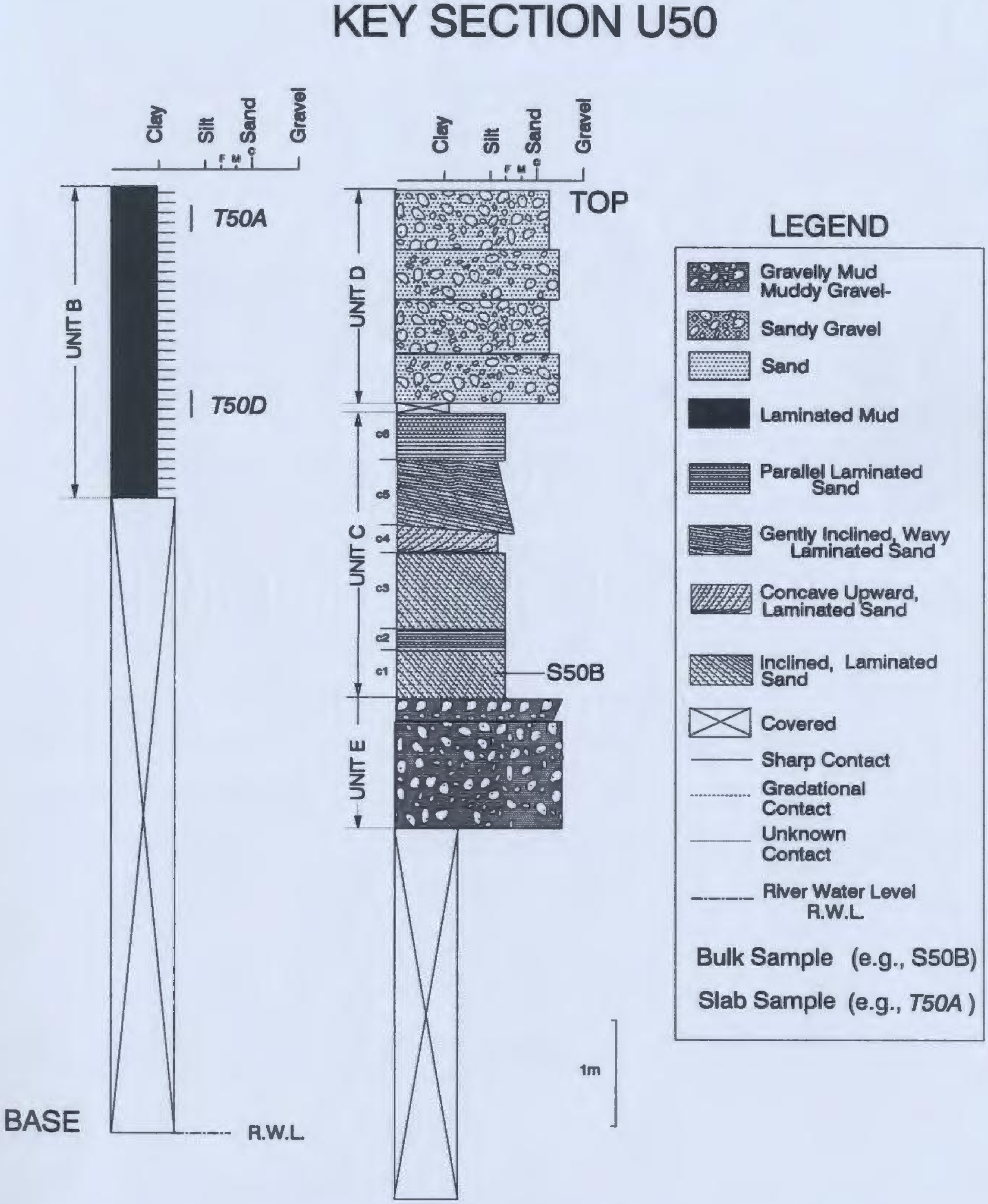


Fig. 3.27. Key Section U50 stratigraphic column. See text for details.

interval and is 1.2 m thick. The base of Unit E is covered. Unit C is 2.65 m thick and sharply overlies Unit E. Unit D, 2 m thick. It overlies a 0.2 m interval of vegetation that is present at the top of Unit C and forms the top of the section.

3.3.1.1. UNIT B

Unit B consists of couplets of fine sand or silt and mud. Colour of the mud is light red (5 R 6/6) to moderate reddish brown (10 R 4/6). Colour of the sand is very light grey (N8) to pinkish grey (5 YR 8/1).

Thin sections made from sample trays T50A and T50D show two types of couplets. The first type of couplet consists of normally graded 6-15 mm-thick sand to silt layers overlain by thin (1 to 5 mm) mud drapes. The lower sand-silt division lacks any clay and the overlying mud has a sharp base (L_{TV} in Table 3.2). The lower contact of the sand-silt division is sharp, planar to slightly irregular, with small scour marks. The mud drapes are massive and do not show any evidence of bioturbation.

The second type of couplet consists of normally graded, 4-12 mm-thick muddy sand laminae or very thin beds that alternate with mud. These couplets differ from the first type in that there is a progressive upward decrease in grain size and an increase in mud so that there is no sharp break in grain size between the sandy part of the couplet and the muddy part (L_{III} in Table 3.2; Fig. 3.28). Particle size in the sandy part is approximately 0.1 mm. The muddy division may be massive or bioturbated. Grain size analyses of bulk samples of Unit B from the upper part (sample S48 and S51 in slab T50A) and the lower part (sample S47 in slab T50D) show that this unit consists of 34.5% - 70.1% sand, 13.9% - 38.2% silt and 16% - 31.7% clay. The samples have a polymodal grain size distribution (Appendix B) with a coarse mode of 3 to 4.5 ϕ.

3.3.1.2. INTERPRETATION OF UNIT B

The first type of couplet lacks tractional structures. The well developed normal grading in the lower sandy part of each couplet, sharp mainly planar contacts and the break in grading between the lower sand-silt part and the overlying mud part are believed to limit the interpretation

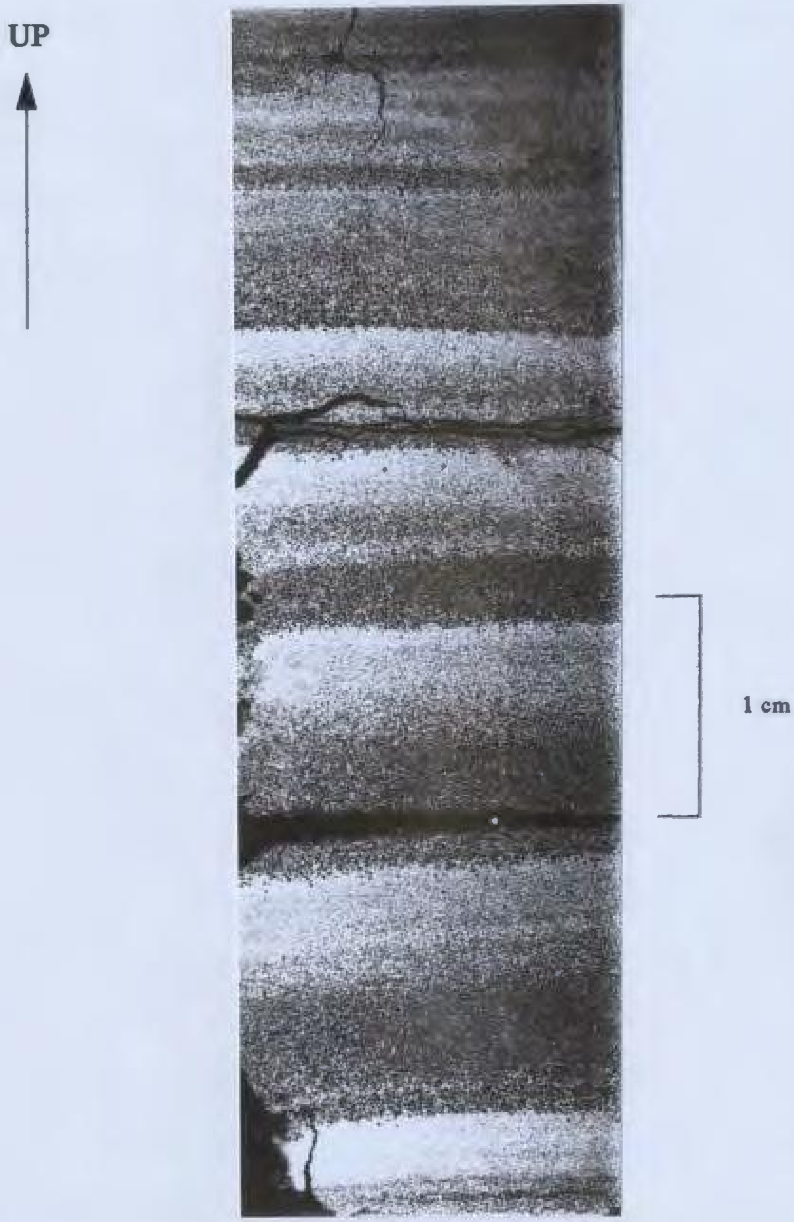


Fig. 3.28. Normally graded laminae in Section U50. Sand is dark and mud is light. These laminae are assigned to type L_{III} in Table 3.2.. The laminae are interpreted as turbidity current deposits, underflow deposits, or proximal overflow deposits (cf. Edwards, 1986; his Fig. 13.10).

of these couplets to underflows or turbidity currents.

The sharp break in grading between the lower sandy division and the overlying mud drape suggests that the deposition of each sand/mud or silt/mud couplet was not a continuous event (Ashley, 1975). The lack of sedimentary structures typical of turbidites and the break in grading does not favour a turbidity current origin for these laminae. An underflow origin may be more appropriate; the break in grading (i.e., sharp contact between the lower sand-silt part and the upper muddy part in each couplet) is more typical of underflow deposits (Gilbert, 1983; Mackiewicz *et al.*, 1984). Normal grading may also be developed in underflow deposits (Smith and Ashley, 1985). The somewhat thick nature of these laminae and their high sand-silt/clay ratio suggest deposition close to the underflow source (Powell, 1981), which was probably a continuously flowing fresh water outlet (Mackiewicz *et al.*, 1984). In such a proximal setting, the salinity of the basin waters may have been considerably diluted so that underflows formed easily.

Features of the second type of couplet in Unit B limit the interpretation to two possibilities. The gradual normal grading suggests that each lamina was deposited during one event. These may be turbidites deposited from density currents of very low concentration. Alternatively, these laminae may have been deposited by suspension fall out from surface sediment plumes. Settling from surface sediment plumes produces normally graded layers with sharp basal contacts (Mackiewicz *et al.*, 1984; Edwards, 1986; his fig 13.10b). If deposited from surface sediment plumes, grain size and thickness of the laminae suggest deposition close to source (Powell, 1981; Mackiewicz *et al.*, 1984). Similar graded beds have been termed "cyclospasms" by Mackiewicz *et al.* (1984).

3.3.1.3. UNIT E

The diamictons of Unit E consist of two beds of consolidated, gravelly mud - muddy gravel. The base of the lower bed is covered: only 1 m is exposed. The upper bed is 20 cm thick. The contact between the upper and lower beds is marked by a planar surface across which there is a subtle colour change and a slight upward decrease in the concentration of clasts.

The clasts in the lower bed range in size from 2 to 20 cm. They are subrounded to very well rounded. Most clasts are spherical but some are ellipsoidal. Bullet shaped clasts are also present. The percentage of clasts ranges from 40-70%. The matrix is sandy mud - muddy sand, in which the sand grains are coarse to fine and angular. Colour of the whole bed on fresh wet surfaces is greyish red (5 R 4/2) to dark reddish brown (10 R 3/4).

The lower gravelly bed lacks internal structure or grading. The clast fabric appears random. Texture ranges from framework supported to matrix supported.

The upper gravelly bed is characterised by reverse grading. Maximum clast size increases upwards from 2 to 20 cm. Some large clasts are concentrated at the top of the bed and project into the overlying sand beds of Unit C. The clast shapes and matrix are similar to those of the underlying gravelly bed. Colour of this upper gravel bed on wet fresh surfaces is moderate red (5 R 5/4) to greyish red (10 R 4/2).

3.3.1.4. INTERPRETATION OF UNIT E

The diamictons of Unit E share features with examples of glaciomarine (ice-rafted) diamictons, tills and sediment gravity flow deposits. Features such as lateral extent, lower bounding surface and thickness that are used by others (e.g. Anderson, 1983) could not be evaluated in the field due to the limited exposure. The clast shape and roundness suggest that the lower bed is till or a debris flow deposit, while the bullet shape of some clasts clearly suggests a glacial source.

The upper bed is interpreted as a debris flow deposit. This is based on the inverse grading and projecting clasts (*cf.* Shultz, 1984)

3.3.1.5. UNIT C

Unit C is divided into six different subunits based on the different thickness of the laminae observed in each subunit and the angle of dip of each set of laminae in each subunit. Unit C is quite different in this section to Unit C in other sections in that it consists mainly of moderate to well sorted, fine to medium grained sand. The colour of the sand ranges from light brown (5 YR 6/4) to greyish orange (10 YR 7/4).

3.3.1.5.1. Description of subunit c1

This subunit is 45 cm thick and consists of laminated (each 0.5 - 0.8 cm), fine to medium grained, moderate to well sorted sand. One bulk sample (individual laminae could not be sampled) is fine sand (mean = 2.27 ϕ), moderately well sorted (σ = 0.67 ϕ) and near symmetrically skewed (sample S50B, Appendix B).

Individual laminae show a crude normal grading from medium to very fine sand. The laminae dip at an angle 20°-30° towards the E-SE. The lower contact of this subunit is sharp and slightly irregular. The individual laminae meet the lower bounding surface at a sharp angle.

3.3.1.5.2. Interpretation of subunit c1

This subunit is interpreted to represent a single planar tabular cross bed set. The angular contact of the foresets with the lower bounding surface suggests this set formed by the migration of a 2D dune (Harms *et al.*, 1982).

The anomalous character of Unit C in this section, including its fine grain size, lack of clay drapes and moderate to good sorting suggests that these are aeolian sands (Glennie, 1970). Aeolian sand is common in glaciated and recently deglaciated areas where abundenet unconsolidated sediment can be mobilized by winds blowing off nearby glaciers (Ashley, 1985).

The normal grading observed in each lamina suggests that grainfall may have been the principle processes by which these dunes were formed; grainfall processes are closely associated with avalanches and sand flow processes that form dunes (Ahlbrandt and Fryberger, 1982).

3.3.1.5.3. Description of subunit c2

This subunit consists of two beds, each 10 cm thick. Each bed is internally stratified. The 0.5 - 1.5 cm-thick strata show crude normal grading from medium to fine sand. These strata have sharp, planar lower and upper contacts. Upper and lower contacts of subunit c2 are sharp and planar. This subunit is flat lying to gently dipping (< 5°) towards the W-SW.

3.3.1.5.4. Interpretation of subunit c2

This subunit is also interpreted as an aeolian deposit based on the overall sediment texture of Unit C. The two beds may represent sheet sands produced by ripple migration or grainfall (Hunter, 1977; Ahlbrandt and Fryberger, 1982; Ashley 1985). Crude normal grading suggests that these are products of grainfall rather than saltation transport during wind ripple migration (Ahlbrandt and Fryberger, 1982).

3.3.1.5.5. Description of subunit c3

This subunit is 70 cm thick and consists of laminated fine to medium grained sand. Individual laminae show well developed normal grading from medium to fine sand. The laminae are planar to slightly wavy. Isolated lenses (maximum thickness 3 cm and 10-15 cm long) of coarse sand are present in this unit. These lenses are draped by the other laminae and are conformable with the underlying laminae. These lenses are the cause of the waviness observed in the other laminae in this subunit. All laminae dip at an angle of 20°-30° towards the E-SE. They meet the lower bounding surface of this subunit at a sharp angle. Upper and lower contacts of this subunit are sharp and planar and slightly inclined (<5°) in the E-NE.

3.3.1.5.6. Interpretation of subunit c3

This subunit is interpreted as a planar tabular cross-bed set like subunit c1. The well developed normal grading suggests that the laminae were produced by grain saltation during wind-ripple migration (Ahlbrandt and Fryberger, 1982; Ashley, 1985). The isolated lenses observed may represent sandflow lenses formed as a result of avalanches down the lee side of a dune, or sand accumulation behind an obstacle of vegetation (Hunter, 1977; Ahlbrandt and Fryberger, 1982).

3.3.1.5.7. Description of subunit c4

This subunit is 20-25 cm thick and consists of well sorted, cross bedded, fine to very fine sand. The foresets are inclined 20°-25° towards the W-NW, opposite to the dip of cross-beds in the underlying subunits. The dip decreases towards the base of the set, and the contact with the lower bounding surface is tangential. The foresets are composed of 1-2 cm-thick strata that are separated by concentrations of dark grains. This subunit was 5 cm in thicker on one side of the trench than on the other side (ca. 1 m wide trench).

<u>3.3.1.5.8. Interpretation of subunit c4</u>

This subunit is also interpreted as an aeolian deposit based mainly on its texture and structure. The variation in the thickness of the subunit across the width of the trench suggests that this subunit represents a wedge-shaped cross bed set (Ahlbrandt and Fryberger, 1982). The lack of deformational structures suggests that sandflow rather than slumping was the main processes of foreset sedimentation (Ashley, 1985; Halsey *et al.*, 1990). The concave upward form of the foreset strata suggests that cohesion and the presence of moisture may have been factors influencing deposition of the cross set (Halsey *et al.*, 1990). The W-NW dip is interpreted to indicate a wind direction different to the prevailing direction that formed subunits c1, c3 and c5.

<u>3.3.1.5.9. Description of subunit c5</u>

This subunit is 60 cm thick. In the lower part, it consists of gently inclined beds,each 5-10 cm thick, that dip < 10° NE. The beds consist of fine to coarse, poorly sorted sand. Individual beds are internally laminated (< 1 cm). Each lamina is normally graded from coarse to medium sand in the lower part, to medium to fine sand in the upper part. The contacts between individual laminae are sharp and planar.

In the upper part, this subunit consists of faintly stratified (0.5-1.5 cm) fine sand. The strata are wavy and have a gentle (<5°) dip towards the northeast. Isolated lenses (2 cm maximum thickness and 5-10 cm wide) of medium grained sand are present between the strata. The contact between this subunit and subunit c4 is sharp and slightly irregular with small scours (1-2 cm deep and 3-5 cm wide). This contact is inclined approximately 5° towards the NE. This subunit increases in thickness to 40 cm (across the ca.3 m width of the outcrop) in the NW direction.

3.3.1.5.10. Interpretation of subunit c5

This subunit is interpreted as an aeolian deposit based primary on its texture. The lamination may have been produced by saltation during wind ripple migration or by grainfall (Ahlbrandt and Fryberger, 1982). The well developed normal grading observed in individual laminae suggests that the laminae were deposited by saltation during wind ripple migration.

This subunit may represent a transverse section of a dune with the laminae produced by avalanching down the lee face (Ahlbrandt and Fryberger, 1982; Ashley, 1985). This interpretation would require that the observed dips of < 10° are only apparent dips in the plane of the trench wall.

In the upper part, the poorly developed normal grading and the draping of the laminae over the medium grained sand lenses suggest that grainfall rather than wind ripple migration was the dominant process. The medium grained sand lenses are interpreted as scour fills that formed around obstacles or vegetation, or they might represent individual grainfall deposits (Hunter, 1977; Ahlbrandt and Fryberger, 1982). The association of sandflow and grainfall deposits is common in aeolian deposits (Ahlbrandt and Fryberger, 1982).

3.3.1.5.11. Description of subunit c6

This subunit is 45 cm thick and consists of flat lying laminae (0.8 cm) of fine to medium, moderate to well sorted sand. The laminae are planar and straight with sharp contacts. Laminae are distinguished by the concentration of dark heavy minerals. This subunit sharply overlies subunit c5 and underlies a vegetation cover (0.2-0.3 m thick).

3.3.1.5.12. Interpretation of subunit c6

This may represent a transverse section of a dune, or a sheet sand at the edge of the dune area. The well sorted texture is the primary evidence for an aeolian origin. The sharp and distinct laminae suggest that ripple migration was the main processes of deposition rather than grainfall (Ahlbrandt and Fryberger, 1982).

3.3.1.6. UNIT D

Unit D is 190 cm thick and consists of four beds, each 40-50 cm thick, of crudely stratified gravel with a coarse sand matrix. Clasts are 10-20 cm across, subrounded to very well rounded. Contacts between the beds are sharp and planar. The beds are differentiated by the size of the clasts within each bed. The beds vary from clast-supported to matrix-supported.

This unit forms the top of the terrace. Its lower contact was not observed and there is no cover on top of the bed.

3.3.1.7. INTERPRETATION OF UNIT D

This unit is interpreted in a similar way to Unit D in Section M85, as gravel bars that formed in shallow channels of a river characterized by high but variable flow velocity (indicated by the large clast size).

3.3.1.8. DISCUSSION OF SECTION U50

This section exposes a sequence of sediments that shows variable depositional mechanisms and environments. The fine sand-silt/ mud couplets in the lower part of the sequence (Unit B) are inferred to have been deposited in the proximal end of a marine embayment. The salinity of the embayment was probably reduced by fresh water input, enabling the formation of underflows. Proximity to the source of fresh water is suggested by the presence of thick couplets that may have been deposited from surface sediment plumes. Such thick couplets are recognized in areas close to fresh water input points in many fjords (Mackiewicz *et al.*, 1984; Cowan and Powell, 1990).

The diamictons of Unit E are problematic. If they are of glacial origin, there is an inconsistency with the observation that no ice-rafted material (pebbles or cobbles) was observed in any of the sections in the thesis area. A glacial re-advance would be required. Sediments inferred to have been deposited in a glacial setting are observed upstream of the area, but the relationship between these glacial sediments and the marine sediments is not known due to the limited exposure (see discussion of Section U32). Batterson *et al.* (1988) suggested that there may have been a period of ice re-advance. This re-advance may have been minor (M.J. Batterson, personal communication, 1992). The upper diamicton bed of Unit E is interpreted as a debris flow based on its inverse grading. Debris flows are certainly common in ice-proximal areas (Lawson, 1981, 1982), but also occur in many other settings.

Both beds of diamicton may have been deposited by debris flows unrelated to glacial ice, either subaerially or subaqueously. There is no conclusive evidence regarding this last issue. The bullet shaped clasts observed in these diamictons suggest that the original material was derived from till.

Sediments of Unit C are interpreted as aeolian deposits. The restriction of these deposits to this section suggests that local topography may have played an important role in the formation of the aeolian deposits. Section U50 is surrounded by many hills that may have baffled down-valley winds. On aerial photographs, many sand dunes (especially in the middle part of the area) are localized on the lower slopes of hills which probably acted as a wind barrier. There are no modern dunes within the area of Section U50, however. The small thickness of cross-bed sets and the presence of parallel laminated intervals in Unit C suggests deposition at the margin of a dune field or as interdune aeolian sands (Ahlbrandt and Fryberger, 1982).

Finally, Unit D is inferred to have been deposited by a braided or a wandering fluvial system carrying glacial meltwater to the Labrador Sea. The fluvial system was probably larger than the present fluvial system (Klassen and Thompson, 1993), with greater velocity as indicated by the large clast size. Many of the terraces observed in the Kanairiktok Valley may have been formed by such a fluvial system.

Thus, sediments in Section U50 reflect a transition from marine deposition in a fjord-type setting to fluvial deposition. Unlike the other sections in the middle and downstream part of the area, this transition is believed to have been interrupted by a period of aeolian sedimentation and, perhaps, a minor glacial re-advance.

3.3.2. ASSOCIATED SECTION U45

This section is located approximately 6 km upstream of Section U50. Sediments of units

B and E are exposed in an approximately 3.9 m high terrace. The base of the exposure is located 15 cm above the river water level (Fig. 3.29).

3.3.2.1. DESCRIPTION OF SEDIMENTS IN SECTION U45

Unit B consists of interstratified sand and mud that is 85 cm thick. The sand beds decrease in thickness from 3-5 cm in the lower part to 0.5 cm laminae in the upper part. The sand strata consist of sharp based, normally graded, fine to very fine sand that grades into the overlying mud. The mud beds are 1 cm thick and massive. In the upper part of the unit, the sand laminae are faint and have diffuse upper and lower contacts. Grading is not well defined and the sand laminae are fine to very fine grained.

