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Late Quaternary glaciation and sea-level change along the Atlantic coast of Nova Scotia: correlation of land and sea events

by

Rudolph Raphael Stea

Submitted in partial fulfilment of the requirements for the degree of Doctor of Philosophy at

Dalhousie University
Halifax, Nova Scotia
1995

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ABSTRACT

The inner continental shelf off Nova Scotia can be subdivided into 5 coast-parallel terrain zones which record glaciation, and sea level rise and fall during the Late-glacial period. At the last glacial maximum an ice mass centred in the Gulf of St. Lawrence extended to the shelf edge (Ice Flow Phase 2). During an initial deglaciation phase (ca. 17-15 ka) ice was drawn out of the Gulf of Maine, isolating an ice mass over Nova Scotia which later became an active centre of outflow (Scotian Ice Divide-Ice Flow Phase 3). The Scotian Shelf End Moraine Complex (SSEM C) lies at the seaward margin of the inner shelf. The surface SSEM C till is correlated with an autochthonous, stony till on land formed under the Scotian Ice Divide. The SSEM C formed at the ramp margin of a buoyant ice sheet. The Phase 3 ice margin was stabilized by isostatic recovery from Phase 2 ice loads.

Landward of the end moraines is a series of linear depositional basins termed the Basin Zone which contain a complete sedimentary record of events after the deposition of the SSEM C. The lowest depositional sequences (Sequences 1+2; Emerald Silt facies A) in these basins are interpreted to be ice proximal sediment, deposited by overflow and interflow meltwater plumes emanating from a retreating ice margin. Periodic stillstands of this ice mass are marked by small, incipient moraines within the Basin Zone. DeGeer moraines are found over a wide area (Morainal Zone) landward of the Basin Zone. Two morphological varieties were defined; symmetrical and asymmetrical moraines. These formed underneath and at the edge of a rapidly advancing local ice cap centred in northern Nova Scotia (Ice Flow Phase 4; ca 13 ka). From 13-12 ka sea-level dropped rapidly, due to general isostatic recovery. Ice receded out of marine areas depositing ice distal glacial marine Sequences 3 and 4 (Emerald Silt facies B), in the Basin Zone. A lowstand
shoreline (ca 11.6 ka) at -65 m is marked by the landward transition from morainal
topography to a shoreface ramp, terrace and wave-cut platform. Increased storminess, sea
ice and icebergs during the Younger Dryas climatic cooling (ca 11-10.0 ka), produced
Sequence 5 in the Basin Zone.
LIST OF AbbREVIATIONS

IKU: Continental Shelf Institute (Norway) grab sampler
SSEMC: Scotian Shelf End Moraine Complex
SMB: South Mountain Batholith
LGM: Last glacial maximum
RSL: Relative sea level
BSL: Below present mean sea level
TESS: Truncated Emerald Silt subzone
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This Ph.D thesis was born out of the frustration I felt when losing the trail of glaciers at the last striated outcrop above high tide. During my 2 years at Dalhousie and with the support of the Geological Survey of Canada, I have learned to be a marine geologist, and to develop a larger view of glaciers, both on land and at sea.

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CHAPTER 1: INTRODUCTION

"The subtlety of nature is greater many times over than the subtlety of argument"

Sir Francis Bacon

The Atlantic Coast of Nova Scotia is a boundary between two vastly different environments; land and sea. During the Quaternary, ocean, land and ice battled for territory. The present coastline is a transient feature, recording the continuing advance of the ocean onto the land before the next major glacial cycle.

Scientists who study the Quaternary in the marine and terrestrial environments have differing goals and methods. The purpose of this thesis is to establish a correlation between Quaternary landforms and sedimentary units on the Atlantic Coast of Nova Scotia and "bridge the gap" between the land and sea disciplines. The ultimate goal is to develop a glaciation/sea-level model for Nova Scotia and the inner continental shelf that is consistent with both the marine and terrestrial records.

Paleoclimatic research over the last two decades has been focussed on atmosphere, ocean and cryosphere interactions (Crowley and North, 1991). Linking the deep ocean Quaternary records with land glacial records is not an easy task. The land stratigraphic record consists of deposits from relatively recent glacial advances and retreats. In Atlantic Canada, pre-Illinoian strata have not been identified, except for the problematic "Bridgewater Conglomerate" (Grant, 1989). In fact, Tertiary sediments are not known on land and the youngest pre-Quaternary unconsolidated strata are Cretaceous in age (Fowler and Stea, 1979).
Erosion, sporadic deposition and exposure, and lack of suitable dating methods make the history of pre-Wisconsinan glaciations less resoluble (Clark et al., 1993).

Deep ocean sediment faunal and isotopic records provide a continuous Quaternary climatic evolution but only give indirect clues to ice sheet activity, and the precision is limited by bioturbation, slow sedimentation rates and oceanic mixing (Kennett, 1982). The resolution of any glaciation/sea-level model is also limited by the "inertia" of the ice masses themselves. Small mountain glaciers are extremely sensitive and respond to changes in global climate in time spans of $10^2$ years. Larger ice masses such as the Greenland Ice Sheet are sluggish, with response times in the order of $10^4$ to $10^5$ years (Sugden and John, 1976). Late Quaternary ice caps in the Maritime provinces were intermediate in size between these two extremes (Prest and Grant, 1969).

Cooke (1972) attempted a global correlation of onshore and offshore glacial events for the last 2 million years by matching major glacial and interglacial periods in diverse land and sea paleoclimatic records. More recently, Clarke et al. (1993) attempted to match the terrestrial record of glacial advances in North America with the oceanic record for the last 100,000 years. Such correlations are flawed because they rely on age assumptions rather than direct lithostratigraphic correlation. Radiocarbon dating provides an accurate chronology for deposits younger than 30,000 years BP. A lack of suitable methods beyond that range hinders temporal correlation (Clark et al., 1993). The chronology of oceanic records has been refined based on the assumed causal relationship between Milankovitch cycles and climatic proxies in deep ocean cores (Hays et al., 1976). This orbitally "tuned" chronology is currently in widespread use (SPECMAP; Imbrie et al., 1984). Recent dating of the Devils Hole speleothems has provided an accurate terrestrial chronology that
is in conflict with the SPECMAP chronology (Winograd, 1992; Broecker, 1992). The application of a direct correlation of land and sea glacial deposits will help to explain discrepancies in land and sea glacial chronologies and refine the role of internal feedback mechanisms forcing glacial-interglacial climatic cycles.

The continental shelf of Nova Scotia is an interface between the deep ocean and the land. Glaciers eroded land areas and transported the sediment to the edge of the continental shelf (King and Fader, 1986; Piper et al., 1990). If erosional and depositional events on land and on the shelf can be linked, then the first step in the linkage of the glaciers with offshore abyssal plain sediments will be established. Ice rafted debris (Conolly and Ewing, 1965; "Heinrich" layers; Heinrich, 1988; Bond et al., 1992; Grousset et al., 1993), recognized within the deep ocean basins, complete the sedimentary cycle from land to deep ocean.

This study will focus on Wisconsinan events on the inner shelf of southern and eastern Nova Scotia. This was a time of worldwide glacial advances and retreats, and fluctuating climates.

The detailed objectives of the project include:

(1) Development of lithic and seismic criteria for the correlation of offshore and onshore Quaternary landforms and formations.

(2) Determination of the flow patterns, offshore extent and recessional history of glaciers.
(3) Dating of the Quaternary formations to establish a temporal framework for glacial events in the region.

(4) Evaluation of the extent of the sea-level variations on inner shelf, including low sea-level stands, residual beaches and erosional unconformities.

(5) Elucidation of the glacial, proglacial and marine processes responsible for the sedimentary units.

Glacial history on the continental shelf of Nova Scotia is inextricably linked to local sea-level variations (Scott et al., 1987a). The goal of this research is to develop comprehensive, empirical model of glaciations and sea-level change in Maritime Canada.

1.1 STUDY AREA

The oceanic area selected for this study is a portion of the coastline and inner Scotian Shelf from Halifax to Country Harbour, extending 80 km seaward of the present shoreline and including a major morainal system termed the Scotian Shelf End Moraine Complex (King et al., 1972, Fig. 1). The terrestrial area includes a strip of land approximately 40 km wide along the Eastern Shore of Nova Scotia from Halifax to Country Harbour (Fig. 1).
Figure 1. Location of the study area, seismic tracklines and terrestrial reference sections. In this thesis study only the Halifax, Sheet Harbour and Country Harbour transects were mapped.

Terrestrial reference sections:
1. West Lawrencetown Section
2. Mushaboom Section
3. Smith Cove Section
4. Wine Harbour Section.
1.2 GEOLOGICAL SETTING AND PHYSIOGRAPHY

Nova Scotia is an amalgamation of disparate bedrock terranes brought together by the dynamics of plate tectonics (Fig. 2). On mainland Nova Scotia, the Cobequid-Chedabucto Fault system separates the Avalon Zone to the north from the Meguma Zone to the south (Williams, 1978). The Avalon Zone consists of highland massifs of Precambrian to Carboniferous volcanic, metasedimentary and igneous rocks flanked by basinal sedimentary rocks of Carboniferous and Triassic age. The Meguma Zone consists largely of Cambro-Ordovician metasedimentary rocks (Meguma Group) intruded by Devono-Carboniferous granitoid rocks.

1.2.1 Meguma Zone

The Meguma Group has been interpreted as the remnants of a North African continental slope sedimentary prism that was amalgamated onto the North American continent during the Devonian (Schenk, 1971). The age of these rocks ranges from Cambrian to Ordovician (O'Reilly et al., 1992). These sedimentary rocks were regionally metamorphosed and deformed into upright, northeast-trending folds during the Lower Devonian Acadian Orogeny (Keppie, 1977). Regional metamorphism associated with the Acadian Orogeny has been dated radiometrically at 390-410 Ma (Muecke et al., 1988). The regional metamorphic grade ranges from chlorite and biotite grade in the central region to andalusite; staurolite-cordierite and sillimanite grade toward the east and southwest (Keppie and Muecke, 1979).
Figure 2. Simplified regional bedrock geology (after Sanford et al., 1991).
Figure 2.
Intrusive into the metasediments of the Meguma Group is a cogenetic suite of peraluminous granitic rocks which range in composition from tonalite to granodiorite to monzogranite (MacKenzie and Clarke, 1975; Albuquerque, 1977, MacDonald et al., 1992). All the granitic units contain the major minerals quartz, plagioclase, alkali feldspar, biotite and muscovite in varying degrees (O'Reilly et al., 1992). A large composite granitic pluton (South Mountain Batholith) underlies much of the area west of Halifax, while a number of smaller plutons occur along the Eastern Shore. The age of the granitic rocks has been established radiometrically and by intrusive relations with fossiliferous Lower Devonian sediments. Radiometric age dates indicate a range of ages from 374 Ma for the older biotite granodiorite and biotite monzogranite to 360 Ma for leucogranite dykes and two-mica monzogranite plutons (Clarke and Halliday, 1980).

The rock units of the Meguma Zone are part of an upland surface called the Southern Upland (Goldthwait, 1924; Roland, 1982; Map 1). The Southern Upland is an undulating, southeastward-dipping surface ranging in elevation from 300 m at the northern boundary with the Carboniferous Lowlands to sea level at the Atlantic coast. The upland surface is irregular with numerous strike-ridges and mounds formed of resistant rocks and constructional glacial features. The terrain is dominated by bedrock with thin drift cover. In the higher elevations, eastward-trending strike ridges are common following the regional Appalachian structural trend (Keppie, 1982). Glacially-modified, southeastward-trending fault valley systems (Cameron, 1956) create a rugged, irregular coastline. Major harbours such as Halifax, Sheet Harbour and Country Harbour (Fig. 1) and inland lakes such as Porters Lake east of Halifax, have fjord-like longitudinal profiles with overdeepened basins and sills (Hopkins, 1985). These silled basins attain 100 m in water depth (McKay, 1981).
A series of Carboniferous and Triassic basins are found within the Meguma Zone. These occur as fault-block basins with intervening upfaulted horst blocks (Boehner et al., 1986). The Carboniferous basins consist of continental sedimentary rocks (Horton Group) overlain by a largely marine sequence of evaporites, limestone, dolomite and clastic sedimentary rocks (Windsor Group). In the Triassic basins red bed conglomerates, sandstones and mudstones of the Fundy Group are overlain by several basalt flows collectively termed the North Mountain Basalt. The Carboniferous and Triassic basins form prominent lowlands including the Annapolis Valley, and the Hants-Colchester lowland (Map 1).

The North Mountain is a cuesta ridge that forms the edge of the Bay of Fundy for 200 km. The ridge attains heights of 230 m and slopes towards the Bay of Fundy.

1.2.2 Avalon Zone

The Avalon Zone found north of the Cobequid-Chedabucto fault system, is defined by an overstep sequence of Cambrian to Ordovician shallow marine rocks that contain a unique Acado-Baltic fauna (Murphy et al., 1991; Fig. 2). The terrane was formed by closure of the Iapetus Ocean during the Silurian to Early Devonian (Keppie, 1977). The Cobequid Highland is a prominent upland feature within the Avalon Zone attaining heights of 300 m (Fig. 2). It is composed of Pre-Late Carboniferous strata bounded by the Cobequid fault on the southern edge and unconformities and faults on the northern edge (Donohoe and Wallace, 1985). These rocks include Hadrynian basement gneisses, quartzites and metavolcanics surrounded by Silurian to Devonian metasediments (Donohoe and Wallace, 1985). Granitic plutons from Hadrynian to Devono-Carboniferous intrude sedimentary strata in this region. These plutons form a bimodal suite of diorite and syenogranite and
tend to be normative corundum free or corundum poor (Clarke, et al., 1980; Pe-Piper et al., 1989). In some areas they exhibit a foliation (Donohoe and Wallace, 1982). The top of the Cobequid Highlands is a plateau which is cut by youthful, V-shaped streams on the southern side and larger, U-shaped valleys on the northern side (Donohoe et al., 1992).

The Antigonish Highlands is another distinctive upland block on the western end of the northern mainland (Fig. 2; Map 1). It consists largely of Precambrian mafic and felsic volcanic rocks and turbidites, unconformably overlain by lower Paleozoic red conglomerates, arkoses, felsic and lithic tuffs, and red slates (Murphy, et al., 1991). Intrusive rocks are rare, compared to the Cobequid Highlands, and consist of appinites, gabbros, alaskite and granites. Appinites are found in isolated plutons in the Cobequid (H. V. Donohoe, pers. comm, 1993) and Antigonish Highlands (Murphy et al., 1991). They are ideal as indicator erratics (Stea et al., 1989). The upland surface of the Antigonish Highlands is more irregular than the Cobequid Highlands and dips gradually from 315 m at the steeper, northeastern fault-bounded edge to 180 m along the contact with the Devono-Carboniferous St. Mary's graben to the south.

1.2.3 Offshore Bedrock Geology

In the study area, the Scotian Shelf is divisible into three physiographic zones. The inner zone extends from the present shoreline out to the contact with the Scotian Basin (Fig. 2). The Scotian Basin is part of an accreted wedge of Mesozoic and Cenozoic sediments that were deposited on the eastern flank of the Appalachian Orogen, after the breakup of Pangea (Wade and MacLean, 1990). The inner zone is a zone of irregular topography with thin glacial drift and transgressive sediments underlain by metagreywacke of the
Goldenville Formation (King and MacLean, 1974). The middle zone consists of a series of broad basins filled with Quaternary sediments and underlain by Cretaceous and Tertiary marlstones, mudstones, sandstones and conglomerates (Wyandot and Banquereau Formations (King and MacLean, 1974). The outer zone is characterized by a series of broad, flat-topped banks composed of thick, sandy Quaternary sediments underlain by Mesozoic and Cenozoic rocks of the Scotian Basin.

1.3 CONCEPTS OF GLACIATION IN NOVA SCOTIA

As early as the late 1800s when the glacial theory was born, the nature of glaciation in Nova Scotia was being debated. Did former glaciers arise from local areas, originating in uplands and confined to the province, or was the ice part of a great continental mass which crossed the Bay of Fundy? The Reverend D. Honeyman, curator of the Provincial museum in the late 1800's, noted the long distance transport of amygdaloidal basalt boulders found along the Atlantic coast near Halifax and used the observation to support the concept of a continental-based ice movement that crossed the Bay of Fundy.

Robert Chalmers (1895) carefully mapped glacial grooves and striations in Nova Scotia and proposed that northern Nova Scotia had been glaciated largely by local glaciers with floating ice a secondary agent in low-lying areas. In contrast to Honeyman, he did not believe that a continental glacier had crossed the Bay of Fundy. Chalmers (1895, p. 95m) stated:

The depression of the Bay of Fundy was not crossed by land ice from southern New Brunswick . . . Neither has Nova Scotia been glaciated by extra-peninsular ice from the north or northeast.
L. W. Bailey (1898) and W. H. Prest (1896), working in mainland Nova Scotia, observed erratics that supported both previous views. Bailey stated the compromise position (1898, p. 26m):

As in other parts of southwestern Nova Scotia the facts connected with the glaciation of Digby Neck are, in the opinion of the writer, best explained upon the supposition of submergence beneath a continental glacier moving southward and bringing debris even from the other side of the Bay of Fundy, followed by a period of more local and restricted distribution, when the higher portions of the peninsula became themselves the centre of the movement, the latter now occurring in all directions.

J. W. Goldthwait (1924), in his treatise "The Physiography of Nova Scotia" dismissed all evidence regarding local glaciers in Nova Scotia. He envisioned a major ice mass moving southeastward, stemming from a Labrador-Laurentide source, and a subsequent southward-directed ice movement stemming from ice in the Gulf of St. Lawrence called the "Acadian Bay Lobe".

Pleistocene mapping in the Annapolis Valley initiated at Acadia University, Wolfville (MacNeill, 1951; Purdy, 1951; Swayne, 1952) revived the concepts of local glaciers (MacNeill and Purdy, 1951). Synchronously, Flint (1951) advanced the idea of local highland centres of outflow based largely on the previous work of Chalmers (1895). Hickox (1962) confirmed that granitic erratics on the North Mountain were derived from the South Mountain Batholith in Nova Scotia and not from New Brunswick as Goldthwait (1924) had previously suggested.
Regional analyses of air photographs prompted V. K. Prest and Grant (1969) to further develop the local glaciation model of eastern Canada. They defined several local centres of ice flow, termed the Appalachian Ice Complex that were sufficiently large to hold off the encroaching Laurentide ice sheet.

King (1969) concluded that the Scotian Shelf End Moraine Complex was the terminus of a Late Wisconsinan continental ice advance. Grant (1975) proposed a southward flood of ice prior to 39,000 years B.P. followed by a retreat about 38,000 years ago and re-expansion of Nova Scotia glaciers during the Late Wisconsinan. Grant (1977) and Grant and King (1984) further developed the hypothesis that Late Wisconsinan ice of Labradorean (Laurentide) origin never crossed the Bay of Fundy. This became known as the "minimum model" (Fig. 3). Grant (1977, p. 247) stated:

"Evidence from scattered stratigraphic sections, from the relationship of a sequence of ice flow indicators to a raised interglacial marine platform, together with the limits of freshly glaciated terrain against weathered bedrock areas, indicates that late Wisconsinan glaciers spread weakly toward, and in many areas not beyond, the present coast. These were fed by a complex of small ice caps located on broad lowlands and uplands. The limiting factor was the deep submarine channels that transect the region".

Thus in Grant's view, Laurentide ice was limited to the northern Gulf of St. Lawrence. Today most continent-wide interpretations of the Late Wisconsinan glaciation show a relatively simple, monolithic ice sheet centred in Hudson's Bay and extending to the shelf
Figure 3. Summary of previous Late Wisconsinan glaciation models (after Stea et al., 1987 and King et al., 1972) and the location of major morainal systems on land and offshore.
edge off Nova Scotia (Flint, 1971; Denton and Hughes, 1981; Budd and Smith, 1987). This is termed the "maximum" model and is believed to represent the equilibrium condition. The contrasting "minimum" model is one with local ice centres and thinner ice, extending only slightly off the coast of Nova Scotia and with some highland areas bare of ice (munataks) (Dyke and Prest, 1987). These two extreme models are not in accord with recent mapping of glacier erosional and depositional landforms in New Brunswick and Nova Scotia (Wightman, 1980; Rampton et al., 1984; Rappol, 1989; Foisy and Prichonnet, 1991; Stea and Finck, 1984; Stea et al., 1986; 1989). These authors envision a time transgressive, ice flow phase evolution of ice caps in the region. Late Wisconsinan ice caps covered all highland areas and extended to the continental shelf edge (Mosher et al., 1989; Gipp and Piper, 1990). The complex, but extensive, glaciation model is essentially a compromise between the earlier models and it represents the current state of the debate. Stea (1982; 1984), Stea and Finck (1984) and Stea et al. (1987; 1992) developed a 4 ice flow phase model of terrestrial glacier flow events in Nova Scotia (Fig. 4).

1.3.1 Ice Flow Phase Model

Complex glacier flow patterns have been imprinted on the landscape of Nova Scotia (Stea, 1984; Stea, 1994). Many outcrops reveal three or four crosscutting sets of striae and bevelled facets that can be traced regionally. Boulders are dispersed east, west, north and south of their bedrock sources in Nova Scotia. Sections can reveal four or more stacked till sheets of differing provenance. From these facts it is hard to conclude that one ice flow or glaciation (through topographic streaming) is responsible. Stea et al. (1984) developed an evolutionary model of flow events defined by discrete, regionally mappable
Figure 4. Ice flow patterns deduced from striae and drumlin directional data. Ice Flow Phases 1a and 1b from Appalachian or Laurentide source areas, flowed eastward and southeastward across the Bay of Fundy and onto the Continental Shelf. Ice Flow Phase 2 stems from the Escuminac Ice Centre or Divide. Ice Flow Phase 3 represents formation of an ice divide (Scotian Ice Divide) across the mainland of Nova Scotia. During Ice Flow Phase 4 remnant ice caps (Chedabucto Bay Glacier, Chignecto Glacier, South Mountain Ice Cap) flowed into marine basins adjacent to the province.
ICE FLOW PHASE IA

ICE FLOW PHASE IB

ICE FLOW PHASE II

ICE FLOW PHASE II

ICE FLOW PHASE III

ICE FLOW PHASE IV

LEGEND

ICE FLOW

ICE FLOW (LATER)

CHIGNECTO GLACIER

ANTIGONISH HIGHLANDS/

CHEDABUCTO BAY GLACIER

SOUTH MOUNTAIN ICE CAP

Figure 4.
ice-flow trends termed "erosional stratigraphy" (W.W. Shilts, personal communication, 1986). Each event is called an "Ice Flow Phase”. Flow lines relating to each phase can be traced over broad regions of Nova Scotia using striae and drumlin orientations (Stea and Finck, 1984). Crosscutting relations and the preservation of older, weathered striations on lee-side surfaces are used in the development of an "erosional stratigraphy". The patterns of ice flow mapped by striae are verified by till fabric, dispersal studies and the orientation of glacial landforms such as eskers and drumlins (Stea, 1984; Stea et al., 1989). The sequence of ice-flow phases has been discerned from superimposed striae sites, and through correlation with stacked till sheets of known provenance (Stea, 1984). Each of the ice-flow phases produced at least one recognizable till sheet, sometimes with basal and englacial facies.

Striae patterns, distinctive erratics, till fabric, and striated boulder pavements suggest that the earliest and most extensive ice-flows in Nova Scotia were eastward and southeastward (Fig. 4). These have been designated Ice Flow Phases 1a and 1b. Several widely-spaced striae sites reveal evidence of a distinct eastward flow, later superseded by a southeastward flow. The eastward ice-flow may represent a separate, older phase of glaciation. Erratic trains of igneous rocks from the Cobequid Highlands and trains of basaltic rocks from the North Mountain are oriented southeastward and can be traced to the Atlantic Coast, up to 120 km down-ice (Nielsen, 1976). This phase may represent Laurentide ice-flow but more probably is related to a centre in New Brunswick (Rampton et al., 1984). Anorthosite erratics in western Prince Edward Island and the Gulf of St. Lawrence, derived from the Canadian Shield, suggest that Laurentide ice did cross the region at some time (Loring and Nota, 1969; Prest and Nielsen, 1987, Stea, 1991).
The second major ice-flow trend (Ice Flow Phase 2) was southward and southwestward from the Escuminac Ice Centre in the Prince Edward Island region (Prest, et al., 1972; Rampton et al., 1984; Fig. 4). This flow phase was called the "Acadian Bay Lobe" by Goldthwait (1924) and the "Fundian Glacier" by Shepard (1930). Goldthwait (1924) envisioned southward flow from a Laurentide source across the Gulf of St. Lawrence. Ice-flow trends in Prince Edward Island (Prest, 1973) and adjacent New Brunswick (Rampton et al. 1984) do not reflect a pervasive southward flow, but suggest radial flow from a local centre. This event is recorded by southward striae crossing earlier southwestward-trending striae at many localities on the upland regions of Nova Scotia and New Brunswick. Carboniferous red beds from northern mainland Nova Scotia and Prince Edward Island region were eroded and transported southward onto the metamorphic and igneous Cobequid and Meguma Terranes of mainland Nova Scotia along with distinctive Cobequid Highland erratics (Grant. 1963; Stea et al., 1989). Schnikter (1987) reported evidence of southward dispersal of red clastic material from the Bay of Fundy into the Gulf of Maine.

During the next event (Ice Flow Phase 3), ice flow was northward and southward from a divide (Scotian Ice Divide) across the axis of the Nova Scotia peninsula. Granitic debris from the South Mountain Batholith were transported northward onto the North Mountain basalt cuesta (Fig. 4; MacNeill, 1951; Hickox, 1962). Erratics from the Cobequid Highlands can be found dispersed throughout the Carboniferous lowlands to the north (Stea and Finck, 1984). Northward-trending striae can be traced across the northern mainland of Nova Scotia (Fig. 4). This well-documented northward ice-flow was a response to the development of an ice divide in southern Nova Scotia. This divide may have formed as a result of marine incursion into the Bay of Fundy (Prest and Grant, 1969) or increased
precipitation after recession of the Escuminac Ice Cap (MacNeill and Purdy, 1951; Hickox, 1962). Northward ice-flow from the Scotian Ice Divide was probably synchronous with the ice dome off Cape Breton Island proposed by Grant (1977). The divide can be traced from areas south of the Antigonish Highlands (Myers and Stea, 1986) offshore into Chedabucto Bay. Here, ice flow was funnelled through the submarine channels out of St. Georges Bay and the Cape Breton Channel towards the Gulf of St. Lawrence. Northward ice flow was less intense in the western Cobequid Highlands. As marine drawdown began to take effect, the ice may have been increasingly directed into the Bay of Fundy to merge with southwestward ice streams from New Brunswick.