Unit E sharply overlies Unit B with an irregular contact. It is 2.9 m thick. The diamictons in this unit are similar to the other diamictons of Unit E that are present in the upstream part of the area (Fig. 3.2a). They consist of clasts that range in size from 2-20 cm and are set in a sandy mud matrix. The clasts are rounded to very well rounded. Some are elliptical and bullet shaped. Percentage of clasts ranges from 40-70%. The unit ranges from matrix-supported to clast-supported. This unit is massive (i.e, lacks grading, intercalations or bedding planes).

3.3.2.2. INTERPRETATION OF SEDIMENTS IN SECTION U45

The thicker sand beds in the lower part of Unit B are interpreted to have been emplaced by turbidity currents or underflows, based on their grading and sharp lower contacts. The mud separating these sand beds was probably deposited from suspension settling from surface sediment plumes and dilute turbidity currents. The faint laminae in the upper part of Unit B are interpreted to have been emplaced either by dilute turbidity currents or as a result of suspension settling from surface sediment plumes, or both. The well developed upward decrease in the thickness of the sand beds suggests that the source of the turbidity currents or underflows became less active with time during the deposition of this unit.

Unit E has a similar interpretation to Unit E in Section U50: till or debris flow deposits.

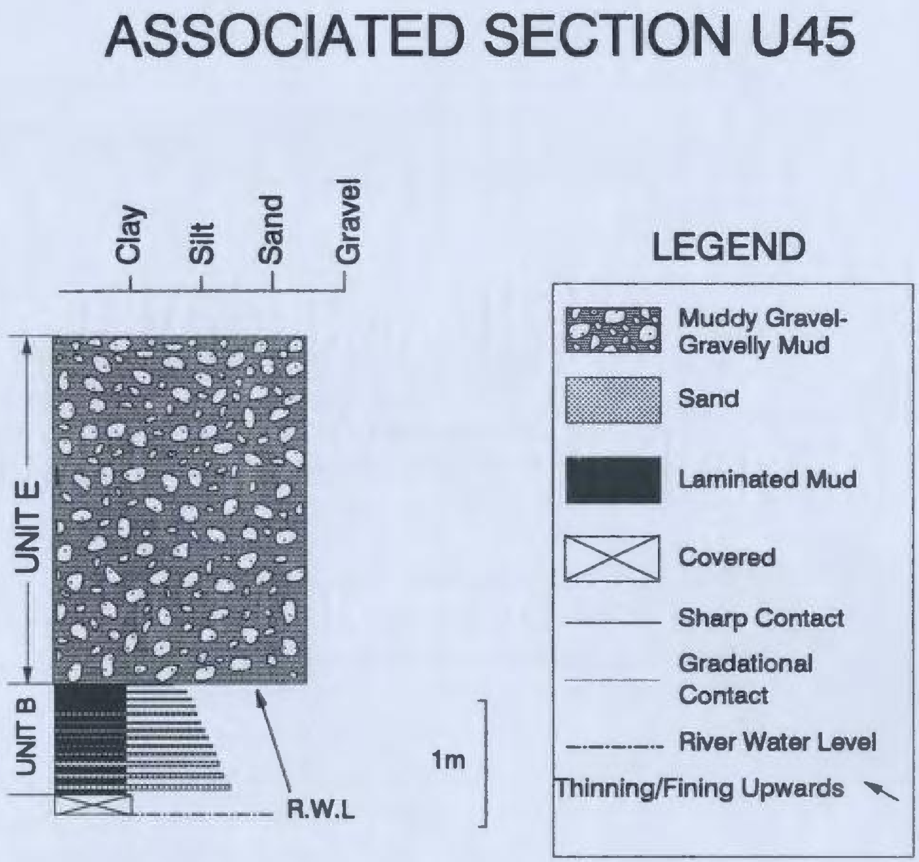


Fig. 3.29. Associated Section U45 stratigraphic column. See text for details.

The sharp and irregular basal contact is believed to rule out ice rafting, a: d no ice rafted material has been recognised in Unit B.

3.3.2.3. DISCUSSION OF SECTION U45

Unit B is interpreted as delta bottomsets or fjord bottom deposits. Unit B only occurs twice further upstream (sections U32 and U35), where it is even thinner than this occurrence (Fig. 3.2c). Either Unit B was never very thick in the upstream part of the study area due to the short duration of time of maximum flooding (Chapter 2), or Unit B has been largely eroded in this proximal area, or Unit B occurs below the river water in the upstream area.

The interpretation of Unit E has many consequences for the history of the area. If Unit E diamictons were deposited by debris flows, then the hypothesis that glaciers had receded onto dry land before deposition of any of the later Quaternary sediments in the thesis area needs no modification. If Unit E diamictons are tills, then a glacial re-advance is required. The magnitude of any such glacial re-advance was probably minor, with a small quantity of ice flowing down the valley (M.J. Batterson, personal communication, 1992). This hypothesis will discussed in § 3.3.6.

3.3.3. KEY SECTION U32

This section is located approximately 7 km upstream of Section U45. This section was described from the downstream end of a sinuous N/S trending ridge (0.6 km long and approximately 200 m wide, based on aerial photographs and topographic maps). The exposed sediments at this location are approximately 10.3 m thick. Units B, C and E are present in this section (Fig. 3.30).

3.3.3.1. UNIT B

This unit is present at the base of the section. It is 50 cm thick and consists of interstratified sand and silty mud. Sand beds range in thickness from 2-7 cm separated by thin layers of silty mud (0.5-2 cm). The sand beds have flat, sharp bases. They are normally graded

KEY SECTION U32

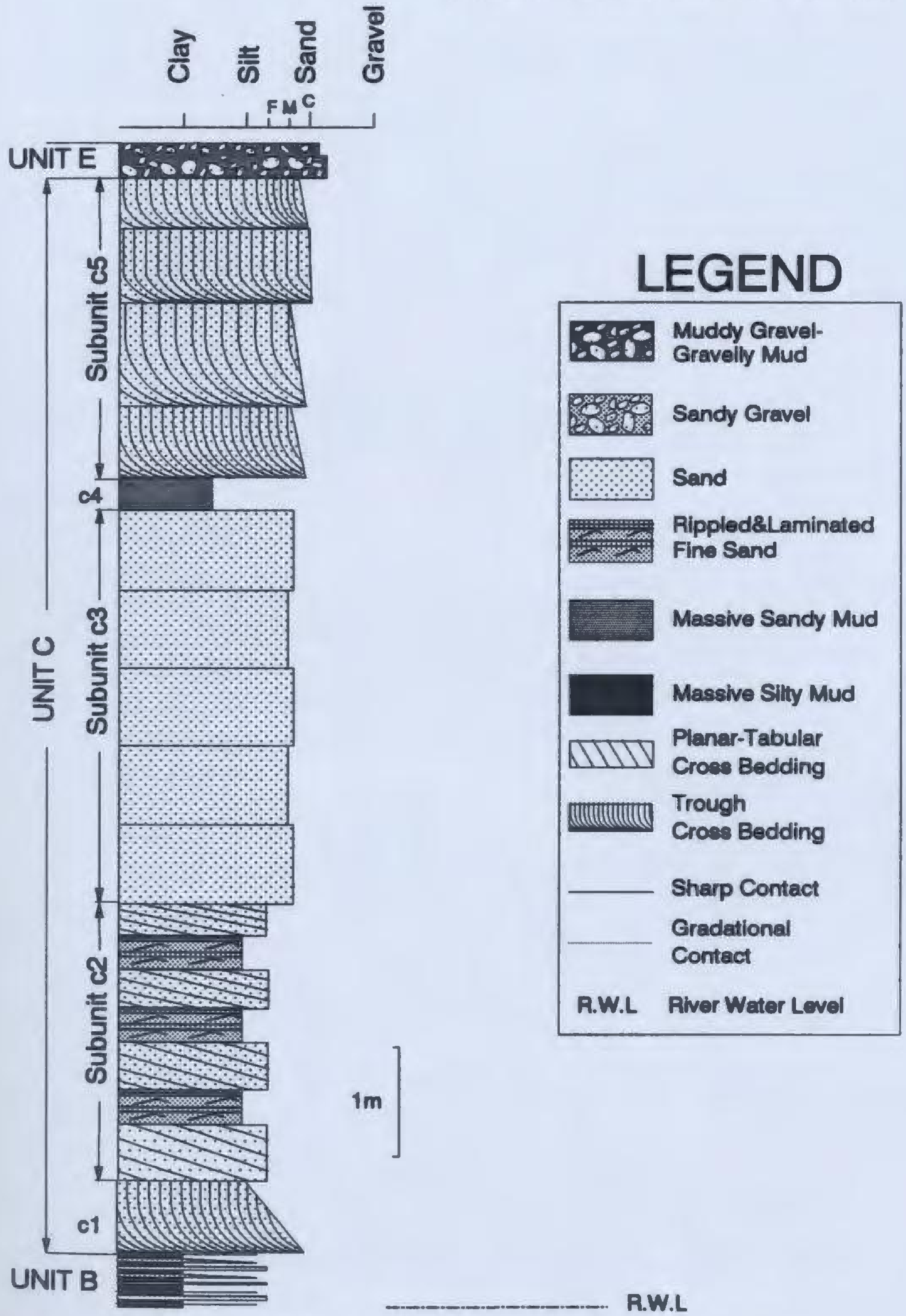


Fig. 3.30. Key Section U32 stratigraphic column. See ext for details.

from fine sand into the overlying mud, or massive with sharp upper and lower contacts. The mud interstrata are massive. Colour of the mud is moderate red (5 R 5/4 - 5 R 4/6) while the sand beds are light grey (N7) to medium light grey (N6).

3.3.3.2. INTERPRETATION OF UNIT B

The graded sand beds are interpreted to have been deposited from underflows, or turbidity currents or high concentration surface plumes. The mud layers were probably deposited from suspension. The massive sand beds probably were deposited from grain flows or liquefied flows or underflows. Justification for these interpretations is given for Unit B exposures in the other parts of the area.

3.3.3.3. UNIT C

This unit is dominated by sand and forms most of the exposed sediments in this section. Unit C is divided into 5 subunits based on sedimentary textures and structures. Subunits are numbered and described from the base upward (Fig. 3.30).

3.3.3.3.1. Description of subunit c1

Subunit c1 is 65 cm thick and consists of a single set of cross bedded sand. The basal contact of this set with the underlying Unit B is sharp and irregular. The upper contact with the overlying subunit c2 is sharp and planar. The sand is coarse to very coarse grained and moderate to poorly sorted. A few pebbles (0.5-1 cm) are present near the base of this subunit. Foreset strata are 1-3 cm thick and dip 20° - 35° towards N 5°E. The angle of dip of the foresets decreases to 15°-20° near the base of the exposure.

3.3.3.3.2. Interpretation of subunit c1

This set of cross bedded sand is interpreted to have been deposited by a migrating bedform. The 65 cm minimum height of the bedform suggests that it was a large or very large dune (*sensu* Ashley *et al.*, 1990). The downward decrease in the angle of dip of the foresets suggests that the bedford was a 3D dune (Harms *et al.*, 1982).

3.3.3.3.3. Description of subunit c2

This subunit is approximately 2.6 m thick and consists of alternating (a) moderately to poorly sorted, planar-tabular cross bedded, coarse to medium grained sand, and (b) stratified (0.5-1.5 cm) and ripple cross laminated fine sand. Internal contacts are planar to slightly undulatory. The cross sets show an upward decrease in thickness from 45 cm to 20 cm. Foreset strata are 1-3 cm thick, planar, and dip 20°-35° towards N 10°E - N 30°E (five palaeocurrent measurements of the trend of the maximum foreset dip in each set were N 10°E, N 17°E, N 19°E, N 24°E, N 30°E). The interbedded planar stratified and ripple laminated sand occurs in 20 cm intervals between the planar cross bedded sets. Some of these intervals consist entirely of planar stratified fine sand and laminae. The sand is poorly sorted. Contacts between the planar bedded sand and the ripple laminated sand within these intervals are sharp and may be planar or undulatory, preserving the ripple form below. These ripples are asymmetrical, sharp crested and show a palaeoflow direction towards N 22°E - N 43°E (eight palaeocurrent measurements were N 22°E, N 25°E, N 27°E, N 31°E, N 33°E, N 37°E, N 41°E, N 43°E). These ripples have heights of 1-3 cm and lengths of 10-15 cm.

3.3.3.3.4. Interpretation of subunit c2

The planar-tabular cross bedded sand was deposited by migration of 2D dunes (Harms *et al.*, 1982). The small range in palaeoflow direction suggests unidirectional flow. The planar foresets suggest a lower current speed than for the 3D dunes that generated the other subunits.

The planar stratified and cross laminated fine sand intervals were formed either by a variation in the current speed or the available grain size, both of which can cause a transition between ripples and plane beds (Simons and Richardson, 1961; Southard, 1991).

Intervals which consist entirely of plane beds of fine sand formed in the upper flow regime (Harms *et al.*, 1982). The presence of planar stratified sand alternating with cross laminated sand with or without ripple preservation suggests that parts of subunit c2 were deposited beneath currents of variable speed or shallow depth or both (Shaw, 1972).

3.3.3.3.5. Description of subunit c3

This subunit is 3.5 m thick and consists of moderately sorted, coarse to very coarse, thick bedded sand (each bed 70 cm). Beds are massive and have sharp planar contacts.

3.3.3.3.6. Interpretation of subunit c3

Massive beds that lack tractional sedimentary structures were probably deposited rapidly either from suspension or by sedimentary gravity flows (Arnott and Hand, 1989). The coarse grain size of individual beds suggests that if they were deposited from suspension, a sudden drop must have occurred in the velocity of a high speed current (Saunderson, 1975). Alternatively, these might be cohesionless debris flow deposits (*sensu* Postma, 1986).

3.3.3.3.7. Description of subunit c4

This subunit is 25 cm thick and consists of massive fine sandy mud. Colour on wet fresh surfaces is pale reddish brown (10 R 5/4) to moderate reddish brown (10 R 4/6).

3.3.3.3.8. Interpretation of subunit c4

This subunit is interpreted to have been deposited rapidly from suspension so that fine sand and mud were deposited together (Shaw, 1972).

3.3.3.3.9. Description of subunit c5

This subunit is 2.75 m thick and consists of cross sets, 50 - 70 cm thick, of coarse to very coarse, moderately to poorly sorted sand. The foresets are asympotitic to the lower bounding surface of each set. Set boundaries are sharp, planar to slightly undulatory. This subunit overlies subunit c5 with a sharp undulatory contact and underlies Unit E with a sharp contact marked by a sudden change in grain size.

3.3.3.3.10. Interpretation of subunit c5

The cross sets in this subunit, with their undulatory contacts, and the asymptotic nature of the foresets are inferred to have been deposited by the migration of large or very large 3D dunes (Harms *et al.*, 1982; Ashley *et al.*, 1990). Part of each set may have been eroded before the deposition of the overlying set, so that the preserved set thickness provides a minimum estimate of the original bedform height.

3.3.3.4. UNIT E

Unit E is 30 cm thick and consists of two diamicton beds that form the top of the exposed section. The texture is muddy gravel to gravelly mud. The upper bed is 10 cm thick and consists of 2-5 cm-diameter, well rounded pebbles set in a sandy mud matrix. Some of the clasts have a bullet shape. The pebbles form 40-60% of the bed. The framework is variably matrix-supported to clast-supported.

The lower bed is 20 cm thick and consists of material similar to that in the overlying bed. However, this bed has a lower percentage of clasts (20-50%) and the framework is mainly matrix-supported. The contact between the two beds is diffuse and only recognized by a difference in clast percentage. Colour of the beds is moderate reddish brown (10 R 4/6).

3.3.3.5. INTERPRETATION OF UNIT E

These two diamicton beds might have been deposited from thin debris flows. Glacial till are typically thicker than debris flow deposits (Anderson, 1983). The bullet shape of some of the pebbles suggests that the original material was probably derived from a glacial till. The interpretation of these diamictons is pursued in the next section.

3.3.3.6. DISCUSSION OF SECTION U32

The deposits of Unit C have many similarities to sequences interpreted as fluvial channel deposits (Shaw, 1972). The sinuous shape of the ridge where the section was measured and the mainly unidirectional palaeocurrent data suggest that Section U32 represents part of an ice tunnel deposit (i.e., an esker) (Banerjee and McDonald, 1975; Saunderson, 1975; Shaw, 1985).

The dominantly sandy texture of Unit C suggests that this section represents the flank of the esker or the downstream end of the esker (Shaw, 1972; Banerjee and McDonald, 1975; Saunderson, 1975). The lack of deformation due to the collapse and melting of the ice wall, typical of esker flank deposits, suggests that the deformed part was eroded by the modern river or that the section is not right at the esker margin. The lack of deformation also suggests that the esker was formed in a subglacial tunnel rather than an englacial tunnel (Boulton, 1972).

The sequence shows conditions of variable current speed and water depth. The lower subunit c1 may represent bars that formed at the base of the tunnel. Planar tabular cross bedded sand alternating with planar stratified and ripple cross laminated fine sand (subunit c2) indicate variations in current speed and water depth. There may have been a period of sudden decrease in current velocity causing the deposition of the sandy mud of subunit c3 (Banerjee and McDonald, 1975).

An increase in current speed is suggested by the presence of the massive "structureless" sand beds. These sands were probably deposited under upper flow regime conditions, with rapid rates of suspension deposition (Arnott and Hand, 1989). The upper subunit c4 was formed as 3D dunes migrated along the floor of the ice tunnel.

Unit E may have been deposited as a melt-out till from the roof of the ice tunnel. Given that it sits on the top of a ridge, a debris-flow origin from the slopes of the river valley is unlikely. Origin as a flow till from melting of the ice around the esker is certainly reasonable, however. The lack of deformation below the diamictons and their small thickness suggest that the ice roof was originally thin or that these diamictons were emplaced as a flow-till due to slumping as the ice melted (Shaw, 1972).

The occurrence of an esker above inferred marine deposits of Unit B proves that there was a glacial re-advance after the time of maximum marine flooding, near the time when this part of the area began to emerge from the sea.

3.3.4. ASSOCIATED SECTION U31

This section is located approximately 0.1 km upstream of Section U32 on the same esker ridge. Sediments of this section consist mainly of gravels of units D and E (Fig. 3.31).

3.3.4.1. DESCRIPTION OF SEDIMENTS IN SECTION U31

The lower 3.5 m consists of 60-70 cm-thick gravel beds of Unit D. Clasts are 15-20 cm and are set in a coarse sand matrix. Clasts constitute 50-70% of each bed. Beds are mainly clast-supported. Clasts are rounded to very well rounded and some are bullet shaped. Beds are differentiated by a variation in clast size. Contacts between individual beds are sharp and planar.

The lower part of Unit E is a 3.0 m-thick, inversely graded diamicton bed. Clast size changes from 5 cm in the lower part to 20 cm in the upper part. The clasts are well to very well rounded and some are elongated and have bullet shapes. Clasts form 35-65% of the bed. The bed is matrix-supported to clast-supported. The matrix is sandy mud. The bed is compact. Upper and lower contacts are sharp and planar.

The upper part of Unit E is 1.0 m thick and consists of thin to medium gravel beds. Clasts are 2-5 cm and are rounded to very well rounded. The matrix varies from sandy mud to muddy sand. Clasts form 20-70% of the beds. The beds are matrix-supported to clast-supported. The beds are differentiated either by a variation in the clast size or their percentage.

The upper two beds are each about 15 cm thick, have a coarse sand matrix and are normally graded from gravel to coarse sand. This is typical of Unit D (i.e., no muddy matrix).

3.3.4.2. INTERPRETATION OF SEDIMENTS IN SECTION U31

The lower gravel beds of Unit D are interpreted to have been deposited under high flow conditions in shallow channels, based on the tabular shape of the beds, large clast size and lack of features suggesting deposition from sediment gravity flows (Shaw, 1972; Hein, 1984). The 60-70 cm thickness of each bed suggests that each high energy event lasted for a considerable time (Shaw, 1972). These high flow conditions may have been variable as suggested by the different clast sizes in each bed. The sand matrix was probably deposited by infiltration during periods of

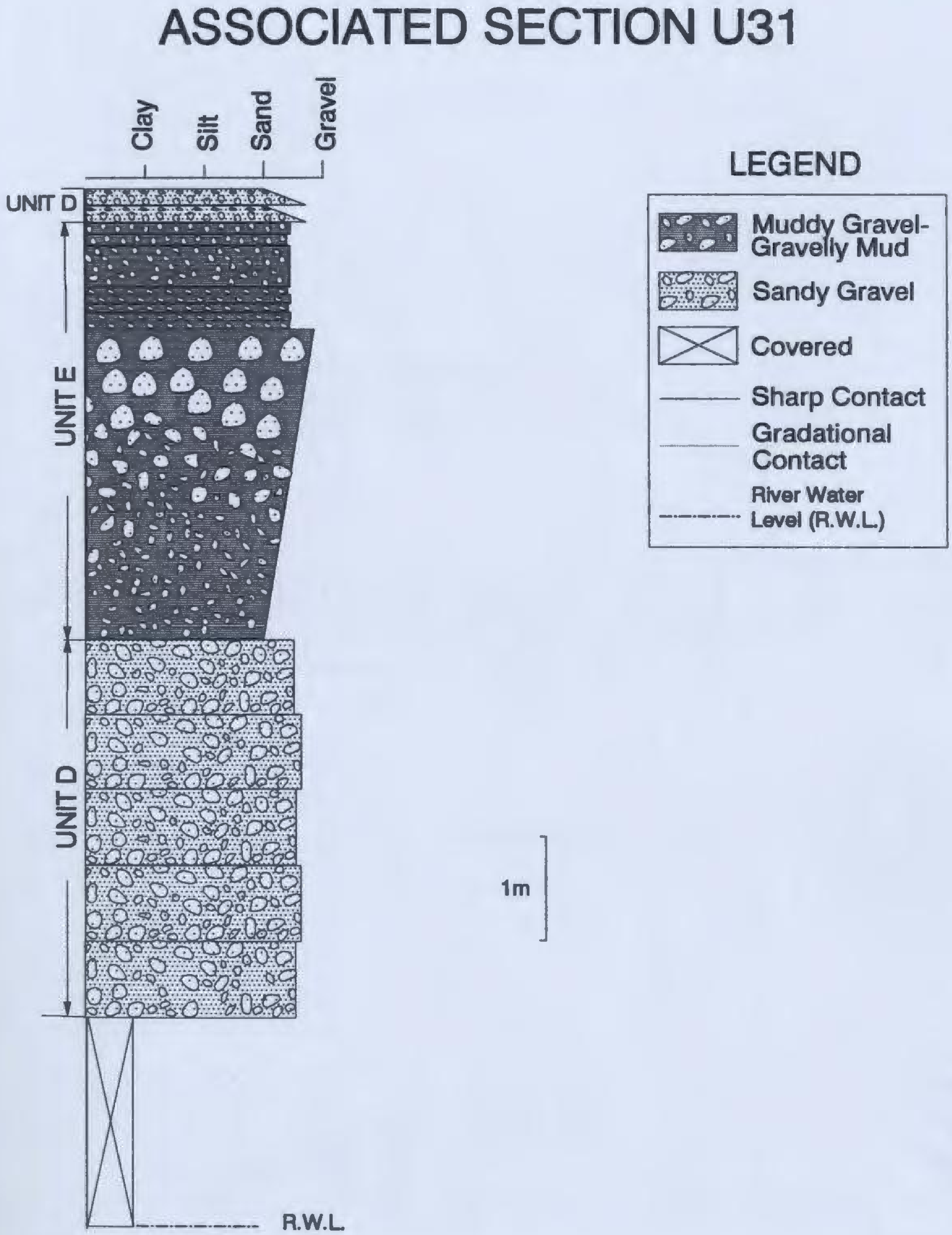


Fig. 3.31. Associated Section U31 stratigraphic column. See text for details.

weaker flow (Smith, 1985).

The inversely graded very thick diamicton of Unit E was probably deposited by a single debris flow. The inverse grading points to clast interaction during flow (Middleton and Hampton, 1976), and suggests that this is not a till. The upper thin to medium bedded gravel beds of Unit E that have a sandy mud matrix also may have been deposited by debris flows.

Beds at the top of the section that have a sandy matrix may have been deposited under high flow conditions that were relatively weaker than those that deposited the main part of Unit D. These flows may have been of short duration, depositing only thin beds.