During this final phase (Ice Flow Phase 4), remnant ice caps developed from the Scotian Ice Divide (Fig. 4). Eskers and striations cut across features formed by earlier ice flows. Ice caps or glaciers that formed over the Chignecto Peninsula and southern Nova Scotia had recessional margins on land, marked by hummocky moraine, glaciofluvial deposits, and the gradual pinch-out of till sheets. Ice flow during this last phase was funnelled into the Bay of Fundy and into the Atlantic Ocean. Erosional features and deposits relating to these late-glacial ice caps are restricted to low-lying areas.

1.3.2 Sea Level

Pronounced sea-level changes accompanied the last glacial cycle in Nova Scotia. Relict shorelines can be found above and below present sea level (Fig. 5). On land, a wave-cut platform 4-6 m above sea level was formed during the last interglacial interval (Grant, 1980, 1989; Stea et al., 1987; 1992d). This shoreline is a product of climatically-induced sea-level rise and isostatic equilibration from former ice loads. Raised marine strandlines,
Figure 5. Plot of elevation isolines of raised marine features and submerged shorelines in the maritime provinces and Maine (data from Welsted, 1971; Fader, 1977, 1989; Grant, 1980; Scott and Medioli, 1980; Kelley et al., 1986; Rampton et al., 1984; Stea, 1983; Stea et al., 1994). Modified after Stea et al. (1987).
formed by crustal rebound after the last glacial maximum, are found in the Bay of Fundy region of Nova Scotia. The complex variability of elevations, tilts and ages of these post-glacial strandlines has been interpreted as a response to the rate of ice retreat and the effects of local and regional ice caps (Wightman and Cooke, Quinlan and Beaumont, 1981; 1982; Scott et al., 1987b; Stea et al., 1987; Grant, 1989).

Evidence for sea-level rise has been found offshore of Nova Scotia. On the outer continental shelf, terraces and erosional features cut into bedrock and glacial deposits, 110-120 m below present mean sea level (BSL) have been interpreted as lowstand shorelines (Amos and Knoll, 1987; Amos and Miller, 1990; King, 1970; King and Fader, 1986; Fader, 1989a; Piper et al., 1990). McLaren (1988), however, reported maximum sea-level lowering of only 75 m (75 m BSL) in the Sable Island area, based on deltaic clinoforms and an associated unconformity. Scott et al. (1989) defined a continuous sea-level curve from 11 ka to the present based on dating of several salt and freshwater peat horizons. From this curve they extrapolated sea-level lowering to 78 m BSL at 15 ka. Summarizing much of the previous data, they emphasized the complexity of the sea-level history and stressed that sea-level histories and glaciation models are inextricably linked.

Contradictory evidence has been submitted regarding the sea-level history of the inner shelf, a portion of the continental shelf extending offshore from the present coastline to water depths of 150 m (Fig. 5). Fader (1989) proposed a shelf-wide low stand of sea-level at -110 to -120 m BSL based on terraces, unconformities and cross-shelf textural variations. King (1970) mapped the surficial geology of the Halifax to Sable Island map area using textural information from samples and morphological and acoustic characteristics from echograms. He defined the boundary between two surficial
formations - the Sambro Sand and the Sable Island Sand and Gravel, as representing a low sea-level stand. The Sambro Sand was described as a sublittoral muddy sand and the Sable Island Sand and Gravel as a clean well-sorted basal transgressive deposit formed through the reworking of glacial materials during the late Wisconsinan-Holocene marine transgression. Surrounding the offshore banks, this boundary consistently occurred at approximately 110-120 BSL. On the inner shelf, however, the depth of this transition varied, and in the area off Halifax, the boundary was mapped at 90 m BSL. Wang and Piper (1982), and Piper and Fehr (1991) proposed that a change from rolling to planar physiography at -50 to -60 m BSL indicates a lowstand and subsequent transgression. LaPierre (1985) mapped a sand body at -65 m which he interpreted as a relict beach. Forbes et al. (1991) found estuarine back-barrier deposits on the inner shelf dated at 10 ka, implying that sea-level fell to at least -50 m. They also defined coast-parallel depositional and erosional landform "zones". Stea et al. (1992b, c; 1994) reported evidence for a -65 to -70 m lowstand on the inner shelf off Nova Scotia, dated at 11.6 ka.

1.3.3 The Thesis Problem(s)

The main problems to be addressed in this dissertation are the glacial history of the inner shelf, the extent and direction of various ice flow phases on the inner shelf and sea levels associated with these glaciation events. Correlation of onshore-offshore glacial sediments and landforms is essential to accomplish this task.
CHAPTER 2: METHODS OF STUDY

To correlate onshore and offshore sedimentary units and landforms the author has employed three basic methods:

1. Mapping of surficial sedimentary units and landforms
2. Definition of the stratigraphy of surficial units.
3. Determination of sediment properties: clast lithology, shape and surface features, grain size.

Two governing paradigms are used:

1. Much of the land stratigraphy is mirrored in the inner shelf, especially in areas not affected by the transgression.
2. The glacial depositional record increases in complexity towards the margin of ice sheets (Sugden and John, 1976).

2.1 MAPPING AND STRATIGRAPHIC CONCEPTS

The marine geologist relies primarily on geophysical techniques and point sampling for basic mapping whereas the land geologist relies on field investigation, aerial photography, and more recently, satellite imagery. It is easier to conduct seismic surveys at sea than on land because water is an excellent conductor of seismic signals. Transverse (shear) waves cannot be transmitted through water so they do not produce interference effects as they do on land (Savit, 1983). A continuous cross-section of the seafloor terrane under the ship is produced because the detector arrays are in continuous motion. Sampling Quaternary
units in the marine realm, however, is difficult and costly. The advantage of land mapping is in the ease of sampling and identifying the Quaternary units in naturally exposed sections and by relatively inexpensive drilling. Another advantage of land surveys is the ability to get a "birds" eye (i.e. aerial photography) view of the terrain. This larger view is becoming available in the marine realm with the advent of digital multibeam sounding (Loncarevic et al., 1992; Costello et al., 1993). The Nova Scotia coast provides an unique opportunity to utilize the advantages of both the terrestrial and marine realms to solve the problems outlined earlier.

Geological maps portray orders of organization in the landscape derived from a remote aerial view (air photo-landsat imagery--multibeam-seismics) to a detailed one (stratigraphic and acoustic sections and samples). The terminology in the terrestrial realm is taken from the glacial-morphologic classification of Goldthwait (1989) utilized in the Surficial Map of the Province of Nova Scotia (Stea et al., 1992a). Mapping essentially consists of identification, classification and interpretation of landform assemblages or terrain "zones". The seafloor topography and subbottom stratigraphy are a result of glacial, periglacial and marine processes acting over time. Distinct landscapes are produced by each erosional and depositional event. Land-sea correlation is accomplished by topographic and stratigraphic comparison of marine and terrestrial landforms. Forbes et al. (1991) and Stea et al. (1993; 1994) defined offshore terrain "zones" as distinct terrain types based on unique sea bottom morphology (landforms) and seismic sequences and facies (stratigraphy).

A major portion of this thesis concerns stratigraphic analysis of the offshore seismic reflection profiles and stratigraphic sections on land. Stratigraphy is obtained in the
offshore by the generation of seismic waves from towed or hull-mounted acoustic sources. These emit regular pulses which are transmitted through the water column, reflected by the seafloor and the underlying geology, then intercepted by hydrophones which convert the acoustic pressure waves to electrical signals. These signals are processed and displayed on an onboard recorder which burns a small mark on paper, then advances while keying another seismic signal. These marks vary in intensity depending on the strength of the returning signal. The reflection of an acoustic signal at the air-water, water-sediment or sediment-sediment interface is a result of the acoustic impedance differences between these mediums. Acoustic impedance is an intrinsic property of all substances and is defined as the product of the compressional wave velocity and the bulk density of the material (Sylwester, 1983). The amplitude or strength of the seismic signal depends on the ratio of acoustic impedances (reflection coefficient) of the media on either side of a reflecting interface. The air-water interface, for example, is nearly a perfect reflector, having a reflectivity coefficient equal to -1.

In essence, the seismic reflection record is an acoustic cross-section of the geology (Sylwester, 1983). The main sources of acoustic reflections in the geological section are stratal surfaces and unconformities. Stratal surfaces are defined as bedding contacts or relict depositional surfaces (Brown and Fischer, 1982). Unconformities are surfaces of erosion and nondeposition which represent gaps in the geological record. Stratigraphic analysis is based on the method of Vail et al. (1977), termed seismic sequence analysis. This analysis is based on the recognition of unconformity-bound packages of strata. Boundaries between seismic sequences are recognized on seismic profiles by terminations of reflections. The boundaries are either erosional, lapout (hiatal) or conformable. Within
each sequence, seismic "facies" can be defined (Mitchum et al., 1977; Belknap and Shipp, 1991). The facies are based on the continuity, amplitude and frequency of the reflections. These acoustic properties are an indication of the lithic properties of the strata. Acoustic reflectivity, the degree of coherence of the incoming reflected pressure pulse, is also a useful tool in determining lithic properties of the seabed (Parrot et al., 1980). Seismic sequence and facies analysis permit the interpretation of the depositional history of the region including sea-level changes and changes in depositional styles during ice advance and retreat. King and Fader (1986) used seismic and lithofacies analysis to differentiate acoustic units on the Scotian Shelf. Boyd et al. (1988) were the first to utilize the sequence analysis approach on the Scotian Shelf in stratigraphic analyses of Sable Island Bank. Gipp (1989, 1994) defined seismic sequences within Emerald Basin.

King and Fader (1986) subdivided marine basin-fill sediments on the Scotian Shelf into three formations: the Scotian Shelf Drift, Emerald Silt and LaHave Clay. These terms are in wide use (cf. Piper et al., 1990) and will serve as a basis for further stratigraphic subdivision in this study. A formation is defined as a mappable body of sedimentary strata which is distinguished and delimited on the basis of lithic characteristics and stratigraphic position (North American Stratigraphic Code, 1983). King and Fader (1986) and Fader (in Williams et al., 1985) used a combination of lithic and acoustic attributes to define Formations and facies on the Scotian Shelf. The oldest formation is the Scotian Shelf Drift overlain by the Emerald Silt and LaHave Clay. These formations are described in Table 1. King and Fader (1986) never stipulated a single type acoustic section or cores for each of the surficial formations but described several piston-cored reference acoustic sections. The Emerald Silt is subdivided into three facies (A, B, and C) based on lithostratigraphy and
<table>
<thead>
<tr>
<th>Formation</th>
<th>Facies</th>
<th>Lithic Properties</th>
<th>Seismic Character</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sable Island Sand</td>
<td></td>
<td>Fine to coarse, well-sorted sand grading to subrounded to rounded gravels.</td>
<td>Highly reflective seabed, closely spaced, continuous reflections.</td>
</tr>
<tr>
<td>and Gravel</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sambro Sand</td>
<td></td>
<td>Silty-sand grading to gravelly-sand.</td>
<td>Similar to above.</td>
</tr>
<tr>
<td>LaHave Clay</td>
<td>LaHave Clay</td>
<td>Greyish-brown, soft, silty-clay grading to clayey-silt, confined mainly to basins and depressions in the shelf.</td>
<td>Generally transparent without reflections, some weak, continuous, coherent reflections at the base of the section.</td>
</tr>
<tr>
<td>Emerald Silt</td>
<td>Facies C</td>
<td></td>
<td>Discontinuous, coherent reflections, and incoherent zones.</td>
</tr>
<tr>
<td>Emerald Silt</td>
<td>Facies B</td>
<td>Dark greyish-brown, poorly sorted, clayey and sandy-silt, some gravel, poorly developed rhythmic banding.</td>
<td>Medium to low amplitude, continuous, coherent reflections; ponded depositional style.</td>
</tr>
<tr>
<td>Emerald Silt</td>
<td>Facies A</td>
<td>Greyish-brown, poorly sorted, clayey and sandy silt, some gravel, well developed rhythmic banding.</td>
<td>High amplitude, continuous, coherent reflections conformable to substrate irregularities.</td>
</tr>
<tr>
<td>Scotian Shelf Drift</td>
<td></td>
<td>Sandy-clay, matrix-supported, diamicton.</td>
<td>Incoherent reflections, sometimes with scattered, point source reflections.</td>
</tr>
</tbody>
</table>

Table 1. Quaternary formations and seismic facies defined by King and Fader (1986) and their properties. ¹ These reflections are now considered to be part of a separate formation (Yankee Bank Formation) defined later in this dissertation.
seismic stratigraphy (Table 1). A facies is described as a body of rock characterized by a particular combination of lithology, physical and biological structures that bestow an aspect different from the bodies of rock above, below or laterally adjacent (Walker, 1992). In this study I will restrict the use of Emerald Silt facies terms defined by King and Fader (1986) to seismic facies (cf. Mitchum et al., 1977). This frees these facies terms from time constraints. Unfortunately, the LaHave Clay is both a formational and facies term. If referring to the formation the term "LaHave Clay Formation" will be used. When referring to the seismic facies, "LaHave Clay facies" will be used.

2.2 SEDIMENT PROPERTIES

Stratigraphic sections along the coast of Nova Scotia (Stea and Fowler, 1979) provide clues to the makeup of moraines in the offshore. Offshore glacial landforms can only be sampled on their uppermost surfaces while terrestrial counterparts, exposed in sea cliff sections, can be sampled in profile from top to bottom. Lithic data obtained from the cliff sections are used to:

1. Correlate with glacial deposits in the offshore.
2. Assess the genesis and provenance of the sedimentary units exposed in sea cliff sections of onshore glacial landforms.
3. Verify previous correlations with terrestrial regional ice flow patterns determined through landform analysis.
The Maritime provinces are characterized by the juxtaposition of tectonic terranes of unique geology. Changing centres of outflow in Maritime Canada during the Wisconsinan produced compositionally and texturally distinct till facies (Prest, 1896; Grant, 1963; Nielsen, 1976; Stea and Fowler, 1979; Graves and Finck, 1988; Stea et al., 1989). These lithic variations are largely due to erosion of distinctive rock formations as ice centres shifted over the varied geological terranes of the region. The marked contrast between Nova Scotia till facies enhances the potential for correlation with the offshore sections. Piper et al. (1986) used lithology and stratigraphy to correlate tills on the inner shelf with terrestrial till units.

In the past small-volume, lightweight grab samplers (e.g. van Veen) have been used to sample inner shelf surficial sediments and as a basis for the interpretation of seabed and subbottom acoustic stratigraphy (cf. Fader et al., 1977). These samples may not be representative of the underlying acoustic stratigraphy because of a widespread transgressive gravel-lag armor in areas above the lowstand (Fader, 1989a) and current winnowing and iceberg furrowing of the till surface on morainal landforms in deeper water (King and Fader, 1986, p. 31). In this study a heavyweight (1/2 ton) grab sampler (IKU-Appendix 2) was used to provide data on the makeup of seabed acoustic units. It penetrated the seabed to depths up to 1 metre, and obtained a relatively large, undisturbed sample in muddy sediments (Appendixes 3,4; Stea et al., 1993).

Once the onshore-offshore sediment link is established, former glaciers can be modelled from terrestrial flow lines and offshore margins and put into a temporal context. Radiocarbon dating and correlation of glacial marine sediments with moraines in the
offshore basins (King and Fader, 1988a; Gipp and Piper, 1990) has made this possible. The interaction between sea-level change and glacial dynamics can be assessed by landform and stratigraphic analysis of the inner shelf.

2.3 MAPPING METHODS-MARINE GEOPHYSICAL DATA ACQUISITION

See Appendix 1

2.3.1 Marine Data Analysis

The ocean floor was mapped by systematically describing all record types along each seismic trackline. In addition to visual descriptions of landforms and seismic stratigraphy, the areal extent, seabed height and thickness of each stratigraphic unit and landform were measured. These units and landforms were later grouped into a hierarchical classification of terrain zones and subzones, formations and facies. The thickness of basinal stratigraphic units were measured every 5 minutes along tracklines.

2.4 TERRESTRIAL MAPPING METHODS

The surficial geological data used in this dissertation were compiled from the Surficial Map of the Province of Nova Scotia (Stea et al., 1992a). Field compass and altimeter traverses were conducted on specific landforms for more detailed information on surface profiles. Several reference Quaternary sections along the Eastern Shore have been selected for detailed stratigraphic and till-pebble lithological analysis. The sections were traversed
by compass and pace and measured by altimeter and tape. The altimeter surveys were run in loops with a high tide marker as datum. Photographic logs were obtained from small boats. Individual sedimentary units were visually described and logged according to the lithofacies classification scheme of Eyles et al. (1983).

2.5 MARINE AND TERRESTRIAL SAMPLING METHODS AND SAMPLE PREPARATION

See Appendix 2.

2.6 SAMPLE ANALYSIS (MARINE AND TERRESTRIAL)

2.6.1 Pebble Lithology

Grant (1963) and Nielsen (1976) conducted regional studies of pebble lithology in Nova Scotia. Grant (1963) traced 50 rock species to seven source regions in Nova Scotia whereas Nielsen (1976) limited his study to basalt and granitic rocks. Both studies utilized the 5-22 mm fraction of the till sample. Many different size ranges have been used for the purpose of quantitative lithological evaluation of till (cf. Krumbein, 1933; Holmes, 1952; Dreimanis and Reavely, 1953; Marcussen, 1973; Fenton and Dreimanis, 1976; Teller and Fenton, 1980). In this study the entire pebble (4-64 mm) size range was utilized. Important questions about pebble analysis are:

1. Do the ratios of different lithologies change with grain size?
2. Are the number of clasts in the grade size sufficient for statistical analyses?

3. Can the worker identify the lithologies using that grade size, and are certain lithologies comminuted into constituent minerals: e.g., granite?

Dreimanis and Reavely (1953) found no significant change in lithology ratios from one grade to another using the 4.699-6.66 mm, 6.680-9.423 mm, 9.423-13.33 mm, and the 13.33-80 mm grade sizes. Bahnson (1972) concluded that the ratios of different rock groups remained relatively constant in fractions larger than 1.44 mm, but changed radically in the fractions smaller than 1.44 mm. In the latter fractions, igneous and metamorphic lithologies are under-represented, because identifiable rock species have been comminuted to mono-mineralic fragments. The accuracy of provenance determination depends on the identification of the pebbles, so utilization of the larger pebble sizes is necessary. In view of these considerations, the 4-64 mm fraction was chosen as a compromise to obtain a large number of clasts in an identifiable size range, utilizing a manageable 2-4 kg sample.

In this study more than 300 pebbles per sample were counted and divided into three main categories:

1. Metasediments (Meguma Group)
2. Granitoids (South Mountain Batholith (SMB))
3. Foreign (not Meguma Group or SMB granites)
In most samples the "foreign" or non-Meguma components comprise less than 10% of the total amount. The significance of statistical comparisons of point counts less than 10% from samples of 300 is questionable (Patterson and Fishbein, 1989). Because of this, various "foreign" components were ranked in order of abundance, rather than individually counted. To enhance rock identification, selected exotic pebble species from marine and terrestrial samples were thin-sectioned and petrographically analyzed using standard techniques.

The lithologic assemblage of a till depends on the intrinsic factors of till formation, the nature of the substrate over which the glacier passes, and inheritance/overprinting processes inherent in a multi-glaciated region (Finck and Stea, 1990). Lodgement tills are deposited by plastering of glacial debris from the sliding base of a moving glacier by pressure melting or other mechanical processes (Dreimanis, 1989). This process tends to form tills dominated by local lithologies, whereas tills formed through a basal melting or ablation process can contain very high percentages of exotic clasts (Dreimanis, 1989). At the base of a glacier, mutual clast attrition is high and lithologies such as shales are rapidly comminuted (Boulton, 1975). Lodgement tills produced from this material generally show little down-ice dispersal. As material is shunted englacially, the amount of clast to clast attrition decreases and transport potential is increased. Boulton (1978) referred to this process as "mechanically passive transport". Debris can be moved up into an englacial position through these processes:

1. Plucking from rock protuberances high in the ice (Shilts, 1976).
2. Up-glacier shearing as ice passes over obstructions and regelation ice forms underneath (Boulton, 1975).
3. As the ice flow becomes compressive, shear planes are created, guiding the dirty ice above the substrate (Dreimanis, 1976).

In any section of tills deposited by a single ice sheet, material from the base, middle and upper parts of the ice sheet (basal, englacial, and superglacial) can be deposited as a sequence, as the temperature regime of the glacier changes. This ablation-lodgement couple is almost universally present in all sections (Shilts, 1976).

2.6.2 Till Provenance

Ice flow trajectories are determined by linking source regions to sample sites (cf. Meyer, 1983). Foreign lithologies at any site along the Eastern Shore study region can generally be traced to one or more of these source regions.

1. Antigonish Highlands
2. Cobequid Highlands
3. North Mountain

Table 2 summarizes the main lithologies and relative abundances of indicator erratics from these source areas. The sampling density and distribution in the marine and terrestrial environs are sufficient to intersect dispersal "trains" emanating from these large source areas (Fig. 6). In evaluation of provenance, the assemblage of erratic rock types are sometimes more valuable than identification of individual species. The possible glacier flow paths are reduced when more than one indicator rock species is considered. The
<table>
<thead>
<tr>
<th>COBEQUID HIGHLAND LITHOLOGICAL SUITE (in order of total outcrop area.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Granitoid rocks (≈40%)</td>
</tr>
<tr>
<td>(a) red, orange, tan and grey hornblende-biotite monzogranite to alkali-feldspar granite.</td>
</tr>
<tr>
<td>(b) dark grey hornblende-pyroxene diorite</td>
</tr>
<tr>
<td>(c) red and grey to pale orange syenogranite</td>
</tr>
<tr>
<td>(d) sheared (cataclastically-deformed) granitoids</td>
</tr>
<tr>
<td>2. Volcanic rocks (≈35%)</td>
</tr>
<tr>
<td>(a) tan to orange rhyolites, andesites and dacites (tuffs, ignimbrites and agglomerates)</td>
</tr>
<tr>
<td>(b) black to red-brown basalts</td>
</tr>
<tr>
<td>3. Sedimentary and metasedimentary rocks (≈25%)</td>
</tr>
<tr>
<td>(a) siltstone, shale, mudstones, conglomerates, sandstones</td>
</tr>
<tr>
<td>(b) quartzite, schist, gneisses, marble</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>ANTIGONISH HIGHLAND LITHOLOGICAL SUITE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Volcanic rocks (≈45%)</td>
</tr>
<tr>
<td>(a) rhyolites, dacites, andesites (tuffs and ignimbrites)</td>
</tr>
<tr>
<td>(b) basalt</td>
</tr>
<tr>
<td>2. Sedimentary and metasedimentary rocks (≈45%)</td>
</tr>
<tr>
<td>(a) red arkosic sandstone, conglomerate</td>
</tr>
<tr>
<td>(b) blue-grey shale, siltstone</td>
</tr>
<tr>
<td>(c) red siltstone, shale</td>
</tr>
<tr>
<td>(d) fossiliferous, sandstone; green and black, laminated siltstone</td>
</tr>
<tr>
<td>3. Granitoid rocks (≈10%)</td>
</tr>
<tr>
<td>(a) grey granodiorite</td>
</tr>
<tr>
<td>(b) red to pink monzogranites and syenogranites (some altered)</td>
</tr>
<tr>
<td>(c) diorite, appinnite</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>NORTH MOUNTAIN LITHOLOGICAL SUITE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Massive and amygdaloidal (zeolitic) basalt (100%)</td>
</tr>
</tbody>
</table>

Table 2. Lithologies of the major highland source areas in northern mainland Nova Scotia.
Figure 6. Location of the terrestrial and marine samples in study region. Till type classifications are given for all land samples. Sediment classifications are shown for the offshore samples. Note the preponderance of gravel inside the 50 m contour, whereas diamicton and muds are largely found in deeper water (see Appendixes 3-7 for sample data).
presence of diagnostic minerals in the samples can assist in the delineation of provenance. For example, muscovite is not a common accessory in the granitic rocks of the Cobequid Highlands but is an important secondary mineral in the rocks of the South Mountain Batholith (Clarke et al., 1980). Hornblende is not known to occur in the granitic rocks of the South Mountain batholith (MacKenzie and Clarke, 1975), but is common in Cobequid granites (Donohoe and Wallace, 1985).

### 2.6.3 Clast Morphology and Surface Features

Grain shape was qualitatively assessed using Powers (1953) visual scale augmented by identification of characteristic shape types (i.e. glacial flat-iron-"bullet" clasts; discoid beach pebbles) and surface markings, such as striations. The percentages of striated clasts were evaluated as an index to the degree of reworking by wave-erosion and the genesis of the deposit (Appendix 4).

Shape analysis is a powerful tool in discrimination of glacial facies and sedimentary environments (Mazzullo and Anderson, 1987). Glacial depositional environments can be inferred from pebble shapes (Boulton, 1978; Dreimanis, 1989).

### 2.6.4 Texture

Karrow (1976) maintained that fine detail in measurement is probably meaningless as local variations in till texture can be great. Till units in Nova Scotia can be distinguished based on their gravel/sand/mud ratios (Nielsen, 1976). Stea (1982; 1984) correlated till units
in central Nova Scotia based on their gravel/sand/mud ratios. Stea et al. (1989) divided these till units into older, matrix-dominated till units formed near the margin of extensive ice sheets and ice caps and clast-dominated, younger tills formed somewhere near a former ice divide.

For the terrestrial and marine samples (Appendix 7) the author uses a simple clast/matrix ratio, which is the weight ratio of the <64 mm-/2 mm/<2 mm fractions. Grain size data from Nielsen (1976) are also utilized for the correlation of till units. Gravel/sand/mud weight percentages were calculated for samples from core 91018-53 (analyses courtesy of the Atlantic Geoscience Centre). A detailed grain size analysis, using 1/2 phi intervals, was done on selected till samples and marine sediments.

The texture of till depends not only on the energy of the glacier transporting agent, but on the provenance of the till itself (Shilts, 1978). As a glacier passes over unconsolidated sediments, these are incorporated into the till and impart a distinctive signature to the grain size distribution. Joint spacing, weathering and degree of induration, are all parameters which contribute to the susceptibility of rocks to erosion, and subsequently to the texture of the resultant till. The glacier base or "glacial mill" can produce fines by abrasion and crushing. Terminal grades in the silt sizes (Dreimanis and Vagners, 1971) are produced. In many regions, especially at the margins of the former great ice sheets, the fines are inherited from older till material. Meltwater may remove considerable volumes of the comminution debris (Drewry, 1986; Eyles and Eyles, 1992). Material in englacial transport is suspended above the traction zone and undergoes little comminution (Boulton, 1978).
2.6.4.1 Marine Tills

Sampling difficulties in the marine environment make genetic interpretation hazardous. Syvitski (1991) cautioned that on glaciated shelves, seismic facies characterized by incoherent reflections and high reflectivity, generally interpreted as till, may encompass a variety of sediment types, including well-sorted gravels. King (1993) compiled much of the pertinent data on marine till and states (p. 349) "The majority of studies ... interpret the diamictons as till a trend that possibly reflects Alley et al. (1989) (till-deformation) hypothesis that provided a plausible mechanism for subglacial deposition under a marine ice sheet". King (1993) states that it may be impossible to differentiate between primary and secondary till. Hart and Boulton (1991) have extended the concept of deformation till in the marine and terrestrial environment to include most till. They argue that deformed till is a common phenomenon. In the marine environment, clay-rich proglacial glacial marine sediments can be transformed into till by rapidly readvancing ice (Alley et al., 1989; King, 1993), however, subglacial deformation may be less effective in coarser-grained sediments (Hughes in Denton and Hughes, 1981). Grain size analysis of large volume grab samples in this study will help to resolve the problems of till genesis in the marine environment.

2.6.5 Age Dating

Marine shell fragments and foraminifera were radiocarbon-dated using Accelerator Mass Spectrometry and regular beta disintegration methods (Table 3). Foraminifera were hand picked from seived (<64 µm) separates for age dating. All ages are calculated with
<table>
<thead>
<tr>
<th>14C Age date</th>
<th>Lab- Number</th>
<th>Sample No. and Type</th>
<th>Material</th>
<th>Water depth/ core depth (m)</th>
<th>Landform/ Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>3970 +/- 60 AMS</td>
<td>Beta-61398</td>
<td>Pcore 91018-53</td>
<td>shell</td>
<td>168.924</td>
<td>Emerald Silt</td>
</tr>
<tr>
<td>5015 +/- 60 AMS</td>
<td>Beta-50594</td>
<td>IKU 91018-51</td>
<td>shell</td>
<td>165.5</td>
<td>Scotian Shelf Dr.</td>
</tr>
<tr>
<td>5410 +/- 60 AMS</td>
<td>Beta-50593</td>
<td>IKU 91018-50</td>
<td>shell</td>
<td>165.5</td>
<td>&quot; &quot;</td>
</tr>
<tr>
<td>5730 +/- 50</td>
<td>Beta-47892</td>
<td>VCore 91018-40</td>
<td>shell</td>
<td>80.80</td>
<td>Channel Fill</td>
</tr>
<tr>
<td>7320 +/- 200</td>
<td>Beta-50595</td>
<td>IKU-91018-83</td>
<td>shell</td>
<td>82.5</td>
<td>CH Moraine</td>
</tr>
<tr>
<td>8650 +/- 90 AMS</td>
<td>Beta-47889</td>
<td>Vcore 91018-22</td>
<td>shell</td>
<td>65/1.65</td>
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Table 3. Compilation of 14C ages from this study.
respect to the Libby half-life of 5570 years and reported with the value of one standard deviation error resulting from the random nature of the disintegration process (Table 3). The reported dates are all adjusted for $^{13}$C fractionation calculated relative to the PDB-1 international standard. They are not adjusted for oceanic reservoir effects, unless stipulated, because of the regional variability of the effect and poor mixing of oceanic water within the shallow, closed basins of the Scotian Shelf (Piper et al., 1990).
CHAPTER 3: MARINE QUATERNARY GEOLOGY

3.1 THE STUDY AREA

The marine study area on the inner continental shelf off Nova Scotia, extends 40-60 km south from the present coastline to a prominent set of sea-bottom ridges termed the Scotian Shelf End Moraine Complex (King et al., 1972) (Fig 1). It is a region of irregular relief and highly variable sediment thicknesses (Forbes et al., 1991). The inner shelf trends NE-SW, parallel to the strike of Meguma bedrock basement and the continental shelf break (Piper et al., 1986).

3.2 OCEANOGRAPHY

The main oceanic circulation pattern over the inner shelf is a southwestward flow known as the Nova Scotian Current (Bigelow, 1927). Tides are semidiurnal with an average range of 1.4 m (Hopkins, 1985). Bottom drifter experiments have shown that fair weather tidal current velocities are not sufficient to transport sediment except when focussed into constricted inlets (Hopkins, 1985). The surface wave climate is seasonally variable, with larger waves during winter storms and rare hurricanes (Forbes et al., 1991). Northeastward-trending storm paths (cyclonic lows) across the mainland generate onshore winds. Wave heights up to 8 m are generated with a median value of 1 m. The offshore extent of sediment transport is best indicated in large embayments where shoreface sand wedges can be traced to -20 m (Hall, 1985, Boyd et al., 1988a). Pebble-sized gravel is mobilized during winter storms at 30 m water depth (Forbes and Drapeau, 1989). Combined wave-tidal current activity may entrain silts at depths of 100 m (Forbes et al., 1991).
3.3 ACOUSTIC TERRAIN "ZONES"

Cok (1970), King (1970) and Drapeau and King (1972) divided the Scotian Shelf into three physiographic zones: an inner zone (Inner Shelf) characterized by rough topography, a middle zone (Central Shelf) consisting of broad shore-parallel depressions, and an outer zone (Outer Shelf) consisting of wide, flat banks. Seismic methods have advanced considerably since the 70's enabling further detailed physiographic subdivisions. Five major coast-parallel zones have been designated in the inner shelf (Fig 7). These are (seaward to landward):

1. Scotian Shelf End-Moraine Complex (King et al., 1972),
2. Basin Zone (this study),
3. Outcrop Zone (Forbes et al., 1991),
4. Morainal Zone (Stea et al., 1993a, b), and
5. Truncation Zone (Forbes et al., 1991, Stea et al., 1993a, b)

Kelley et al. (1989) mapped similar shore-parallel zones off the coast of Maine. The inner shelves of both Nova Scotia and Maine exhibit a nearshore zone of sediment accumulation (sand-gravel) grading into a denuded zone with a thin veneer of residual material over rock and till. The author has utilized these previous informal mapping concepts and fused them into a formalized zonal concept.

"Terrain Zones" can be defined as areas of unique bottom morphology (landforms) and seismic sequences and facies (stratigraphy). The zones are a consequence of bedrock type and structure, glacial erosional and depositional processes and sea-level rise and fall. They include assemblages of glacial landforms such as moraines and drumlin fields, areas of
Figure 7  Mapped terrain zones and subzones on the inner shelf. These zones can be defined as areas of unique bottom morphology (landforms) and seismic sequences and facies (stratigraphy). The zones are a consequence of bedrock type and structure, glacial erosional and depositional processes and sea-level rise and fall. Tracklines and the location of the grid model of the Sheet Harbour area (Fig. 8) are also indicated on this figure.
bedrock outcrop, and deep, sediment-filled basins and channels. Subzones are based on landform and stratigraphic commonalities at a smaller scale than zones. The lowest level in the mapping hierarchy is individual stratigraphic units (sequences and facies) definable in basins and valleys within zones and subzones. These seismic stratigraphic units are based on the amplitude, frequency and pattern of internal reflections and their relationship to bounding surfaces (e.g., King and Fader, 1986, Vail et al., 1977).

The Scotian Shelf End-Moraine Complex (SSEMC) is a series of large ridges that occur 30-40 km offshore, parallel to the present-day Nova Scotian coastline (King et al., 1972, Fig. 9). The Basin Zone is a series of closed, sediment-infilled basins in water depths greater than 145 m. The "Outcrop Zone" was defined by Forbes et al. (1991) as a broad area of bedrock outcrop largely devoid of surficial sediment. The "Morainal Zone" was defined as a region of sub-parallel ridges and intervening areas filled by conformably-bedded and ponded seismic units (Stea et al., 1993a, b). The Truncation Zone was defined as a region of the inner shelf with muted topography and planar erosional surfaces truncating bedrock and surficial cover (Stea et al., 1993a, b). A grid model of the bathymetry off Sheet Harbour (Appendix 1), displays four of these zones (Fig. 8). A SSEMC morainal ridge (Eastern Shore Moraine) can be clearly seen, with a linear form and asymmetric profile. The Basin Zone is a flat topographic area north of the moraines characterized by conformably-bedded acoustic units with high amplitude, coherent reflections (Table 1, Emerald Silt) and transparent, ponded facies (Table 1, LaHave Clay). Landward of the Basin Zone are ridges and mounds composed of acoustically incoherent material (Morainal Zone). The Truncation Zone is marked by the change to muted topography around 80-90 m. These topographic zones are described in detail in the next sub-chapters starting with the Scotian Shelf End-Moraine Complex.
Figure 8  Digital grid model of the bathymetry of the inner shelf in the Sheet Harbour transect area (Fig. 1) It was digitized from 12 kHz seismic records of the Canadian Hydrographic Service. The tracklines were 93 km apart and depth measurements were taken every 5 minutes along lines. Note the continuity of the Scotian Shelf End Moraine Complex (SSEMC). The view is towards the southwest from the northeastern end of the Eastern Shore Moraine (see Fig. 9). The arrow marks the location of a major channel also mapped in Huntex records.
Legend:
- Sable Island
- Sand and Gravel
- Alluvial Clay
- Mermaid Silt
- Coastline Shelf
- Ice drift
- Sediments

Figure 8.
Figure 9  The Scotian Shelf End Moraine system (after King et al., 1972)
3.4 SCOTIAN SHELF END-MORAINE COMPLEX

3.4.1 Introduction

The Scotian Shelf End-Moraine Complex (SSEMC) is a series of large sea bottom ridges that occur 30-40 km offshore and parallel the present day Nova Scotian coastline (Fig. 9). The ridges occupy the landward margin of a trough (Basin Zone) that extends more or less continuously along the entire Scotian Shelf with depths of 140-200 m and a width of 40-50 km (King, et al., 1972). They are composed of the Scotian Shelf Drift, defined as an incoherent acoustic unit with occasional diffractive point sources interpreted as boulders (King and Fader, 1986) (Table 1). On the southward-facing, deep water side these ridges interfinger with the Emerald Silt in a series of wedge-shaped projections called till-tongues. Till-tongues are defined as wedge-shaped deposits lacking in internal reflections, interbedded with stratified glacial marine sediment and rooted in large constructional moraines (King and Fader 1986, King, et al., 1991, King, 1993). On the proximal (landward) side of the SSEMC moraines Emerald Silt laps onto the moraines.

3.4.2 Halifax Moraine

3.4.2.1 Location and Stratigraphy

The Halifax Moraine is located 50 km southeast of Halifax Harbour (Figs. 10, 11 and 12) and consists of a series of 2-4 km wide, discontinuous ridge segments which protrude above the seabed. The main segment is oriented NE-SW and is 12 km long. Several shorter segments to the northeast of the main ridge are oriented NW-SE. The total length is 30 km. The ridges...
Figure 10  Legend for the seismic geology-seismic landform maps of the inner shelf transect areas (Fig 1). These maps depict the distribution of acoustic units and their associated landform assemblages. In most cases, stratigraphic formations and facies (King and Fader, 1986, Table 1) or units (cf Forbes et al., 1991) are mapped, but where the unit distribution is too complex, or erosion is dominant, then subzones are depicted on the maps. Schematic diagrams illustrate the geometry and lithostratigraphy of the seismic bodies. Stratigraphic classifications are from King and Fader (1986) and Forbes et al., 1991 (Table 1).
<table>
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Figure 10
Figure 11  Surficial geology-landform map of the seismic tracklines in the southern Halifax transect region of the inner shelf of Nova Scotia (Fig. 1). See Figure 10 for the legend to this map. Core locations are marked by an X, IKU samples are marked by a box symbol. Reference sections, depicted in later figures, are shown with solid black lines and large letters beside the interpreted tracklines. Major southwest-trending channels can be interpolated between tracklines.
Figure 12  Sparker acoustic profile and interpretation of a segment of the Halifax Moraine (Sections B-A, Fig 11) Three till tongues can be delineated in the southern or distal portion of the moraine  Close up of the distal portion of Till Tongue 3 showing the “feather edge” The water depth is shown in metres on the side of the section  These depths are calculated from two-way travel times using 1500 m/sec as the velocity in water and the uppermost sediments
have an asymmetric profile, steeper on the landward (north) side. The moraine segment reference section (Section A-B, Fig 12) is a 4 km wide, broad rise in the seabed. Sediment thicknesses vary from 15 m on the landward (north side) to greater than 40 m on the seaward side, whereas the ridge itself has 30 m of seabed relief. The upper moraine surface ranges between 150 m and 170 m water depth. The reference section ridge is composed entirely of acoustically incoherent material, coherent reflections are not visible even on the low frequency sleevegun record (Appendix 2). Three till-tongues can be resolved on the seaward side of the type section (Fig 12), facing southeastward. These tongues are irregular in shape when compared with the reference tongue from the Gulf of Maine (King and Fader, 1986, p. 14). The lowest wedge-shaped protrusion (400 m long by 10 m thick) is isolated from the morainal fragment by irregularly-shaped, high amplitude reflections that extend well into the morainal segment. It has a convex-upward shape and terminate sharply against the Emerald Silt. In section A-B (Fig 12), a package of Emerald Silt reflections appear to be closer together at the edge of the till-tongue, than either underneath or beyond the tongue (Fig 12). The second tongue is thicker with little apparent thinning at its extremity (Fig 12). The highest tongue is only a few ten's of metres long and few metres thick. On the seaward (southern) side of the moraine the uppermost Emerald Silt reflections appear to grade into, rather than lap onto, the incoherent material that makes up the ridge. The surface of the Halifax Moraine is irregular, probably furrowed. Numerous point hyperbolic reflectors within the moraine may indicate boulder-sized clasts. On the landward side, Emerald Silt reflections appear to lap onto the moraine.

3.4.1.2 Lithology

IKU grab samples (28, 50, 51 and 52, Figs 11, 13, Appendixes 3, 4) were taken from the
Halifax Moraine  IKUs 28, 50 and 52 consist of a homogenous, olive-grey, sandy, matrix-supported diamicton (Fig 13)  IKU 50 contains grey sand and reddish-brown clay inclusions  The basal samples of IKUs 28, 50, 51 and 52 vary between 23.8 and 38.2 wt % gravel (Appendix 7)  IKU 52 contains 51% sand and 23% mud (Fig 14)

The clast shapes of these samples generally fall into three populations, angular to subrounded, unstriated (70-95%), subangular to subrounded, striated (1-30%), and rounded (<5%)  Significantly, some rounded, unstriated pebbles in the bottom sample of IKU 52 (91018-52-5, Appendix 4) have angular faces indicating recycling (Fig 15)  There does not appear to be any significant variations in clast shape and striation percentages between the top, middle and bottom subsamples of all the IKU samples from the moraines (Appendix 4)

The pebbles consist primarily of Meguma Group metagreywacke (84-90%)  SMB granite fragments make up 1-5% of the pebble fraction and foreign lithologies comprise 5-11%  The foreign lithologies are dominated by mafic intrusive rocks (diorite), pink, K-feldspar and hornblende-bearing granitoid intrusive rocks, porphyries, felsic volcanic rocks and red clastic sedimentary rocks  Again, there is no apparent systematic variation between the top and bottom subsamples of each IKU grab (Appendix 4)

3.4.2.3 Age

The Halifax Moraine may be somewhat younger than the Sambro Moraine which is located
Figure 13. Photographs of the IKU grab samples. A: IKU 52 (M) Halifax Moraine, B: IKU 55 Sheet Harbour Moraine, C: IKU 49 Morainal Zone. Note that the diamictons are matrix-supported, and some clasts exhibit stoss and lee form.
Figure 14. Detailed grain size analysis of selected samples from the marine and terrestrial portions of the study region. LT-Lawrencetown Till; ST-Beaver River Till (Stony Till Plain); HT-Hartlen Till; TT-Till-Tongue (Emerald Silt?); RM-Morainal Zone (Scotian Shelf Drift); IKUs 52, 55 from Scotian Shelf Morainal Complex (Scotian Shelf Drift).
Figure 14.
Figure 15  Photographs of pebbles in IKU grab samples from the Scotian Shelf End Moraine Complex. Three pebble shape populations can be distinguished in most samples from the moraines: (A-right) IKU 52, glacial, flat-iron with striations, (A-middle) well-rounded with fracture indicating recycling, (A-left) angular, unstriated, (B-left) IKU 55 striated glacial clasts, (B-right) IKU 85 (Till Tongue) - well-rounded clasts.
15 km seaward (Fig. 9), assuming that the landforms were deposited sequentially during landward-retreat. The age of the distal till-tongue on the Sanbro Moraine (Fig. 9) has been estimated at 15 ka by Gipp and Piper (1990, age not corrected for reservoir-age effect). During the 1991 survey, core 91018-53 was obtained in a basin just north of the Halifax Moraine (Fig. 11). The core penetrated into glacial marine sediments which onlap onto the proximal part of the Halifax Moraine. The apparent age near the base of the core is 13,050 ± 140 14C years on a fragment of *Nucula* sp (τ3 C corrected) (Table 3). The age can be as much as 440 years older with reservoir age uncertainty (Mangerud, 1972). Gipp (1989) calculated sedimentation rates of 10-30 m/kyr for basal glacial marine sediments in the Emerald Basin. Using 20 m/kyr as a sedimentation rate for the lower part of the core and an age of 14.1 ka for the core base (corrected for reservoir age-effect), the interpolated age for the top of the moraine surface at the core site is 14.5 ka. This is considered to be a minimum age only.

### 3.4.3 Eastern Shore Moraine

#### 3.4.3.1 Location and Stratigraphy

The Eastern Shore Moraine is located 45 km south of Sheet Harbour (Fig. 9). It is a continuous, NE-SW-trending ridge (Figs 8, 9, 16), 1-3 km wide and 23 km long. This morainal segment has a marked asymmetry, thicker and steeper on the north (landward) side and thinner on the south (Fig. 17). It has 20 m of seabed relief in water depths of 128 to 148 m and ranges in thickness from 45 to 70 m. The moraine consists of Scotian Shelf Drift divisible into three massive acoustic units on the sleevegun record (Fig. 17), with a
Figure 16. Surficial geology-landforms of the Sheet Harbour Transect area (Fig. 1). See Figure 10 for the legend to this map. Core locations are marked by an X; IKU samples are marked by a box symbol. Reference sections, depicted in later figures, are shown with solid black lines and large letters beside the interpreted tracklines. The geology is interpolated between tracklines by matching morainal landforms on the Huntex records with topographic highs on a contoured bathymetric map derived from echograms (Map 2). Boxed region is a region of closely-spaced tracklines over the Morainal Zone. Note also a major channel emanating from the Morainal Zone.
Figure 17  Sleevegun record and interpretation across main segment of the Eastern Shore Moraine (Fig 16, A-B)  The moraine is comprised of three massive acoustic units (Units 1-3)  The units taper in a seaward direction  Unit 3 appears to truncate Units 1 and 2 on the landward side of the moraine
Figure 17
prominent reflection between each of them. Unit 3 appears to truncate Units 1 and 2 on the landward and seaward sides of the moraine. On the seaward side of the moraine is a unit with higher frequency reflections, interpreted as Emerald Silt (Fig. 17) and the acoustically incoherent units taper towards the Emerald Silt, but the resolution is inadequate to determine the precise contact relationships. Huntec records show that Emerald Silt reflections are draped on the landward side and interfinger with Unit 3 on the seaward side. The distal surface of the moraine is rough with ridge-swale amplitudes of 2-6 m, while the proximal slope is relatively smooth. The sidescan record of the distal slope reveals a series of subdued, straight grooves 30 to 50 m wide with elevated edges or "berms." The roughness of the moraine surface apparent on the reflection records, is attributed to these linear furrows rather than isolated mounds and hollows. These furrows resemble those described by Fader (1989b) and Josenhans and Barrie (1989) attributed to iceberg scour. These appear to be best developed in water depths above 145 m. On the Sambro moraine, iceberg furrows are found above 180 m water depth.

3.4.3.2 Lithology

IKU grab samples (55 and 56 Fig. 16, Map 2) were taken from the top of the moraine. They consist of olive-grey, sandy, matrix-supported diamicton, with angular to subrounded clasts. Both IKU samples have higher clast/matrix ratios in their top or surface samples. IKU 56 contains clayey inclusions. The basal samples of IKUs 55 and 56 vary between 18.8% and 20.4% gravel whereas the top sample of IKU 56 contains 34.8% gravel. IKU 55 (basal subsample) contains 37% gravel, 49% sand and 14% mud (Fig. 14).
As in the Halifax Moraine samples, three distinct populations of clast shapes can be discerned. In IKU 55, most of the larger clasts are unstriated and subangular, whereas in IKU 56 they are largely subrounded. In both samples a minor proportion of the clasts are striated (16 to 27%) and facetted with flat-iron shapes indicative of glacial erosion. A few clasts are rounded to well-rounded with no surface features. There are no significant variations in pebble shapes and percentages of striated pebbles between the surface and bottom IKU subsamples. On sidescan sonograms the surface of the moraine is variegated with white and dark areas. The strong reflections (dark regions) and acoustic shadows (white areas) suggest randomly spread objects 1-4 m in diameter (boulders).

Metagreywacke of the Cambro-Ordovician Meguma Group (85-90%) is the dominant rock type in the pebble fraction of the IKU samples. Meguma Zone granites and foreign lithologies are only minor components (Appendix B). Foreign lithologies include pink, K-feldspar and hornblende-bearing, intrusive rocks, felsic volcanic rocks and red and grey sandstones. Metagreywacke percentages do not vary significantly between the top and bottom IKU subsamples.

The age of the Eastern Shore Moraine could not be determined either directly or indirectly.

3.4.4. Country Harbour Moraine

3.4.4.1 Location and Stratigraphy

The Country Harbour Moraine, located 30 km south of Country Harbour, is the most complex landform in the thesis area (Figs. 9, 18). It has an arcuate shape, the western
Figure 18  Surficial geology-landforms of the Country Harbour Transect area (Fig. 1). See Figure 10 for the legend to this map. Core locations are marked by an X; IKU samples are marked by a box symbol. Reference sections, depicted in later figures, are shown with solid black lines and large letters beside the interpreted tracklines.
segment trends NE-SW parallel to the other SSEMC moraines, whereas the eastern portion trends NW-SE, nearly perpendicular to the main moraine complex (Fig 9). The western segment is 3 to 8 km wide, similar to the other SSEMC moraines. The eastern section widens to 23 km (King and Maclean, 1974). Like the two previously described moraines, profiles through the western section of the Country Harbour Moraine reveal a marked landward asymmetry and are composed of 3 sub-units of acoustically incoherent material (Fig 15). The Huntec profiles across the Country Harbour Moraine reveal an irregular surface above 125 m water depth resembling the iceberg furrows of the other SSEMC moraines but of lesser amplitude. These were not visible on the sidescan sonograms from the 91018 cruise, but in previous surveys G. Fader (pers. comm., 1994) has noted small-scale iceberg furrows on the truncation surface south of the moraine. South of the Country Harbour Moraine are numerous wedge-shaped, acoustically massive units intercalated with Emerald Silt, Fig 20). These units are rooted in the moraine ridges and are therefore defined as till-tongues. These tongues are several hundred metres to 10 or more kilometres long (Fig 20), but till-tongue lengths can be exaggerated in oblique and transverse sections.

The Country Harbour Moraine differs from other SSEMC moraines in four respects:

1. Shallower water depths (the moraine segment with the greatest seabed relief ranges from 84 to 115 m water depth)
2. Multiple till-tongues within an exceptionally thick glacial marine section (King and Fader, 1980, p. 22)
3. An extensive, planar erosional surface seaward of the moraine which truncates till-tongues and Emerald Silt
4. Lack of an extensive basin or trough on the north side of the moraine
Figure 19  Sleevegun record and interpretation of transect C·D  Section is NW-SE across the Country Harbour Moraine (see Fig 18)  Three massive acoustic units can be differentiated in this moraine segment  The asterisk refers to the truncated root of the till-tongue (cf King et al, 1991)
Figure 20  Sleevegun and Huntac acoustic records and interpretations of transect A-B (see Fig 18 for locations)  Till-tongues are numbered 1 through 6 (at their distal ends)  These tongues are separated by Emerald Silt of variable thickness  The scale on the left is water depth in metres  Note the truncation of acoustically incoherent till tongues and Emerald Silt reflections by the seafloor surface  The proximal end of Till Tongue 3 has an arcuate, asymmetric profile with a steep and gentle slope and may represent a former morainal position
The moraine is a complex, interdigitating succession of till-tongues and intercalated Emerald Silt facies A. King and Fader (1986) counted 9 till-tongues in a vertical section across the NE-trending section of the moraine. 6 till-tongues were identified along with their truncated "roots" (King et al., 1991) in Section A-B (Fig. 20). They are 10 to 50 m thick, less than 1 km to 20 km long, and are horizontal, or nearly so, at their distal end. The orientation of these tongues with respect to the main morainal segments is not certain, so these distances can be misleading. A planar erosional surface truncates the roots of distal tongues 5 and 6 (Fig. 20). Interpolating the tongues and intercalated Emerald Silt above the truncation surface, a former morainal fragment can be reconstructed (Fig. 20). The erosional surface can be traced to the distal part of a distinctly asymmetrical morainal ridge (Fig. 19). The morainal ridge itself does not appear to be extensively eroded, as it exhibits some variable local relief. It has 18 to 25 m of seabed relief. The sleevegun record shows a series of acoustically incoherent units each linked to a till-tongue and truncated by the proximal slope of the morainal ridge (Figs. 19, 20).