3.3.4.3. DISCUSSION OF SECTION U31

The sediments exposed in this section suggest variable processes from high energy flow conditions to resedimentation by debris flows. The sediments are exposed in the same ridge as section U32 and are likewise interpreted to represent a part of an esker complex. The gravel beds of Unit D may represent a more central part of the esker where current speed may have been sufficiently high to transport gravel. The overlying inversely graded diamicton bed of Unit E may have been emplaced after the collapse of the ice walls and roof that were confining the esker. Similar deposits have been termed flow till by Shaw (1972). The lack of deformation at the base of this diamicton bed supports a gravity flow origin (Shaw, 1972). The upper thin gravel beds may have been deposited by several processes including thin gravity flows or high discharge fluvial transport following the retreat of the tongue of glacial ice responsible for the minor re-advance.

3.3.5. KEY SECTION U25

This is the most upstream section studied in detail. It consists entirely of muddy gravel or gravelly mud (diamicton) of Unit E (Fig. 3.32). The lithologic section was measured in an approximately 10 m-high terrace.

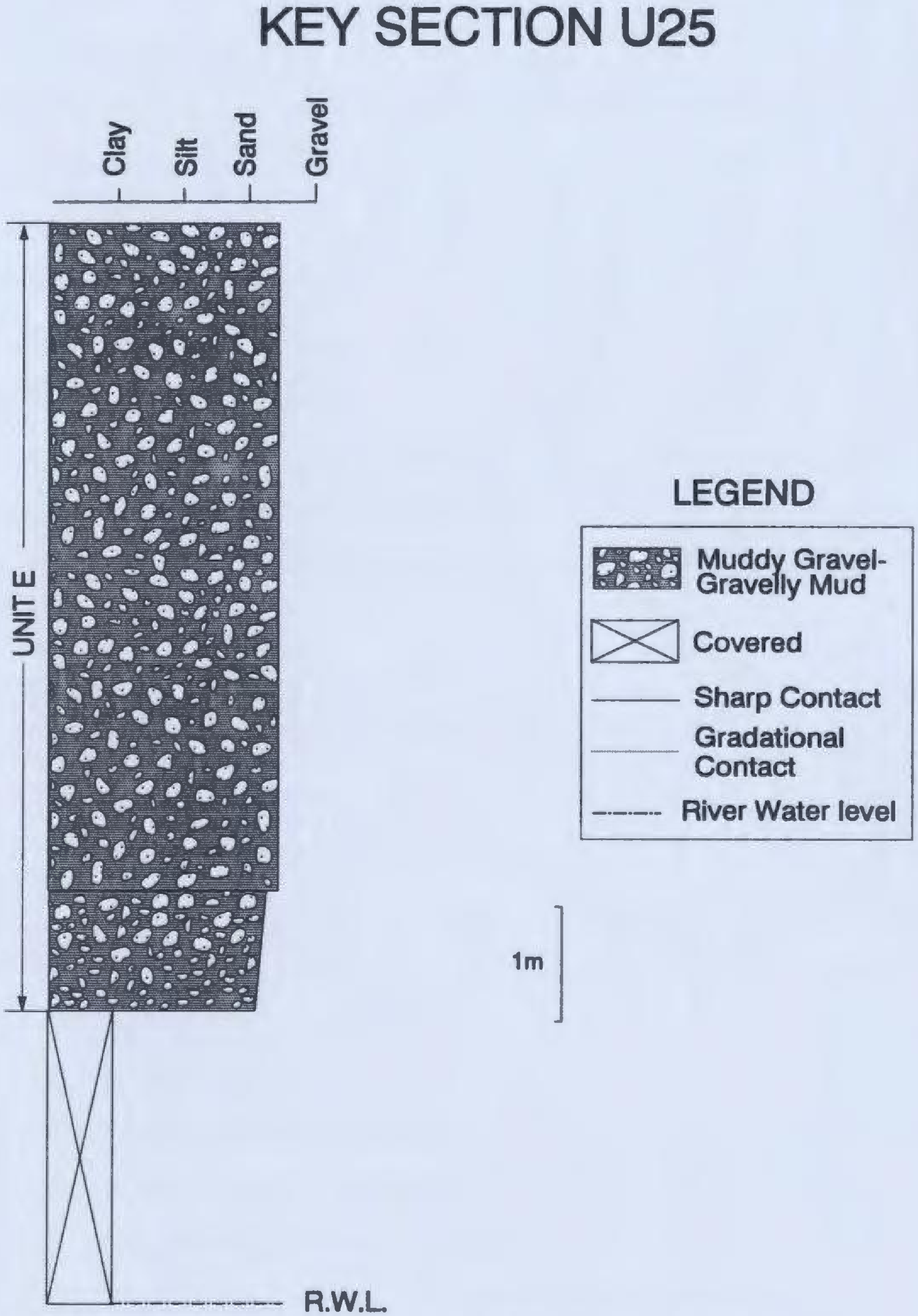


Fig. 3.32. Key Section U25 stratigraphic column. See text for details.

3.3.5.1. UNIT E

This unit consists of two beds of muddy gravel - gravelly mud (diamicton). The contact between the two beds is sharp and planar (as seen in a 1 m wide trench). The upper bed underlies a thin vegetation cover. The lower bed is 1 m thick and overlies 2.5 m of covered sediments. Clasts are of variable sizes; the maximum size is about 30 cm. The clasts are rounded to very well rounded. Most of the clasts are elliptical to spherical but some are rod and bullet shaped. These clasts sit in a matrix of sandy mud. Colour of the matrix on fresh wet surfaces is moderate red (5 R 5/4 to 5 R 4/6). The texture varies from clast-supported to matrix-supported. The clasts show a crude upward coarsening. Clasts form 40-60% of the bed.

Overlying the inversely graded bed is a 6 m-thick bed of gravelly mud - muddy gravel (diamicton). This bed in ungraded. The colour of the matrix on fresh wet surfaces is pale reddish brown (10 R 5/4). This bed is similar in texture to the underlying inversely graded bed.

3.3.5.2. INTERPRETATION OF UNIT E

These deposits may represent either till or debris flow deposits. The lack of any internal laminations or dropstone structures rule out a glaciomarine (ice rafted) origin. The inverse grading in the lowest bed suggests that this is a debris flow deposit (Middleton and Hampton, 1976). The bullet shape of some clasts suggests that the original material was till. The inferred debris flow probably originated by collapse and dilation of wet sediments on steep slopes. The upper bed may be till or a debris flow deposit.

3.3.5.3. DISCUSSION OF SECTION U25

The sediments in this section probably formed as a result of collapse of older till on steep slopes. However, it is possible that the upper thick diamicton is till that formed during the minor glacial re-advance or earlier. The age of this sediment relative to the marine inundation that formed Unit B is unknown, because Unit E occurs in isolation here. Its elevation relative to the upstream sections (Fig. 3.2c), however, suggests that it may be part of the Holocene sequence.

3.3.6. SEDIMENTARY ENVIRONMENTS IN THE UPSTREAM PART OF THE AREA

The sediments exposed in this area show the greatest lateral and vertical variability. The upward coarsening trend that is observed in the other parts of the study area is not developed here. The lower fine grained sediments (Unit B) are sharply overlain by coarse sand or gravel beds. The sedimentary stages below (Fig. 3.33) describe the suggested sedimentary environments in which each unit was deposited. As before, these stages have no time connotation. Each stage represents deposition under relatively comparable depositional conditions, but these conditions may have occurred at different times in the different sections in the area.

In stage 1, deposition was in a presumed marine embayment. This embayment is inferred to have existed during a period of relatively high sea level when glacial ice had retreated from the area (Fig. 3.34). Sedimentation was dominated by suspension settling from surface sediment plumes, by underflows, and by turbidity currents. The salinity of the water in the embayment is not known due to a lack of fossils, although other factors such as water turbidity may explain the lack of microfossils. Ice probably was not in contact with basin waters during the deposition of Unit B because this unit lacks any gravel clasts that may be interpreted as ice-rafted debris.

In stage 2, ice re-advance may have taken place in the upstream part of the area. The ice probably moved down the valley and eroded some of the underlying sediments. The lack of proximal deltaic deposits over the fjord bottom sediments, similar to what is observed in the other parts of the area, suggests that these deposits were eroded by the ice re-advance, or by rivers that left no depositional record.

The magnitude of the glacial re-advance is not known, although it is believed to be minor (M.J. Batterson, personal communication, 1992). An ice tunnel existed in the upstream end of the area. Glaciofluvial deposits (esker deposits) form a ridge in the area of sections U31 and U32. Tills associated with this ice advance may have been deposited in the surrounding areas (U25, U45 and U50). This re-advance may have only affected the upstream part of the middle part of the study area (§ Fig. 5.1). Alternatively, these diamictons may have been deposited by debris flows. These mass flows may have been related to the ice re-advance (e.g. flow till) or may be unrelated.

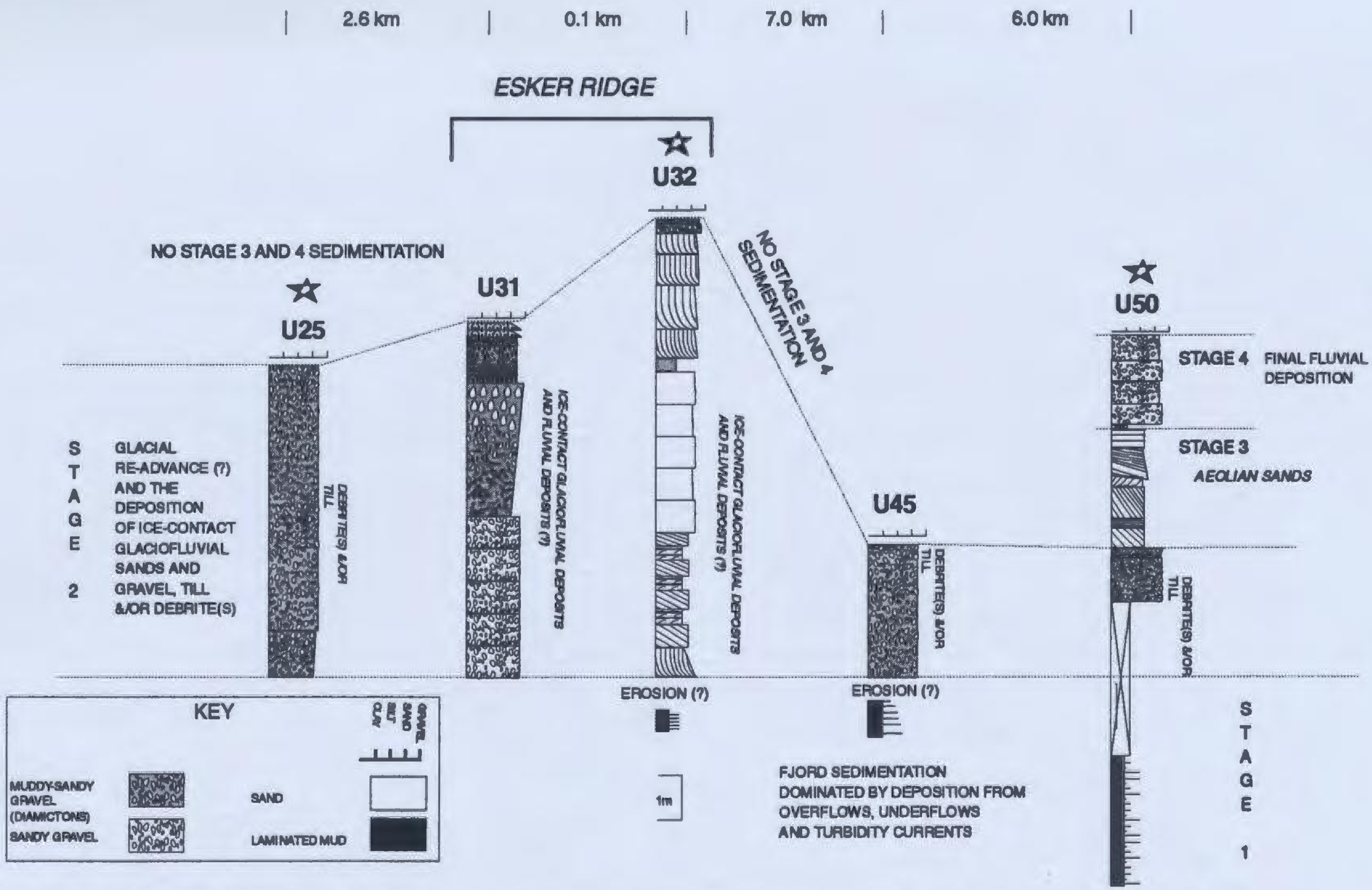


Fig. 3.33. Hypothetical stages during which the different sedimentary units in the sections of the upstream part of the study area were deposited. Key sections are indicated by stars. The positioning of the sections within these stages is based on the inferred depositional environments in which each unit was deposited. This is not a correlation chart since the lateral relationships between the units are unknown and there is no chronological control due to the lack of fossils. Individual sections are shown in a previous part of this chapter.

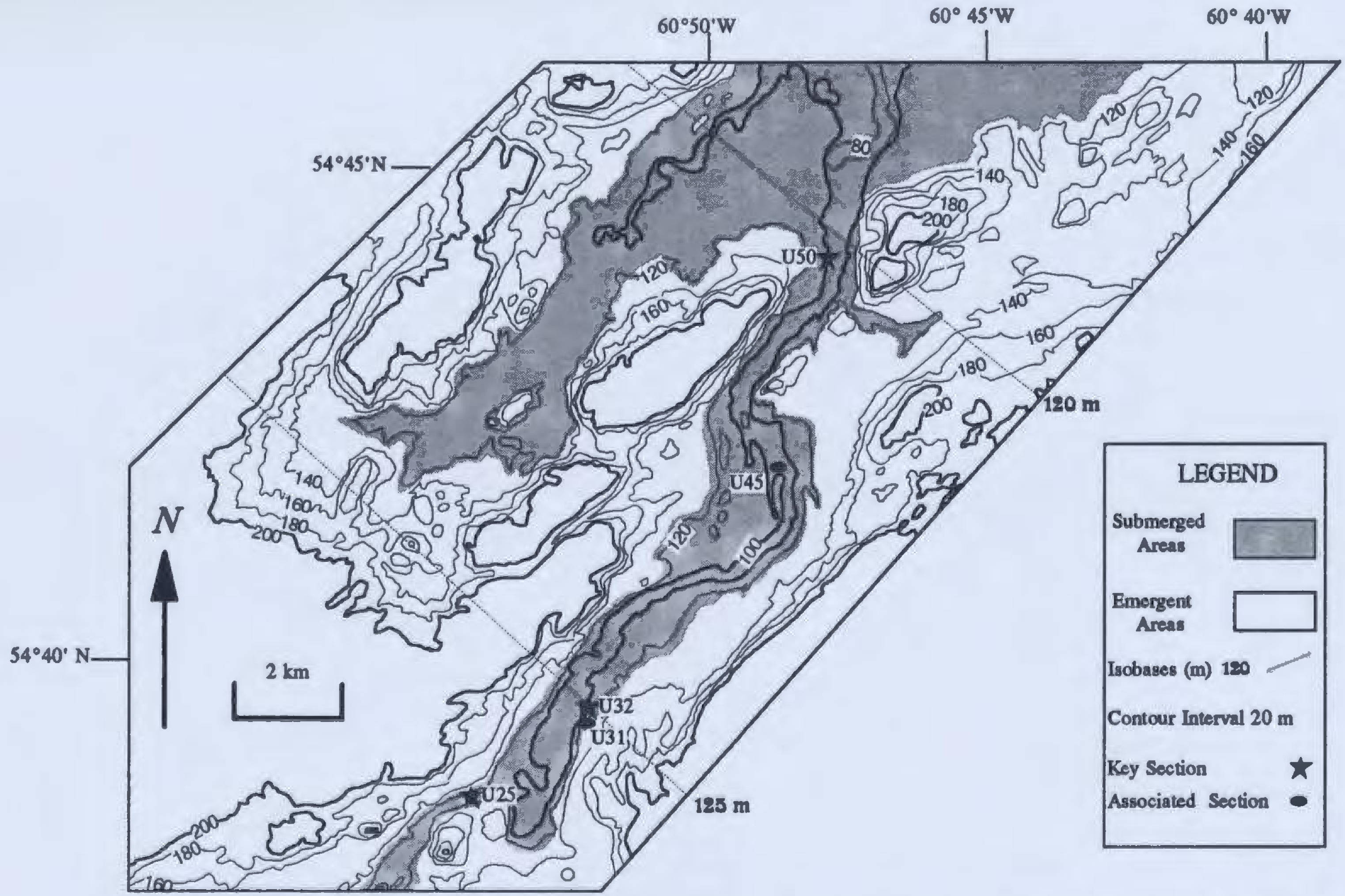


Fig. 3.34. Suggested shape of the basin or embayment in the upstream part of the thesis area at 8 ka. Shoreline position is based on the marine limit estimates for the entire area and the inferred crustal deflection (Chapter 2). Location of the key sections and associated sections is also shown. Note that the modern elevation contours for areas above 200 m have been omitted. Ice may have been present in the area but its position are unknown.

In stages 3 and 4, sedimentation was spotty. Deposition of sands of Unit C in the area of Section U50 was probably controlled by the topography. Many modern dunes fringe hills in the area (as seen on aerial photographs) and it is suggested that similar topographic control may have played a role in the localization of aeolian sands in the area of Section U50. The inferred fluvial sediments of Unit D indicate that a river developed in the area of Section U50 after the aeolian deposition. The lack of similar fluvial gravels in other sections may be a consequence of topographic control on the path of this fluvial system.

These four stages are very simplified and preliminary and other depositional conditions may have influenced some units and subunits in the upstream area. Relatively poor exposures, complex vertical and lateral sediment changes, and no chronological control make a more refined analysis impossible at this time.

CHAPTER 4

PALAEONTOLOGY

Macrofossils and microfossils are small in number, low in diversity and are limited in their distribution. Only limited information can be obtained from these fossils. This information assists in the interpretation of the depositional conditions. It should be noted here that the stratigraphic relationship between the sedimentary units in different sections is not known and units with the same type of sediment (e.g., Unit B) in different parts of the area cannot be considered time equivalent. Thus, the information obtained from a specific section applies only to the area in which the section occurs. Palaeontological data will be used in conjunction with sedimentological data from Chapter 3 to deduce the depositional conditions that prevailed when the sediments were deposited.

4.1. MACROFOSSILS

Shells were collected from one outcrop of Unit B in the middle part of the area (Section M66). Separation and identification of these shells are explained in Appendix A. The shells were small in number and fragmented, which limited the number of species that could be identified (Plate 4.1). Some shell fragmentation took place during sampling and separation of the shells from the muds. The shells were identified by Mr. John E. Maunder, Newfoundland Museum. The discussion and interpretation in the following section is the sole work of the author.

Many of the shells collected from Section M66 have been reported elsewhere from Quaternary sediments (Foster, 1946; Boss and Marill, 1965; Wagner, 1977; Macpherson, 1971; Clarke, 1974; Gilbert, 1982; Vilks *et al.*, 1982; Dale *et al.*, 1989; Syvitski *et al.*, 1989; Aitken, 1990). These occurrences and the environmental significance of the shells are summarized in Table 4.1. Most of the species suggest shallow water (<200 m) and nearly fully marine salinity (32‰ - 34‰). The tolerance of these species to temperature and type of substrate is variable. The only species that indicates a specific environment is *Portlandia arctica*. It is a mobile, deposit-feeding bivalve that has been widely documented in the North Atlantic (Macpherson,

Plate 4.1. Marine shells (facing page).

A: *Mya arenaria*, A-external view of the right valve (2.6X).

B: *Spirorbis spp.* (upper left sample (5.5X), upper right sample (7X), middle lower (6.5X)).

C: *Panadora glacialis*, C1-external view of the left valve (2.3X), C2-internal view of the left valve (2.5X).

D: *Trichotropis borealis*, D1-oral view (4.4X), D2-aboral view (3.9X).

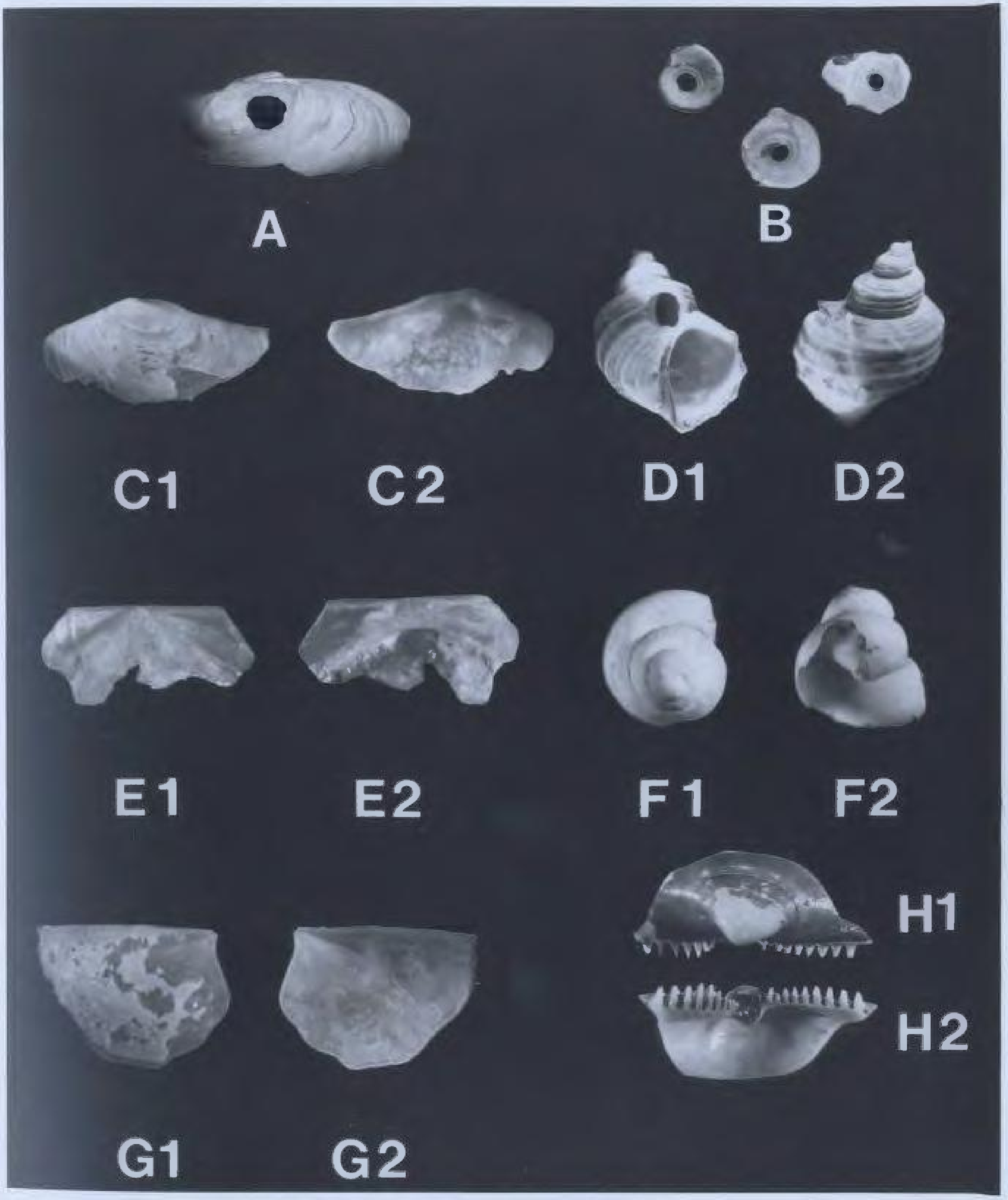
E: *Delectopecten groenlandicus*, E1-external view (5.5X), E2-internal view (6.2X).

F: *Buccindae* ? (~6X).

G: *Balanus crenatus*, G1-external view (4.8X), G2-internal view (5.8X).

H: *Portlandia arctica*, H1-external view of the right valve (8.3X), H2-internal view of the right valve (8.2X).

Borings observed in some shells (*Mya arenaria* (A), *Spirorbis spp.* (B) and *Trichotropis borealis* (D)) may have been caused by predatory gastropods (Thomsen and Vorren, 1986; Aitken and Risk, 1988).