3.4.4.2 Lithology

IKU grab samples 82, 83 and 84 were taken from the upper surface of the Country Harbour Moraine. IKUs 85 and 86 were obtained from a till-tongue exposed on the erosional surface (Fig. 18). IKUs 82, 84 and 86 consisted of olive-grey sandy-mud, matrix-supported diamicton. IKU 83 grades from an olive-grey sandy-mud diamicton at the surface to a grey clay with few stones at the base. The top sample of IKU 83 contains 41.7% gravel whereas the basal sample of IKU 84 contains 32.2% gravel comparable to other SSEMC samples (Appendix 9). IKUs 85 and 86 are from a till-tongue exposed on the erosional surface. IKU 85 grades from an olive-grey sandy-mud, matrix-supported diamicton at the surface to a grey
clayey mud with few stones at the base (gravel content 3.5%). IKU 86 is a homogenous, olive-grey, sandy-mud diamicton. It contains only 8% gravel, 68% sand and 24% mud (Fig. 14). Core samples through a till-tongue from the Country Harbour moraine by King and Fader (1986, p. 38) revealed a structureless silty-clay with an average composition of 1% gravel, 7% sand, 43% silt and 49% clay.

IKUs 82 and 83 both display subangular to well-rounded clasts with 7-9% striated clasts in the basal samples, respectively. In IKU 83 the percentages decrease from 7 at the base to 0 at the top. IKU 84 has a higher percentage of rounded stones, some of which appear to be recycled, and has 3% striated pebbles. There is a significantly greater degree of rounding in these IKUs than previous SSEMC diamictons (Fig. 15). Striated pebbles were not found in IKUs 85 and 86 but most of the clasts were angular, soft marlstones, not amenable to retaining striae.

IKUs 82, 83, and 84 contain 82-88% Meguma metagreywacke, 1-6% SMB granitoids and 2-6% foreign and 5% Tertiary or Cretaceous brown marlstones. The foreign lithologies are dominated by volcanic rocks and some pink, probably K-feldspar-bearing intrusive rocks. Also noted were andalusite? and cordierite?-bearing schists derived from high grade metamorphic terranes north of the section. IKUs 85 and 86 on the till-tongues contain 6-17% Meguma metasedimentary rocks, 70-90% buff-brown soft Mesozoic marlstones, 1-3% SMB granites, and 3-17% foreign lithologies.

3.4.4.3 Age

Radiocarbon dates were obtained from two sites in the morainal complex, the morainal crest.
IKU 83) and the erosional surface south of the moraine (IKU 89). The base of IKU 83 contained a shell hash of pelecypods and barnacles including *Chalmys islandicus* and *Placopecten magellanicus* giving an age of 8320±200 yrs BP. IKU 89 was taken from truncated Emerald Silt B (Table 1) facies above a till-tongue on the erosional surface. Shell fragments include *Placopecten magellanicus* and *Pallitium subimbrifer*. Barnacles (*Balanus crenatus*) were most abundant and were radiocarbon dated at 11,770 ± 200 yrs BP.

The erosional surface, found distal to the morainal ridge, formed sometime during or after deposition of Emerald Silt glacial marine strata overlying the youngest till-tongue. The surface appears to truncate reflections from part or all of Emerald Silt Facies B and older till-tongues intercalated with Emerald Silt facies A. In a basin west of the Country Harbour Moraine, King and Fader (1988a) dated the top of the Emerald Silt Formation at 10,580 ± 110 yrs BP. The erosional surface must therefore must be 10.5 ka or older if the correlation with the adjacent basin is valid. The exceptional thickness of Emerald Silt south of the moraine makes correlation with the adjacent northeast Emerald Basin difficult. Facies may be repeated in successive advance-retreat depositional seismic sequences. The age of the erosional surface can only be estimated at somewhere between 10.5 ka and 14.2 ka (base of Emerald Silt Facies B at a reference section in the northeast Emerald Basin, King and Fader, 1988). The 11.7 ka date obtained from the surface of the moraine may represent a "maximum" age for this erosional event if one assumes that the barnacles lived during sediment deposition and younger material above the dated horizon was removed. Alternatively, the barnacles may have grown on the erosional surface during a lowstand at 11.7 ka (Stea *et al.*, 1994 - see discussions of the Country Harbour Moraine in Truncation Zone subchapter).
The early Holocene age for the shell hash on the moraine crest (IKU 83) may be explained by accretion of marine muds, epifaunal growth and subsequent (later) Holocene erosion by submarine currents.

3.5 DEPOSITIONAL MODEL FOR THE SCOTIAN SHELF END MORAINES (SSEMC)

3.5.1 History of Ideas

The origin of these ridges has been debated since their discovery (King, 1969). They were originally interpreted as grounding-line moraines at the terminus of a Late Wisconsinan glacier either in advance or retreat (King, 1969; King et al., 1972). This interpretation was based on the location of the moraines; poised on the edge of a deep trough, at a presumed critical buoyancy depth.

MacDonald (1982) developed a conceptual model of four consecutive northward-retreating ice margins represented by the thickest part of the ridges, and readvance positions marked by till-tongues. King and Fader (1986) reinterpreted the mode of formation of the SSEMC as a subglacial ice-shelf moraine, based on the discovery of till-tongues on the northern edge of some moraines, notably the Country Harbour Moraine.

King et al. (1991) synthesized work done on shelf ice margins off Norway and Nova Scotia and reassessed the sub ice-shelf origin of the SSEMC. They devised a classification of moraines relating to floating ice fronts based on the presence of till-tongues rooted in morainal ridges. The SSEMC was classified as a "linear grounding-zone moraine" deposited
at the grounding line of a floating ice front Powell (1984) and Syvitski (1991) pointed out that ridge formation is unlikely under the constraints of an ice ceiling King et al. (1991) met this objection by proposing that the ridges built up synchronously with rising sea level.

To recapitulate, the SSEMC moraines have these salient characteristics:

1. Asymmetric ridge morphology, steep landward side and gentler seaward slope.
2. Stacked, incoherent acoustic facies that, at least on the surface, are composed of an olive-brown, gravelly-sandy-mud (27% gravel, 73% matrix, Table 4), matrix-supported diamicton with striated (16%, Table 4) and facetted clasts of predominately local origin (88% Meguma Group, 7% foreign, 3% Meguma Zone granite, Table 4).
3. South-(seaward) facing, stacked till-tongues with morainal roots cut off by the steep, north-facing slope. Tongues composed of muddy diamictons and mud (see also King and Fader, 1986, p 15).
4. Ice-berg furrowing above 180 m (Fader, 1991).
5. Absence of closely-spaced reflections and sorted sediments on the surface of the moraine such as would characterize subaqueous outwash-grounding line fan-lobes (Powell, 1990).
6. Anastomosing character in plan view (Fig. 9).

The incoherent internal configuration and absence of high frequency reflections within the morainal segments and till-tongues suggests that massive diamicton-till or unlaminated ice-loaded glacial marine sediments are present and (Belknap and Shipp, 1991, Syvitski, 1991).
### Table 4  Average values of lithic parameters for marine sediments in the study region

**Scotian Shelf End Moraine Complex** (MZ)  **Morainal Zone** (UB)  **Unit B** (UC)  **Unit C** (SSSG)  **Sable Island Sand and Gravel** (ES)  **Emerald Silt**
Piper (1988) and Syvitski (1991) suggested that coarse waterlain deposits may also be present, but undistinguishable from till. Surface sampling of the SSEMC, however, produced only massive diamictons and muds. Matrix-supported diamictons were most common on the moraine ridges, with striated and facetted clasts indicative of subglacial erosional processes (Dreimanis, 1989). Sorted, waterlain deposits were not found, although well-rounded clasts were found in poorly-sorted muds and diamictons of the Country Harbour Moraine (Fig. 15). Grab samples of the Scotian Shelf Drift (largely from the Gulf of Maine) taken by King and Fader (1986, p. 31) have sorting (sigma) values $>3$ (very poorly sorted) and average textural analysis of 38% gravel, 41% sand, 11% silt and 10% clay. Similarity in composition and texture between the SSEMC samples and samples of the Morainal Zone (see p.) and the textural and lithologic homogeneity between top and bottom IKUs suggest that ice-berg turbation (ice-keel turbate, Vorren et al., 1983, Josenhans and Fader, 1989) is not a major factor in the formation of these diamictons.

Grounding line fan lobes, common at the mouth of channel conduits at tidewater margins (Powell, 1984, 1990, Ashley et al., 1991) are not apparent either from the digital model of echogram profiles from the Eastern Shore Moraine (Fig. 9) or from surficial mapping across the region (King and MacLean, 1970)

The absence of high frequency bedding or laminations within the moraine as evidenced in the high resolution Huntec and NSRF sparker records (e.g. Fig. 12) strongly suggests homogeneity of composition. Grounding line fans, on the other hand, exhibit stacked, complex stratified sequences with variable bed thicknesses, often less than 2 m (Ashley et al., 1991).
Isostatically-elevated, tidewater moraines in Maine and New Brunswick (Fig 3) are characterized by complex sedimentology dominated by coarse glaciofluvial and fine-grained glacial marine facies (Smith, 1982; Nicks, 1988). Bedded units within these moraines with large grain-size contrasts i.e. (clay-sand rhythmites; sand-gravel beds) have strong impedance contrasts and would produce reflections on seismic records. The morphology of some of these moraines (Fig. 21) also differs substantially from the SSEMC moraines.

3.5.2 Modern and Pleistocene Analogues

The SSEMC moraines appear to have elements in common with both the modern polar ice shelf and temperate valley glacier settings. Table 5 is a comparison of the characteristics of landforms and deposits produced at ice marginal grounding lines in Alaskan temperate tidewater glaciers and polar glacier ice shelves. Ice shelves inhibit the formation of morainal banks (Powell, 1984) but wedge-shaped, ice-marginal diamicton bodies (tilt-tongues) have been postulated and observed (Alley et al., 1989; Anderson and Bartek, 1992; Drewry and Cooper, 1981; Hambrey et al., 1992; Powell, 1991). Diamicton facies are dominant in ice shelf environments (Anderson et al., 1991; Hambrey et al., 1991) while tidewater termini are characterized by complex sedimentology and structure (Powell and Molnia, 1989). The tidewater environments described in Alaska are dominated by diamictons and proximal glaciofluvial sediment fans (subaqueous outwash, Rust and Romanelli, 1975, Powell, 1984; Powell and Molnia, 1989). The implied absence or rarity of bedded fluvial facies in the SSEMC suggests an incompatibility with the tidewater-morainal bank model.

The moraine created by the surging Bråsvellbreen glacier in Spitsbergen is a possible analogue to the SSEMC (Elverhøi et al., 1989). The scale and morphology of the morainal
### Table 5. A sediment and landform comparison of temperate tidewater and polar ice shelf glacial margins.

<table>
<thead>
<tr>
<th>Margin type/ Property</th>
<th>Tidewater/ Temperate</th>
<th>Ice Shelf/ Polar</th>
</tr>
</thead>
<tbody>
<tr>
<td>Topography</td>
<td>Linear, asymmetric ridges, (morainal banks) elongate, slump scars on the distal side, mud flow fans at the base of the ridges, subaqueous outwash fans also prevalent.</td>
<td>Flat or low relief, lobate fan-shaped areas tens of kilometres long.</td>
</tr>
<tr>
<td>Sediments</td>
<td>Muddy diamictons, poorly-sorted regions of sandy-gravel and sand with boulders common. Rhythmically-laminated muds and sand.</td>
<td>Muddy diamictons predominately, often texturally homogenous.</td>
</tr>
<tr>
<td>Stratigraphy</td>
<td>Glaciotectonic deformation; interdigitating diamictons, laminated muds and outwash, distal muds.</td>
<td>Deltaic clinoforms, made up of diamicton strata (till deltas).</td>
</tr>
</tbody>
</table>
ridge is similar to the SSEMC moraines (Fig. 21) but a surge hypothesis cannot explain the volume of glacial marine sediment interfingering with the SSEMC moraines. After a surge, the glacier profile is markedly lowered and large areas become stagnant as a new active front is initiated upstream (Paterson, 1981). Meltwater production at the surge front is reduced or curtailed as the hydrostatic head is reduced. In the case of the SSEMC moraines, the onlap and interfingering relationships between the moraine and laminated, coarse and fine-grained glacial marine sediment (Emerald Silt facies A) imply a quasi-stable, active ice margin and a substantive meltwater flux.

Holtedahl (1989) described a 60 km long submarine morainal feature (Ra Moraine—Fig. 21) off Norway that can be traced onto land. It has an asymmetric shape (steeper side facing landward) and is composed of diamicton with ice-push structures. Holtedahl (1989) interpreted it as a feature formed at the margin of an advancing ice-front. Barnett and Holdsworth (1974) studied lacustrine moraines from the Barnes Ice Cap, Baffin Island. They postulated that moraine formation occurs in a wedge-shaped space beneath a tapered section of ice (ramp) by meltout.

3.5.3 Till-tongues

Till-tongues are wedge-shaped deposits lacking in coherent internal reflections, interbedded with stratified glacial marine sediment and rooted in large constructional moraines (King and Fader 1986; King et al., 1991; King, 1993). They were originally interpreted by King and Fader (1986) as the result of buoyancy-line fluctuations of a grounded ice shelf. King (1993) later envisioned formation by combined processes of sediment-gravity flowage from the grounding zone of a floating-front glacier and upstream deformation during advance and
Figure 21  Profiles of moraines in the study are compared with other worldwide examples. Note in all cases a marked asymmetry, steeper on the landward side, and gentler slope on the seaward-down glacier side. Data from Elverhøi (1989), Holtedahl (1989), Nicks (1988).
Bràsvellbreen Moraine, Spitsbergen
St John Moraine, New Brunswick
Ra Moraine Norway
Wilkes Land Moraine, Antarctica
Scotian Shelf End Moraines

1 Eastern Shore Moraine
2 Halifax Moraine
3 Country Harbour Moraine

Figure 21.
retreat of the grounding zone. The deformation till-tongue model is analogous to the till-delta hypothesis (Alley et al., 1989). In this model, deforming till under an ice stream, responsible for the motion of the glacier, forms horizontal till-topset beds over inclined, foreset gravity-flow deposits at the grounding line. The overlying ice shelf does not accommodate morainal ridge formation. Banks or ridges imply a cessation of motion or stability and sediment accumulation (Powell, 1984). Bank or ridge formation in King's (1993) model postdates till-tongue formation and occurs during ice retreat. Deposition occurs when sea-level rises and the glacier separates from the bed. In the case of the Scotian Shelf, however, isostatic recovery during deglaciation would cause initial sea-level fall and grounding of the glacier.

Alternative origins for till-tongues have been presented which include:

2. Glacial marine deposition (Vorren et al., 1991)

The details of the mass movement processes in hypothesis 1 have not been elucidated. A slump is a type of slide whereby the movement is essentially rotational, and the mass of sediment is not deformed or fluidized. Debris flow is a moving fluidized mass that when deposited, can exhibit lateral grain-size variations (Lee et al., 1993). Grain-size variations between till-tongues and their morainal roots (Fig. 14) suggest a mass-flow process rather than slumping. The morphology of tongues from this study area (Figs. 12, 19, 20) and the type sections in the Gulf of St. Lawrence (King and Fader, 1986, p. 14) differ from classic debris or mud flows (Edwards et al., 1993, Fig. 22). Head scarps are not apparent in the sections although it could be argued that these were eroded by subsequent ice movements. Instead
Figure 22  Contrasting the morphology of till tongues and mudflows  

Mudflow morphology (Edwards et al., 1983)  
1. A main scarp at the head of the slope failure that marks the point of detachment of the flow (Head Scarp)  
2. An arcuate amphitheatre-shaped region of erosion (Evacuation Zone) whereby sediment is removed and constitutes the main body of the mudflow downslope  
3. The edge of the flow is generally marked by a thickening bulge of sediment (Toe Bulge) 

Till tongue morphology  
1. Abrupt contact with distal sediments, undulating under surfaces and flat upper surface near the thin or "feather edge" (King and Fader, 1986, p 48)  
2. Horizontal distal margin  
3. Moraine-ward thickening
MUDFLOW MORPHOLOGY

Head scarp

Toe bulge

Evacuation Zone

TILL TONGUE MORPHOLOGY

Morainal ridge

Feather edge

Till tongue "root"

Figure 22.
of marginal thickening (toe bulge), till-tongues generally thicken towards the moraine source, sometimes quite markedly (Figs 12, 19, 20, King and Fader, 1986, p 14, 48) An evacuation zone is not apparent in any of the moraine profiles although some of the tongues have steep surface slopes at the margins. The volume of sediment required for the tongues and the low repose angles also mitigate against a simple mass-wasting origin. Tongues with a more complex shape than the idealized forms, may have been modified by minor syn-or post-depositional slumping, but in general these features do not display the characteristics of either slumps or debris flows (King et al. 1991).

Vorren et al. (1991) argue for a suspension-fallout marine origin for transparent, wedge-shaped facies in the Barents Sea based on preservation of relict iceberg furrows underneath the features and a proximal to distal decrease in backscatter along the "tongues." Core samples from these wedges produced overconsolidated and fissile, muddy diamictons which Vorren et al. (1991) attributed to post-depositional glacial modification.

Wedge-shaped, incoherent-chaotic, seismic facies have been observed on the continental slope (Mosher et al., 1989) but are not rooted in morainal ridges. Some of these may be true mudflows or slides. The term till-tongues should be restricted to those forms that are clearly rooted in morainal ridges and exhibit marginal thinning ("feather edge", King and Fader, 1986).

King and Fader (1986 p 38) cored a till-tongue (till-tongue 7) south of the Country Harbour Moraine. They found that the tongue was texturally identical to Emerald Silt facies. A Samples from the eroded surface of a distal to proximal till-tongue from Country Harbour (IKU 85-86, Fig 19) consist of a muddy diamicton and mud. Minor deformation of Emerald
Silt strata is evident on underside of some till-tongues on an erosional surface that truncates till-tongues off Country Harbour (Fig. 23). Selective "homogenization" of Emerald Silt was also observed on this truncation surface. (Fig. 23). Fader (pers comm, 1994) cored till-tongues off Newfoundland and found that they were largely composed of mud, and less consolidated than morainal banks, but with higher shear strength than Emerald Silt.

3.5.4 Genesis of the Scotian Shelf End Moraine Complex

The author found King et al. (1972) and MacDonald's (1982) original interpretations of the SSEMC compelling and in accord with most of the evidence. The asymmetric profiles of the SSEMC moraines have not been stressed in the previous literature (cf. King and Fader, 1986; Piper et al., 1990), but indicate that the ice front was closer to the proximal side than the distal. The Alaskan tidewater and polar ice shelf models do not satisfy many of the SSEMC characteristics, for reasons discussed earlier. Barnett and Holdsworth's model appears to be a compromise between ice-shelf and vertical tidewater fronts. Ridge formation occurs under an ice ramp rather than proximal to a vertical ice front (Fig. 24). Diamicton is deposited directly on the ice ramp-till bed by conveyor-belt recycling and lodgement (Powell, 1984) or push-squeeze processes (Boulton, 1986). Rock outcrop, thin drift and basins behind the SSEMC, and the steep morainal slope suggests erosion behind the grounding line. Erosion of previous till was accomplished through freeze-on (Powell, 1984). Changes in mass balance, sea level drop or buildup of the morainal bank (Powell, 1984) dictated further forward movement. Till-tongues were formed by grounding-line advance and smearing-out of the moraine in the proximal part, then deformation of Emerald Silt in the distal regions (cf. Boulton, 1990; Hart and Boulton, 1991). Increased buoyancy in deeper water caused the ice sheet to decouple from the substrate forming the "feather edge" till-tongue margin.
Figure 23  Huntec seismic profile and interpretation of a till-tongue cut by an erosional surface south of the Country Harbour Moraine (see diamicton zone in section A-B, Fig 18) Note apparent transformation of Emerald Silt reflections into a "massive zone" and deformation of some Emerald Silt reflections
NW

SEISMIC RECORD

SE

INTERPRETATION

0 700 m

TILL TONGUE

Figure 23
Figure 24  Moraine and till-tongue formation a schematic model

1 Quasi-static ice front at approximately 300 m water depth (critical buoyancy thickness)  Ice erodes the proximal side of the moraine and builds-up the distal side by "conveyor-belt" transportation and melt-out

2 A change in mass balance or sea level drop? causes ice to advance into deeper water and truncate the former moraine surface, smearing-out moraine diamicton (A) in the proximal part and forming a deformation till in the distal part (B) of the tongue with no debris in transit

3 Ice retreat and secondary bank formation, with deposition of Emerald Silt
Figure 24.
Following this decoupling phase the ice either spread out as an ephemeral ice shelf or rapidly retreated through calving. A new margin was established in shallower water near the region of thicker morainal deposits left behind at the former margin (Fig 24). Implicit in an ice-advance model for till-tongues is synchroneity of tongues across the shelf. Age dating of the tongue margins will provide a test of this hypothesis.

There is an apparent lack of sorted gravel and sand within and distal to the SSEMC moraines. Grounding-line fans have not been identified (Fig 9). This is not compatible with a temperate-tidewater margin as described earlier (Table 5). A limited meltwater flux is indicated by glacial marine mud intercalated with the moraines. High porewater pressures can result when glaciers override till and impermeable bedrock substrates like those of the inner Scotian Shelf (cf. Boulton, 1990). Weertman (1966) suggested that meltwater in these environments is relegated to a thin film which emanates from the ice base as sheet flow. The rapid transition from till to laminated muds which characterizes the Scotian Shelf End Moraine Complex can be explained by the interaction of low-velocity sheet flow and subglacial processes. This relationship may also indicate that the glacier is in positive mass-balance; with frontal advance balanced by ice-berg calving. In this case, meltwater generated at the surface of the ice sheet under high hydrostatic pressure would be minimal. The tidewater moraines in New Brunswick and Maine (Fig 3) formed around 14 ka, during a period of enhanced worldwide glacial melting (Fairbanks, 1989).
3.6 BASIN ZONE

3.6.1 Introduction

North of the Scotian Shelf End-Moraine Complex is a series of narrow depositional troughs filled with glacial marine and marine sediments (Figs. 7, 8, 11, 16, 18). This region is termed the Basin Zone. Within the Basin Zone are two primary depositional areas termed the Sheet Harbour and Halifax sub-basins.

3.6.2 Sheet Harbour Sub-Basin

The Sheet Harbour sub-basin is a 7 km x 30 km trough located landward of the Eastern Shore Moraine in water depths greater than 145 m (Fig. 16; Map 2). At the type section (C-D; Fig. 16), six seismic sequences (SH1-SH6) with a total thickness from 12 to 30 m overlie acoustic basement (Fig. 25). Over most of the Sheet Harbour sub-basin, acoustic basement (internal reflections cannot be detected within this unit) occurs beneath a strong reflection with irregular relief marked by numerous side-echoes which extend 20 to 50 ms down-record. It is interpreted as Meguma Group bedrock (King and Fader, 1986). Bedrock is sporadically overlain by a unit with incoherent backscatter interpreted as the Scotian Shelf Drift (Table 1).

At the type section, the lowest sequence (SH1; Fig. 25) consists of hummocky, coherent reflections of low continuity and low to moderate amplitude. Emerald Silt facies A and C are represented in this sequence (Table 1). These reflections are draped over the underlying bedrock and drift topography. Isopach data show that SH1 is thickest (12 m) on the north
Figure 25. Huntec seismic profile and interpretation of the type section (C-D; Fig. 16) of the Sheet Harbour sub-basin. Seismic sequences (SH1-SH6) and erosional sequence boundaries (SHA-SHC) are shown. Water depths are in metres below sea surface.
SEISMIC PROFILE

INTERPRETATION

Figure 25
side of the Eastern Shore Moraine, and thins markedly in a northward direction to the till-tongue moraine on the northern boundary of the sub-basin (Fig. 16; 26).

Reflections within the overlying sequence (SH2) exhibit moderate to high continuity and amplitude. SH1 and SH2 appear conformable at the type section, but SH2 laps onto SH1 at the basin margin. Small and large scale undulations on the bounding surface between SH1 and SH2 are largely concomitant with the dips and rises in underlying bedrock. In the Emerald Basin, Gipp (1989, p. 116) reported steep, erosional features on the Emerald Silt surface marked by hyperbola and depressions and interpreted these as buried pockmarks and iceberg furrows. These features are rare or absent in the inner shelf Basin Zone. Isopach data shows that SH2 has variable thicknesses, thickening on the northern and southern boundaries of the basin (Fig. 26). The reflection configurations of SH1 and SH2 are compatible with the Emerald Silt facies "A" as defined by King and Fader (1986) (Table 1).

An erosional sequence boundary (SHA) separates SH2 and SH3. This sequence boundary is best developed where it merges with an erosional surface that truncates sequences SH1 and SH2 on the northern side of the Eastern Shore Moraine (Fig. 25). This boundary in the rest of the basin is formed by downlap of Sequence SH3. It can be traced across the Sheet Harbour sub-basin and the entire Basin Zone.

Sequence SH3 is characterized by moderate reflection amplitude and continuity and high frequency. It is ponded in broad hollows within the underlying sequence (SH2). SH3 and SH4 are separated by an erosional sequence boundary (SHB). SH4 is characterized by low amplitude, discontinuous reflections which lap onto Sequence SH3 at the basin margins. SH1 through SH4 drape over underlying topography to a lesser degree for each successively
Figure 26. Isopach data from Sequences 1 through 6 in the Sheet Harbour sub-basin. The contour interval is 2 m. Sample locations along tracklines are indicated as small boxes. See Fig. 16 for trackline locations.
younger unit SH3 and SH4 are composed of Emerald Silt facies "B" (Table 1). Isopach data shows a broad area of SH3 that exceeds 4 m in thickness just north of the Eastern Shore Moraine, while SH4 is thickest in the middle of the basin (Fig 26).

The SH4/SH5 sequence boundary (SHC-Fig 25) is marked by erosional truncation of the underlying units. Sequence 5 (SH5) is characterized by a return to high amplitude, coherent and continuous reflections (Emerald Silt facies A) which lap on broad hollows in the SH4/5 boundary surface. Above SH5 is a thick, transparent unit with some barely discernable, low amplitude reflections (SH6). This unit appears conformable with SH5 but the contact relationships are not well resolved. SH5 and SH6 are thickest in the middle of the Sheet Harbour sub-basin (Fig 26). SH6 is also very thick (>20 m) in channels located in the southwest part of the Sheet Harbour map area (Fig 16, 26, Map 2).

SH1 and SH2 (Emerald Silt facies A) pinch out on the landward side of the Sheet Harbour sub-basin (Fig 16, Map 2). Where bedrock lies close to the seabed, SH1 and SH2 are truncated by an unconformity at the sea floor. SH6 (LaHave Clay Formation) overlies Emerald Silt in basins between topographic highs in water depths from 137 to 124 m. At the landward boundary SH6 (LaHave Clay) pinches out abruptly against a rise in acoustic basement.

3.6.2.1 Lithology

This basin was not cored or otherwise sampled.
3.6.3 Halifax Sub-Basin

The Halifax sub-basin is an irregular trough landward of the Halifax Moraine in water depths greater than 150 m (Fig. 11). The type section comprises a series of sparker records (Fig. 27) located several km north of a segment of the Halifax Moraine (D-E-F, Fig. 11). Six seismic sequences can be recognized in the NSRF sparker and Hunttec seismic profiles at the type section. These units are separated by sequence boundaries of erosional and lapout nature. The lowest unit (Scotian Shelf Drift) is characterized by incoherent backscatter, high reflectivity, high surface relief and occasional point source diffractions. It has some internal structure and less surface relief than the acoustic basement of the Sheet Harbour sub-basin. A coherent reflection found underneath the Scotian Shelf Drift may be bedrock.

The Scotian Shelf Drift is overlain by Sequence H1 which has a draped geometry, and is characterized by moderately continuous to discontinuous reflections of low to medium amplitude. Point-source diffractions (boulders) are visible within H1. The contact relations between acoustic basement, the Scotian Shelf Drift and H1 are obscure. H1 is composed of both Emerald Silt facies "A" and "C". Isopach data show that H1 is thickest in proximity to morainal segments (Halifax Moraine) (Fig. 28). In topographic hollows in the basin, H1 is overlain by a unit with reflections of higher amplitude and continuity (Sequence H2). Reflections within H2 lap onto H1 at basin edges. Sequence H2 also appears to be thickest adjacent to morainal fragments especially in the eastern part of the Halifax sub-basin (Fig. 28).

Sequences H2 and H3 are separated by a high intensity reflection which represents a major sequence boundary (HA). The boundary is predominately lap out in nature, although
Figure 27. Three segments representing the type section for the Halifax sub-basin. Sequences are indicated by boxed numbers shown on the figure. The locations of segments D, E, and F are on Figure 11. Sequences H1-H7 and boundary erosional unconformities (HA-HC) are described in the text. The arrow marks a prominent unconformity near top of Sequence 4 (HC) probably formed during the deposition of Sequence 5.
Figure 28. Isopach data for Sequences H1 to H6 from the Halifax sub-basin. Contour intervals are 2 m. Tracklines from the Basin Zone area are included for clarity (see Fig. 11 for trackline geology).
evidence of erosional truncation can be found at basin margins. The magnitude of the acoustic impedance boundary represented by the unconformity is indicated by a "double" reflection which may indicate an abrupt grain size change or the presence of gas (Belknap et al., 1991, Bacchus, 1993).

Sequences H3 and H4 are differentiated from H2 by lower amplitude reflections with less continuity. H3 contains a higher percentage of point-source diffractions (boulders) and slightly higher amplitude reflections than H4. These sequences also display a ponded, onlap-fill style of deposition. The base of H4 is characterized by a thin zone of high amplitude, continuous reflections. The rhythmic reflections evident in sequences H2, H3 and H4 may result from acoustic impedance contrasts between sand and silt layers and massive muds. The sequence boundary that separates H3 and H4 (HB) is another high amplitude "double" reflection. H3 is truncated at the sequence boundary (HB) (Fig. 27). Isopach data shows that H3 and H4 are thickest in the middle of the basins north of the Halifax Moraine (Fig. 28).

An erosional and hiatal (lapout) boundary (HC-Fig. 27) separates sequences H4 and H5. This hiatal boundary can be traced to a planar erosional surface developed on the basin margins, where H3 and H4 are clearly truncated (arrow-Fig. 27).

Sequence H5 consists of extremely high amplitude reflections (Emerald Silt Facies A) in marked contrast to H4, with a conformable depositional style. The correlative unit in the Sheet Harbour sub-basin (SH5) has a contrasting onlap-fill morphology. H5 is 4 m thick in the east and west end of the Halifax subzone north of the Halifax Moraine (Fig. 28). South of the moraine the unit appears to thin out.
H6 has reflections of low amplitude and continuity which lap onto H5 and boundary HC. It is the seismostratigraphic equivalent to the LaHave Clay Formation of King and Fader (1986). It is thickest in the northern end of the Halifax sub-basin concomitant with the thinnest regions of Emerald Silt (H1-H4, Fig 28) A sequence boundary H6/H7, recognizable on the sparker records in the Halifax sub-basin is not seen in the Huntec records of the Sheet Harbour sub-basin H7 is marked by reflections of slightly higher amplitude than H6

3.6.3.1 Lithofacies

The Halifax sub-basin was piston cored at a location 25 km north of the Halifax Moraine (Fig 13) The core is located 46 km south of the present coastline in 168 m water depth (Fig 13, 29) A distinctive, high amplitude seismic horizon just below the LaHave Clay and above the Emerald Silt was targeted because of suspicions that it could relate to the Younger Dryas climatic event (G Fader, pers comm, 1992) The core penetrated 936 m of sediment and 820 m of core was recovered The core lithostratigraphy is described in Figure 30

3.6.3.2 Biofacies and Ages

A total of 72 samples from this core were qualitatively analyzed by D Scott (pers comm, 1992) and Costello (1994) for foraminiferal content and their results are summarized below with some modifications to defined biofacies and with additional data on marine pelecypod fauna. Seven major biofacies were defined (Fig 31) Radiocarbon dates were obtained from intact shell valves, or valve fragments, and samples of total foraminifers
Figure 29  Huntco seismic profile at core 91018-53 location (Fig 13) Core placement within the seismic section is based on matching high amplitude reflections with coarse sediment beds in core 91018-53 and age dating core sediments. Numbered boxes refer to seismic sequences described in the text. The core lithostratigraphy is shown on Figure 30.
Figure 29.
Figure 30. Lithology-lithofacies and sand percentages of core 91018-53. Dashed pattern refers to predominately clay-silt strata. Dotted pattern refers to sandy-mud horizon (Lithofacies 5). Gravel percentages in the core are uniformly less than 1 percent. Silt and clay (mud) make up 95 to 100% of the sample weight.
Lithofacies 7. Olive-grey (5Y 3/2) massive mud.

Lithofacies 6. Dark greyish-brown (2.5Y 4/2) mud. Bioturbation at 100 cm.

Lithofacies 5. Olive grey (5y 3/2) massive stiff mud. Bioturbated from 100-240 cm. White silty layer at 230 cm.


Lithofacies 3. Olive grey (5y 3/2) slightly mottled, bioturbated mud.

Lithofacies 2. Dark greyish brown (2.5Y 3/2), heavy black mottling, oxidation to reddish hue.

Lithofacies 1. Dark greyish brown (2.5Y 4/2) compacted mud. Red silt and sand in matrix.
Figure 31  Graph showing percent foraminifera versus depth in core 91018-53 (after Costello, 1994)  A-I are biofacies represented by foraminiferal assemblages (modified after Costello, 1994)  See text for assemblage description
Figure 32  Summary diagram of core 91018-53 showing rates of sedimentation and the correlation among foraminiferal assemblages (biofacies), lithofacies and seismic sequences. Included are the interpolated ages of the boundaries between lithofacies-biofacies (foraminiferal assemblages) and seismic sequences (modified after Costello, 1994).
3.6.3.2.1 Biofacies G, H and I

The upper metre of the core revealed a high diversity assemblage of calcareous foraminifera (planktic and benthic), with 4-14,000 individuals per 10cc (Biofacies G-I, Fig 31). The dominant species are *Bulimina margnata*, *Nonionellina labradorica*, *Cassidulina laevigata* and *Bolvina subaenanensis*. These biofacies represent Holocene warm water basin faunas similar to the present day Emerald Basin (Scott et al., 1984). A radiocarbon date of 9220 ± 90 14C yrs BP was obtained from total foraminifera at 98-100 cm (Fig 30). A shell fragment (*Chnocardium* sp) from 24 cm depth was dated at 3970 ± 60 14C yr BP. The interpolated age at the base of biofacies G based on an age-depth curve is 9840 14C yrs BP (Fig 32).

3.6.3.2.2 Biofacies E-F

Abruptly, at 130 cm to 260 cm the fauna changes to one dominated by *Elphidium excavatum f. clavatum* and little else. The total foraminiferal count drops to 66-207/10cc concomitant with the influx of sandy sediment (Figs 30, 31). Diatoms are common with this faunal assemblage, occurring in pulses of abundance. Diatoms have been reported in association with sea or pack ice (Kennett, 1982). In this core interval there is an increase in reddish-brown organic fragments (algae? twigs? and one spruce needle?) and pyritized organic fragments. From 260 to 340 cm (Biofacies E) total foraminiferal amounts remain low (17-132/10cc) but several other species can be identified including *Nonionellina labradorica*, *Islandiella teretis*, and *Cassidulina reniforme*. The base of Biofacies E corresponds to the base of sandy lithofacies 5 (Figs 30, 31). A single shell valve (*Yoldia thraciaeformis*) at 287 cm produced an age date of 11,340 ± 70 14C yr BP. This shell fragment was probably reworked from older deposits, as erosion of previous deposits is clearly evident in seismic profiles at this stratigraphic level (Fig 27). The interpolated age for the base of Biofacies E-
Lithofacies 5 is 10680 $^{14}$C yrs BP (Fig. 32)

3.6.3.2.3 Biofacies D

Between 340 cm and 370 cm the total foraminiferal count increases to 600/10 cc and then decreases to 300/10 cc. *Elphidium excavatum f. clavatum* is still the dominant species. Biofacies D was duplicate-dated to assess sampling variability. Ages of 10,740 ± 90 and 10,870 ± 70 $^{14}$C yrs BP were obtained from separate shell fragments (*Yoldia thraciaeformis*) at 345 cm. The age at the base of this interval is interpolated as 10,900 $^{14}$C yrs BP (Fig. 32).

3.6.3.2.4 Biofacies C

From 370 to 520 cm the total foraminifera increase to average about 2000/10cc. The fauna changes back into a cold water, normal-salinity fauna with the return of abundant *Nonionellina labradorica, Islandiella teretis* and other species of outer Labrador current affinities (cf. Scott *et al.*, 1984) (Biofacies C). *Elphidium excavatum f. clavatum* is the dominant species. The faunal assemblage of Biofacies C persists to about 390 cm where *Cassidulina reniforme* begins to increase. *Nonionellina labradorica* peaks and then decreases markedly towards the base of Biofacies C and total foraminiferal counts also decrease. A age of 12,420 ± 80 $^{14}$C yr BP was obtained from this zone from bulk foraminifera. This date is slightly older than the shell dates and is offset on the age-depth curve (Fig. 32). "Old carbon" contamination may account for this discrepancy because these samples cannot be etched and pretreated. Nielsen *et al.* (in press) describes a systematic offset of as much as 5000 years, between total foraminiferal dates and mollusc ages in the same core. They interpret the offset as a result of reworking. The interpolated age for the
3.6.3.2.5 Biofacies B

Biofacies B from 520 to 620 cm depth is characterized by a marked drop in total foraminiferal abundances to 103-500/10 cc, similar to the values in Biofacies E-F. *Elphidium excavatum f. clavatum* remains the dominant species but *Islandiella teretis* begins to decline and *Nonionellina labradorica* disappears. Between 530 and 570 cm the counts drop to less than 300/10 cc. There is an influx of detrital material including reworked organics and reddish clastic grains between 500 and 600 cm. Shell (*Yoldia thraciaeformis*) ages of 12,020 ± 90 (590 m) and 12,230±70 14C yrs BP were obtained from this foraminiferal unit. The interpolated age for the base of this interval is 12,203 14C yr BP.

3.5.3.2.6 Biofacies A

Below 620 cm the dominant foraminiferal species are *Elphidium excavatum f. clavatum* and *Cassidulina remiforme*. The total counts increase to average about 2000 individuals/10 cc. A single pelecypoda valve (*Nucula* sp) from the base of the core (750 cm) was dated at 13,050 ± 140 14C yr BP. Sandy sediment residues in this unit were reddish-brown in colour. The age at the base of the core is interpolated at 13,500 14C yr BP.

3.6.4 Core-Seismic Correlations

The discrepancy between the apparent penetration (clay on the outer core barrel) and actual core recovery is 1 16 m. This discrepancy may be due to
1 Sediment being forced away from the core barrel by the force of core penetration and core acceleration (bypassing, Buckley et al., 1994)

2 Sediment compaction

3 Incorrect scope length The piston did not start moving at the correct time (D J W Piper, pers comm, 1994)

Assuming hypotheses 1 or 3 to be correct the core can be plotted on the seismic column displaced under the sea surface by 1.16 m. The position of the core on the seismic line is shown on Figure 29 although there may be up to 100 m of horizontal uncertainty (Fehr, 1991). The trigger weight core did not deploy with Core 91018-53, hence valuable information about the surface metre of sediment was lost.

The core and seismic stratigraphy can also be matched by correlating a unique event horizon in the core with a recognizable event in the seismic record. The uppermost sandy horizon within lithofacies 5 at 2.3 m can be matched neatly with the top of Sequence H5 (Emerald Silt facies A) consisting of high amplitude reflecting horizons. If this match is made, then the core is displaced under the sea surface by 1.2 to 1.8 m. This is close to agreement with the penetration-recovery discrepancy.

Fehr (1991) tested the sediment-bypassing hypothesis (1) by extrapolating the age-depth curve from a dated core to the zero age of the core assuming a constant sedimentation rate. Extrapolation of the age-depth curve to the zero axis in Core 91018-53 (Fig 32) indicates that the seafloor is 1 m higher than the core surface. This is also consistent with the 1.16 m penetration-recovery discrepancy and indicates that sediment bypassing did indeed occur in the coring process.
Using the previously-defined core placement within the acoustic section, the core seismic sequences, facies, lithofacies and biofacies can be correlated. This correlation is shown on Figure 32. Sequence H6 (LaHave Clay) can be correlated with lithofacies 5, 6 and 7 and Biofacies G-H-I and parts of F (Fig 31). H5 (ES facies A) matches well with the lower part of lithofacies 5 characterized by silt horizons. Biofacies E-F comprise most of Sequence H5. H4 (LaHave Clay seismic facies) is correlated with Lithofacies 3 and 4 and Biofacies C and D. H3 is correlative with Lithofacies 2 and Biofacies B and A. The boundary between Sequences H3 and H4 is marked by a drop in total foraminifera (Biofacies B-probably a response to sediment influx). The lowest part of the core, Lithofacies 1, matches with H2 (ES facies A) and Biofacies A.

3.6.5 "Proto" Moraines

At several locations in the Basin Zone north of the Halifax Moraine, mounds or ridges of Emerald Silt facies C (Table 1) rest on bedrock topographic highs and are draped and surrounded by acoustically laminated sedimentary infill of the Halifax sub-basin. These features are asymmetric in form, steeper on the north (landward) side, 50 to 70 m high and 200-400 m wide (Fig 33). H1 interfingers with the mound on the south side, and overlies it on the north side (Fig. 33). H1 is approximately 30% thinner north of the feature than south. Sequence H2 drapes over the feature. These forms are called "Proto" Moraines. Emerald Silt facies C is interpreted as an ice-proximal, massive, glacial marine diamicton formed by coarse clast deposition from icebergs in association with fines from meltwater plumes at the glacier tidewater terminus (Powell, 1984; Powell and Molnia, 1989; Syvitski, 1991). The position of the former terminus is placed just north of the moraine (Fig. 33). These features differ from lift-off moraines described by (King and Fader, 1986, p. 15) in
Figure 33  Sparker acoustic profiles and interpretation through the Halifax sub-basin (B-C, Fig 13) and close-up of "proto" moraines. The Halifax Moraine is located on the south side of a small basin. The "proto" moraine close-up reveals a zone of acoustically incoherent material, till or deformed glacial marine sediment that separates two sub-basins with varying thicknesses of Emerald Silt. Emerald Silt reflections appear to curve over the feature on the north side and interfinger with it on the south side. This feature is interpreted as a short-lived ice marginal stand where till was formed or deposited through melt-out or deformation.
SEA LEVEL

Figure 33.
two respects lack of symmetry and unequal thicknesses of Emerald Silt on either side of the feature

3.6.6 Local and Regional Correlations

The consistency of stratigraphy between the Halifax and Sheet Harbour sub-basins suggests that the six sequences found in each basin can be correlated one to one. Sequence H7 is an exception but was only resolved in the higher contrast sparker records in the Halifax sub-basin. The uppermost sequences representing Holocene marine deposition (SH6, H6) in both basins are correlative with the LaHave Clay Formation as defined by King and Fader (1986, 1988a; Table 1). The interpolated basal date of 9.2 ka (Fig. 32) agrees well with the basal date of the LaHave Clay (9.4 ka) obtained in the northern Emerald Basin by King and Fader (1988a). In the southwest Emerald Basin, Gipp (1989), Fehr (1991) and Piper and Fehr (1991) defined five sequences designated in this report as SW1-SW5 (Fig. 34). Piper and Fehr (1991) correlated SW4 and SW5 with the LaHave Clay Formation. The age of the base of SW4 is estimated to be 14 ka (Gipp, 1989). The 4000 year age discrepancy between the LaHave Clay in the inner and outer Scotian Shelf has been attributed to the time-transgressive nature of the boundary (Piper and Fehr, 1991). Deglaciation generally proceeded from the southeast to the northwest on the Scotian Shelf, therefore all glacial marine sequence boundaries will be diachronous to some extent. A 4000 year age difference between the stadial/interstadial boundary in two adjacent basins, however, seems excessive. This paradox can be resolved with a reassessment of the correlations. The difficulty in correlation is due to the acoustic similarity between Sequences SW4 and SW5 in the Emerald basin and between Sequences 4 and 6 on the inner shelf. Some of the acoustic similarities between Sequence 4 and 6 may be gain effects, but the lithic and biotic attributes are
Figure 34. Correlation of seismic sequences, facies and formations in the Gulf of Maine and Scotian Shelf. Seismic sequences are numbered 1 through 7. Sequence boundaries are shown as bold lines. Marine inner shelf stratigraphy 1 is from Belknap and Shipp (1991). Maine outer shelf stratigraphy 2 is from Bacchus (1993). Generalized Scotian Shelf stratigraphy 3 is from King and Fader (1986). Emerald Basin seismic stratigraphy 4 is from Gipp (1989) and Piper and Fehr (1991).
MAINE

NOVA SCOTIA

INNER SHELF

OUTER BANKS

SCOTIAN SHELF

GENERALIZED

EMERALD BASIN

INNER SHELF

FACIES LEGEND

LaHave Clay-Marine facies

Distal Glaciomarine

ES facies B-
Transitional Glaciomarine

ES facies A-
Proximal Glaciomarine

Stratified sediment

ES facies C-
massive glaciomarine

Acoustic Basement

LaHave Clay

Truncation Event

Yankee Bank

Scotian Shelf Drift
analogous. The abrupt and distinctive break in the acoustic record marked by Sequence 5 dated between 10.7 ka and 9.4 ka is correlative with an unconformity at the top of the Emerald Silt Formation, in the northeastern Emerald Basin, dated between 10.0 and 10.5 ka. The boundary between SW4 and SW5 in the southwest Emerald Basin dated at 10.2 ka is a single strong reflection that Piper and Fehr (1991) attributed to the Younger Dryas climatic event. Sequence 5 (inner shelf) characterized by Emerald Silt facies "A" with high amplitude reflections, has not been described in the southwest Emerald Basin, but is a marker horizon on the inner shelf. The author proposes that the lower boundary of the LaHave Clay Formation be designated as the top of Sequence 5 in the inner shelf and its correlative unconformity (SW4/SW5 boundary) in the outer shelf basins. The Emerald Silt Formation should be redefined as Sequences 1 through 4 in both areas comprising seismic facies A, B and C. Sequence 5 is regionally mappable, has unique seismic, lithic and biologic characteristics and should be conferred with formational status. I propose the name Yankee Bank Formation (the nearest geographical name to the type section) to apply to Sequence 5 - Lithofacies 5 - Biofacies E-F, bounded on the bottom by the Emerald Silt Formation and on the top by the LaHave Clay Formation (Table 6).

The Yankee Bank Formation can be correlated with event horizons in other cores along the inner shelf. Marsters (1988) described a core (77-15) off Mahone Bay (see also Piper et al., 1986) with an oscillation in foraminiferal assemblages from an ice-marginal fauna (Elphidium-Cassidulina) (5 4-3 7 m) to a zone with low foraminiferal counts (2 5-3 m) and then to a Post-Glacial assemblage (0-2 m) characterized by increasing numbers of Islandella teretis. Marsters (1988) interpreted the "barren zone" as a period of rapid deposition. The similarity in biostratigraphy and depth suggests that the "barren zone" is correlative with the Yankee Bank Formation. The seismic record of this core site does not resolve any distinct
<table>
<thead>
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<th>SEQUENCE</th>
<th>FORMATION</th>
<th>INTERPRETATION</th>
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<tr>
<td>6</td>
<td>LaHave Clay</td>
<td>Holocene marine clay</td>
</tr>
<tr>
<td>5</td>
<td>Yankee Bank Formation</td>
<td>&quot;paraglacial&quot; deposit</td>
</tr>
<tr>
<td>4</td>
<td>Emerald Silt Formation</td>
<td>Glacial marine deposition-distal</td>
</tr>
<tr>
<td>3</td>
<td>Emerald Silt Formation</td>
<td>Glacial marine deposit-proximal</td>
</tr>
<tr>
<td>2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Scotchian Shelf Drift</td>
<td>Till</td>
</tr>
<tr>
<td></td>
<td>Acoustic Basement</td>
<td>Bedrock-Meguma Group metasediments</td>
</tr>
</tbody>
</table>

Table 6. Seismic sequences, Quaternary Formations, their correlation and interpretation for the study region.
horizon, but is of poor quality (Piper et al., 1986). Souchen (1986) described a core in St. Ann's Basin north of Banquereau with a turbidite horizon between muddy sediment facies with a Late-Glacial faunal assemblage and a Post-Glacial normal-salinity assemblage. In this case an excellent seismic record revealed a distinct acoustic horizon characterized by high amplitude reflections on top of Emerald Silt facies "B" and below the LaHave Clay. On Banquereau, Amos and Miller (1990) interpreted a conformably-bedded sand unit dated at 11 ka as a proglacial unit based on the presence of an ice marginal foraminiferal fauna.

Belknap and Shipp (1991) divided the inner shelf glacial marine sediments in the Gulf of Maine into ponded and draped units. Bacchus (1993), working in deeper water, used unit subdivisions based on seismic reflection configurations equivalent to seismic "facies" (Table 7). Figure 34 is summary correlation diagram of formations, seismic sequences and seismic facies within the inner and outer Scotian Shelf and Gulf of Maine basins. Each region displays a general progression from Drift (massive-glacial marine) to Emerald Silt facies A and facies B (proximal to distal glacial marine) (cf. Syvitski, 1991). The distal glacial marine unit of Bacchus (1993) is a weakly stratified, ponded unit, similar to the uppermost seismic sequence (4) of the Emerald Silt which varies from acoustically transparent to weakly stratified. It may be correlative with the lower portion of the LaHave Clay or the uppermost Emerald Silt Formation (Fig. 34). Separating the transitional and distal glacial marine facies in the Gulf of Maine is a strong double reflection which has been termed the Truxton Event (G. Fader pers. comm., 1992). In some areas in the Gulf of Maine this horizon is series of high amplitude reflections. Cores through this unit revealed a distinctive, laminated, red-brown, gravelly-sandy mud (Bacchus, 1993, p. 172). The "Truxton Event" in the Gulf of Maine was estimated to be between 14.4 and 12.5 ka based on amino acid racemization dating but the method has a 2000 year uncertainty (Bacchus, 1993). Possible
<table>
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<th>Description</th>
<th>Interpretation</th>
<th>Designation</th>
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<tr>
<td>6</td>
<td>Acoustically transparent</td>
<td>Postglacial</td>
<td>M</td>
</tr>
<tr>
<td></td>
<td>Weakly stratified, conformable to unconformable</td>
<td></td>
<td>DGM</td>
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<td>Transitional</td>
<td>TGM</td>
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<tr>
<td></td>
<td>reflectors</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Densely stratified, conformable internal</td>
<td>Ice-proximal</td>
<td>PGM</td>
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<tr>
<td></td>
<td>reflectors</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Dense chaotic returns</td>
<td>Till</td>
<td>T</td>
</tr>
<tr>
<td>1</td>
<td>High intensity surface returns</td>
<td>Bedrock</td>
<td>B</td>
</tr>
</tbody>
</table>

Table 7. Seismic facies and interpretation from the Gulf of Maine (after Bacchus, 1993).
correlative horizons in the inner shelf study area are the Sequence 3/4 boundary and the Yankee Bank Formation (Fig. 34) Syvitski (1991) compiled the stratigraphic sequences from twenty glaciated shelves around the world and found a common theme linking many of the sections. His "composite section" consists of the following units summarized into genetic categories:

1. Till, drift or ice contact sediments
2. Glacial marine sediment-proximal
3. Glacial marine sediment-distal
4. Paraglacial prodelta muds
5. Post-glacial basin fill

The first three genetic units are comparable to the sequence and facies progressions in the Nova Scotia inner shelf and the Gulf of Maine. The "paraglacial" unit consists of deltaic wedges and may be more prominent in shelf areas close to the lowstand shoreline (see Truncation Zone).

### 3.6.7 Depositional Processes

Sequence boundaries as defined have no genetic connotations but are purely descriptive (Mitchum et al., 1977). In models developed by Vail et al. (1977) sequence boundaries are considered to result from varying eustasy, tectonics and sediment supply producing changes in relative sea-level (RSL). The basins in the inner shelf were always below sea-level, hence the sequence boundaries must have formed from marine or glacial erosion and submarine onlap. Sea-level fall may have indirectly played a role in regions of the inner shelf basins. As
sea level falls, storm wave reworking may become an increasingly important erosional agent. RSL fall may have been sporadic rather than continuous, responding to advances of ice, and meltwater loading (Belknap et al., 1987, Stea et al., 1994). Resuspension by turbidity, tidal and storm wave-generated currents are probably the main mechanisms generating lapout boundaries and ponded geometries (Barrie and Piper, 1982). Erosional truncation surfaces can be produced by currents, density flows and by direct glacial erosion. Wedge-shaped sediment geometries are indicative of tidal currents (Piper et al., 1983).