A
B
C1
C2
D1
D2
E1
E2
F1
F2
H1
H2
G1
G2

SHELL NAME	DEPTH RANGE (m)	TEMPERATURE (°C)	SALINITY (‰)	SUBSTRATA	OCCURRENCE	REFERENCE
Mya arenaria	Intertidal-Shallow	Temperate	———	———	North Atlantic Ocean.	Foster, 1946.
Panadora glacialis	5.5 to 238	———	———	Mud, sand, muddy sand	Hudson Bay, Tyrrell Sea. SE Beaufort Sea.	Boss and Marrill, 1965.
	66, 6 to 35	-1.5 to +1.0	33	———		Wagner, 1968, 1977.
Portlandia arctica	60 to 100	-0.6 to -1.7	32 to 34	Ice-margin	Baffin Island fjords.	Gilbert, 1982;
				Sandy mud	Baffin Island fjords.	Syvitski *et al.*, 1989;
					Baffin Island fjords.	Aitken, 1990.
Trichotropis borealis	62 to 74	+2	34	———	Labrador Shelf.	Vilks *et al.*, 1982;
	120 to 140	<0	33	———	Hudson Bay, Tyrrell Sea.	Wagner, 1968;
	403 to 575	———	———	Silt-clay,	Iceland Shelf, Baffin Bay.	Clarke, 1974;
	5 to 95	<0	33	gravel-mud	Arctic Ocean.	Macpherson, 1971.
Balanus crenatus	73 to > 100	Cold?	———	———	Champlain Sea.	Aitken, 1990.
Delectopecten groenlandicus	———	———	———	Subtidal; Fjord sill, head & mouth	Baffin Island Fjords.	Dale *et al.*, 1989.

Table 4.1. Ecological significance of the different shell species collected from the study area.

1971), Beaufort Sea (Wagner, 1984), Baffin Island fjords (Dale *et al.*, 1989; Syvitski *et al.*, 1989; Aitken, 1990) and Oslo Fjords (Spjeldnæs, 1978).

The abundance of *Portlandia arctica* has been attributed to variable environmental factors and it has been used as an arctic water indicator (Wagner, 1984). Recent studies in fjords of Baffin Island and Scandinavia have documented the presence of this species near tide-water glaciers or fresh water outlets on a substrate composed of sandy mud or muddy sand (Spjeldnæs, 1978; Gilbert, 1982; Syvitski *et al.*, 1989; Aitken, 1990). Syvitski *et al.* (1989) have proposed two faunal associations of this species: a pioneer *Portlandia* association, and a mature *Portlandia* association. The pioneer *Portlandia* association develops in sediments proximal (< 1 km) to the retreating tide-water glacier or fresh water outlets in areas of high rates of sedimentation. In the mature *Portlandia* association, *Portlandia arctica* occurs with other species like *Hiatella arctica*, *Mya truncata*, and *Macoma calcarea*, in areas of lower sedimentation rates after retreat of the glacier onto land or following a decrease in the fresh water discharge. The *Portlandia arctica* shells collected in this study are of the pioneer *Portlandia* association based on the absence of the other species representing the mature association (J.P.M. Syvitski personal communication, 1990). Thus, these shells indicate that the sediments in which they occur where probably deposited in proximity (< 1 km) to a cold fresh water source. A radiocarbon date of 7950 ± 95 yr B.P. (Beta-28885) was obtained from these shells (see Appendix A).

Borings observed in some of the collected shells (*Mya arenaria*, *Spirorbis spp.* and *Trichotropis borealis*, Plate 4.1) were probably caused by predatory gastropods (Thomsen and Vorren, 1986; Aitken and Risk, 1988). These borings are similar to borings described by Aitken and Risk (1988) who attributed them to predatory naticid and muricacean gastropods. The presence of these borings in only some of the collected shells may have been due to the selectivity of the predatory gastropods (Aitken and Risk, 1988).

4.2. MICROFOSSILS

A small number of benthic foraminifera, pollen, spores, algae and dinoflagellates were separated from units A and B in different parts of the area. Their presence may provide information on the climate and environmental conditions prevailing during the deposition of the enclosing sediments. Benthic foraminifera were identified by the author and revised by B. Deonarine and G. Vilks at the Atlantic Geoscience Centre. Pollen, spores, dinoflagellates and algae were identified and counted by P. J. Mudie at the Atlantic Geoscience Centre. The discussion and interpretation in the following sections is the sole work of the author.

4.2.1 BENTHIC FORAMINIFERA

Of the more than forty samples taken from units A and B, only ten samples contained benthic foraminifera tests while all the other samples contained less than ten tests or were barren. These ten samples yielded only seven calcareous species (Plate 4.2) and no agglutinated species. Only two samples (samples T68B and T99B; the location of these samples is shown in Chapter 3) contained more than 50 tests per gram of sample while the other eight samples contained less than 30 tests per gram of sample (Appendix B). Only the two samples that contained a significant number of benthic foraminifera tests will be discussed. The number of tests in the other samples are too few to give a reliable idea of the relative abundances of species, and even these tests may have been reworked (G. Vilks, personal communication, 1990). The scarcity of benthic foraminifera suggests that suitable living conditions for benthic foraminifera may have been very short and limited in certain areas and probably affected by sedimentation rates and frequent fluctuations in salinity due to multiple points of fresh water discharge and its fluctuation (G. Vilks, personal communication, 1990).

Two species dominate the two samples: *Elphidium excavatum* f. *clavata* and *Cassidulina reniforme*. These two species were separated from samples taken from the lower parts of Unit B in Section M68. In Section D99, these species are present with other species in the middle - upper part of Unit B. The other species in sample T99B occur in very low numbers (less than 10 per sample); they are *Islandiella helenae*, *Cibicides lobatulus*, *Epistominella takayanagii*, *Pseudopolymorphina novangliae* and *Protelphidium orbiculare*.

Elphidium excavatum f. *clavata* is widely distributed in early post-glacial sediments on the Labrador Shelf (Vilks *et al.*, 1984, 1987), and has been reported in areas with calving glaciers

PLATE 4.2. Benthic foraminfera (facing page).

A: *Cibicides lobatulus*, A1-dorsal view (X85), A2-ventral view (X95).

B: *Cassidulina reniforme*, B1-apertural view (X130), B2-side view (X175).

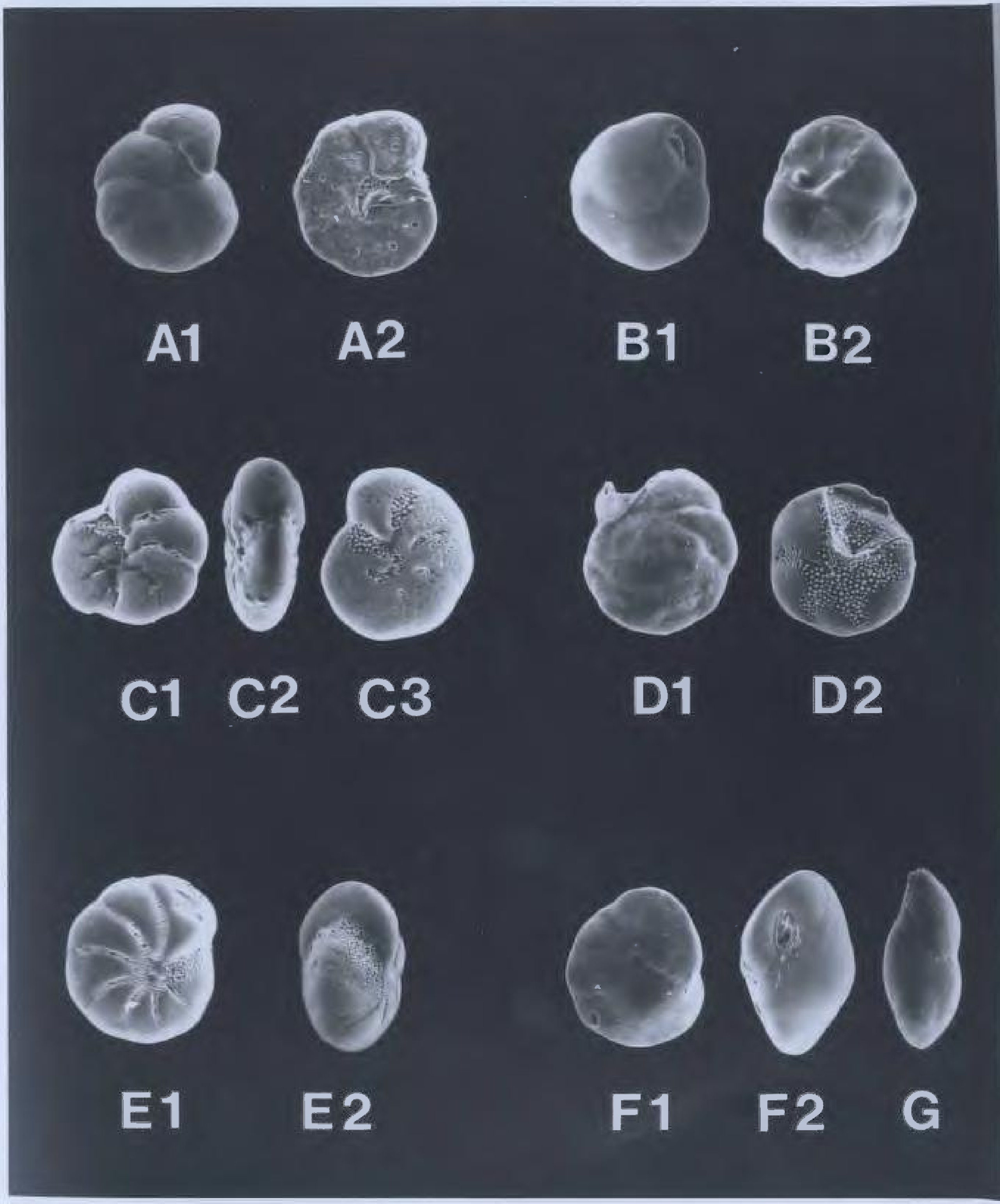
C: *Elphidium excavatum* f. *clavata*, C1-side view (X90), C2-apertural view (X95), C3-side view (X115).

D: *Epistominella takayanagii*, D1-dorsal view (X120), D2-ventral view (X125).

E: *Protelphidium orbiculare*, E1-side view (X65), E2-apertural view (X95).

F: *Islandiella helenae*, F1-side view (X85), F2-apertural view (X80).

G: *Pseudopolymorphina novangliae*, side view (X50).


A1
A2
B1
B2
C1
C2
C3
D1
D2
E1
E2
F1
F2
G

in Spitsbergen (Elverhoi *et al.*, 1980). This taxon is absent in modern surface sediments on the Labrador and Scotian shelves (Mudie *et al.*, 1984; Williamson *et al.*, 1984). It is believed to indicate surface waters of reduced salinity (as low as 18‰ according to Barrie, 1980) and temperature, particularly in areas of extensive melting of continental ice at a shoreline or where the ice is grounded on the seabed (Corliss *et al.*, 1982; Vilks *et al.*, 1984, 1987; Williamson *et al.*, 1984). Where present in large numbers, the species indicates a greater nearshore influence or shallow water depths (Vilks *et al.*, 1987).

Cassidulina reniforme is also widely reported in cores collected on the Scotian and Labrador shelves and is considered to indicate arctic nearshore environments (Mudie *et al.*, 1984; Williamson *et al.*, 1984). It is also considered to indicate proximity to the ice margin and a high concentration of suspended particulate matter (Schafer and Cole, 1982).

The species found in the samples taken from the lower part of the Unit B in the central part of the study area (sample T68B in Section M68) are similar to Zone C assemblages in Lake Melville in that they are dominated by *Elphidium excavatum* f. *clavata* and *Cassidulina reniforme* (Vilks *et al.*, 1987; G. Vilks, personal communication, 1990). No assemblages equivalent to Zone B and Zone A have been observed in the overlying sediments. In Section D99 in the downstream part of the area, *Elphidium excavatum* f. *clavata* and *Cassidulina reniforme* occur with small numbers of *Islandiella helenae*, *Cibicides lobatulus*, *Epistominella takayanagii*, *Pseudopolymorphina novangliae* and *Protelphidium orbiculare* in the middle - upper part of Unit B (sample T99B). Although *Elphidium excavatum* f. *clavata* and *Cassidulina reniforme* are the dominant species, their numbers are small (less than 50 tests per gram of sample) compared to abundances in Unit B in Section M68. These species suggest more saline conditions during deposition. This interval may correspond to the Zone C-B transition observed in Lake Melville (G. Vilks, personal communication, 1990).

The lack of assemblages representing Zone B and Zone A in the sediments overlying sediments with *Elphidium excavatum* f. *clavata* and *Cassidulina reniforme* in the central part of the area may indicate unfavourable living conditions in terms of salinity and sedimentation rates or a fall in sea level below the elevation of the section by the time that these younger zones developed elsewhere (Chapter 2). A reduction in salinity may be due to (a) diminished exchange between fresh and saline waters as a consequence of the isostatic recovery of bedrock sills, or (b) a strongly stratified water column (P.J., Mudie personal communication, 1990). The algae and dinoflagellate data support the first interpretation (see below). In the middle - upper part of Unit B in Section D99, algae and dinoflagellates suggest estuarine conditions. This points to a somewhat better exchange of fresh and salt water in the downstream part of the area during the deposition of the middle - upper part of Unit B.

Benthic foraminifera data suggest that, during the initial stages of ice retreat, a dense, salt wedge may have penetrated a considerable distance up the fjord. The *Elphidium excavatum* f. *clavata* and *Cassidulina reniforme* assemblage, probably equivalent to Zone C in Lake Melville (Vilks *et al.*, 1987), lived in proximity to fresh water outlets from a melting glacier (Elverhoi *et al.*, 1980). This may have been due to the area being isostatically depressed and the sills that may have been present were submerged. Isostatic recovery of the area may have isolated the central part of the fjord as a result of emergence of some sills that hindered the intrusion of salt water. Continuous input of fresh water progressively reduced the salinity of the central area and did not permit the survival of Zone C fauna or the development of Zone B and Zone A during the deposition of the middle - upper parts of Unit B.

The base of Zone B in the west of Lake Melville is younger than 7 ka. At 7 ka, the middle part of the thesis area was probably emergent and only the downstream part may have been submerged (using the inferred emergence rates in Table 2.1). In the downstream part of the area, the effect of fresh water input was not high, perhaps due to the partial diversion of fluvial discharge into the Kaipokok Valley area (Batterson *et al.*, 1988; M.J. Batterson, personal communication, 1990), or because of efficient mixing with normal seawater from the Labrador Sea.

4.2.2. DINOFLAGELLATES AND ALGAE

Dinoflagellates and algae separated from thesis samples can be grouped into three categories: marine, estuarine-fresh water, and *Brigantedinium* cysts. Three genera of algae were also identified. Twenty samples were analyzed for their algae and dinoflagellate content. Only eleven samples contained algae and dinocysts (Table 4.2). The samples were selected from the slab trays taken from units A and B in different parts of the area. The location of the samples is shown in Figure 4.1. In some sections, more than one sample was analyzed (e.g., Section M85) to search for a vertical change in the dinocyst and algal assemblages within those sections. Due to the scattered and unsystematic nature of the sampling, only very preliminary results can be obtained from these samples. For a complete palynological study, a more systematic sampling and analysis is necessary. Such an undertaking is beyond the scope of this study.

The total number of dinoflagellates varies from 0 to 453 cysts per gram of sample. Algae numbers vary from 0 to 688 algae per gram of sample. The abundance of each type of dinoflagellate and alga was assigned a numerical rating, in the categories rare (1 to 5), occasional (6 to 10) and common (> 10) (Table 4.2). The absolute counts of dinocysts, algae, and pollen and spores are shown in Figure 4.2.

The small number of dinoflagellate cysts per sample may reflect rapid sedimentation, turbid surface water or very cold water (Scott *et al.*, 1984). Areas where steep slopes probably existed (fjord side walls and sills) are sites susceptible to frequent sediment failures and probably turbulent waters and thus would be expected to have few or no dinoflagellate cysts (Samples T90 and T91 in Table 4.2). The interpretation of the processes responsible for the deposition these sediments supports rapid sedimentation and high sediment concentrations in surface waters (see Chapter 3).

Fresh water algae are present in different parts of the area and are more frequent than dinocysts (e.g. T68B, T75B, T90, T91). The presence of the algae suggests that fresh or brackish water conditions were generally more common during the deposition of units A and B or that there was significant vertical mixing between the surface and bottom waters.

Brigantedinium cysts are commonly or occasionally present in the lower parts of sections U54, M67, M75 and M85 while they are absent or rare in the middle - upper parts of the same sections (except for Section M67 where abundance is unknown because only one sample was analyzed). *Brigantedinium* cysts indicate saline conditions (Vilks and Mudie, 1983; Scott *et al.*,

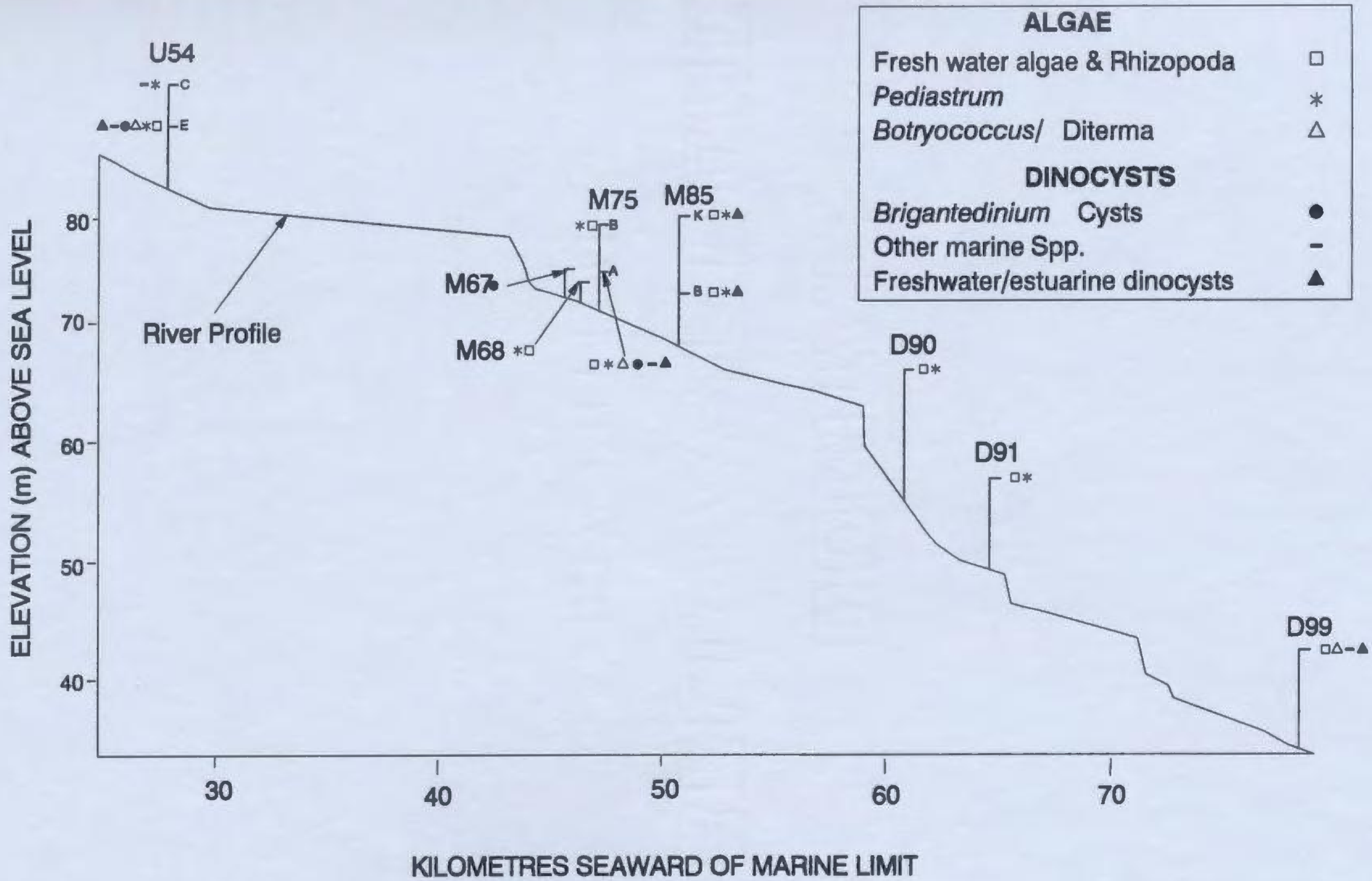


Fig. 4.1. Location of the different types of dinocysts and algae.

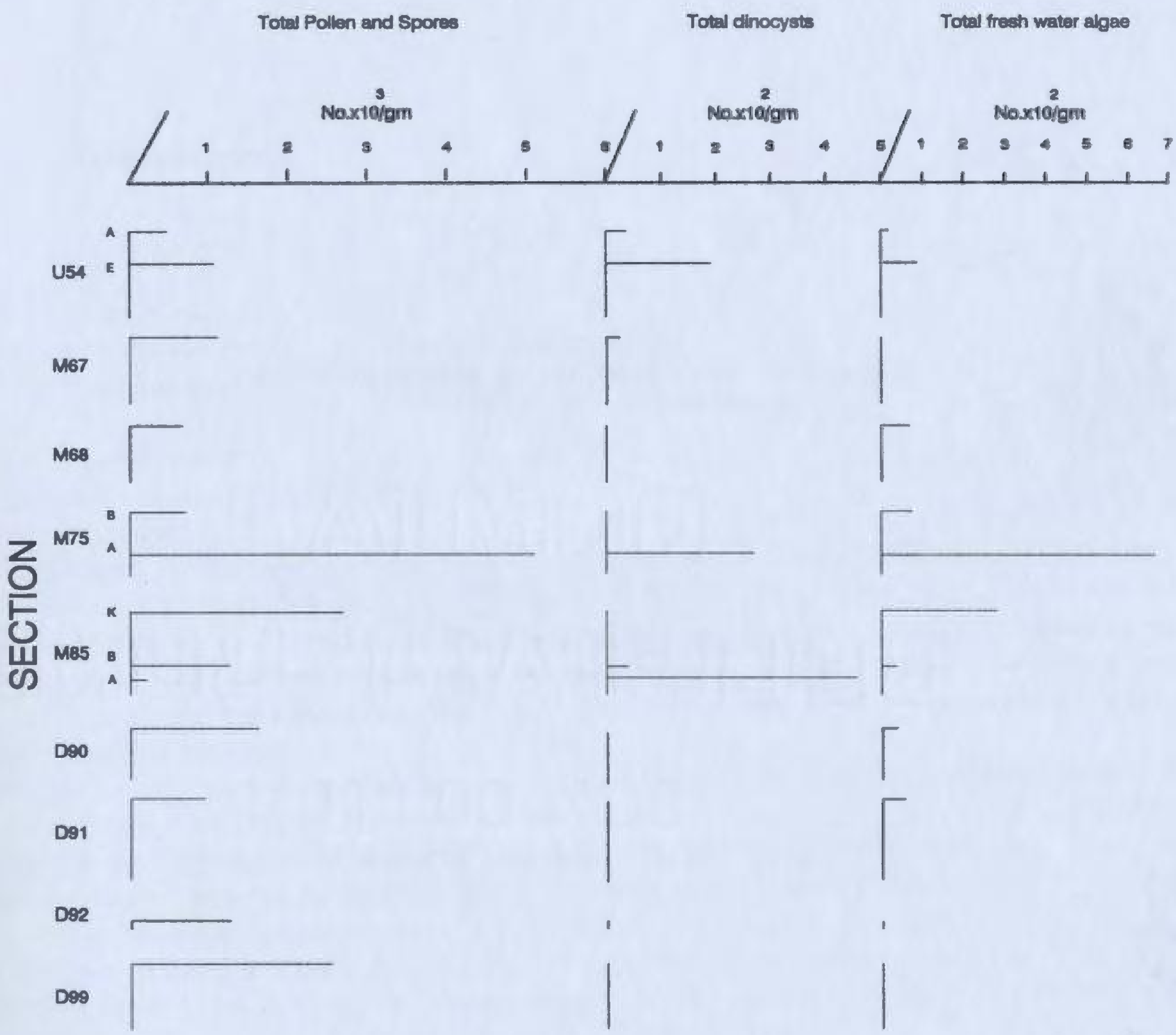


Fig. 4.2. Diagram showing the absolute numbers of algae, dinocysts, pollen and spores take from different parts of the study area. The stratigraphic position and location of each sample is shown in Figures 4.1 and 4.3.