3.6.7.1 Sequence 1

Sequences 1 and 2 consist of Emerald Silt facies A and C and are thickest at the southern end of the basins nearest the moraines (Figs. 26, 28). It is assumed that they have common lithic characteristics with Emerald Silt facies A and C in the deeper basins south of the Scotian Shelf Morainal Complex (Table 1). In this study, Emerald Silt facies A was only sampled on erosional surfaces that were stripped of overlying seismic facies. These samples were clay-rich but contained a significant amount of gravel (23.8% in IKU 51-Fig 4, Appendix 7). Much of this gravel, however, may have been winnowed from overlying sediments by the erosional processes responsible for the unconformity.

Seismic facies A and C are characterized by draped geometries. Reflections traceable for tens of kilometres are unique to facies A. Two conflicting interpretations have been proposed to explain the properties of these acoustic units.

2 Suspension settling from overflow and interflow plumes away from a stationary ice front (Oldale et al., 1990)

It is generally agreed that a draped geometry results from rapid suspension fallout and high sedimentation rates (Barrie and Piper, 1982, Belknap and Shipp, 1991, Syvitski, 1991). Models 1 and 2 both produce draped geometries if the sedimentation rate is rapid enough. Facies "C" without bedding, has rarely been sampled and is interpreted either as a grounding line proximal deposit or a till (King and Fader, 1986). Rhythmic bedding can be produced by episodic deposition enhanced by low water temperatures which delay settling of fines but not coarse sediments (Cowan and Powell, 1991). Episodic deposition can also be explained in both models by periodic changes in sediment delivery, by tidal action, and seasonal (varve-like) variation in sediment supply (Phillips et al., 1991). The continuity of glacial marine sediments can be explained by both models.

Powell (1984), Oldale et al. (1990), Anderson et al. (1991) and Syvitski (1991) argued that ice shelves cannot form in temperate zones. They presented these cogent arguments against ice shelf formation:

1. There are no modern examples.
2. Warm water under ice shelves would rapidly cause their demise by undermelt and enhanced calving.
3. Cold ice is required because ice at the pressure melting point has lower tensile strength and experiences easier fracture propagation.
4. Ice shelves could not withstand the onslaught of open-sea storms with breaking waves.
These arguments have limited validity because they are based on modern, not ice age environments. An ice shelf was postulated in the Gulf of Maine based on the lack of planktonic foraminifera (Schnitker, 1975) and diatoms (Schnitker and Jorgensen, 1986) in glacial marine sediments. Diatoms are common in ice marginal environments today and absence implies light suppression. Scott et al. (1984) proposed that an ice shelf existed in the Notre Dame Channel off Newfoundland based on the presence of the deep water estuarine foraminifer *Fursenkoma fusiformis*. Belknap and Shipp (1991) countered the anti-shelf arguments 2, 3 and 4 by suggesting that Gulf of Maine ice shelf was fed by fast moving ice streams and buttressed by numerous pinning points. Wide, shallow banks in the Scotian Shelf may have attenuated storm waves (Argument 4). Gipp (1989) argues for a short-lived ice shelf in the Emerald Basin based on these properties of the lowest seismic sequence (0):

1. Sparse benthic foraminifera (8 tests/g)
2. Lack of evidence of current or storm wave resuspension
3. The absence of in-situ molluscs (cf Powell, 1984)
4. The absence of buried ice-berg scours

The most likely pinning points of a hypothetical ice shelf on the Scotian Shelf would be the outer banks. King and Fader (1986, p. 55) proposed that the Scotian Shelf End Moraine Complex (SSEMC) was formed underneath the ice shelf, at a sub-shelf pinning point, while till-tongues and interfingering glacial marine sediment formed simultaneously on the north and south sides of the moraines. In this study, till-tongues were found only on the deep water sides of the SSEMC. If the ice shelf was pinned on the outer banks and Emerald Silt formed simultaneously with the SSEMC, then sediment thicknesses should be approximately equal in basins north and south of the SSEMC, or perhaps thicker on the northern (glacier-
proximal) inner shelf basins. Sequence 1, however, markedly thins towards the north in the inner shelf basins (Figs 26, 28). Seismic records show a diachronous, onlap relationship between the moraines and the inner shelf basin sediments. The "proto" moraines which separate sub-basins with varying thicknesses of Emerald Silt are interpreted as ice marginal stands. Ice-rafted boulders within Sequences 1 and 2 also lend support to a proximal-ice marginal setting rather than sub ice-shelf origin. The lack of iceberg scours in the basin sediments of the inner shelf can be related to water depths. Iceberg furrowing seems to be restricted to the top surface of the SSEMC in the study area in water depths less than 170 m (Fader, 1991). Sequence 1 strata were deposited during sequential retreat and slight readvances of the ice margin from various pinning points represented by the "proto" moraines.

3.6.7.2 Sequence 2

Sequence 2 can be distinguished from Sequence 1 in the Halifax and Sheet Harbour sub-basins by reflections of higher amplitude and continuity. It can be traced northward as far as the Morainal and Outcrop zones (Fig. 7). The synchronous, basin-wide extent and onlapping nature of this unit suggests deposition from overflow-interflow meltwater plumes by suspension fall-out followed by wave resuspension when ice was somewhat removed from the region. The strong reflections imply impedance contrasts between mud and sand and/or gravelly-sand strata. The unit may represent an enhanced period of melting. This sedimentological interpretation is corroborated by biological data from core 91018-53. Sequence 2 records an ice marginal *Elphidium-Cassidulina* fauna (Fig 32).
3.6.7.3 Sequences 3-4

Sequences 3 and 4 are made up of Emerald Silt facies B and LaHave Clay seismic facies and are correlative with Biofacies B-D (Figs 32, 34, Table 1). Dominance of *Elphidium* within Biofacies B-D suggest an ice marginal environment (Scott *et al.*, 1984), but Biofacies B-D also exhibits an increasing up-section trend to a cold water, normal salinity fauna similar to the lower part of the LaHave Clay (Fig. 32). In the Newfoundland shelf foraminiferal assemblages associated with the upper part of Emerald Silt facies B could not be easily differentiated from the lowest parts of the LaHave Clay (Scott *et al.*, 1984). Correlative transitional glacial marine units in the Maine inner shelf (Fig. 34) have a similar lithology to the Presumpscot Formation on land (Belknap and Shipp, 1991). The onlap depositional style is due to resedimentation by currents and wave action (Barrie and Piper, 1982, Piper *et al.*, 1983, Gipp, 1989, Belknap and Shipp, 1991) or from reduced rates of deposition (Vorren *et al.*, 1990). Asymmetric thickness distributions in the Basin Zone suggest wedging by tidal currents (Gipp, 1989). Higher percentages of point-source diffractions (boulder or boulder-dumps), higher-amplitude reflections and increased sediment thicknesses in a seaward (ice-distal) or southward direction (Fig. 28) suggest that Sequence 3 was deposited closer to a glacial terminus than Sequence 4. The frequency of point-source diffractions implies that iceberg rafting played an important role in the deposition of this unit (Belknap and Shipp, 1991, Bacchus, 1993). In the Halifax sub-basin, Sequence 4 has a uniform thickness, and a low frequency of point source diffractions suggesting that the ice source had retreated north of the area, possibly onto the land areas (Fig. 28). The boundary between Sequences 3 and 4 in the Halifax sub-basin is marked by erosional truncation, several high amplitude reflections, a drop in total foraminiferal count and species composition (Biofacies B; Fig. 31). The increased sedimentation rate and erosion at the sequence boundary may be due to:
1. Falling sea levels and increased influence of storm waves with suspension and transport of material from shallower areas to the Basin Zone

2. Glacier readvance, sea-level rise and increased sedimentation and current erosional activity due to the proximity of a grounding line. A sequence boundary may be formed by subglacial erosion, in the case of truncation, or by proglacial erosion and by slump-driven turbidity currents adjacent to the grounding line. Renewed crustal depression by advancing ice forms basin-wide sequence boundaries by changing the accommodation space available for sediment fill.

Hypothesis 1 is favoured because sea level had dropped to its lowest point (−70 m) at about 116 ka (see Stea et al. 1994). Sequence 3/4 boundary occurs in water depths from 140 to 170 m, 70 to 100 m below the inferred lowstand. As stated in the introduction to this chapter, Forbes et al. (1991) observed that combined wave-current activity can entrain silts to depths of 100 m. Erosion and deposition during the lowstand may have also been enhanced by the proximity of the paleo-shoreline to the Basin Zone.

Hypothesis 2 is less likely, because at the time of formation of the Sequence 3/4 (117 ka) boundary, ice had retreated from the land areas of Nova Scotia (Stea and Mott, 1990, Map 1).

### 3.6.7.4 Sequence 5

Sequence 5 is composed of conformably-bedded, Emerald Silt facies "A", marked by an influx of coarse material and a drop in total foraminiferal counts. Two hypotheses put forward to
explain the sequence 3/4 boundary are also applicable to Sequence 5. Hypothesis 1 (sea-level change) is unlikely because at the time of Sequence 5 deposition (10.8 ka) sea-levels had already dropped to their lowest levels (11.5 ka) and were rising (Stea et al., 1994). A glacial marine environment (Hypothesis 2) may be indicated by the *Elphidium*-dominated foraminiferous assemblage (Fig. 31) but the absence of *Cassidulina* is enigmatic as the two forms are normally found together in these Pleistocene glacial marine environments (Scott and Medioli, 1980a; Vilks, 1981; Seidenkrantz, 1993). Lowered salinities may have excluded *Cassidulina* (D. B. Scott, pers. comm., 1995). Diatom fluxes within Biofacies E-F may point to increased winter sea ice cover (Kennett, 1982). The ice source probably was not a calving margin because point source diffractions (boulders) are lacking. Sequence 5 thickens towards the south, away from the extensive Younger Dryas marine ice margin proposed by King (1994) and the less extensive terrestrial margins postulated by Stea and Mott (1989; Map 1). Erosional surfaces synchronous with Sequence 5 (Fig. 27) did not form by ice advances because lake sediment records indicate adjacent land areas were cleared of ice by 12 ka and Younger Dryas ice was restricted to northern Nova Scotia (Stea and Mott, 1989; 1990; Map 1).

I interpret this event horizon as a response to a combination of increased storminess (cf. Gipp, 1989), debris-laden sea ice and increased sediment supply from the devegetated coastline. The coastline was about 20 km seaward of present about that time (Stea et al., 1994). Permanent and semi-permanent snowpacks and small glaciers, terminating in shallow water provided plumes of low-salinity, sediment-laden water which may have spread over much of the inner shelf (cf. Map 1). Terrestrially-derived organic debris may have been borne by sea ice.
If correlations with the outer bank turbidite unit of Souchen (1986), the conformably-bedded sand of Amos and Miller (1990) and Gulf of St. Lawrence "paraglacial" units of Syvitski (1991) are correct then a widespread climatic event is implicated. Ice advances have been documented in the land areas adjacent to the Gulf of St. Lawrence (Stea and Mott, 1989, Grant, 1989, LaSalle and Shilts, 1993).

3.6.7.5 Sequences 6 and 7

A major sequence boundary can be resolved in the LaHave Clay Formation at the Halifax sub-basin type section (Fig. 27). This boundary may be a response to changing climates and sea-levels during the mid-Holocene hypsithermal interval.

3.7 MORAINAL ZONE

3.7.1 Introduction

King et al. (1972) first described morainal ridges on the inner shelf north of the Scotian Shelf End-Moraine Complex. LaPierre (1985) also identified transgressed till-cored ridges on the inner shelf off Petpeswick, Nova Scotia using sparker seismic profiles, sidescan sonograms and grab samples. Fader et al. (1991) and Stea et al. (1992b, 1993) described a zone off Sheet Harbour characterized by morainal mounds and ridges. The Morainal Zone is located 40 km northeast of Halifax along the Eastern Shore (Fig. 7, 8). Huntec seismic profiles reveal a unique acoustic morphology characterized by steep, irregular mounds and ridges 2-30 m in height and 100 to 300 m wide. They are scattered throughout a shore-parallel zone in water depths of 65 to 145 m. Continuous, coherent reflection traces cannot be resolved within the
landforms, they are acoustically homogeneous. Sidescan sonograms of the mounds reveal high intensity backscatter (gravely-bouldery) areas of irregular pattern and ridge shapes (Fig 35). Large boulders (1-4 m diameter) can be seen scattered randomly across the seabed. These boulder patches occur in ribbons on the sonograms and appear to be uniquely associated with the moraines. These landforms were originally interpreted to be glacial in origin, either drumlins or moraines (Fader et al., 1991).

The ridge morphology of the landforms in the Morainal Zone was defined by sidescan sonograms (Fig 35), mapping of closely-spaced tracklines (Fig 36) and a digital terrain model developed from echograms of the Canadian Hydrographic Service (Fig 8). Interpolation between widely-spaced tracklines was based on sea-bed depth contours obtained from the digital grid model (Fig 16, Map 2). Many of the Morainal Zone ridges are oriented northeast-southwest, parallel to the northeast-southwest orientation of the Scotian Shelf End-Moraine Complex. In the northern part of the Morainal Zone ridge orientations change from NE-SW to NW-SE (Figs 16, 36).

Five main morphologic and stratigraphic categories of ridge-landforms were discerned from the seismic and sidescan sonar records:

1. Till-Tongue Moraine
2. Symmetrical Moraine
3. Hummocky and "Doughnut" Moraines
4. Bedrock-Controlled (Ground) Moraine
5. Asymmetrical Moraine
Figure 35  Sidescan sonograms of "doughnut" (A) and ridge-like (B) landforms in the Morainal Zone. The scan width of the sonogram is approximately 400 m. The morainal segments in the sonogram are 20-40 m wide and 300-400 m long. Some of the moraine segments are buried under LaHave Clay, so are longer than they appear.
Figure 36  An expanded view of a region of the Sheet Harbour Transect area (see Figure 16 for location) showing mapping details of closely-spaced tracklines. Patterned areas are morainal ridge segments mapped using Huntec profiles and sidescan records. The intervening white areas are underlain by LaHave Clay. Note the change in orientation of the ridges in the northern part of the transects from E-W to NE-SW.
3.7.2 Till-Tongue Moraine

Immediately north of the Sheet Harbour sub-basin (Fig 16, Map 2) are several mounds of acoustically-incoherent material equivalent to the Scotian Shelf Drift (Table 1). The largest feature, found in 124-135 m water depth, is a steep, asymmetrical ridge with 30 m of sea-bed relief connected to a lower, muted ridge (Fig 37). The ridges are composed of material lacking in coherent internal acoustic reflections. A reflection representing acoustic basement can be recognized underneath the landform. A till-tongue is rooted in the muted ridge and projects into Sequence SH1 where it terminates in a feather edge. Sequence SH2 appears to drape over the tongue and is not found on the landward side of the moraine. Sequence SH1 decreases in thickness from 11 m on the seaward side of the feature to 6 m on the landward side. Sidescan records show a linear, high-intensity backscatter zone suggesting a gravel-boulder cover and ridge morphology. These moraines are essentially smaller versions of the larger Scotian Shelf End Moraines and are transitional to the Morainal Zone in shallower water. The main morainal fragment is likewise interpreted as formed by accretion of basal debris at a tidewater margin and the tongues by forward movement in the ice front which was halted by buoyancy forces.

3.7.3 Symmetrical Moraine

The most common type of mound-ridge identifiable in the Huntec profiles has been termed "Symmetrical Moraine". These are defined as mounds or ridges with symmetrical sides consisting of acoustically-incoherent material (Fig 38). These are commonly found in water depths of 85-135 m (Fig 39). The symmetry of some of these landforms is quite remarkable (Fig 38). The shapes vary, however, from sharp-crested forms to those with...
Figure 37. Huntec acoustic profile and interpretation of the "Till-Tongue" Moraine in the Sheet Harbour transect area (see location F on Fig. 16). Water depth in m on the left.
SEISMIC RECORD

INTERPRETATION

- LaHave Clay
- Emerald Silt (Facies B)
- Emerald Silt (Facies A)
- Scotian Shelf Drift
- Bedrock

Figure 37
Figure 38. Huntac acoustic profile (external) and interpretation of symmetrical moraines in the Sheet Harbour transect area. Section G-H is located on Figure 16. Inter-moraine seismic sequences M1 and M2 are described in the text. Acoustically incoherent material within the Morainal Zone has been designated as Scotian Shelf Drift after King and Fader (1986; Table 1).
SEISMIC PROFILE

Depth below water surface (m)

INTERPRETATION

Scotian Shelf Drift (Unit 3?)

ACOUSTIC BASEMENT

Figure 38.
Figure 39  Notched box and whisker plots of the water depth relationships of the various morphological types of moraines  The box and whisker plot is an effective way to demonstrate summary statistics graphically  The notched plot (McGill et al 1978) corresponds to the confidence interval for the median, whereas the width of the box is proportional to the square root of the number of observations in the data set  If the two notches on the boxes do not overlap (hatching) the median values of the data distributions do not significantly overlap at a 95% confidence level
Depth Relationships of Moraine Types, Eastern Shore Region.

Figure 39
plateaus on top. The seabed surfaces of these features exhibit moderate to high reflectivity. A horizontal reflection representing acoustic basement can be traced through many of the landforms although some appear to have a "shadow zone" that is propagated downwards from the sea surface and virtually eliminates all information below. This zone often appears to be best developed in the most symmetric forms (Fig. 38). This "shadow zone" may be due to the dissipation of acoustic energy by a concentration of boulders on the moraine crests and/or energy scattering on the steep sides. The intervening regions between ridges and mounds contain two acoustic sequences (Sequences M1, M2, Fig. 38). Sequence M1 consists of low amplitude, draped reflections truncated by an onlap fill sequence (M2) of moderate to high amplitude reflections. In deeper water areas a sequence consisting of LaHave Clay seismic facies (M3) infills inter-moraine depressions. The lateral relationships between Sequence M1 strata and the moraines are uncertain. In some cases reflections appear to lap onto the moraines while in other areas a gradational contact is observed.

Late in this thesis study, multibeam bathymetric imagery became available (Costello et al., 1993). A region of the inner shelf within the sampling area east of Lunenburg was surveyed (Stea and Fader, 1993, Loncarevic et al., 1994, Figs 1, 40). The image reveals three main topographic zones, the eastern area characterized by irregular topography, the middle region by closely-spaced, linear ridges and the western region dominated by crenulated, curvilinear ridges with intervening flat, basin regions. These ridges generally trend northeast-southwest and are regularly spaced. Huntec profiles across the ridges (Fig. 40) show symmetrical moraines 4-18 m high. Many of these moraines are found on bedrock highs. The linear ridges in the central region were found to be bedrock-controlled forms. A later survey by the NSC Frederick G. Creed (John Hughes-Clarke, pers. comm. 1993) over the moraines included camera reconnaissance and grab samples. The moraines were composed of a
Figure 40. Multibeam sonograph image of a region southeast of Lunenburg, Nova Scotia (Fig. 1) and interpretation. The sonograph area shows two types of ridges; anastomosing, curvilinear, ribbed moraines and straight bedrock ridges. The Huntec profile in the upper left corner is an acoustic cross section of a typical ribbed moraine in the sonograph area (from D. J. W. Piper, pers. comm, 1993). Huntec records across these features show that they are identical to symmetrical moraines (Fig. 38) and relate to topographic highs composed of acoustically incoherent material, underlain by a reflection representing acoustic basement.
matrix-supported, sandy-mud diamict, compared with silty-mud in the intervening flat regions.

### 3.7.4 Hummocky and "Doughnut" Moraines

Another moraine type consists of ridge forms or mounds of acoustically incoherent material with two or more minor mounds superimposed on the upper face. These are termed hummocky moraines. On sidescan sonograms some of these hummocky forms appear as closed loops of high backscatter regions with boulder shadowing ("doughnuts", Fig. 35 A).

### 3.7.5 Bedrock-Controlled Moraine

Some bathymetric highs consisting of acoustically incoherent material appear to be conformable with the undulating trace of acoustic basement interpreted to be the bedrock surface. These features are called bedrock-controlled moraine.

### 3.7.6 Asymmetrical Moraine

Moraines in the Sheet Harbour transect area in water depths less than 82 m, and not less than 65 m (Fig. 39) have a distinctly asymmetric shape, with a steeper, landward side (Figs. 16, 39, 41). The external Huntec profile resolves indistinct clinoform beds on the seaward side of some of the asymmetrical moraines that are truncated by the landward slope. The intervening areas between ridges consist of more transparent, incoherent acoustic material (M2-Fig. 41). These moraines are similar in scale and morphology to terrestrial moraines in Nova Scotia (Fig. 41).
Figure 41. Huntec seismic profile and interpretation of asymmetrical moraine ridges in the Sheet Harbour transect area (Location I- Fig. 16). The bottom figure is a profile of terrestrial ribbed moraines in the vicinity of Cobrielle Lake, southern Nova Scotia (Stea et al., 1992a), plotted using the same vertical exaggeration as the Huntec records.
SEISMIC PROFILE

INTERPRETATION

Metres below sea level

60
70
80
90

Scotian Shelf Drift (Unit 3?)

ACOUSTIC BASEMENT

TERRESTRIAL MORAINES, SOUTHERN NS

Metres above datum

0  200 m

Beaver River Till

Figure 41.
3.7.7 Lithology

Six samples were obtained from ridges or mounds north of the Scotian Shelf End-Moraine Complex (IKUs 24, 25, 29, 49, 59, 64 - Appendixes 3, 4) Only one sample site produced complete recovery (IKU 29), the rest contained lag gravel (cobbles-pebbles) and lumps of olive-grey, matrix-supported, sandy-mud diamicton. It is assumed that the lumps are bits of the "true" moraine material from underneath a lag armour of cobbles and boulders.

The Morainal Zone grab samples contain subangular to subrounded clasts, with 8-47% striated, averaging 15% (Table 4). IKU 29 shows a significant increase in percentages of striated pebbles from the top to bottom subsamples (17%-Top, 47%-Bottom, Appendix 4). Sample 64, from 64 m water depth, contains subrounded to rounded clasts, none of which are striated. Morainal Zone samples vary from 82 to 92% Meguma Group metasediments and average 86% (Table 4). They consist of an average of 45% gravel and 55% matrix weight percentages (Table 4).

3.7.8 Truncated Emerald Silt Subzone (TESS)

Another feature of the Morainal Zone is a large (approx 900 sq km) area of Emerald Silt outcropping at the seabed, truncated by planar erosional surfaces (Fig 8). This extensive area of erosion is found in water depths shallower than 140 m, at the contact between the Morainal and Basin Zones off Sheet Harbour, and widens to the northeast towards the Country Harbour Moraine (Figs 8, 42). Southeast of Country Harbour the surface is remarkably flat and featureless (Fig 19). Off Sheet Harbour, Sequences 1 and 2 (Emerald Silt facies A and C) are found at the seabed at water depths of 130-150 m (Fig 42).
Figure 42  Huntec acoustic profile of a region of eroded Emerald Silt 30 km south of Sheet Harbour. Vibra-core 91018-67 was obtained from this area, and revealed an olive-brown, sandy diamicton overlying a clayey-silt and silty-clay. Seismic sequences, shown on the left of the acoustic profile, are extrapolated from the type section (Fig. 25).
Figure 42
Sequences 3 and 4 (Emerald Silt facies B) lap out on the earlier units before the TESS surface. Several channels were cut into the Emerald Silt and later infilled with LaHave Clay. At several locations an onlap unit (Emerald Silt facies B) is truncated by the erosional surface. King and Fader (1983a) correlated the erosional surface south of the Country Harbour Moraine to an unconformity at the top of the Emerald Silt Formation dated at 10.8 ka.

3.7.8.1 Lithology

An IKU sample from this subzone in the Sheet Harbour transect region (IKU 57) recovered an olive-grey, muddy, matrix-supported diamicton with subangular to subrounded clasts. Eighteen percent of the clasts are striated and several have distinct, bullet-like shapes with distal facets indicative of erosion in the basal traction zone of a glacier (Krüger, 1984; Clark and Hansel, 1989). Eighty-seven percent of the clasts are derived from Meguma Group metasediments, 5% from Meguma Zone granitoids, and 8% are foreign to the Meguma Zone. Core 91018-67 (Fig. 42) taken on the TESS south of Sheet Harbour revealed an olive-brown, sandy diamicton over a grey cohesive clay with few pebbles. IKUs 85, 89 and 93 were obtained from the Emerald Silt on the erosional surface in the Country Harbour Transect region (Appendices 3, 4, Fig. 18). IKUs 89 and 93 consisted largely of a grey, massive, silty clay. IKU 85 revealed a more complex stratigraphy, with a cobble lag overlying an olive-grey, sandy diamicton and a greyish mud with a few stones. Samples 89 and 93 contained only a small percentage of striated stones (2-5%).
3.7.9 Age of the Morainal Zone

In the deeper parts of the Morainal Zone, Sequence M1 (Emerald Silt facies B) onlaps the symmetrical moraines. Sequence M1 correlates with either Sequence 3 or 4 in the Basin Zone which were dated between 12.75 and 10.7 ka (Fig. 32). The Till-Tongue Moraine interfingers with Sequence 1. Therefore, the Morainal Zone formed between 14.0 ka (interpolated base of Sequence 1) and 10.7 ka (top of Sequence 4) (Fig. 32). Sequence 2 can be traced as far landward as the distal side of the Till-Tongue Moraine and formed sometime after the moraine. In the Halifax sub-basin the top of Sequence 2 is dated at 12.75 ka (Fig. 32).

The TESS seabed surface (Fig. 42) appears to have formed during or after deposition of Sequences 3 and 4 (Emerald Silt facies B). Sequences 3 and 4 lap onto the TESS surface (Fig. 42) in the Sheet Harbour transect area. However, off Country Harbour, erosional truncation of Emerald Silt facies B has clearly occurred (Fig. 19). King and Fader (1988 a, b) and King (1994) proposed that the TESS surface was formed by a Younger Dryas ice advance. In the northeast Emerald Basin they correlated the erosional surface with the Sequences 4/6 boundary dated at 10.8 ka.

Sea-levels reached their lowest point along the inner shelf (-65 to -70 m) around 11.6 ka after ice had retreated landward of the present coastline (Stea et al., 1994). Moraines or any other unmodified glacial landforms do not appear above the -65 m level, so it is assumed that they were destroyed by the post-glacial transgression. The last glaciation therefore pre-dated 11.6 ka across much of the inner shelf. By this reasoning the Morainal Zone formed sometime between 14.0 ka (deposition of Sequence 1 in the Basin Zone by the retreating glacier) and
11.6 ka, the time of the lowstand. Sequence 2 is interpreted as the depositional response to the glacial advance that formed the Morainal Zone; so the time of formation of the Morainal Zone may be further bracketed between 13.5 ka and 12.75, the duration of Sequence 2 (Fig. 32). These delicate morainal features probably represent the last glacial advance to have affected the inner shelf.

3.7.10 Genesis of the Morainal Zone

Surface samples from the Morainal Zone ridges are composed of an olive-grey, matrix-supported diamicton. In the Morainal Zone diamicton samples, Meguma pebbles average 86% whereas foreign pebbles average only 8% (Table 4). The gravel fraction averages 45% and the matrix 55%. These diamicton samples average 15% striated clasts. This diamicton is similar in lithology and seismic character to the uppermost unit (3) of the Scotian Shelf Drift of the main morainal complex (Fig. 41, Tables 1, 4). Morainal Zone ridges are similar in scale and morphology to land ribbed moraines (Fig. 41) and I therefore interpret these constructional landforms as ice-formed. The term "moraine" in the marine environment is defined as "positive morphological features in a glacially-dominated region" (Von Haugwitz and Wong, 1993). DeGeer Moraines are defined as low-straight to arcuate ridges of loose, sandy diamicton or sand/clay formed parallel to the former ice cliff edge (Goldthwait, 1989). Theories about the formation of DeGeer Moraines fall into two main groups:

**Group 1** The moraines represent successive, annual glacier-marginal stillstand or advance features with chronological significance (Smith, 1982; Larsen et al., 1991)
Group 2 They are formed simultaneously by till-squeeze or meltwater inflow into basal crevasses some distance behind the ice front, and have no chronological significance (Elson, 1957, Hoppe, 1959, Zilliacus, 1989, Lundqvist 1989, Beaudry and Prichonnet, 1991)

The definition of DeGeer Moraine may be too broad to be useful for genetic classification. For example, Elverhøi et al (1989) document two types of DeGeer Moraine resulting from a surge of the Bråsvallbreen glacier in 1936, "squeeze up" ridges (Group 1) and annual "push" moraines (Group 2). Group 1 moraine formation is indicated by these features:

1. Parallel and linear ridges
2. Perpendicular to the regional ice flow direction
3. An asymmetric cross-sectional profile
4. A complex sedimentology, glaciotectonic deformation, and inter-fingering relationships with adjacent sediments

Group 2 moraines are more likely to display a simpler stratigraphy (i.e. till), and symmetric external form, but may be more variable in plan view as are the crevasse patterns in surging, marine glaciers. The surging Bråsvallbreen glacier produced a rhombohedral pattern of crevasse-fill moraines (Elverhøi et al, 1989).

Parallel ridges ("Lift-off" Moraines) have been described in the basins south of the Scotian Shelf End Moraine Complex (King and Fader, 1986, p. 15, Gipp, 1989). Lift-off Moraines vary in height from a few metres to 20 m and are composed of acoustically incoherent material overlain and interfingering with Emerald Silt facies A. King and Fader (1986)
proposed that Lift-off Moraines form simultaneously when buoyant ice lifts off the seabed and till is injected into subglacial fractures. Lift-off Moraines fall into the Group 2 genetic category.

The Morainal Zone moraine types fall within both of the two broad genetic categories. The till-tongue moraine of the Basin Zone is similar to the Scoxian Shelf End Moraines which fall within Group 1. The "asymmetrical" moraines may also fall into this genetic category. The most abundant type, "symmetrical" moraines belong in Group 2. They appear to be identical in form to "Lift-Off" moraines without the cover of overlying Emerald Silt. The lack of Emerald Silt (facies A) between symmetrical moraine ridges in the Morainal Zone favours a simultaneous, subglacial origin, rather than as marginal features (Fig. 43).

The Truncated Emerald Silt Subzone (TESS) is found in similar water depths to the Morainal Zone (Figs 8, 16, 18). Emerald Silt facies A, C and B have been eroded. Formation of the TESS by surf-zone erosion is ruled out because of the depth of the erosion (130-150 m; see Truncation Zone subchapter). Direct subglacial erosion is unlikely especially in the case of the erosional surface south of Country Harbour because sediment was removed without widespread disturbance of the underlying beds (Fig. 20). The planar nature of the unconformity also mitigates against direct subglacial erosion. Subglacial deformation of the Emerald Silt has been proposed as a mechanism of the formation of till-tongues exposed on the surface, but these are clearly truncated by the TESS surface. The age of the erosional surface associated with the Country Harbour Moraine was discussed earlier. If it is synchronous with the -65 m lowstand (11.7 ka) then a sub-wave-base erosional origin may be plausible. Storm-generated bottom traction currents associated with the lowstand shoreline may have eroded soft, muddy sediments while leaving the till-cored landforms.
Formation of the Morainal Zone 1 Formation of "symmetrical moraines" by surges or readvances and stagnation. Till is squeezed into subglacial crevasses as the ice settled back on the substrate after the rapid movement (cf. Zilliacus, 1989). These features seem to be located on acoustic basement highs, where a compressive flow regime favors vertical fracturing.

2 Formation of "asymmetrical moraines" during shoaling of the ice mass. Moraines are formed proglacially by seasonal readvances, similar to "push" moraines of Boulton (1986).
Symmetrical Moraines

Asymmetrical Moraines

Till

Bedrock

Debris in transit

Crevasses

Figure 43.
such as in the Morainal Zone and SSEMC intact. Erosion seems to be prevalent above 140 m or 60 m below the lowstand. It is also possible that in areas where only Emerald Silt A and C outcrop on the seabed (Fig. 42), erosion was accomplished either proglacially or subglacially, during the readvance of ice that produced the Morainal Zone. After retreat of this ice, and isostatic uplift, deposition was inhibited by stronger bottom currents. The absence of significant thicknesses of Holocene mud from these surfaces implies erosion or perhaps non-deposition, resulting from strong bottom currents active at present, that may be sufficient to erode earlier glacial marine sediments. More data has to be collected from these surfaces in order to assess the validity of erosional models.

3.8 OUTCROP ZONE

The Outcrop Zone was first reported by King and Fader (1986) and described in detail by Piper et al. (1986), Forbes et al. (1991) and Stea et al. (1994). It is a broad area of high-relief acoustic basement, largely devoid of surficial sediments, extending from 80 to 120 m water depth south and southeast of Halifax (Figs. 13, 44). The Outcrop Zone merges with the Morainal Zone southeast of Halifax as the two zones occupy similar water depths (Figs 8, 45). The contact between the two zones was not surveyed.

3.8.1 Unit B

Within the Outcrop Zone, acoustic basement is sporadically overlain by a thin (1-10 m), incoherent, transparent seismic unit termed Unit B by Forbes et al. (1991). It can be found at the base of channels and isolated valleys, or as muted, ridge-like features visible in the multibeam image of the sea floor near the contact with the Truncation Zone (Fig. 46).
Figure 44  Huntec profile and location of the type section of the Outcrop Zone (see Figs 10, 11 for legend and figure locations) Note the high relief and lack of sedimentary infill in the southeast part of the transect, located within the Outcrop Zone. In the Truncation Zone, the amplitude of surface relief of the seabed surface lessens markedly.
TRUNCATION ZONE

0 1400 m

OUTCROP ZONE

Metres below sea level

Figure 44
Figure 45. Surficial geology-landforms of the Halifax north transect area (Fig. 1). See Figure 10 for legend. Locations of IKU (boxes with numbered leaders) and core samples (crosses with numbered leaders). Note the southwest-trending channel delineated by the 100 m bathymetric contour. Dashed box indicates multibeam survey area (Fig. 48). Core 91018-38 is located over the Sambro Delta.
samples taken from Unit B include nos 14, 17, 32, 33, 63 and 91 (Figs 7, 16, 18, 45; Appendices 3, 4) in water depths ranging from 65 to 106 m. The samples often consisted of gravel (pebble-cobble) and "lumps" of diamicton. The lumps consisted of an olive-grey, sandy diamicton with subangular to subrounded clasts, with an average value of 10% striated clasts (Table 4, Figs 46, 47). It is assumed that these lumps represent remnants of an unmodified unit underneath a pebble-cobble armour on the sea bottom. In IKUs 32 and 91, the "lag" samples were devoid of striated pebbles while percentages in the diamicton "lumps" varied from 3-40%. Meguma pebble percentages vary from 74 to 98% with an average of 84% (Table 4) and foreign lithologies vary from 2-24%, averaging 11% (Table 4). IKUs 32 and 33 have a higher percentage of Meguma lithologies in the diamicton lumps (Appendix 4) than the lag samples (74%/82%, 74%/83%). If the lag samples have been reworked through transgression one would expect a greater percentage of metagreywacke clasts in the lag gravels because of the durability of Meguma metagreywacke (Piper et al., 1986). The decrease in metagreywacke pebbles in the surface subsamples of the IKUs may be due to primary glacial depositional processes including englacial transport and melt-out deposition of far-travelled "foreign" components. Unit B is interpreted as a till because of ridge morphology, diamicton sedimentology and striated pebbles.

### 3.8.2 Origin of the Outcrop Zone

Three erosional mechanisms have been advanced to explain the widespread outcrop zone on the inner shelf (Fader, 1989a, Forbes et al., 1991):  
1. Coastal transgression and wave-erosion under lower relative sea-level  
2. Locally-intensified glacial erosion  
3. Subglacial meltwater erosion
Figure 46. Type Huntex profiles of "Unit B" in the northern Halifax transect area (Figs. 1, 45). Middle Panel-Huntec profile of IKU 32. Note the undulating sea-bed topography. Left-bottom and left panels-Huntec profile and stratigraphy of vibra-core 91018-36. Right bottom-right panels- Huntex profile and stratigraphy of vibra-core 91018-40. Unit B at this locality underlies the LaHave Clay.
Figure 47. Sea bed photographs of Unit B in the vicinity of core 91018-99 (Figure 18-location). Scale is 10 cm long. Note the surface armor of subangular to subrounded clasts.
Hypothesis 3 was discussed at some length by Forbes et al. (1991). They suggested that the smaller Nova Scotian ice caps and glaciers were not thick enough to produce or store the vast quantities of water required to produce catastrophic subglacial outbursts. Wave-base erosion in the intertidal zone can be discounted as a primary erosional mechanism because of the water depth in which the outcropping bedrock occurs (120 m) and the interpreted low-stand position of -65 m (Stea et al., 1994, Loncarevic et al., 1994). The high relief and dearth of sediment within some valleys in the outcrop zone suggests that this region was not planed off by surf-zone erosion and then transgressed.

Glacial erosion is most effective in a wet-based ice zone and erosion is closely linked to deposition (Sugden and John, 1976, Boulton, 1979, Hughes, 1981). Linear fluting evident on the outcrop zone (Fig. 48) implies erosion by a wet-based glacier (Boulton, 1979) but the reason for the lack of deposits over this broad region is unclear. Large areas of bedrock exposed in land areas along the Eastern Shore of Nova Scotia (Stea et al., 1992a) are surrounded by till-covered terrain. These barren areas are generally constrained to local topographic highs. On land, strike-ridges developed from Goldenville Formation rocks have similar amplitudes to the ridges in the outcrop zone and are free of intervening sediment in some areas. Many of the bedrock areas are located near a Wisconsinan ice divide that straddled the axis of the Nova Scotia peninsula (Stea et al., 1989). Freezing and frozen bed conditions inhibit both erosion and deposition (Hughes, 1981). Glacier flow, deflected upwards by slight variations in the elevation between bedrock facies of differing competency, can create freezing or frozen bed conditions (Nobles and Weertman, 1971). It is suggested, therefore, that the outcrop zone is a result of non-deposition rather than erosion. This region of non-deposition can be explained by subglacial freezing zones at the margin of the former Scotian Ice Divide (Stea et al., 1989, 1992a) localized over the Nova Scotia.
Figure 48. Multibeam image and interpretation in the north Halifax Transect area (see Figs. 7 and 45 for location). Note the muted, topography in the Truncation Zone compared with the starker, strike-ridge topography of the Outcrop Zone. Morainal ridges formed of Unit B are found south of the Truncation Zone. A major southwest-trending channel cuts across the map area. Southeast-trending lineations that cross-cut structures are interpreted as large scale glacial grooves or fluting.
INTERPRETATION

TRUNCATION ZONE

OUTCROP ZONE

Legend

- Transition Subzone (Shoreface-deltas)
- Moraines
- Glacial fluting
- Location of Sambro Delta

Strike ridges
Terrace scarp
Contact between terrain zones

Figure 48.
peninsula during the last glacial maximum, or the relative competency of the metagreywacke
strike ridges.

3.9 TRUNCATION ZONE

3.9.1 Introduction

The Truncation Zone is defined as a region of the inner shelf from 90 m water depth to the
present shoreline, characterized by muted acoustic topography, terraces and sub-bottom
erosional surfaces truncating bedrock and surficial cover. The type sections for the
Truncation Zone are acoustic profiles K-J and E-F (Fig. 49). These profiles, the trackline
maps (Figs. 16, 18, 45) and the Halifax multibeam map area (Fig. 48) show a similar
landward progression of depositional and erosional landform subzones. These are (seaward
to landward):

1. Valley Subzone
2. Transition Subzone (shoreline)
3. Platform Subzone
4. Estuarine Subzone

3.9.2 Valley Subzone

Sediment-infilled valleys are commonly found in water depths of 65-90 m. These valleys
trend southwestward and southeastward parallel to drainage systems on land (Figs. 16, 18,
45, 48). They display a distinctive seismic architecture consisting of several seismic
Figure 49  Hunted profiles K-J (Fig 16) and E-F (Fig 18) northwest-southeast off Sheet Harbour and Country Harbour. These are the type transects of the Truncation Zone. The transect shows a parallel progression of landforms from sediment-infilled valleys (Valley subzone) to a ramp and terrace (Transition subzone) and finally a platform surface with truncation of both bedrock and surficial deposits (Platform subzone) (modified after Stea et al., 1994)
MORAINAL ZONE

TRANSITION SUBZONE

PLATFORM SUBZONE

TRUNCATION ZONE

ACOUSTIC BASEMENT

CORE

GRAB

DEPTH BELOW WATER SURFACE (m)

Figure 49
sequences with a ponded style of deposition, truncated by erosional unconformities in the subbottom and at the sea floor (Figs 49, 50). In the valleys, a basal, acoustically incoherent unit (Unit B), is overlain by a succession of ponded, onlap-fill acoustic units with moderate to high amplitude reflections. In the Sheet Harbour map region several southeastward-trending valleys or basins have been mapped (Fig 16, Map 2). The valleys are 700 to 1000 m wide and 40 to 70 m deep and incised into acoustic basement (Meguma Group bedrock).

At the type section off Sheet Harbour (L- Fig 16, Fig 50), seven sequences are found within the Valley Subzone. A conformable massive acoustic unit with occasional indistinctly stratified zones (Unit B) overlies acoustic basement. Overlying and infilling channels within Unit B are a sequence with horizontal reflections of moderate to high amplitude (V1). V1 is overlain by a sequence with reflections of higher amplitude and frequency (V2). V2 is overlain by two ponded, inclined seismic units, with low amplitude reflections varying in reflection amplitude and continuity (V3 and V4). These units are truncated by a major horizontal unconformity (VA) which is overlain by an acoustically transparent unit (V5). V5 is incised by channels infilled with V6.

### 3.9.2.1 Lithology

Grab samples and cores were obtained within the Valley Subzone at various locations. Samples taken within the Valley Subzone in Sequences V5 and V6 produced cobble lags with subrounded to well-rounded, discoid clasts (IKU 62, 76 Figs 16, 18) and no striated pebbles.
Figure 50. Type section (L; Fig. 16) of the Valley Subzone. The ponded units (V1-V4) within this valley are truncated by an unconformity. Units V5 and V6 rest on this major subbottom unconformity and appear to be themselves truncated by an erosional unconformity represented by the sea bottom reflection.
Figure 50.

SEISMIC PROFILE

INTERPRETATION

Figure 50.
Two vibracores were taken in the Valley Subzone at the type sections (Cores 91018-78; 91018-66, Fig 16, 18) The recovery was poor, but both revealed a cobble-lag of subrounded to rounded metagreywacke pebbles overlying coarse to medium sand and grey clay. Samples of the bottom clay in both cores revealed a foraminiferal assemblage dominated by *Elphidium excavatum* and *Cassidulina reniforme* (E. Collins, pers. comm., 1992) This benthic foraminiferal assemblage is similar to that described for the Emerald Silt, formed near a temperate ice margin (Scott *et al.*, 1984)

### 3.9.3 Transition Subzone

Landward of the Valley Subzone the sea floor topography abruptly changes from an undulating, irregular surface to a planar one (Platform Subzone) This transition between these two subzones is marked by a relatively steep ramp leading to a terrace and sometimes a terrace scarp (Fig 51) The surface of the Transition Subzone ramp can be an erosional unconformity, cut into Unit B, as in transect K-J off Sheet Harbour (Fig 49), or a depositional surface formed by clinoform beds of Sable Island Sand and Gravel (Table 1). A enlarged multibeam bathymetric image of the Transition subzone reveals a relatively narrow zone with a gentle dip, resembling the modern shoreface (Fig 51) The ramp at the type section of the Transition Subzone (Sambro-Delta Fig 51) extends from 75 to 60 m water depth Topographic profiles through the type region show the distinctive ramp-terrace morphology (Fig 52) Profiles C and D and the multibeam image reveal a terrace scarp at 63 to 65 m water depth Similar terraces have been found on the Maine inner shelf at 55 to 60 m water depth (Kelley *et al.*, 1989, Shipp *et al.*, 1989, 1991) and are interpreted as lowstand shorelines
3.9.3.1 "Sambro Delta"

A progradational feature, the "Sambro Delta" is located in the Transition Subzone in the Halifax trackline map region, a region also covered by multibeam bathymetry (Figs 45, 48, 51). It is interpreted as a delta primarily because of oblique terminations of clinoform reflections against the upper sea bed, wedge-like geometry, and its location in an embayment of the Truncation Zone (Figs 51, 53). The Sambro Delta surface can be seen as a flat depositional surface, roughly triangular in shape, on the enlarged multibeam image (Fig 51). Along the axis of a channel in deeper water the delta merges into a valley.

A strong, coherent and serrated reflection with point and cross hyperbolic diffractions, interpreted to be bedrock, occurs at the base of the progradational section (Fig 53). Above this reflection is an incoherent seismic unit (Unit B), 1-10 m thick. Unit B is incised and infilled with another unit (Sequence 1) with moderate amplitude reflections and a ponded style of deposition. Downlapping on unit 1 on the flank of a low hill are clinoform reflections that are obliquely terminated at the sea floor (Sequence 2). On the upper terraced surface another sequence (3) with chaotic, steeply-dipping reflections becomes prevalent. Sequences 1, 2 and 3 are interpreted as the topset, foreset and bottomset parts of a delta. The clinoform reflections of sequence 2 could be interpreted as shoreface progradational beds or deltaic foresets, but oblique terminations are more characteristic of deltaic systems (Mitchum et al., 1977). From a submersible, Forbes (1990) noted a pebbly sand on the foreset slope. Sequence 3 overlies or grades into an acoustically incoherent Unit (C) (Forbes et al., 1991) that outcrops at the sea floor over much of the Halifax transect area (Figs 10, 45). Forbes et al. (1991) interpret Unit "C" as glacial outwash, subaqueous in origin, based on the complex internal structures and gravelly-sand lithology. They noted that the unit is truncated.
Figure 51  (A) A digitally-enhanced color SWATH image, enlargement of Figure 48, looking northeastward up an embayment in the Truncation Zone, revealing morphology and subdivisions. The Valley Subzone is in the topographic low. The Transition Subzone is a ramp leading to a planar surface (Platform Subzone). A is at the approximate location of the Sambro Delta. B depicts the interpreted extent of the topset part of the paleo-delta (after Stea et al. 1994). (B) A terrain model of an offshore drumlin field described in Loncarevic et al. (1994) located SE of Lunenburg Nova Scotia (44° 19'N, 63° 49'W) compared with the digital terrain model of a similar-sized drumlin field from the Lunenburg drumlin field (Map 1). Note the truncated surface on the drumlin (arrow) at -65 m. Ribbed moraines are superimposed on the drumlins. The colours in the diagram B are in 10 m increments. The offshore drumlin diagram keys are deep red - 70-60 m BSL, orange 70-80 m BSL, yellow 80-90 m BSL.
A
SHOREFACE-DELTA COMPLEX
HALIFAX NORTH TRANSECT AREA

B
TRUNCATED DRUMLINS OFFSHORE
 COMPARED WITH LUNENBURG
 DRUMLIN FIELD

Figure 51.
Figure 52. Topographic profiles across the embayment in the Truncation Zone (Figs. 46, 51). Note the distinctive ramp/terrace morphology and the terrace scarp at the landward end of the embayment. Four topographic profiles across the embayment show the ramp/terrace morphology of the Transition Subzone. The break from the ramp to the upper terrace is inferred to be the former lowstand. This occurs at -65 m (modified after Stea et al., 1994).
Figure 52.
Figure 53. Nova Scotia Research Foundation sparker profile and interpretation of the Truncation Zone in the Halifax area (Figs. 1, 4; Forbes, 1987). Description of core 91018-38 on the right. Note the coarse surface layers underlain by fine sand (dotted pattern). A shell hash was found consisting entirely of mussel valve fragments (*Mytilus edulis*). One of these fragments was radiocarbon dated and an age of 11,650 +/-110 ¹⁴C years BP. Unit stratigraphy from Forbes *et al.*, (1991). Core depths in centimetres (modified after Stea *et al.*, 1994).
SEISMIC PROFILE

INTERPRETATION

Topset/foreset contact?

ACOUSTIC BASEMENT (Unit A)
by the sea floor at depths less than 70 m. Forbes (1990) also noted large boulders on the surface of Unit "C" on the upper slope of the "Sambro Delta".

3.9.3.1 Lithology and Age

Core 91018-38 was obtained from Sequence 2 (Fig 53) at 73 m water depth. The top 70 cm of the core consists of grey, medium to coarse sand capped by a lag surface of well-rounded cobbles. At 70 cm depth a shell hash was found consisting entirely of mussel valve fragments (Mytilus edulis). One of these fragments was radiocarbon dated and an age of 11,650 ± 110 14C yr B.P. was obtained, adjusted for isotopic fractionation. The reservoir effect correction has not been assessed in this region, but a correction of 400 years can be applied representing a world-wide mixed layer oceanic average (Mangerud, 1972). The corrected age is the same as the pre-13C adjusted age of 11,250 14C yr B.P.

The upper 40 cm of core 38 revealed abundant (>1000/10 cm³) nearshore-type benthic foraminifera dominated by Islandiella islandica, Cibicides lobatulus, and Islandiella teretis. The bottom 50 cm, however, is barren of foraminifera, probably implying higher sedimentation rates or lowered salinity typical of a glacial front deltaic system (Swift and Borns, 1967; Stea and Wightman, 1987).

3.9.3.2 Country Harbour "Delta"

Huntrec records revealed progradational clinoforms at the contact between the Platform Subzone and Outcrop Zone in water depths of 68 to 75 m in the Country Harbour transect region (Fig 18). The Transition Subzone shoreline can be traced NE of Country Harbour in
water depths of approximately -70 m where Hunttec and 3.5 kHz seismic profiles indicate a marked topographic transition from Platform Subzone to Outcrop Zone. In water depths greater than 70 m, the seabed is marked by moderate to high relief bedrock ridges and intervening basins filled with Emerald Silt. At -70 m a terrace/scarp is developed. Above -70 m the bedrock-dominated terrain becomes muted.

### 3.9.4 Platform Subzone

The Platform Subzone of the Truncation Zone is defined as a low relief erosional surface with a seaward gradient of approximately 1°, cut into bedrock and surficial deposits, in water depths of -65 to -70 m. The multibeam bathymetric images reveal a region of muted topography broken by occasional bedrock highs that rarely exceed 5 m of seabed relief (Figs 45, 51). The type section for the Platform Subzone is transect E-F which extends southeastward off Country Harbour (Fig 49). This remarkably flat erosional surface cuts bedrock and unconsolidated deposits.

#### 3.9.4.1 Unit C

A widespread, incoherent seismic facies (Unit C, Fig 45) overlies acoustic basement (Unit A) and Unit B in the Platform Subzone. It varies from a metre to 10 m in thickness and has a blanket-like morphology. It is best developed on the inner shelf near Halifax (Forbes et al., 1991). The surface of Unit C is flat and featureless (Fig 54), in sharp contrast to Unit B (Fig 47) which can have undulating, or ridge and basin topography. Two subunits can be resolved within Unit C, one distinguished by chaotic incoherent reflections, and the other with arcuate coherent reflections (Fig 54). These reflections are truncated by the sea bed in
Figure 54 Huntec seismic profiles of Unit C (locations on Fig 45) Contrast these records with Unit B (Fig 47) Surficial Unit C reflects more of the incoming acoustic energy than Unit B, hence the unit is darker, and sometimes obscures the trace of acoustic basement in the subbottom The lower section of Unit C (from Forbes et al., 1991) reveals some coherent, arcuate reflections
UNIT C TYPE SECTIONS

Depth below sea surface (m)
water depths between 50 and 70 m (Forbes et al., 1991) Sidescan sonograms show a variably reflective, light and dark (salt and pepper) surface interpreted as a cobble veneer. Bottom photographs of this surface from the Halifax region show that it is armored with a cobble lag of well-rounded boulders (Sable Island Sand and Gravel, Fig 55). Eight IKUs taken from Unit C ranged in texture from diamictons to gravelly-sand, with relatively little matrix (12%, Table 4) and abundant gravel (88%, Table 4) Pebbles are subrounded to well-rounded with an average of 4% striated pebbles compared to an average of 10% for Unit B (Table 4, Fig 56) The lag armour on Unit C can be considered to be equivalent to the Sable Island Sand and Gravel Formation of King and Fader (1986-Table 1) The boundary between Units B and C in the Halifax map area (Fig 45) occurs between 70 and 80 m water depth Unit C may be a composite of outwash, ice-contact stratified drift, or melt-out till overlying basal till, these deposits may not always be resolvable on seismic profiles The surface lag-armour probably reflected much of the incoming acoustic energy, so low-impedance contrast, subbottom reflections were not resolved

3.9.4.2 Valley Fill

Isolated valleys or basins are found in the Platform Subzone incised into the acoustically basement (Meguma Group bedrock) These valleys are several hundred metres to 2 km wide and 4-20 m deep and contain a sedimentary infill with a similar facies geometry to the Valley Subzone At the base of the valley sequence is a draped, acoustically incoherent unit (Unit B) Incised into Unit B are channels filled with ponded sediments consisting of several sequences of moderate to high amplitude reflections These high amplitude sequences are truncated by a horizontal erosional unconformity often near the sea bed Overlying the unconformity are two incoherent or indistinctly-bedded seismic facies
Figure 55. (A) Rounded gravels obtained from the Sable Island Sand and Gravel above Unit C in the vicinity of IKU 44 (Location Fig. 45). (B) Sable Island Sand and Gravel from IKU 79 (Location Fig. 18). (C) Bottom photograph of Unit C at IKU 31 (Water depth 64 m; Fig. 45). Contrast the pebble shapes evident from these photographs with those from Unit B (Fig. 47) in water depths greater than 80 m.
Figure 56  Plot of striated pebble percentages and sample sedimentology vs water depth for all IKU samples. The sedimentology of the samples is clearly related to water depth, with diamictons and mud found generally below 65 m, whereas sand and gravel is restricted to water depths above this level. Note also the marked reduction of striated pebbles above 65 m, the inferred sea level lowstand position. The landform-unit designations are indicated for each sample (see Fig 10).
Inferred lowstand depth

LEGEND

- Scotian Shelf End Moraine Complex
- Emerald Silt Formation
- Morainal Zone (moraines)
- Unit B
- Unit C
- Sable Island Sand and Gravel

Figure 56
Samples from the sea bed (Sable Island Sand and Gravel) in these valleys typically contain variable amounts of sand and gravel, the gravel fraction is overrepresented (60%, Table 4) due to sand wash-out upon IKU retrieval. Pebbles in the gravel lag are typically subrounded to well rounded, and lacking or devoid of striations (2%, Table 4). Twelve gravel samples from valleys and basins within the Truncation Zone averaged 2% striated clasts (Fig. 56, Table 4, Appendix 4). The uppermost incoherent and transparent seismic facies in these valleys encompass the Sable Island Sand and Gravel (Table 1).