SAMPLE	Unit or Subunit	ALGAE		DINOCYSTS			
		Fresh water algae, Rhizopoda	*Pediastrum*	*Botryococcus*/ Diterma	*Brigantedinium* Cysts	Other marine Spp.	Freshwater/ estuarine
U54C	Unit B	_____	Rare	_____	_____	Rare	_____
U54E	Unit B	Common	Common	Rare	Occasional	Rare	_____
T67	Unit B	_____	_____	_____	Common	_____	_____
T68B	Subunit b1	Common	Common	_____	_____	_____	_____
T75A	Subunit b1	Common	Occasional	Common	Occasional	Occasional	_____
T75B	Subunit b1	Occasional	Occasional	_____	_____	_____	_____
T85B	Unit B	Rare	Rare	_____	_____	_____	Rare
T85K	Unit B	Common	Common	_____	Occasional	_____	Common
T90	Unit B	Rare	Rare	_____	_____	_____	_____
T91	Unit B	Rare	Rare	_____	_____	_____	_____
T99	Subunit b1	Occasional	_____	Occasional	_____	Rare	Rare

Table 4.2. Relative abundances of algae and dinocysts.

1984; Piper *et al.*, 1990). This dinocyst is also present in Zone 4 in Lake Melville and in the three dinoflagellate zones in cores obtained from the Cartwright Saddle (Vilks and Mudie, 1983; Piper *et al.*, 1990). Other marine dinocysts are occasionally or rarely present with the *Brigantedinium* cysts in some sections (samples T54E and T75A). Elsewhere, the other species are present without the *Brigantedinium* cysts (e.g., samples T99B and T54C). Fresh water and estuarine dinocysts are absent or rarely present in some sections (samples T99B, T85B, T75A, T67 and T54E) and only commonly present in one sample (T85A).

Dinoflagellate and algal data suggest that, during the deposition of the lower parts of Unit B in the area of sections U54, M75 and M85, marine waters were present and concentrations of sediment in the surface water were not high enough to exclude dinocysts. The absence of these marine dinocysts in the middle - upper part of these sections suggests turbid and/or cold water or the absence of saline water. In areas where dinocysts are totally absent, sedimentation rates were probably very high or the waters were cold and/or turbid.

In the downstream part of the area (area of Section D99), the presence of marine and fresh water dinocysts with fresh water algae in the middle - upper part of Unit B suggests that marine-estuarine water conditions were present with some degree of vertical mixing in the water column. Very little confidence can be placed on the generality of this data from the downstream part of the area because only one sample from Section D99 was analyzed for palynology.

4.2.3. POLLEN AND SPORES

Twenty samples were analyzed for their pollen content. The samples were selected from the slab trays taken from units A and B in different parts of the area. The distribution of the different samples that yielded pollen and spores is shown in Figure 4.3.

Several samples were taken from a few sections (e.g., Section M85) to search for a vertical change in the pollen and spore assemblages within those sections. Due to the scattered and unsystematic nature of the sampling, only very preliminary results and data can be obtained from these samples. For a complete palynological study, a more systematic sampling and analysis, beyond the scope of this study, would be necessary.

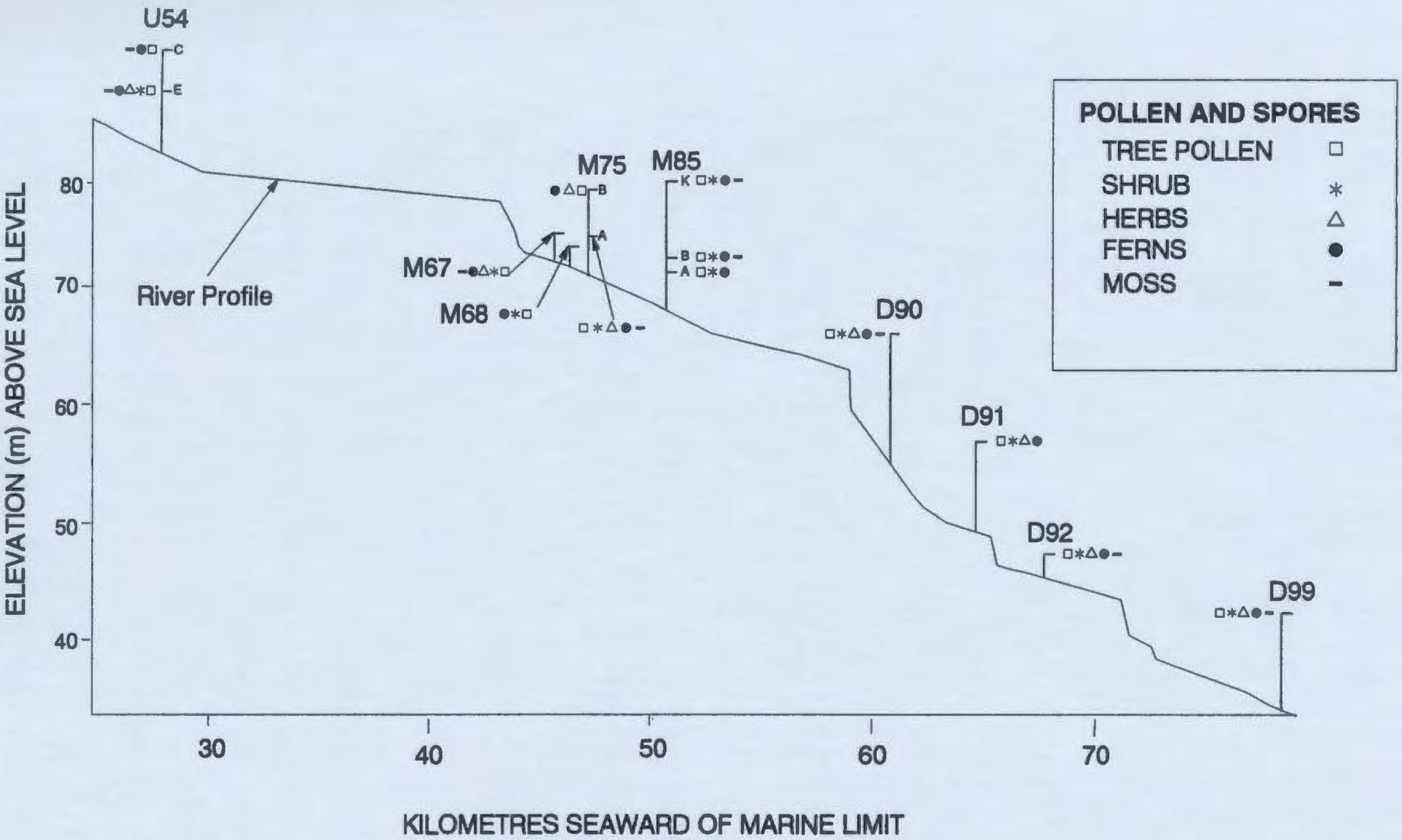


Fig. 4.3. Location of samples with main types of pollen and spores.

Of the twenty samples, only six samples contained a very small number of pollen and spore grains. The number of pollen and spores per gram of sample varies from 0 to 5099 (Appendix B). Nineteen different types of pollen and spores have been identified, representing six main types of vegetation (hardwood and softwood trees, shrubs, ferns, and moss).

Softwood tree pollen (*Picea* cf. *alba*, ranging from 4 to 36%, and *Picea mariana*, 6 to 48%) and fern spores (*Polypodiaceae*, 0 to 20%, and *Lycopodium*, 5 to 42%) are the most dominant types present in the samples. There are small amounts ($\leq 11\%$) of herb pollen and moss spores.

These data, assuming that the pollen and spores were locally derived, suggest pollen assemblages similar to the shrub-tundra association that became dominant in Labrador at about 8.0 ka (Vilks and Mudie, 1978; Scott *et al.*, 1984; P.J. Mudie personal communication, 1990). This association may correlate with Zone 2 of Lamb (1980).

Moss spores and herb pollen and spores like those that would have characterized the polar-tundra vegetation of coastal Labrador from about 10.3-8.3 ka in coastal Labrador (Vilks and Mudie, 1978; Mudie, 1982) are small in number. Also, the small absolute pollen concentration (APC) suggests that the entire sequence was deposited before the development at 5.0 ka of the northern boreal forest vegetation, which is characterized by high APC. Hence, units A and B were deposited after 8.0 ka and before 5.0 ka, consistent with the inferred sea level history (Chapter 2).

Correlation of the pollen data with other studies on lake sediments and marine cores (Jordan, 1975; Vilks and Mudie, 1978, 1983; Lamb, 1980; Piper *et al.*, 1990) should be treated with caution because other factors, especially long distance aerial and aquatic transport, may have affected the concentration of certain pollen and spore species so that the assemblages may not reflect the local vegetation (Jordan, 1975; Mudie, 1982; Catto, 1985). This problem is further compounded by the fact that the pollen and spore species observed in this study are very susceptible to transport. Hence, very little confidence can be placed in the pollen and spore data until more dedicated palynological studies are undertaken.

CHAPTER 5

GEOLOGICAL HISTORY AND CONCLUSIONS

5.1. POST-GLACIAL HISTORY

In this chapter, the data and interpretations presented in the previous chapters will be used to describe the Late Quaternary post-glacial history of the area. The history will be compared with the results of other studies in adjacent areas, and the regional implications of this study will be examined. Suggestions for further studies will be given at the end of this chapter.

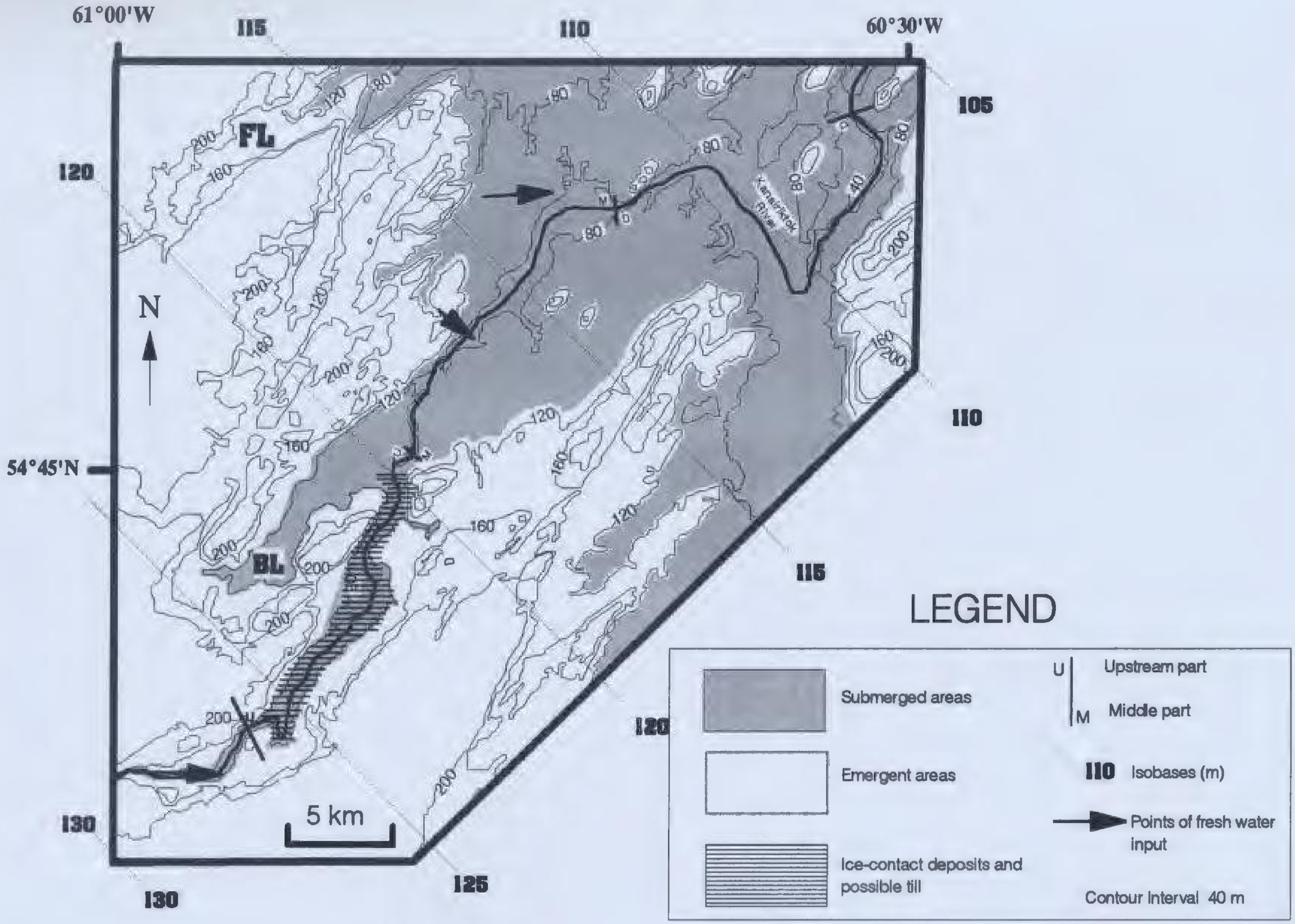
Due to the limited control on the age of the sediments studied, the history of the area will only be discussed for two very broad periods. The first period is from 8 ka to 7 ka when sedimentation in most of the area was in a marine embayment. This embayment probably existed for only a short period in the upstream part of the area due to the rapid rate of isostatic rebound inland. The second period is after 7 ka when a narrow marine embayment existed in only the downstream part of the area (at least for a few hundreds of years after 7 ka) and a fluvial system developed in the remainder of the valley.

5.1.1. HISTORY FROM 8 - 7 ka

Following the retreat of ice, the isostatically depressed area was inundated by sea water (Fig. 5.1). The Kanairiktok Valley was probably a fjord-type embayment with more than two narrow arms in the inland area and a connection between the Kanairiktok and Kaipokok valleys. Marine waters probably reached further inland than the upstream limit of the study area. The marine limit may have been as high as 130 m in the upstream part of the area and was lower by a few tens of metres in the downstream part of the area due to less crustal depression there (Fulton *et al.*, 1980; Fulton, 1986; Batterson *et al.*, 1988, Andrews, 1989).

In areas away from meltwater outlets, sedimentation was dominated by the deposition of fine silt and clay from surface sediment plumes and low concentration turbidity currents generated by sediment failures on steep slopes (Unit A and the lower part of Unit B). Marine currents may have influenced the deposition of some of this sediment. During the deposition of Unit A,

Fig. 5.1. Suggested shape of the basin at 8 ka. Note that the distances between the isobases are equal, because it is assumed that the crustal deflection increases at a constant rate towards the SW. No data are available on the flexural rigidity of the crust in this area, but it probably does not vary over this 50 km-across area. Areas where ice-contact glaciofluvial deposits and till (?) exist were affected by a minor glacial re-advance. This glacial re-advance occurred after the marine waters receded from those areas. Other areas that were not studied may have also been affected by this re-advance. Suggested points of fresh water input are shown as arrows. Other points of fresh water input may have also existed. The locations of two modern lakes referred to in the text are shown (FL = Florence Lake and BL = Beartrack Lake).


61°00'W
115
110
60°30'W
105
120
FL
N
54°45'N
BL
110
115
120
130
130
125
5 km
Kanairiktok River
LEGEND
Submerged areas
Emergent areas
Ice-contact deposits and possible till
U Upstream part
M Middle part
110 Isobases (m)
Points of fresh water input
Contour Interval 40 m

prolonged winter sea ice conditions may have existed. Deposition from underflows may have taken place during periods of high discharge and associated reduced salinities.

In the upstream part of the study area and near the sides of the fjord, sedimentation was affected by the proximity of river outlets. Near the outlets, high concentration surface-sediment plumes, turbidity currents and underflows were active during periods of high river discharge. Grain flows and liquefied flows or cohesionless debris flows may have deposited some thick sand beds. These processes deposited laminated sand and mud with thicker sand intervals (Unit B).

The salinity of the basin was reduced in the middle part of the area so that only a limited assemblage of benthic foraminifera could flourish. In the downstream part of the area, salinity of the basin waters may have reached normal marine values. The upstream sections apparently lack any benthic foraminifera, perhaps due to reduced salinity and high sedimentation rates. A shrub-tundra vegetation may have started to colonize the area during tne deposition of units A and B.

The oldest unit is inferred to be Unit A in the downstream part of the area. However, it may be a lateral time equivalent to Unit B in the other parts of the study area.

Meltwater from glaciers probably entered the embayments though several outlets, some of which may have been larger than others. Major outlets in the upstream and middle part of the area may have been areas presently occupied by the Kanairiktok River, Florence Lake and Beartrack Lake (Fig. 5.1). In the downstream part of the area, additional outlets were probably away from the area presently occupied by the Kanairiktok River. Delta top deposits that might have been associated with the outlet in the Kanairiktok River are apparently not preserved in the study area. These deposits may have been reworked by younger rivers or may be present outside the thesis area.

Land emergence was probably rapid (Chapter 2), causing the sea to recede quickly from the area. In the upstream part of the area, emergence may have taken place in the first few hundred years after deglaciation (if there was no restraining effect on emergence by nearby thick ice). In the middle and downstream part of the area, emergence started earlier but eventually lagged the upstream area, due to slower rates of rebound. After the upstream region became emergent, there was apparently a minor glacial re-advance that resulted in the deposition of ice-contact glaciofluvial deposits (esker deposits) and perhaps till (Unit E in some upstream sections). This re-advance only affected the upstream part of the study area (M.J. Batterson, personal communication, 1992), and probably produced no significant ice load.

Emergence in the middle part of the area restricted water exchange with the Labrador Sea, resulting in the exclusion of benthic foraminifera; only fresh-water algae and dinoflagellates flourished. This restriction may have been due to the emergence of some sills that were present in the study area. Similar restricted conditions are suggested by the marine fauna in Lake Melville (Vilks and Mudie, 1983; Vilks *et al.*, 1987).

Progradation of a fluvial system from the west deposited more proximal and coarser deltaic sediments over the marine and distal deltaic muds. Deltaic sediments were probably deposited from different points where river outlets were reaching the embayment. There may have been several side-entry deltas, two of which were probably near sections M85 and M61. A fjord-head delta was probably prograding down the valley from the west, leading to a general upward coarsening in all sections.

During the emergence of the middle part of the area, there may have been a period of still stand during which delta-top deposits formed at an elevation of 80 - 100 m in the area of Section M61. This may have been the same still stand inferred to have occurred in the Kaipokok Valley area (based on a delta surface at an elevation of 95 m; Batterson *et al.*, 1988).

Progradation of the fluvial-deltaic system down the valley continued through the 8-7 ka period and by 7 ka only a small part of the middle part of the area was still submerged. A fluvial system cut and reworked most of the upper deltaic deposits as rebound continued. Many ages of fluvial deposits may be represented in the Holocene record. The oldest fluvial deposits could have accumulated on the tops of prograding deltas, but other fluvial gravels might be hundreds or thousands of years younger than this. The fluvial environment was characterized by shallow channels and fluctuating discharge, depositing gravel beds of Unit D.

5.1.2. HISTORY AFTER 7 ka

At 7 ka only a small part of the middle part of the area and all the downstream part of the area were still submerged (Fig. 5.2). The emergence of the land surface changed the morphology of the basin to a narrow arm of the sea. The main fluvial-deltaic system prograding from the west supplied sediment to this narrow embayment. Side-entry river outlets probably also supplied sediment to the area. Some of the side-entry channels may have been stabilized by the shape of the bedrock surface.

After the area became emergent, fluvial deposition prevailed. This fluvial system was characterized by shallow channels and strongly fluctuating discharge. This may have been a braided or a wandering fluvial system. Such systems are common in recently deglaciated areas (Church, 1983; Brierley, 1989).

The changing morphology of the area and the changing topography may have affected the type of sediment moved along the river. For example, the sudden change in the texture of the fluvial deposits from gravel in the middle to sand in the downstream part of the area may have resulted from reduced transport efficiency on lower gradients or emergence of a large sill that may have hindered the passage of sediment to the downstream area. Similar causes for variations in sediment texture have been suggested by Church (1983) for the Fraser River area.

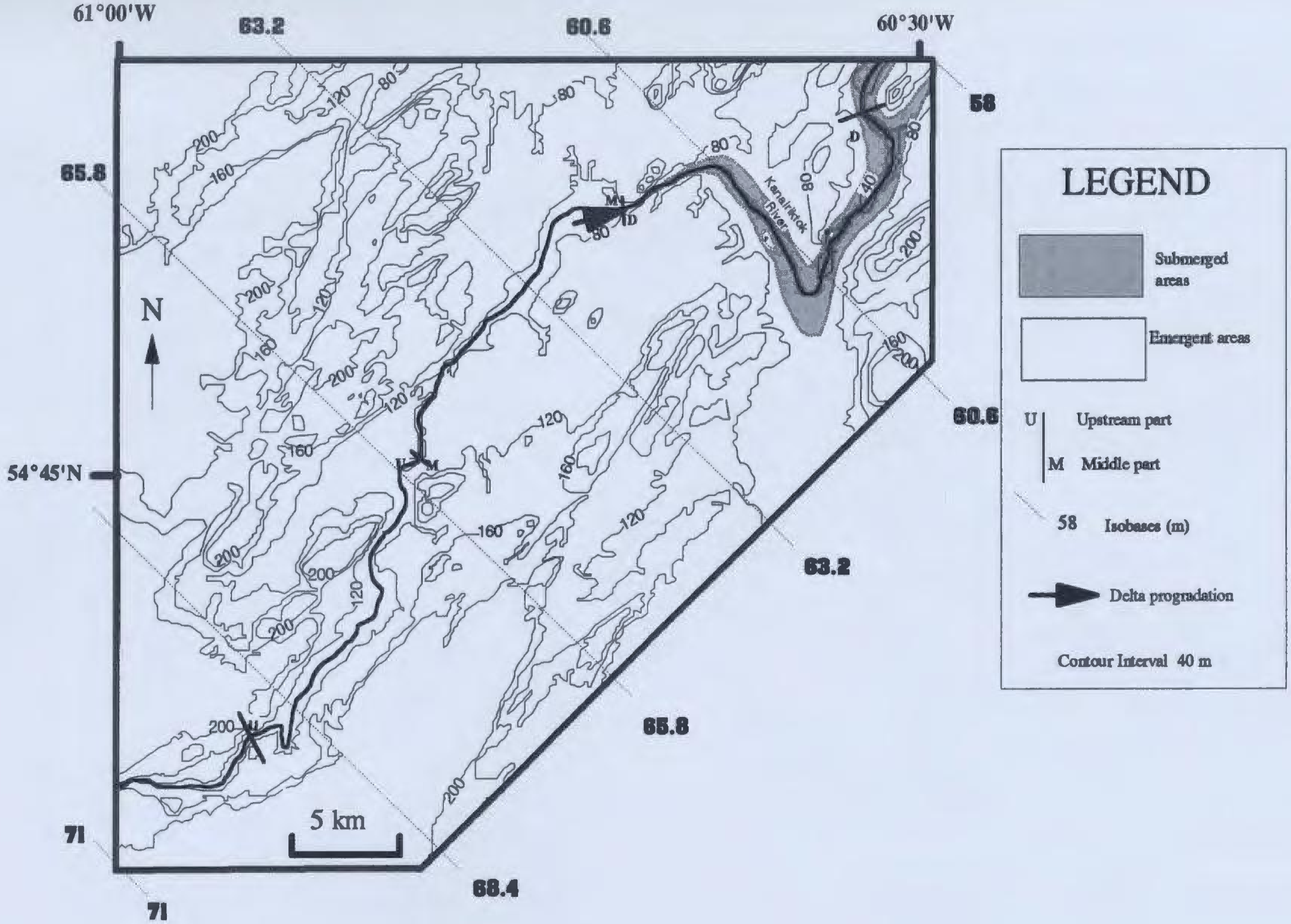
There may have been several stages of fluvial activity in the area as suggested by the several terraces present at different elevations. At least two levels have been observed in the field and several more levels are apparent on aerial photographs. Thompson and Klassen (1986) suggest the former presence of a larger river system in the area. Changes in the discharge and gradients due to rebound could have accounted for the different terraces.

5.1.3. SUMMARY OF DEGLACIATION HISTORY

The following sequence of events is suggested for the deglaciation history of the Kanairiktok Valley, based on the sediments examined.

(1) Sedimentation dominated by suspension settling from surface sediment plumes, originating from river outlets in a fjord-type embayment, took place following deglaciation of the

Fig. 5.2. Suggested shape and limited extent of the marine basin at 7 ka (compare to Fig. 5.1). The distances between the isobases are equal, based on the assumption that the crustal deflection increases at a constant rate towards the SW. No data are available on the flexural rigidity of the crust in this area, but probably does not vary over this 50 km-across area. The main fresh water input was from the southwest (arrow). Other side-entry rivers probably existed.


61°00'W
63.2
60.6
60°30'W
58
65.8
N
54°45'N
60.6
63.2
65.8
68.4
71
71
5 km
Kanairiktok River
LEGEND
Submerged areas
Emergent areas
U Upstream part
M Middle part
58 Isobases (m)
Delta progradation
Contour Interval 40 m

isostatically depressed area at 8 ka or earlier.

(2) A fluvial deltaic system prograded over the entire area. Both side-wall and fjord-head deltas probably prograded over the marine muds. There may have been a period of glacial re-advance that affected the upstream area after it became emergent. A period of still stand also may have occurred during emergence. Most of the upstream and middle parts of the area were emergent by 7 ka while the downstream part of the area was emergent after 7 ka and before 6 ka.

(3) a fluvial system (probably several stages) entrenched and reworked part of the upper deltaic succession.

5.2. COMPARISON WITH OTHER STUDIES

Very few similar studies have been carried out in central Labrador and most of the deglaciation models are based on more regional studies. Most of these studies suggest deglaciation of the area between 8 ka and 9 ka (King, 1985; Dyke and Prest; 1987b; Vincent; 1989), consistent with the results of this study. This study suggests, however, that the Laurentide Ice Sheet probably retreated from the area not later than about 8 ka, contradicting some other studies (e.g. Clark and Fitzhugh, 1990).

When comparing the results of the sea-level history suggested for this area with the other proposed models, differences appear which have implications for the deglaciation history of the area. Although the radiocarbon date obtained from this study cannot be related to the marine limit suggested for the area by Fulton *et al*. (1980), Fulton (1986) and adjacent areas by Batterson, *et al*. (1988), there are certain consequences of this date for deglaciation models. For example, deglaciation by 9 ka, suggested by Vincent (1989) and Dyke and Prest (1987b), seems to be too early, because in a ca. 1000 yr interval from 9 ka to 7950 ± 95 yr B.P. the site where the shells occur would probably have rebounded sufficiently to have resulted in a very shallow water depth, inconsistent with an inferred depth of some tens of metres at 7950 ± 95 yr B.P. It seems reasonable to conclude that rapid emergence started well after 9 ka. Unless nearby ice substantially delayed the onset of rapid emergence, this leads to the conclusion that local deglaciation also occurred well after 9 ka.

The claims of Clark and Fitzhugh (1991) for the sea-level history of coastal Labrador should be re-evaluated, because a marine limit of 119 m at Hopedale at 7600 yr B.P. would imply little or no differential rebound between the inland areas and coastal Labrador. Assuming a crustal deflection of 0.4 m/ km (Andrews, 1989), a marine limit of 119 m at Hopedale would suggest a contemporary marine limit of about 160 m at Moran Lake and in the Kanairiktok Valley area, areas that were apparently ice free at 7600 yr. B.P. The highest marine delta observed in the Moran Lake area is at 125 m (Batterson *et al.*, 1988) and the maximum marine limit in the Kanairiktok Valley area is suggested to be 135 m (Fulton *et al.*, 1980; Fulton, 1986). Additional problems with the claim of a 119 m marine limit at Hopedale are outlined in Chapter 2, and will not be repeated here.

The 8 ka ice margin position of King (1985) is in better agreement with thesis results, although the 8 ka ice margin is believed to have been somewhat further to the west than suggested by King (1985). The isobase maps of Andrews (1989) for 7 ka suggest a higher marine limit at 8 ka than the suggested 135 m value. These isobases are drawn from more regional data and may not accurately portray local variability (J. T. Andrews, personal communications, 1993). Similarly, the most recent emergence curve of W.R. Peltier (W.R. Peltier, personal communication, 1993) should be revised because it would require emergence before 8 ka of the site where shells were collected, in disagreement with the radiocarbon date.

The only nearby studies on Quaternary marine sedimentology are those in modern bays, like Makkovik Bay (Barrie, 1980; Barrie and Piper, 1982; Piper *et al.*, 1983), Kaipokok Bay (Kontopoulos and Piper, 1982) and Lake Melville (Vilks and Mudie, 1983; Vilks *et al.*, 1983). The depositional environments in these inlets were controlled by the falling relative sea level and the fluctuations in the amount of discharge. Units A and B in the thesis area may be lithologically similar to the upper basin fill units observed by Barrie and Piper (1982) while the proximal deltaic and fluvial deposits observed in the Kanairiktok Valley area are absent from the modern marine inlets because these areas did not experience emergence. An appreciation of the potential for sedimentological links between outcrops and marine surveys can be gained by comparison of the characteristics of units from the Kanairiktok Valley outcrops and those of marine surveys of Barrie and Piper (1982), observed in Makkovik Bay. For example, marine surveys allow the tracing and correlating of seismic units over wide areas. The three dimensional geometry of the sedimentary packages is therefore well constrained (although it depends on the density of the survey lines). Land-based surveys like this thesis study allow better observation and description of units which may be indistinguishable on seismic section. For example, sediments with the same texture would have the same seismic character in marine data sets.

5.3. CONCLUSIONS

The primary objective of this study was to describe the postglacial sediments exposed along the banks of the Kanairiktok Valley, so as to deduce sedimentary environments. This objective has been achieved and the sedimentary environments suggested for this study are comparable to sedimentary environments that exist in many modern fjords (Barrie, 1980; Gilbert, 1983, Syvitski and Farrow, 1989; Syvitski *et al.*, 1987; Syvitski and Hein; 1991; Hein and Syvitski 1992). Specifically, the sediments observed in this study are comparable to fjords that are supplied by glacial meltwater outlets at the head and sides of the fjords. As the area emerges coarse sediments are deposited as a result of the progradation of fluvial deltaic system over distal deltaic and marine sediments forming upward coarsening sequences (*cf.* Powell, 1981; 1983; 1984).

The second objective was to provide a local chronology of deglaciation and sea level. This was partly achieved although the chronological data base for this study is very limited. It was necessary to consider data from other studies to constraint the chronology of deglaciation and sea level change. The interpretation of sea level history in the thesis area is very preliminary and should be treated with caution, because other factors may have affected the sea level history of the area (e.g., delayed crustal rebound).

The third objective of the thesis was to use the field data to evaluate the regional models that have been suggested for the deglaciation of the area. This objective has also been partly achieved. Although only one radiocarbon date was obtained, this date and the sedimentary history provides constraints for some of the deglaciation models suggested by other workers (King, 1985; Dyke and Prest, 1987b; Vincent, 1989) and suggests the need for a re-evaluation of other models (e.g., Clark and Fitzhugh, 1990, 1991).

5.4. SUGGESTIONS FOR FURTHER STUDIES

Given the time constraint of a single field season due to funding limitations, the results presented in this thesis need to be followed up in the field. This is the norm for initial studies in remote and unknown areas. Some points that need further study are outlined below.

(1) The area upstream of the thesis area must be investigated for the presence of deltaic or shoreline deposits to verify the maximum marine limit suggested by others for the area. The delta observed by Batterson *et al.* (1988) in the area between the Kanairiktok and Kaipokok valleys should be checked and investigated for the presence of any dateable material that may be tied to the marine limit. If this delta could be dated and the age of the marine limit in the upstream part of the area could be better constrained, then the deflection due to glacial loading and the sea level history could both be more accurately determined.

(2) The relationships of both Unit E and the ice-contact glaciofluvial deposits (i.e. esker deposits) to the marine deposits in the area need to be clarified in order to determine if there really was a glacier re-advance as suggested by Batterson *et al.* (1988). Only poor or limited exposures of ice-contact glaciofluvial and glacial deposits are present in the area and they occur near the suggested marine limit. A good area to check is the area of Beartrack Lake where Batterson *et al.* (1988) suggested the presence of ice-contact glaciofluvial and glacial deposits over marine deposits.

(3) More dateable material should be collected to achieve a better chronological control on deglaciation.

(4) A more systematic sampling for microfossils should be undertaken in order to correlate the Kanairiktok assemblages with the data from Lake Melville and Makkovik Bay (Barrie, 1980; Vilks *et al.*, 1987). The downstream part of the study area may provide more suitable sites for sampling for benthic foraminifera because marine conditions may have existed for a longer period

there. Additional data for dinoflagellates and algae would provide valuable information on the surface water conditions.

(5) The various fluvial terraces should be examined and their elevations and relative chronologies should be evaluated in more detail.

(6) A marine survey should be conducted in Kanairiktok Bay, and the data obtained from this survey should be compared with the data from Makkovik Bay (Barrie, 1980), Kaipokok Bay (Kontopoulos and Piper, 1982) and Lake Melville (Vilks *et al.*, 1987). If bathymetric charts are still unavailable, proper preparation should be made in advance to ensure availability of a sub-bottom profiling system to work from a launch to obtain bathymetry.

(7) On a technical note, although the field study was carried out in a cheap manner from a canoe, this survey method is not efficient. Time for section study is proportionally reduced. A sedimentological study using a helicopter would be more appropriate (but very expensive), allowing coverage of more area, collection of a larger number of samples from different parts of the area, and checking of many surface geomorphological features. A preliminary survey from a helicopter would allow selection of the best localities for detailed description. A canoe traverse and the many areas of water rapids and falls which had to be portaged consumed several days very valuable field time.

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APPENDIX A
FIELD AND LABORATORY METHODS

A.1. FIELD METHODS

Field work consisted of the description, measuring and sampling of lithologic sections exposed along the banks of the Kanairiktok River. Good exposures are rare, so that digging using shovel and pick was necessary to obtain useful sections and samples. Digging was hindered by frequent collapse of material into the trenches. In some cases stiff pebbly mud was impossible to penetrate.

Each section was described, then measured with a tape and a 1 m Jacob staff. Selected intervals in most of the sections were sampled (some sections were impossible to sample due to the steepness and slipperiness of the section or due to the softness of the material).

Samples of selected intervals of the main mud and silt units were taken in slabbing trays (Mosher and Asprey 1986). Each tray (25 cm x 4 cm x 1 cm) was inserted into the cleaned vertical face of the exposed unit (using a hammer when the material was stiff). The area around the edges of the tray was cleaned and the tray was released from the section by passing a taut metal wire along the open side of the tray. The samples were wrapped in thin sheet plastic (plastic food wrap), sealed with cellulose tape to avoid the drying, shrinking and cracking of the samples. The location of slab samples were selected at intervals that exhibit the different types of laminations as seen on the cleaned face of the outcrop. In most cases, two slab samples (one from the lower part of the muddy interval and one from the top) were obtained.

Bulk sand and silt samples (50-100 g) were taken from the uppermost sandy units, and from sandy interbeds in the main lower muddy units. These samples were taken from the entire bed, as a channel sample by inserting one or more slab trays into the wet surface of the outcrop so as to cover the entire bed. Due to the impossibility of sampling entire sections, the texture of the many beds was described in the field using a hand lens and a field microscope with magnification of (X20).

The dimensions and geometry of sedimentary structures, including deformational structures where present, were measured, and palaeocurrent measurements were taken wherever possible. The number of measurements of palaeocurrents and deformational structures was small due to the small number of structures and poor exposures. Therefore, the data were not plotted on a stereonet. The deformational structures are present in soft mud and silt units that are generally covered by a thin surface layer of mixed silt, sand and mud washed down from overlying units. Attempts to remove this surface veneer in order to obtain a more detailed account of these structures only obscured the outcrop more. Even repeated washing of parts of the sections did not greatly improve the clarity of the structures.

Sampling for foraminifera was done with the aid of the X20 field microscope. At every 10 cm interval in marine muds (units A and B), a sample was taken from the muddy material and placed on a glass plate. Then the sample was mixed with a few drops of water and was searched for foraminifera. Intervals with the most foraminifera tests (which were very rare) were sampled using a slab tray. Shells were picked from one of the sections in the central part of the area (see shell separation and treatment below).

A.2 LABORATORY METHODS

Several types of analysis were conducted on the samples collected in the field, to study the sedimentary structures, texture, mineralogy, and micropalaeontology of these sediments. Results from these analyses are used in the interpretation of the late Quaternary history of the area.

A.2.1 SEDIMENTARY TEXTURES AND STRUCTURES

Sedimentary structures and textures present in the slab samples were studied using X-radiographs, thin sections and grain size analysis of selected laminae. Bulk sand samples were sieved to determine their grain size distribution. More than fifty sand samples were analyzed for grain size by sieving and more than seventy samples of mud were analyzed using the Sedigraph. Only the samples mentioned in the text are included in Appendix B.

A.2.1.1 GRAIN SIZE ANALYSIS

Several techniques were used in grain size analyses of the samples. The channel samples of sand were thoroughly mixed by combining the material from all slab trays taken from the bed. The sand samples were then split by wet sieving through a 4 ϕ sieve. The coarse fraction (>4 ϕ) was dried and sieved between -2 ϕ and 4 ϕ at 1/2 ϕ intervals, except for -1.5 and 2.5 ϕ sieves, which were not available. When the fine fraction (>4 ϕ) of the sand samples exceeded 10% of the original sample, this fine fraction was analyzed by pipette down to 9 ϕ at an interval of 1 ϕ, following techniques outlined in Piper (1980).

Selected intervals of the slab samples were analyzed using a combination of wet sieving and a Sedigraph 5100 automated size analyzer. The sedigraph gives results similar to pipette analysis (Stein, 1985; Coakley and Syvitski, 1991). Fourteen duplicate samples of the seventeen mud samples shown in Appendix (BI) were analyzed by the sedigraph (see below).

The samples were first put in a 1% sodium hexametaphosphate (calgon) solution for several hours, then insonified with a B. Braun Melsungen AG (model Braun-sonic 1510) ultrasonic probe at a setting of 400 W for a period not longer than 1 minute. The material was then wet sieved at a 1/2 ϕ interval through a set of sieves spanning 1.5-4.25 ϕ. The fine fraction (>4.25 ϕ) was collected and analyzed using the Sedigraph. In many cases, the suspension of fines collected from the wet sieving was greater than 80 ml, which is the maximum volume permitted in the Sedigraph sample chamber. In these cases, the sample was either centrifuged down to the appropriate volume, insonified and analyzed, or the sample was stirred vigorously with a magnetic stirrer while a 50 ml volume was withdrawn. Centrifuging the sample and insonifying it before analysis is the preferred technique. Samples withdrawn while using a magnetic stirrer had a lower concentration. In many cases this created problems for analysis. In these cases when the concentration was too low for Sedigraph analysis to be reliable, the sample was centerfuged to increase its concentration.

The duplicate mud samples were analyzed to check the precision of the sedigraph technique. The duplicates were new samples taken from the same part of the sample tray. The duplicate samples were only analyzed up to 10.5ϕ.

Duplicate samples that were analyzed many months apart show greater variation, generally < 10% in each size class. The additional variation probably is due to different machine set up and laboratory environment (e.g. temperature, humidity). Duplicate samples that were analysed on the same day had a margin of error of less than 2-3 % with the maximum variation being in the determination of the finer and coarser tails of the distribution. These variations might reflect machine inconsistencies or a slight variation in the composition of the duplicate samples even though they were taken from the same part of the sample tray.

The most important variables that may affect Sedigraph results are (a) the temperature during analysis, (b) sample concentration, and (c) the percentage of coarse to medium silt in the sample. A variation in the temperature of the X-ray sample chamber will affect the density of the sample and the settling velocity. The presence of a large proportion of coarse silt in the sample may affect the analysis because so much silt may settle out that constant concentration is not maintained. The presence of non-magnetic heavy minerals (magnetic minerals are removed with a magnet before analysis) also may affect the analysis.

The results of the duplicate analyses are shown in Figure A.1. Although some percentages of silt and clay are different between duplicate runs, the shape of the duplicate curves are similar, with an imprecision of less than 10%. The variation is greatest in the fraction finer than 9 ϕ. This is probably because this fraction is most susceptible to any slight variation in temperature of sample concentration.

A.2.1.1.1 Calculation of the Mean, Standard deviation, Skewness and Kurtosis

The results from the grain size analyses of most of the samples were open ended and the class size (1/2 ϕ) in some samples was not consistent due to a lack of two sieves. Therefore, the method of moments was not used to calculate statistical parameters. Statistical parameters were instead determined graphically according to the method of Folk (1974). For samples with less than 5% of sediment finer than 12.5 ϕ. the standard deviation, skewness and kurtosis were calculated from cumulative curves using 95% of the population. Samples with 5-16% of sediment finer than 12.5 ϕ, the standard deviation and skewness were calculated according to

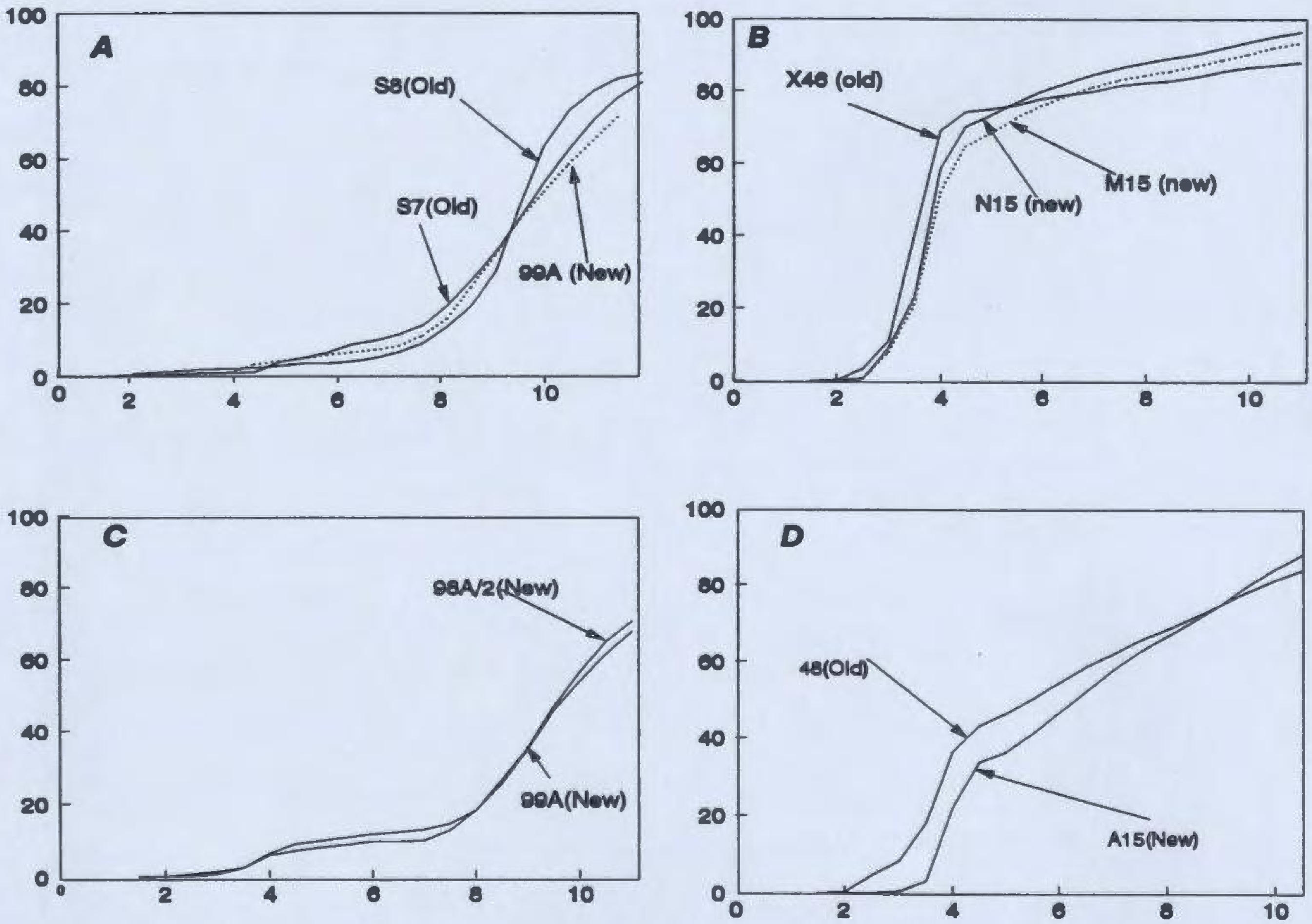


Fig. A.1. Selected examples of duplicate analyses of muddy sediments. Samples in Figure C were analyzed the same day and have the least difference.

the procedure of Inman (1952), while the mean was calculated as outlined by Folk (1974). When the sample contained more than 16 % of sediment finer than 12.5 ϕ no statistical parameters were calculated.

Many of the samples were bimodal to polymodal, so that statistical parameters based on the assumption of log-normality are meaningless (Blatt *et al.*, 1973). For clay-rich samples, the size distribution after dispersion with calgon and ultrasonic disaggregation is unrelated to the size distribution of flocs that would have developed in sediment plumes entering a marine embayment. For such samples, the percentages of sand and medium to coarse silt is probably more useful in deducing proximity or strength of the depositing currents.

A.2.1.2. X-RADIOGRAPHS

X-radiographs were made of all the slab tray samples using a shielded Picker X-ray tube operating at 70 kilovolts and 2.75 milliamperes. Based on trial and error, a 42 second exposure was found to give best results. Theory, principles and methods of X-radiography in the study of sedimentary structures are contained in Bouma (1969).

A.2.1.3. THIN SECTIONS

A number of thin sections were made from selected intervals of the slab trays for studying sedimentary structures. The number of sub-samples taken from each tray varied from one to four, and was based on the presence of sedimentary structures observed visually or on the X-radiographs. The sediment was impregnated with a thermal resin to form a rigid chip for thin sectioning using the method described in Awadallah (1991). These rigid chips were then cut and trimmed using a diamond saw. Samples were cut, ground, and mounted on unfrosted glass slides using normal epoxy prior to thin sectioning. No difficulties in bonding or thin-section preparation were encountered.

Each thin section was studied under the microscope in order to determine the texture and sedimentary structures of the sediments. Laminae thicknesses and their contacts were studied. The grain size of each laminae and its vertical variation was measured. Percentage of matrix (defined here as material <0.01 mm) was estimated visually by comparison with percentage estimation charts of Scholle (1979).

A.2.2. PALAEONTOLOGICAL METHODS

Palaeontological methods included separation of shells for identification and dating, and separation of benthic foraminifera, pollen, spores and dinoflagellates for identification and counting.

A.2.2.1. SHELL SEPARATION AND DATING

Shells representing several species were collected from one section in the middle part of the area. The shells were picked from the outcrop and washed repeatedly in the field with river water to remove the attached mud. Later, further mud removal in an ultrasonic bath was necessary for some of the shells. The shells were then left to dry and were photographed.

Identification of the shell species (Table 4.1) was done by Newfoundland Museum naturalist John Maunder. The level of classification is fairly crude due to the fragmented nature of the shells.

Highly fragmented shells (approximately 5.1 g) were sent to Beta Analytic Inc. (University Branch, P.O. Box 248113, Coral Gables, Florida 33124, USA) for radiocarbon dating. The date was obtained using the accelerated mass spectroscopy (AMS) technique, where the shell fragments were prepared by etching away the outer layer with dilute acid. Carbon dioxide was produced by heating the shells to release carbon dioxide in the presence of a cobalt catalyst. The reported date of 7950 ± 95 years BP is adjusted for total isotope effect using carbon 13 where the adjusted dates are normalized to -25 per mil carbon 13 according to the PDB-1 international standard. The carbon-13 content was measured concurrently with that of carbon-14 and carbon-12 in the accelerator beam, allowing a precise correction (Beta Analysis Inc, written communication 1989).

A.2.2.2 BENTHIC FORAMINIFERA

Benthic foraminifera were separated from selected intervals of slab trays according to the procedure described by Piper (1980). This procedure consists of drying, weighing and then disaggregating the sample with a rubber policeman. The disaggregated sample was placed in 20 ml of 1% sodium hexametaphosphate (calgon) for 24 hours and then sieved though a 62.5μm (4 ϕ) screen. The material remaining on the sieve was dried at room temperature. Foraminifera tests were then picked and counted under a binocular microscope. Scanning electron photographs were made of the different species identified (Plate 3.2). The identification of the species was based on index samples of species from Labrador Shelf cores provided by Dr. G. Vilks of the Atlantic Geoscience Centre in Dartmouth. Drs. Vilks and Deonarine verified the species identification of the author.

A.2.2.3. POLLEN, SPORES AND DINOFLAGELLATES

Approximately 10 g of dried sediment was subsampled from twenty slab trays. Pollen, spores and dinoflagellates were separated according to the following procedure. Two Eucalyptus (spike) tablets were added to each of the samples. Carbonate was removed by treatment with 10% HCl for two hours with frequent stirring until no reaction was observed; the sample was then washed and centrifuged several times (more than three times). Silica was removed by treatment with concentrated hydrofluoric acid and continuous stirring; in some cases the reaction was so violent that acetone and distilled water were added to reduce the temperature of reaction. The samples were kept in the hydrofluoric acid for 12 hours with regular stirring every 1-2 hours. The samples were then centrifuged and washed in order to remove the acid.

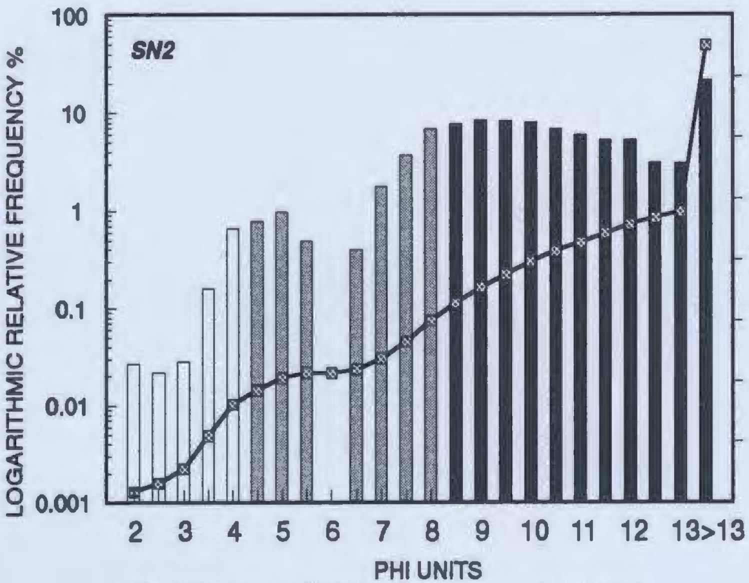
After treatments, the samples were wet sieved through a 10 μm sieve using distilled water. The material was then mounted on glass slides for identification and counting. Identification and counting were done by Dr. P.J. Mudie at the Atlantic Geoscience Centre.

APPENDIX B

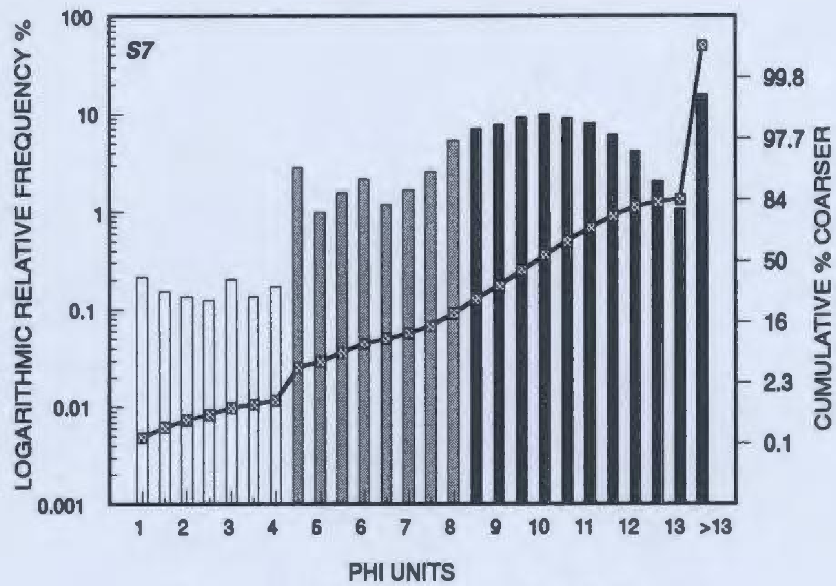
BI. GRAIN SIZE DATA

Table BI.1. Grain size analyses of the mud samples mentioned in the thesis. The graphical grain size statistical parameters are also included.

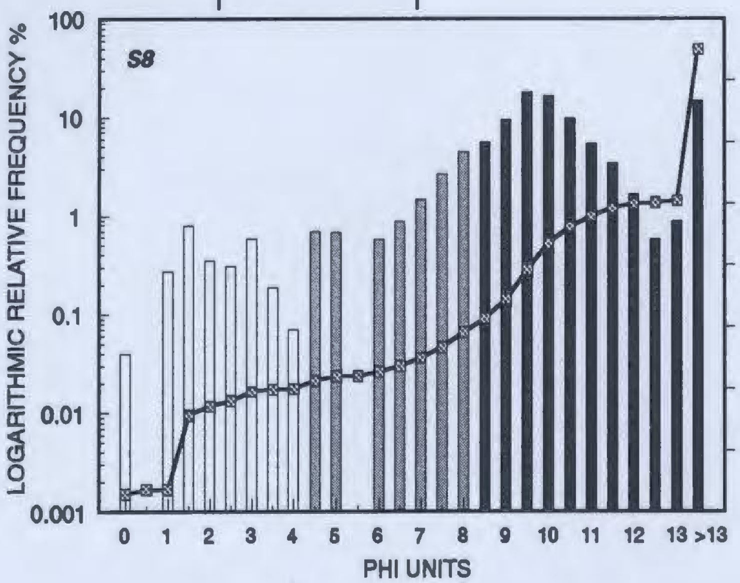
SAMPLE	SAND %	SILT %	CLAY %	MEDIAN $Md\phi$	MEAN $M\phi$	SORTING $\sigma\phi$	SKEWNESS Sk_G
MUD SAMPLES							
S1	3.10	23.40	73.60	7.75	7.25	2.00	-0.25
S2	13.90	24.40	61.70	9.75	----	----	----
S3	3.80	28.90	67.30	10.0	9.75	2.25	-0.11
S7	1.10	31.80	67.10	9.00	8.25	3.25	-0.23
S8	2.30	26.60	71.10	9.25	8.25	2.25	-0.44
S13	11.13	61.25	27.62	4.25	----	----	----
S16	3.70	65.40	30.90	4.00	----	----	----
S17	26.90	25.40	47.80	7.85	----	----	----
S18	1.60	71.20	27.20	7.00	8.75	3.25	0.54
S47	34.50	38.20	27.40	5.50	7.13	3.38	0.48
S48	36.40	31.90	31.70	4.00	6.13	3.38	0.63
S51	70.10	13.90	16.00	3.75	5.88	2.63	0.81
S71	15.50	27.70	56.80	8.80	7.10	3.3	-0.52
S72	11.60	29.30	59.10	9.00	7.83	2.83	-0.42
90F	2.78	48.53	48.69	----	----	----	----
SM	81.50	9.30	9.20	3.25	3.43	0.58	0.30
SN2	0.9	31.3	67.8	10.2	----	----	----



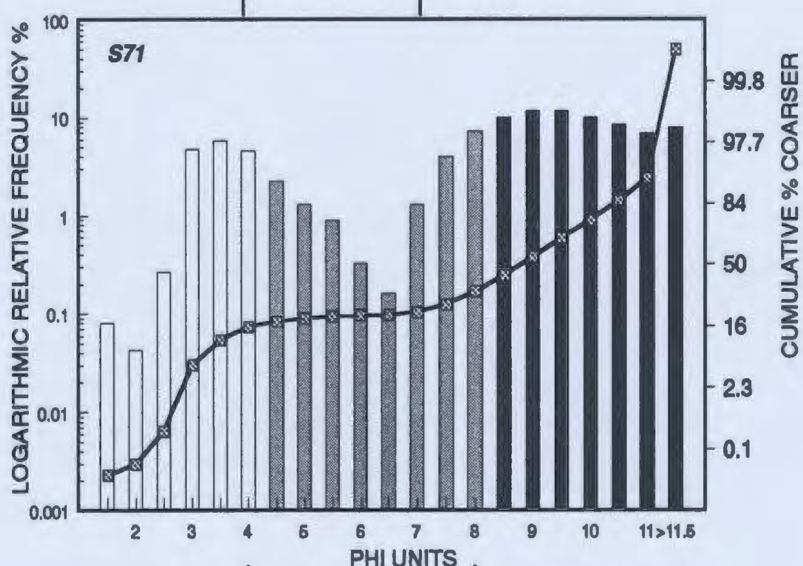
SAND=0.89% | SILT=15.05% | CLAY=84.06%



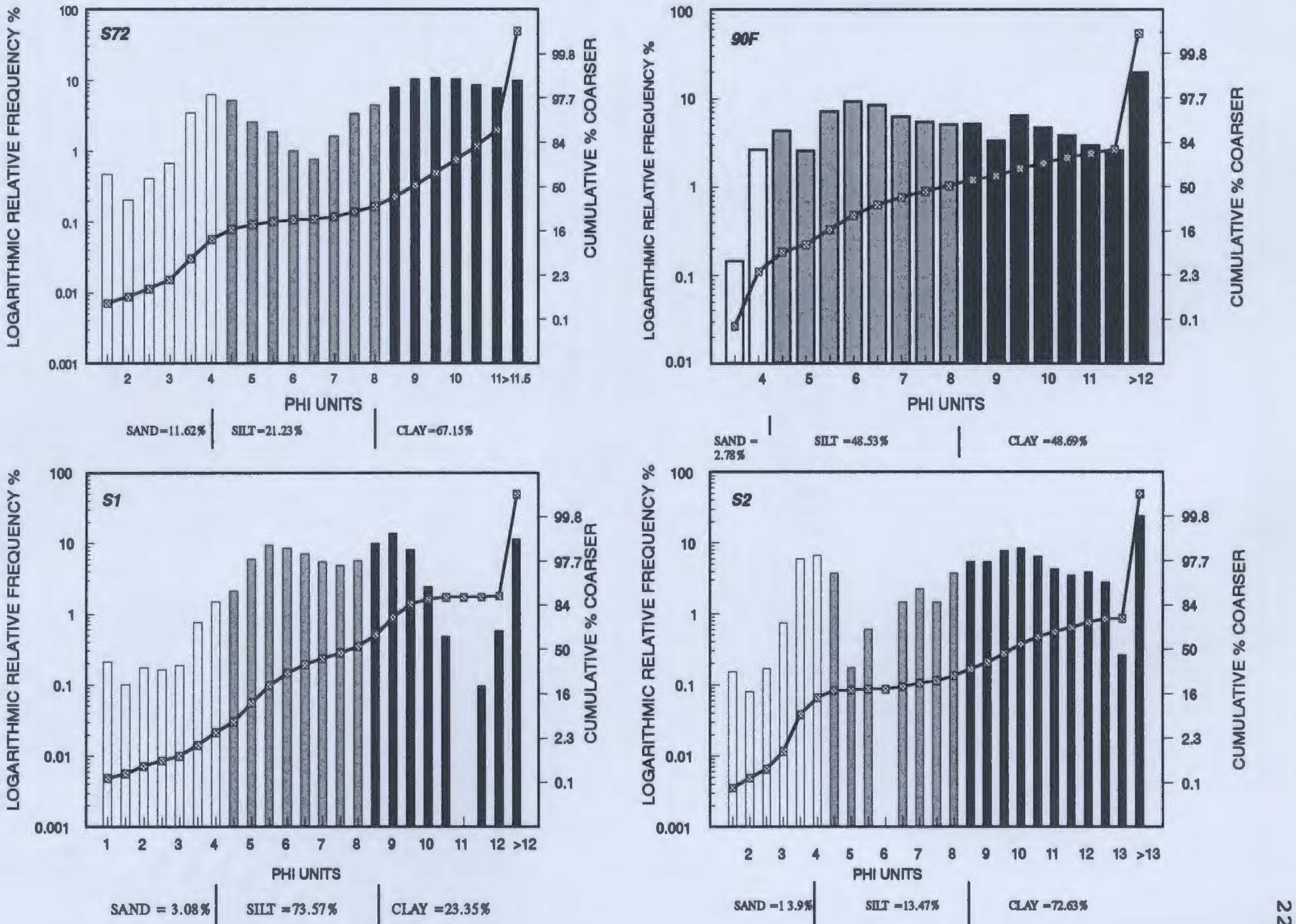
SAND = 1.14% | SILT =18.39% | CLAY =80.47%

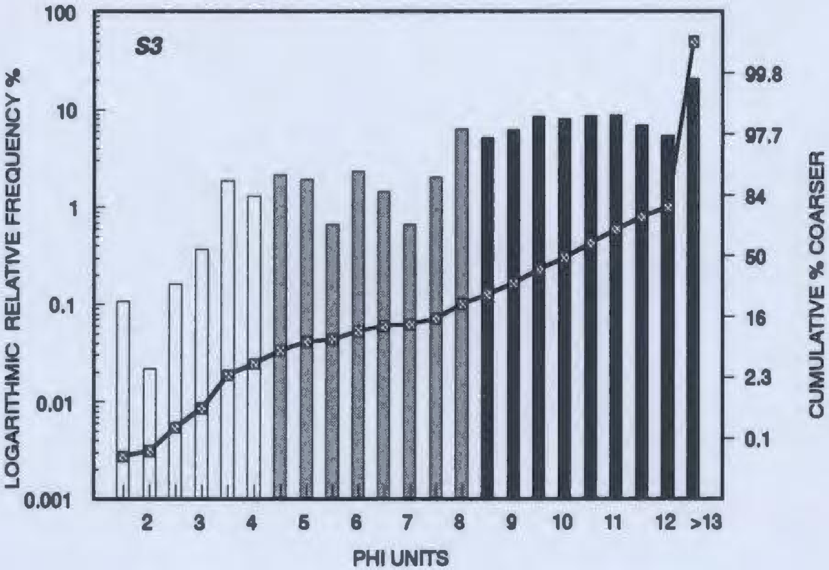


SAND = 2.55% | SILT =11.43% | CLAY =86.02%

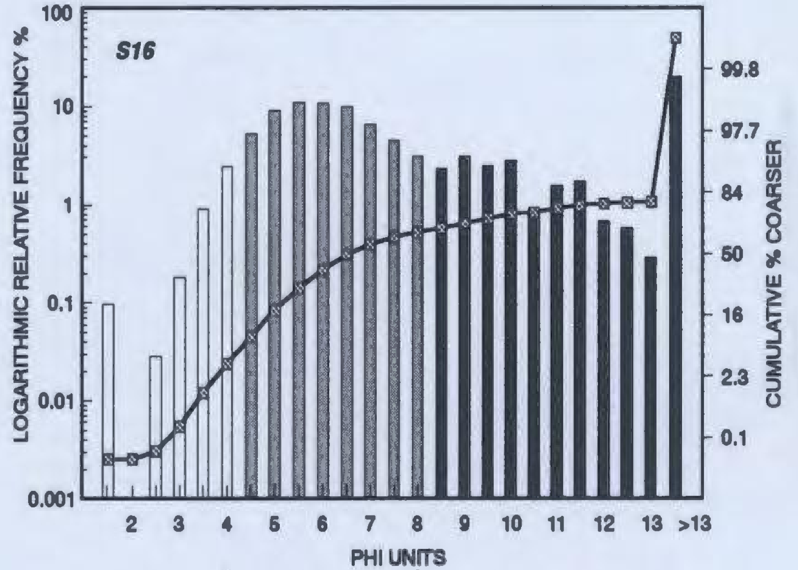


SAND=15.5% | SILT=17.66% | CLAY=66.84%

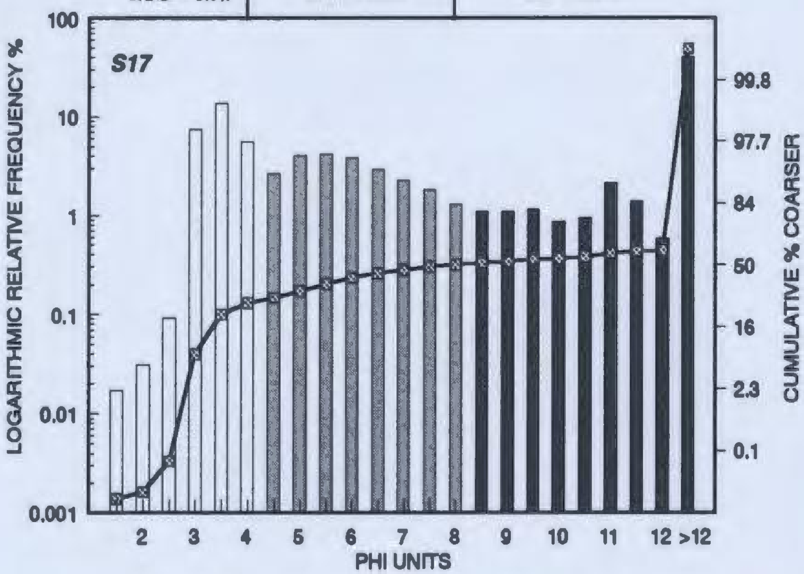

S72
LOGARITHMIC RELATIVE FREQUENCY %
CUMULATIVE % COARSER
PHI UNITS
SAND=11.62%
SILT=21.23%
CLAY=67.15%
90F
LOGARITHMIC RELATIVE FREQUENCY %
CUMULATIVE % COARSER
PHI UNITS
SAND = 2.78%
SILT =48.53%
CLAY =48.69%
S1
LOGARITHMIC RELATIVE FREQUENCY %
CUMULATIVE % COARSER
PHI UNITS
SAND = 3.08%
SILT =73.57%
CLAY =23.35%
S2
LOGARITHMIC RELATIVE FREQUENCY %
CUMULATIVE % COARSER
PHI UNITS
SAND =13.9%
SILT =13.47%
CLAY =72.63%



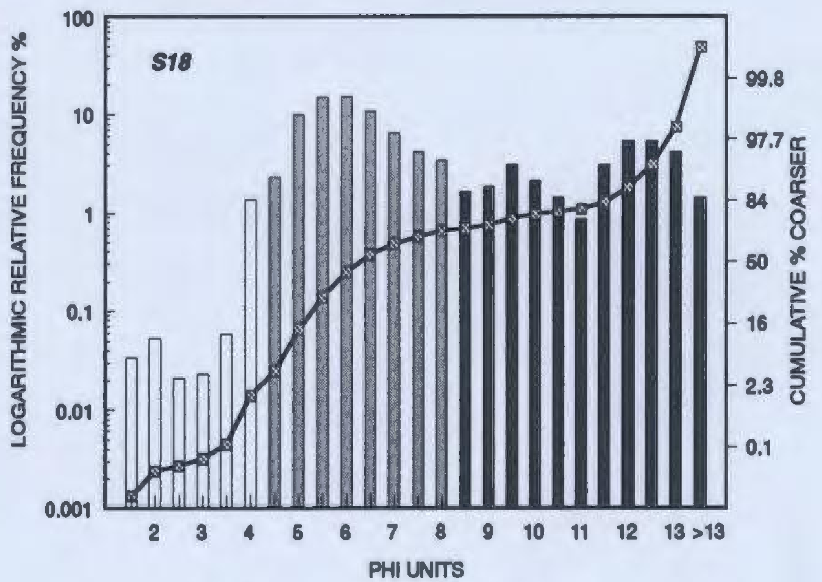
SAND = 3.8% | SILT =17.56% | CLAY =78.64%



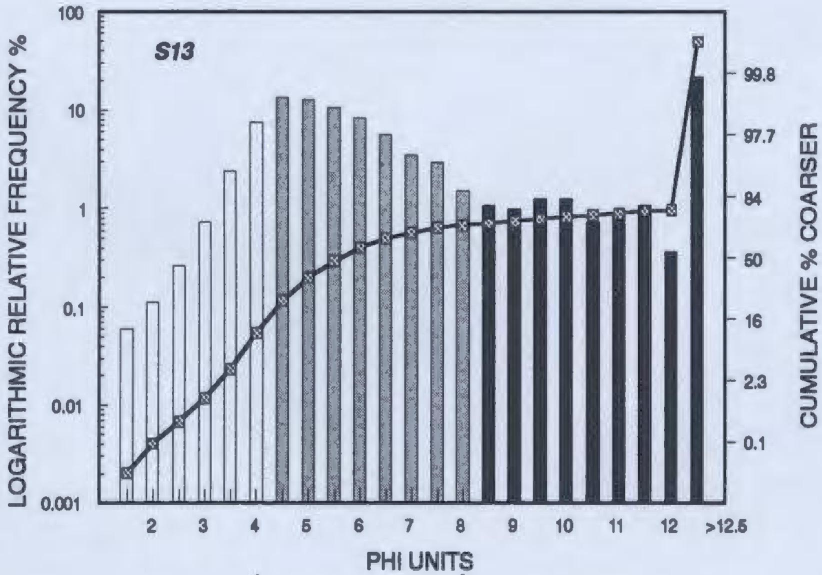
SAND =3.7% | SILT =65.4% | CLAY =30.9%



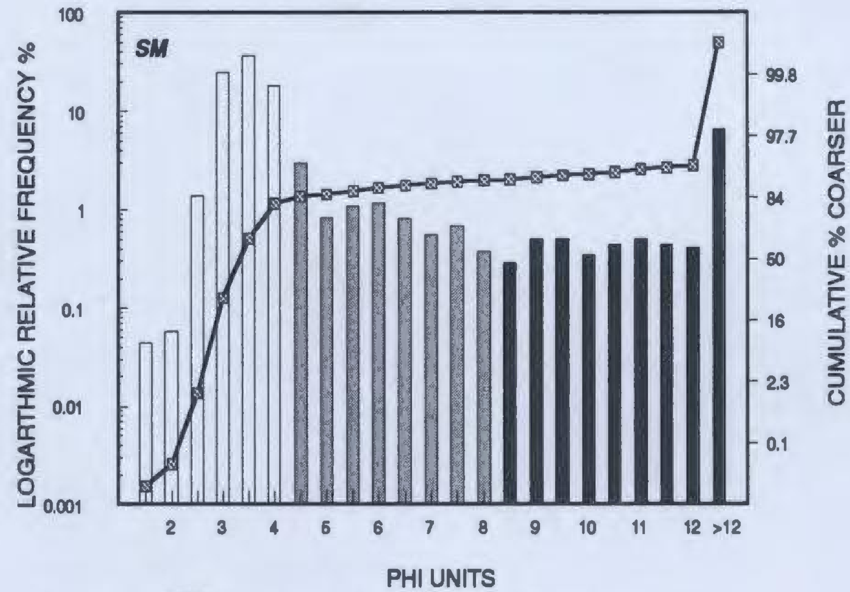
SAND =26.9% | SILT =23.15% | CLAY =49.95%



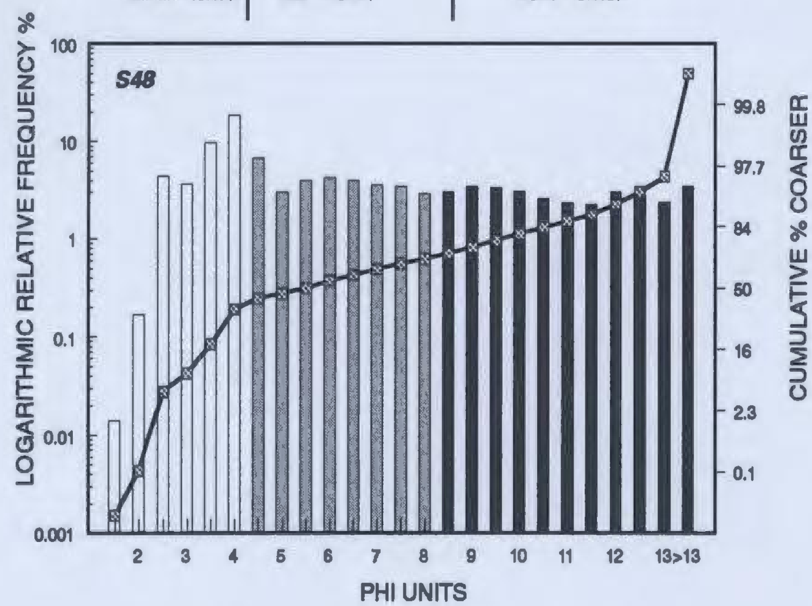
SAND =1.56 % | SILT =67.64 % | CLAY =30.8 %



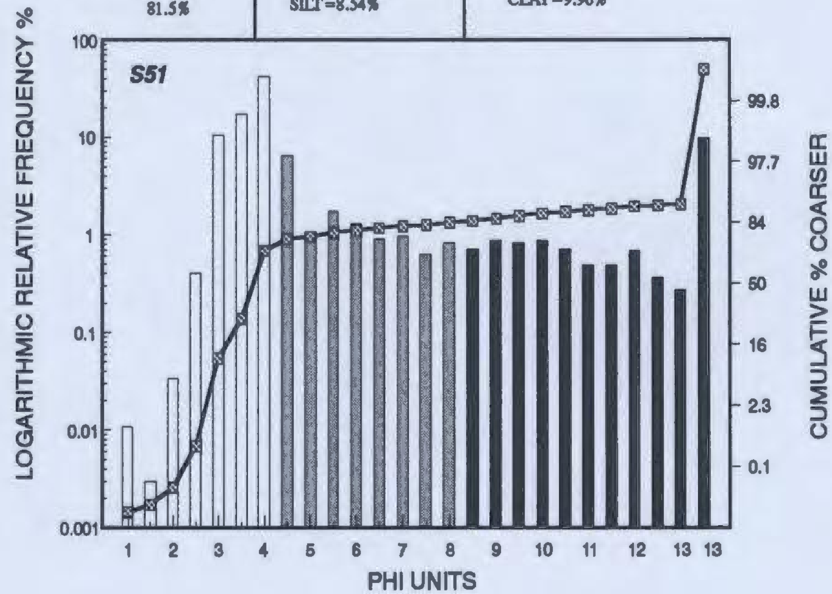
SAND =11.13% | SILT =61.25% | CLAY =27.62%



SAND= 81.5% | SILT=8.54% | CLAY=9.96%



SAND =36.4% | SILT=31.9% | CLAY =31.7%



SAND=70.14% | SILT=13.86% | CLAY=16%

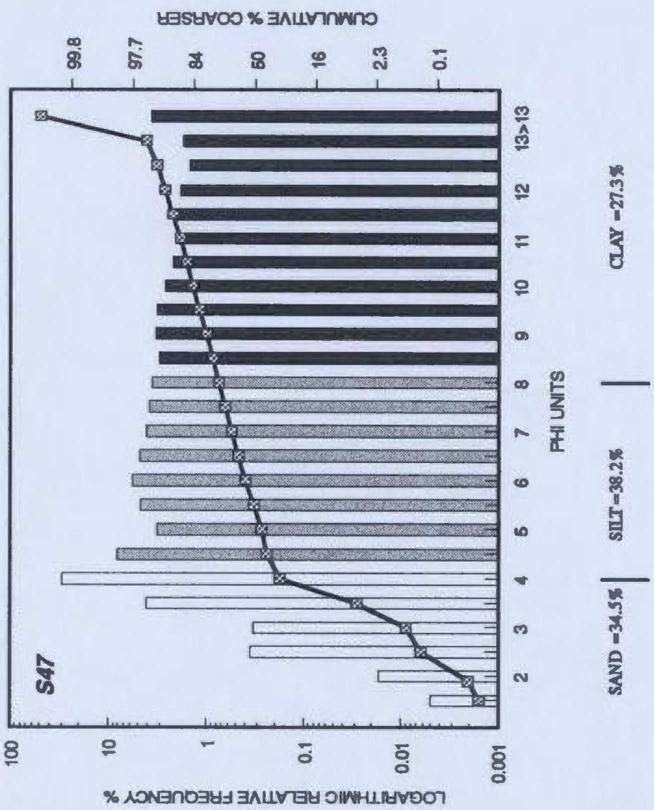
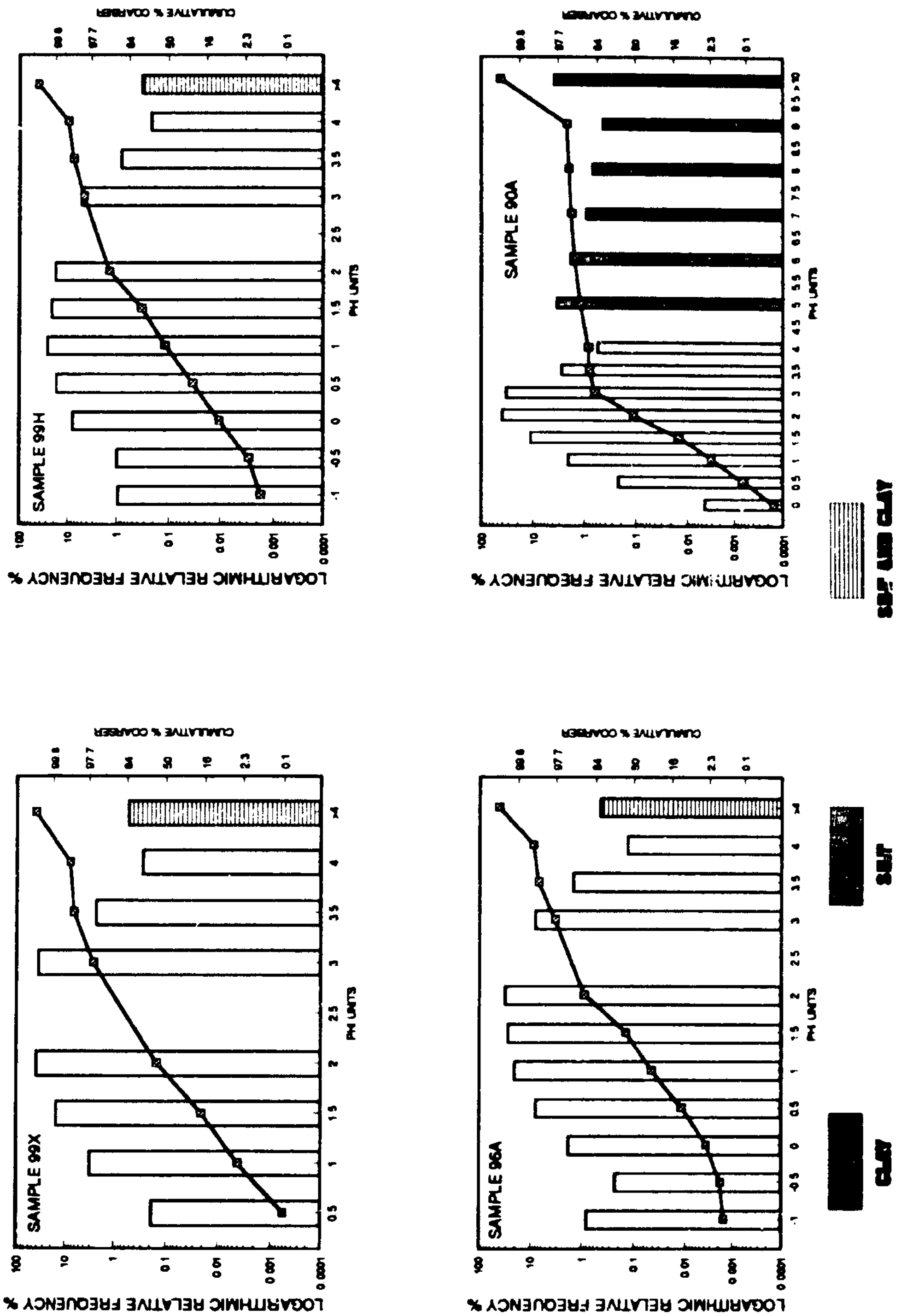

S47
LOGARITHMIC RELATIVE FREQUENCY %
100
10
1
0.1
0.01
0.001
CUMULATIVE % COARSER
99.8
97.7
84
50
16
2.3
0.1
2
3
4
5
6
7
8
9
10
11
12
13>13
PHI UNITS
SAND =34.5%
SILT =38.2%
CLAY =27.3%

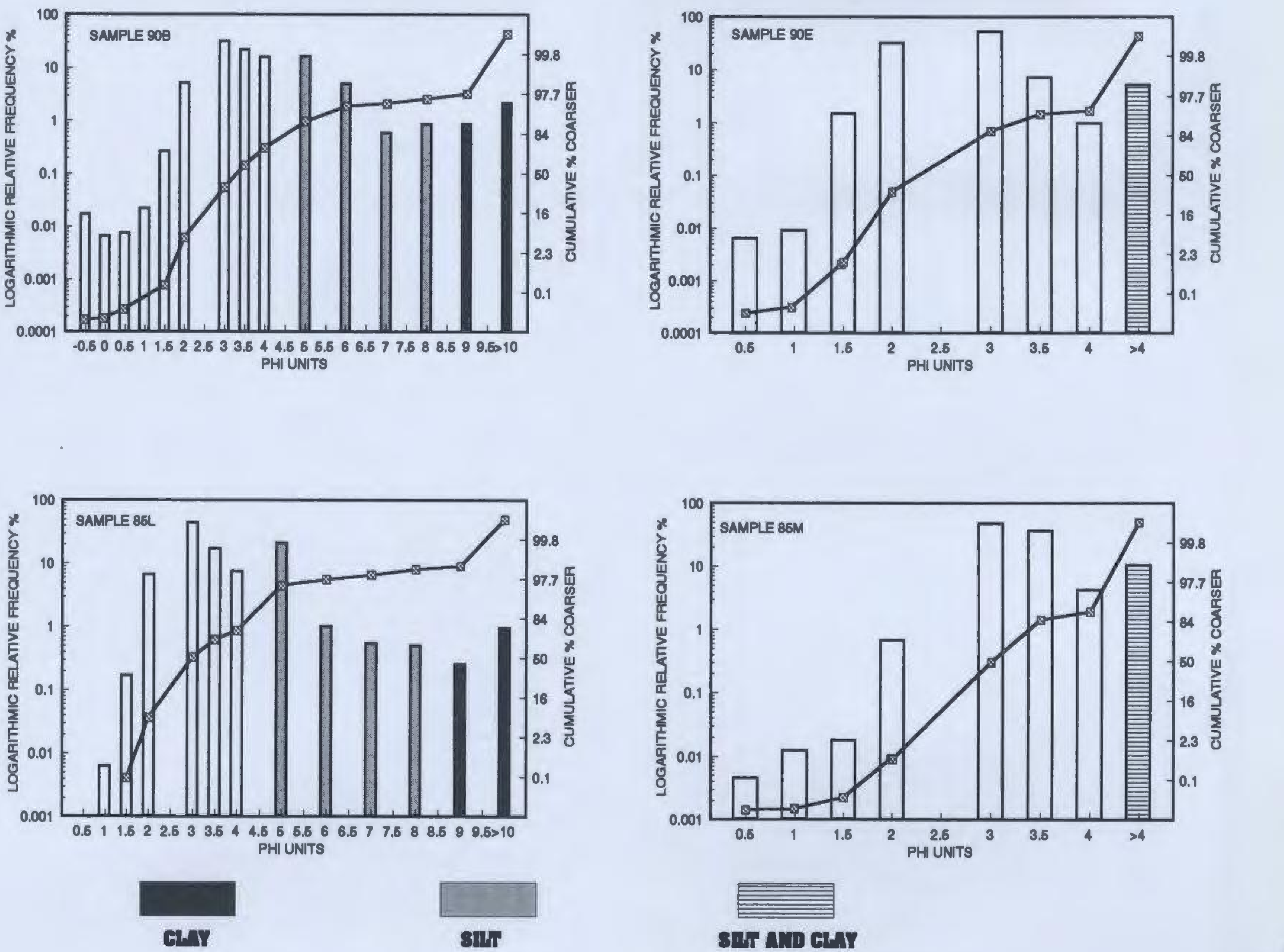
Table BI.2. Grain size analyses of the sand samples mentioned in the thesis. The graphical grain size statistical parameters are also included.

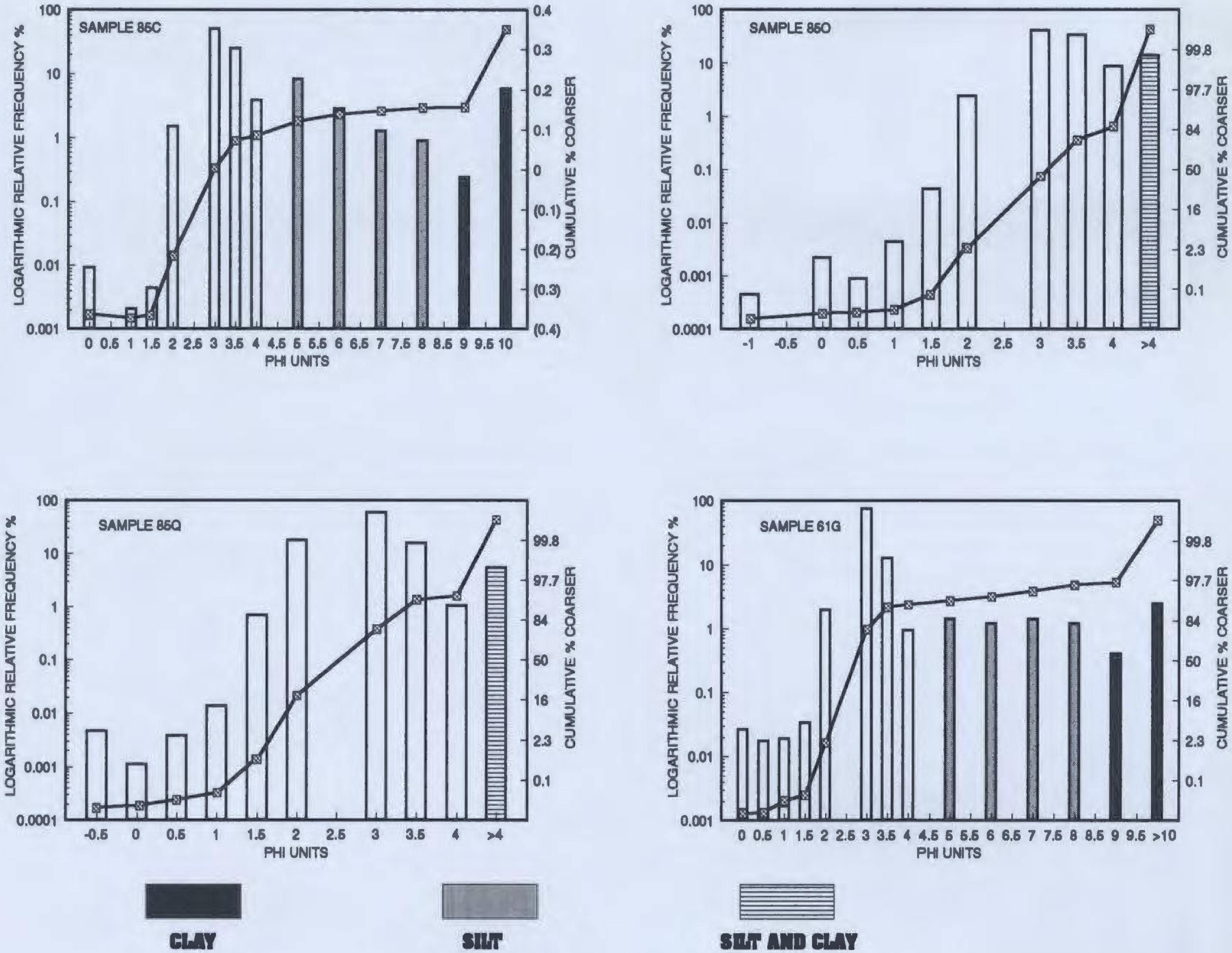
SAMPLE	SAND %	SILT %	CLAY %	MEDIAN $Md\phi$	MEAN $M\phi$	SORTING $\sigma\phi$	SKEWNESS Sk_G	KURTOSIS K_G
SAND SAMPLES								
50B	97.68	2.32*		2.29	2.27	0.67	-0.05	----
61B	97.12	2.88*		2.14	2.21	1.27	0.19	1.32
61G	91.88	5.24	2.88	2.76	2.79	0.43	0.11	4.24
85C	80.61	13.24	6.15	3.01	3.29	0.85	0.33	----
85L	75.63	23.17	1.20	3.12	3.27	1.08	0.21	0.82
85M	89.45	7.37	3.18	3.00	3.04	0.47	0.14	----
85O	86.09	9.78	4.12	3.06	3.12	0.67	0.14	----
85Q	94.50	5.50*		2.47	2.44	0.57	0.088	----
90A	89.04	6.69	4.27	2.06	2.20	0.74	0.29	2.85
90B	74.37	22.55	3.08	3.38	3.48	1.04	0.14	1.32
90E	94.78	5.22*		2.29	2.34	0.56	0.14	----
96A	99.48	0.52*		1.35	1.29	0.62	-0.19	1.17
99H	99.61	0.39*		0.87	0.89	0.71	0.03	0.95
99X	99.42	0.58*		1.89	1.91	0.49	0.07	0.05

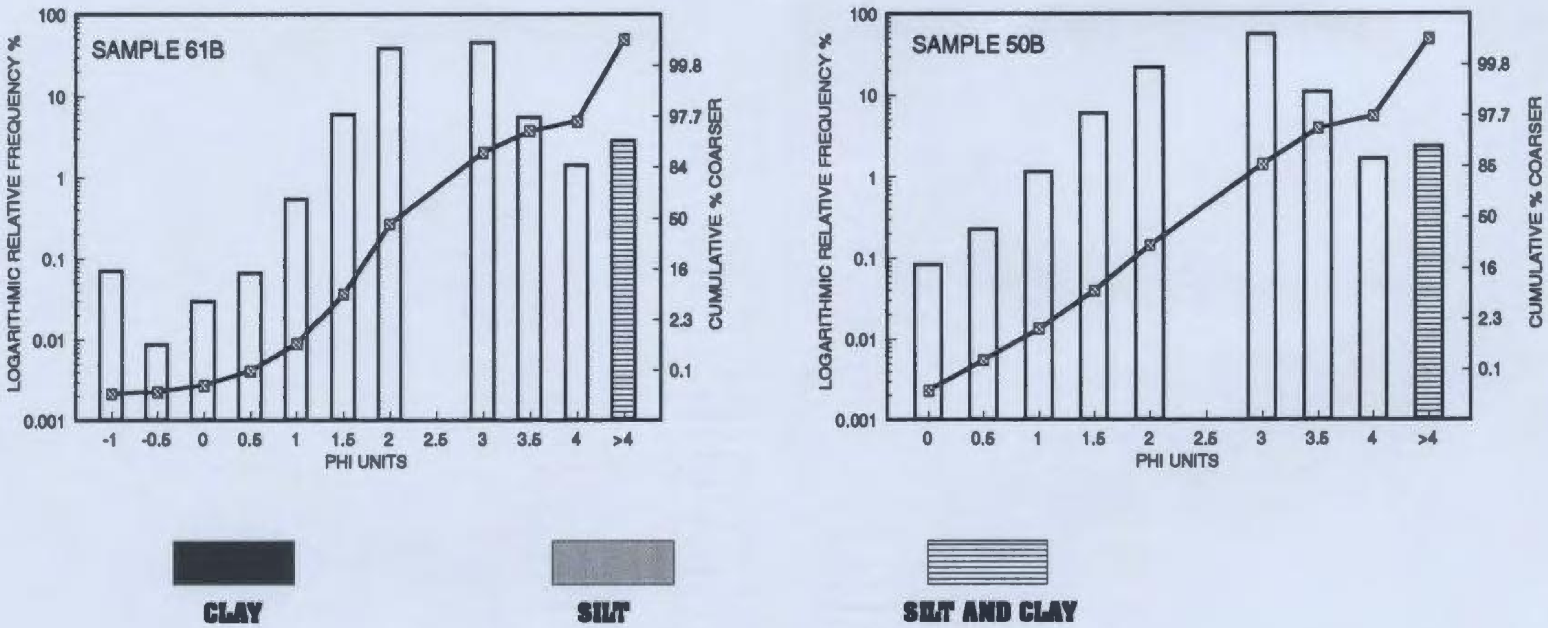
* Silt and clay.



SAMPLE 99H
SAMPLE 90A
SAMPLE 99X
SAMPLE 96A
LOGARITHMIC RELATIVE FREQUENCY %
CUMULATIVE % COARSER
PHI UNITS
SILT AND CLAY
SILT
CLAY


SAMPLE 90B
SAMPLE 90E
SAMPLE 85L
SAMPLE 85M
LOGARITHMIC RELATIVE FREQUENCY %
CUMULATIVE % COARSER
PHI UNITS
100
10
1
0.1
0.01
0.001
0.0001
99.8
97.7
84
50
16
2.3
0.1
-0.5 0 0.5 1 1.5 2 2.5 3 3.5 4 4.5 5 5.5 6 6.5 7 7.5 8 8.5 9 9.5 >10
0.5 1 1.5 2 2.5 3 3.5 4 >4
CLAY
SILT
SILT AND CLAY


SAMPLE 85C
SAMPLE 85O
SAMPLE 85Q
SAMPLE 61G
LOGARITHMIC RELATIVE FREQUENCY %
CUMULATIVE % COARSER
PHI UNITS
CLAY
SILT
SILT AND CLAY


SAMPLE 61B
LOGARITHMIC RELATIVE FREQUENCY %
100
10
1
0.1
0.01
0.001
CUMULATIVE % COARSER
99.8
97.7
84
50
16
2.3
0.1
-1
-0.5
0
0.5
1
1.5
2
2.5
3
3.5
4
>4
PHI UNITS
SAMPLE 50B
LOGARITHMIC RELATIVE FREQUENCY %
100
10
1
0.1
0.01
0.001
CUMULATIVE % COARSER
99.8
97.7
85
50
16
2.3
0.1
0
0.5
1
1.5
2
2.5
3
3.5
4
>4
PHI UNITS
CLAY
SILT
SILT AND CLAY

APPENDIX B

BII. PALAEONTOLOGICAL DATA

SAMPLE	54C	54E	67	68B	75A	75B
TOTAL POLLEN AND SPORES	199	1253	1141	797	5099	388
TOTAL DINOCYSTS	36	189	23	0	287	0
TOTAL FRESH WATER ALGAE	18	87	0	67	668	76
TREE POLLEN %						
Abies	0	3	4	0	0	0
P.alba	36	9	20	0	4	35
P.mariana	0	15	6	0	31	47
Pinus	9	6	2	8	5	0
Larix	0	0	0	0	0	0
Betula spp.	9	9	2	8	19	0
SHRUB %						
Alnus	0	15	12	25	3	0
B.nana	0	2	4	9	0	0
Ericales	0	2	2	0	3	0
Salix	0	0	2	0	0	0
HERBS %						
Ambrosia/ *Artemisia*	0 0	0 0	0 2	0 0	0 0	0 0
Cyperaceae	0	1	4	0	0	0
Compositae	0	0	0	0	3	0
Gramineae	0	3	4	0	3	0
Rosaceae	0	1	0	0	0	6
FERNS %						
Polypodaceae	18	9	16	8	4	0
Lycopodium	19	17	18	42	14	12
MOSS %						
Sphagnum	9	8	2	0	11	0

Table BII.a. Number of pollen, spore, algae and dinocysts species collected from sections 54 to 75.

SAMPLE	85A	85B	85K	90	91	92	99B
TOTAL POLLEN AND SPORES	203	1248	2662	1608	983	1263	2533
TOTAL DINOCYSTS	0	38	453	0	0	0	133
TOTAL FRESH WATER ALGAE	0	38	283	36	52	0	0
TREE POLLEN %							
Abies	0	15	2	0	0	0	0
P.alba	20	21	18	16	33	5	12
P.mariana	20	15	16	25	28	25	48
Pinus	10	0	11	4	0	8	3
Larix	0	0	0	2	0	0	0
Betula spp.	0	3	2	2	0	0	5
SHRUB%							
Alnus	20	3	9	2	0	5	5
B.nana	0	0	0	7	11	7	5
Ericales	0	0	0	0	0	0	0
Salix	0	0	0	0	0	0	0
HERBS%							
Ambrosia/ *Artemisia*	0 0	0 0	0 0	0 0	0 0	3 0	0 0
Cyperaceae	0	0	0	0	6	3	3
Compositae	0	0	0	0	0	0	0
Gramineae	0	0	0	4	0	5	0
Rosaceae	0	0	0	0	0	3	0
FERNS%							
Polypodaceae	20	15	4	13	0	5	3
Lycopodium	10	25	27	18	22	23	5
MOSS%							
Sphagnum	0	3	11	7	0	8	11

Table BII.b. Number of pollen, spore, algae and dinocysts species collected from sections 75 to 99.