3.9.4.3 The Estuarine Subzone

Forbes et al. (1991) and Boyd et al. (1992) identified and mapped estuarine deposits in valleys confined to water depths less than 50 m. Above 50 m water depth, north of the type section (E-F) of the Platform Subzone off Country Harbour (Fig. 49), are a series of wedge-shaped acoustic units composed of gravel and sand interpreted as a transgressed barrier system (Stea et al., 1993). Boyd et al. (1992) also identified transgressed barrier-lagoonal systems. The Platform Subzone appears to have largely formed through erosion, whereas depositional processes were dominant within the Estuarine Subzone.

3.9.5 Origin of the Truncation Zone

Erosional processes related to the regression and the post-glacial rise in sea level are largely responsible for the formation of the Truncation Zone and its constituent subzones. The paleoshoreline or lowstand is interpreted to be the top of the Transition Subzone. The Platform Subzone represents the transgression surface (Fig. 46).
3.9.5.1 Valley Subzone

The Valley Subzone is essentially a series of valleys and embayments that merge landward and are attenuated at the lowstand shoreline (Figs. 8, 49, 51). Some of these incisions merge with larger valleys that trend southwestward and cut the SSEMC ridges (Figs. 8, 11, 16). The origin of these incisions is unclear. Some may be meltwater conduits of the ice cap that formed the Morainal Zone, as they appear to start at the Zone (Fig. 16). Others formed before the last glacial advance because they are lined with Scotian Shelf Drift or Unit B, acoustic units believed to be tills. The predominant south-westward trend of these incisions is also problematic, as it is at odds with the primary southeastward, fault-related trend of most river valleys on the Nova Scotia mainland.

Unconformities in the valley fill at the surface and sub-bottom may have formed by erosion at or below wave-base, or even by earlier glacial advances. Erosional unconformities in glacial marine clays contained within these valleys, may indicate direct glacial erosion or current erosion that predated sea-level rise. This is evident along the South Shore of Nova Scotia, where shell dates of 14 and 28 ka have been obtained from truncated glacial marine valley fill at -60 m (Piper and Fehr, 1991). An unconformity at the top of the Emerald Silt in valley fill at 71 m water depth was dated between at 14 ka and 9 ka (Piper and Fehr, 1991, D J W, Piper, pers comm, 1993). This unconformity is similar in age to the TESS surface and may be related to the lowstand at 11.6 ka.

The sand and gravel bodies on the surface of valleys in water depths from -65 to -90 m are a result of wave-current redistribution of shoreface deposits related to the lowstand (Lapierre, 1982; Stea et al., 1994). The seaward margin of the Truncation Zone is essentially the limit.
of sublittoral deposition and wave reworking during the lowstand. The sublittoral Sambro Sand (Table 1) mapped by King (1970) represents the seaward limit of wave-base reworking. King (1970) interpreted the contact between the Sable Island Sand and Gravel (Table 1) and the Sambro Sand as the lowstand. Off Halifax this contact was mapped at -90 m.

3.9.5.2 Transition Subzone (shoreline) and Platform Subzone

Three major lines of evidence suggest that the maximum sea-level lowering was only -65 m.

1. Till-moraines do not appear to have been modified or removed by a regression-transgression cycle as they have retained a distinctive constructional morphology. They are found in water depths of 145 to 65 m, similar to the depth range of the Outcrop Zone.

2. The seabed surface changes markedly from one of low or subdued relief (Platform Subzone) above -65 m water depth to high or undulating relief seaward below -65 m. Unit C and Sable Island Sand and Gravel are only found above -65 m.

3. Sediments below -65 m are largely mud and diamicton whereas above -65 m gravel and gravelly-sand predominate (Fig 6). Pebble roundness changes from predominately subangular to subrounded below -65 m, to subrounded to well-rounded above -65 m and the percentage of striated pebbles decrease markedly above -65 m (Fig 56).

At -65 m to -70 m the Transition Subzone ramp and terrace separates the Platform Subzone from the Valley Subzone (Fig 51). It is interpreted as a product of shoreface erosion during regression and transgression (Belknap and Kraft, 1981), in essence a ravinement surface.
Nummedal and Swift, 1987) or, in part, a deltaic surface. The terrace surface is interpreted to be the paleo-shoreline, although the transgression may have been initiated at a slightly lower elevation. The average ramp height of 10 m found in this study is similar to values of wave-base erosion cited by Suter et al. (1987). At the type profiles (Fig. 49) the base of the ramp represents the lower limit of wave-base erosion in late-glacial time.

From the Sambro Delta type locality, sea level is interpreted to be near the topset/foreset contact (Fig. 52) which is approximately -65 m. The presence of Mytilus edulis in the core indicates deposition of the foresets close to the intertidal zone. Mussel shell hashes have been described in foreset beds of raised glacial marine deltas and moraines in the Saint John, New Brunswick, area (Nicks, 1988). The known deglaciation chronology does not negate a glacio-deltaic hypothesis. Stea and Mott (1989) and Gipp (1994) estimated that the ice margin was close to the present coastline at about 12 ka. Recent AMS radiocarbon dating of the base of Porters Lake, near Halifax, suggests deglaciation by 11.7 ka (D. B. Scott pers. comm., 1994). The age of the Sambro Delta is close enough to the lake date to infer a glacial sediment source, but the evidence for meltwater drainage has been removed by the regression/transgression. Thus, the deltaic hypothesis must remain speculative. An alternative explanation for this zone may be a shoreface-coastal terrace. Boulders noted by Forbes (1990) on the surface of the Sambro delta may be relict from till-cliff erosion.

The Country Harbour and Sheet Harbour type transects (Fig. 49), in water depths of less than -65 m show low gradient erosional surfaces cut into acoustic basement and unconsolidated deposits. These planar surfaces are interpreted as shore platforms, defined as surfaces of low gradient (1-5°) formed by mechanical wave erosion at or near sea level (Trenhaile, 1987). The Country Harbour transect (Fig. 49) is a particularly striking example of a low gradient...
shore platform. These surfaces were cut during the regressive and transgressive phases of the Late-glacial period.

Forbes et al. (1991) and Boyd et al. (1992) describe estuarine facies within a depositional zone (Estuarine Subzone) restricted to water depths of less than 50 m. The reason for the lack of estuaries and relict barriers within the Platform Subzone below depths of 50 m is unclear. One reason could be that they are there but were not found. Another reason may be related to rapid initial transgression rates. Belknap and Kraft (1981), however, suggested that preservation of estuarine sequences is more likely with higher rates of relative sea-level rise. Climate may have also played a role in the production of estuarine sediments.

3.9.6 Revised Sea-level Curve for the Eastern Shore

The recognition of ravinement surfaces, shore platforms and deltas graded to depths of 65 m BSL on the inner shelf, extends the sea-level curve for the region (Fig. 57). Shell dates obtained from this study, and Piper and Fehr (1991) have been added to a sea-level curve of the inner shelf compiled by Forbes et al. (1991) from Scott (1977), Scott and Medioli (1982), Miller et al. (1982), and Honig (1987) (Fig. 57). The curve before 9000 $^{14}$C yr B.P. is based on shell dates. The accuracy and precision of the data are controlled by the uncertainty of reservoir corrections and depth assignments. The data after 7000 $^{14}$C yr B.P. are better constrained because they are based on salt marsh peat cores whose RSL assignments are determined by foraminiferal assemblages accurate to ± 10 cm of RSL (Scott and Medioli, 1978). Upon initial deglaciation, sea levels were higher than -60 m (Piper and Fehr, 1991, Forbes et al., 1991). The maximum RSL at this time is not known although Emerald Silt has been identified in 40 m water depth (Stea et al., 1993). RSL then fell to a lowstand of
Figure 57.

Calculated RSL curves
A ———— "MINIMUM ICE"
B ———— "MAXIMUM ICE"

Field observations
MEAN + 2 SIGMA

DEPTH ERROR
Reservoir Uncertainty
Fairbanks, 1989
-65 m, followed by a rapid rise of 1.5 to 2 m per century until ca. 11 ka. Rates of sea-level rise decrease markedly at 11 ka and then increase starting at about 8 ka. From 9 ka to present, rates of sea-level rise follow a step-like pattern with inflection points (Scott and Medioli, 1982, D. B. Scott, pers. comm. 1995). This curve is characterized by an initial high amplitude, short wavelength isostatic response, followed by step-like sea-level changes. It is remarkably similar to the new curve obtained for the inner shelf off Maine, USA which has a lowstand position of -50-60 m followed by a RSL plateau between 11 and 9 (Barnhardt et al., 1992). All data points are within a 50 km radius, so the complexity of the curve cannot be explained by rebound variations relating to ice thickness or crustal variations. New sea level data presented after this thesis was in preparation by Edgecombe (1994) suggests a more complex sea level curve after 9 ka than presented in Figure 57.

3.9.7 Glaciological and Geophysical Implications

Clark et al. (1978) developed a mathematical model which derived a relative sea level history from the earth's rheological properties, ice sheet thickness and deglaciation history. Quinlan and Beaumont (1981, 1982) utilized this approach to deduce the sea-level response of two contrasting ice models for the Maritime Provinces of Canada. They divided the Maritime Provinces into four zones based on their model-derived RSL history. Zone A (continuous emergence) to Zone D (continuous submergence). According to their model, these zones shift seaward with more ice and landward with less ice. A primary factor governing the shape of the resulting sea-level curves is the peripheral forebulge, which migrates towards the former ice centres after deglaciation. The maximum model represents a uniform ice cover over eastern Canada (1-2 km), thinning to the southeast, and a margin at the outer shelf (cf. Flint, 1971). The minimum model (Grant, 1977, Grant and King, 1984) has much thinner ice.
(<1 km), segregated over Nova Scotia, with margins near the present coastline and no grounded ice in the Gulf of St. Lawrence. Quinlan and Beaumont (1981) calculated theoretical relative sea-level curves for the inner shelf based on the maximum and minimum ice models. At that time they found that the observed RSL field data were bracketed by the theoretical curves derived from the extreme ice models. In a later paper (Quinlan and Beaumont, 1982) constructed a "compromise" ice model based on existing field RSL observations characterized by:

1. No grounded ice in the Gulf of St. Lawrence
2. Margins not substantially seaward of the minimum reconstruction
3. Ice retreat from Nova Scotia by 14 ka

The "compromise" RSL ice model closely resembles the minimum model of Grant (1977) because both models feature an ice-free Gulf of St. Lawrence. Tushingham and Peltier's (1991) recent ice sheet reconstruction (ICE-3G) also based on existing field RSL data, produced a similar ice distribution and retreat pattern to the Quinlan and Beaumont (1982) RSL-derived model.

The new empirical RSL curve of the inner shelf (Fig. 57) differs from the theoretical minimum curves of Quinlan and Beaumont (1981, 1982) and Tushingham and Peltier (1991) primarily in the high amplitude and short wavelength response of RSL change after deglaciation. The maximum sea-level lowering is about 20 m below that predicted for the "compromise" minimum model and 40 m or more lower than predicted in the maximum model. The transition from initial RSL fall to RSL rise occurs over a much shorter time span than predicted.
The most recent theoretical RSL curve obtained from Peltier (pers. comm., 1993) also fails to predict the magnitude of sea-level lowering evident in the empirical RSL curve by 20 m but it does predict the response in the later Holocene rather well. The lowstand can be reproduced by either shifting the ice margin 2° westward or thinning the ice substantially to ice volumes comparable with the minimum model (Peltier, pers. comm., 1993). Muting of the pre-lowstand emergence pattern characterized by RSL fall, however, is concomitant with removal of the ice.

If ice margins existed out on the outer shelf, as recent field observations suggest (McLaren, 1988; Mosher et al., 1989; Gipp, 1994), then observed sea-level curves should be skewed toward the calculated curves for maximum ice (Fig. 57) which also predict raised strandlines along the Eastern Shore of Nova Scotia. Low-profile ice masses (cf. Boulton and Jones, 1979) cannot be invoked to explain the discrepancies between extensive ice and minimal rebound because of the presence of local ice divides (Stea et al., 1992a, b) and the vast areas of Meguma Group metagreywacke outcrop on land and the inner shelf, which would tend to increase basal shear stresses and produce more parabolic ice masses (Forbes et al., 1991). Ice thicknesses in the Maritime provinces of Canada were more comparable to southern Maine, a region on the outer edge of the Laurentide Ice Sheet (Mayewski et al., 1981).

The timing and mode of ice retreat are crucial factors affecting RSL history. A brief summary of the deglaciation history of the Atlantic shore of Nova Scotia is necessary to place the RSL curves in context with the position and timing of retreating ice. Till-tongues along the northern margin of Emerald Basin have radiocarbon ages of 15,000 to 17,000 14C yr B.P. (King and Fader, 1988; Gipp and Piper, 1990; King pers comm., 1994). Based on lake sediment chronologies, Stea and Mott (1989, 1990) and Gipp (1994) estimated that the ice
retreated to the present coastline by 12,000 $^{14}$C yr B P (Map 1)  Southern Nova Scotia land areas, deglaciated by ablation and downwasting (Grant, 1989, Stea and Mott, 1989), are characterized by vast areas of melt-out tills, and hummocky topography without ice marginal moraines. In contrast, the deglaciation of southern Maine proceeded by rapid calving and active-zone retreat marked by extensive morainal systems and was largely complete by 12.8 ka (Smith and Hunter, 1989, Belknap and Shipp, 1991). The glacial marine Presumpscot Formation clay (Bloom, 1963) forms a widespread blanket over southern Maine below the marine limit. The Atlantic coast of Nova Scotia has no raised marine features, although these are predicted to occur if ice extended as far as the outer banks (Quinlan and Beaumont, 1981). The paradox is best explained by sub-ice emergence under a carapace of late-melting ice (Fig 58). Rapid deglaciation of the Nova Scotia inner shelf was probably hindered by the outer banks which acted as bastions to help buttress the ice sheets against sea-level rise (Gipp, 1989). Ice sheet deflation occurred from the top down and large areas of the ice sheet stagnated. Raised features did not form along the Atlantic coast because by the time the ice melted, RSL fell below the ice margin surface (Fig 58).

The drop in RSL between 14,000 and 12,000 $^{14}$C yr B P (Fig 57) on the inner shelf may be attributed to initial, rapid crustal response (rebound) as the ice retreated from the inner shelf. The rate of RSL drop during this period, however, is largely unknown. Differential uplift of various moraines across Eastern Shore indicates initial RSL drop assuming that the moraines were formed by grounded ice at similar water depths (-250 m). 70 m of differential uplift occurred between the Country Harbour Moraine (-84 m) and the Sambro Moraine (-160 m).
Figure 58  Schematic comparison of ice sheet-sea level interactions between Maine and Nova Scotia during the period between 14 and 12 Ka. Rapid ice retreat of an ice stream in the Gulf of Maine allowed the formation of the Presumpscot Formation on the isostatically-depressed landscape. Delayed retreat, ablation by downwasting and sub-ice isostatic equilibration of the Scotian Ice Divide prevented the formation of raised features along the Atlantic coast. Arrows indicate uplift and ice retreat and downwasting.
Figure 58.
The amplitude and wavelength of RSL change in the late glacial period is not predicted by the
geophysical models Barnhardt et al. (1992) and Shipp et al. (1991) made the same
observation for the Maine continental shelf. Josenhans et al. (1992) also noted substantial
sea-level lowering adjacent to the Cordilleran Ice Sheet at 10.6 ka for the Pacific coast. RSL
lowering predicted in the maximum ice model (Fig. 57) is only 20 m between 9 and 7 ka.
Some of this discrepancy can be explained by underestimation of worldwide ice volumes, but
the proximity of the age and depth of the lowstand to the RSL curve of Fairbanks (1989)
suggests that crustal depression was not as protracted as predicted by visco-elastic earth
models. Variations in the strength and thickness of the lithosphere may be responsible for
the disparity between models and data, but this effect was discounted by Tushingham and

The rapidity of RSL rise after the lowstand from 11,500 to 11,000 \(^{14}\text{C}\) yr B P may be due
entirely to melting, as rates of rise are comparable to the Fairbanks (1989) curve for this
interval (Fig. 57). The RSL plateau between 11,000 and 8000 \(^{14}\text{C}\) yr B P can be explained
by crustal uplift due to passage of a peripheral forebulge across the region. This forebulge-
related plateau is evident on both the maximum and "compromise" minimum ice theoretical
curves (Fig. 57). In the Maritime Provinces, glaciers advanced during the Younger Dryas
Chronozone (11-10 ka, Stea and Mott, 1989, Grant, 1989, King, 1994). Local ice advances,
and by analogy local ice centres, are considered to have negligible effects on the RSL record
because of the lithosphere's ability to support loads of small areal extent (Quinlan and

The -65 m shoreline appears to be a level surface over most of the Eastern Shore of Nova
Scotia. J. Shaw (pers. comm., 1993) has documented a lowstand (as yet undated) in
Chedabucto Bay of only -35 m. The amplitude of the RSL inflection is not known in these cases so the significance of these RSL fluctuations is unknown. The discrepancy in Chedabucto Bay may also be explained by late ice. Northern Nova Scotia was the breeding ground for the Younger Dryas glaciers postulated by King and Fader (1988) and Stea and Mott (1988, 1989). Although this area was deglaciated ca. 12.4 ka (Stea and Mott, 1989), evidence of early marine incursion may have been removed by the late re-advance. Local crustal depression caused by the Younger Dryas glacier would decrease the magnitude of lowering after removal of the load.

3.10 SUMMARY OF MARINE GLACIATION AND SEA-LEVEL EVENTS FOR THE EASTERN SHORE

The author presented a preliminary interpretation of marine events on the inner shelf after initial seismo-stratigraphic interpretation (Stea et al., 1992b). Although details have changed, the basic six-phase evolutionary model is still valid (Fig. 59).

The composition and symmetry of the Scotian Shelf End Moraine Complex suggests that it was formed underneath the ramp-margin of a grounded glacier. Diamicton was deposited directly on the ice ramp-till bed by conveyor-belt recycling and lodgement or push-squeeze processes. Meltwater was restricted to basal sheet flow rather than being channelized. The steep landward slope of the moraine probably reflects the former ice contact slope. Till-tongues rooted in the morainal complex represent short-lived readvances of the semi-buoyant ice front, smearing-out till on the morainal front and overriding and deforming previously-
Figure 59  Six phase model of proposed deglaciation and sea level events on the Eastern Shore of Nova Scotia (modified after Stea et al., 1992 b)

1 Scotian End Moraines form at a ramp-margin of a grounded glacier.

2 Retreat to "proto" moraine  Proglacial deposition of Sequence 1

3 Retreat, then a readvance-surge ca 13 ka forming the symmetrical moraines and Sequence 2 in the Basin Zone

4 Retreat and shoaling of the glacier, increased wave-activity-turbulence and dropping sea levels entrain sediments  Sequence 3 deposited in the Basin Zone

5 Lowstand nadir  Erosion at the lowstand nadir produces a shoreline (Truncation Zone-Transition Subzone ) and Sequence Boundary 3/4 in the Basin Zone  Subsequent sea level rise and climatic warming  Sequence 4 deposited in the Basin Zone

6 Climatic cooling, increases storminess, sea ice and ice bergs increase sedimentation rates during the Younger Dryas  Sequence 5 is deposited in the Basin Zone
1. 16-15 KA

2. 15-14 KA

3. 14-12.8 KA

4. 12.8-12 KA

5. 12-11 KA

6. 11-10 KA

LEGEND

GLACIER

ICEBERG

SEA ICE

Meltwater Plume

Current

Erosion

Breaking Waves

SEQUENCE 1

SEQUENCE 2

SEQUENCE 3

SEQUENCE 3/4

SEQUENCE 5 YANKEE BANK FORMATION

Figure 59
Sequence 1 in the Basin Zone (Emerald Silt) is interpreted to be ice proximal, glacial marine sediment deposited by overflow and interflow meltwater plumes, emanating from an ice margin. Landward thinning of the Emerald Silt and the landward decrease in identifiable seismic sequences imply diachronous deposition from retreating ice margin rather than synchronous deposition under an ice shelf. "Proto" moraines represent these short-lived ice marginal stands. Sequence 1 in the Basin Zone was deposited from 14.5 to 13.5 ka.

The Morainal Zone formed after retreat of the glacier out of the Basin Zone and during the deposition of Sequence 2 in the Basin Zone (ca. 13.5-12.7 ka). Symmetrical Moraines probably formed synchronously, as a result of a readvance (surge?). Asymmetrical Moraines formed at the margin of a shallow-water tidewater glacier. After this readvance (13.5 to 12.7 ka), sea levels dropped rapidly, due to general isostatic recovery. Wave turbulence and debris from the melting glacier suspended very large amounts of sediment. Iceberg rafting was also an important mechanism for sediment delivery into the nearshore basins.

Sea levels dropped to their lowest point at -65 to -70 m around 11.6 ka. A shoreline formed during a short period of crustal stability (Transition Subzone). Sea-level rose rapidly after 11.6 ka and a large area of the inner shelf was transformed by erosion into a low-relief surface, barren of depositional features (Platform Subzone). At the lowstand nadir erosion...
was prevalent below the lowstand to depths as great as 140 m above which substantial quantities of Emerald Silt were removed. Synchronously in the Basin Zone an erosional sequence boundary was formed (Sequence 3/4). The rate of relative sea-level rise slowed markedly between 11 and 9 ka, a result of forebulge passage, and depositional features were again established (Estuarine Subzone).

A distinctive seismic and lithic unit (Yankee Bank Formation) was deposited during the period from 10.7 to 10.0 ka. This event horizon was a response to several possible factors: increased storminess, debris-laden sea ice and increased sediment supply from the devegetated coastline. The coastline was about 20 km seaward of present about that time. Permanent and semi-permanent snowpacks and small glaciers in northern Nova Scotia extending slightly offshore provided pulses of low-salinity, debris-laden meltwater plumes and small bergs which may have spread debris over much of the inner shelf.
CHAPTER 4: TERRESTRIAL QUATERNARY GEOLOGY

4.1 INTRODUCTION

A fundamental understanding of the morphology, stratigraphy and lithology of terrestrial Quaternary deposits is essential for correlation with seismic records of the offshore. The scope of the terrestrial study of this thesis is limited. The author will draw heavily on previously published material, especially the Surficial Geological Map of the Province of Nova Scotia (Map 1, Stea et al., 1992c). Several reference sections are examined in detail to provide a basis for land-sea correlation of till sheets.

Stea et al. (1992a) derived regional ice flow lines and the ice flow phase concept from their compilation of all striae, drumlin and esker orientations in Nova Scotia, verified with pebble and boulder trajectories from known source areas (Grant, 1963, Nielsen, 1976, Stea, 1984, Stea and Finck, 1984, Graves and Finck, 1988, Stea et al., 1989). These ice flow phases were correlated with till sheets in stratigraphic section based on detailed structural and pebble-provenance studies (e.g., Stea, 1984, Graves and Finck, 1988, McClenaghan and Dilabio, 1993). The relative chronology of ice flow phases developed through analysis of superimposed striae ("erosional stratigraphy" - Stea et al., 1992a) was verified by correlation with depositional stratigraphy.

About 20% of the Eastern Shore study region can be characterized as bedrock-controlled topography or bedrock (Map 1). This can be subdivided into greywacke and granite types of country (Goldthwait, 1924, Roland, 1982). Areas underlain by Devonian-Carboniferous
granites consist of low, rounded hills with relief amplitudes less than 20 m. The topography is random and irregular. Large, rounded perched boulders are common ranging from less than a meter in diameter to a maximum of 50 m. Valleys between bedrock highs are commonly filled with sediment which may typically be a stony diamicton at the base, overlain by waterlain sediments and peat in low-lying regions.

The dominant bedrock type is metagreywacke of the Meguma Group (Schenk, 1971). The greywacke topography consists of long, low ridges running east to west. Like the granitic areas, the intervening hollows are swampy and filled with a stony diamicton, sand and peat. Trellis drainage is developed in many areas. Small and large scale glacial erosional features are well developed on the bedrock surfaces.

Most of Nova Scotia is covered with glacial drift molded into a variety of landforms. Two common glacial landforms in Nova Scotia are till plains and streamlined drift or drumlins. Bolton (1990, p. 15) suggested that "there is no fundamental reason why subglacial sedimentary processes should be different in kind between glaciers terminating in water and those terminating on land." The sedimentology and stratigraphy of onshore tills provide a model for the offshore. In the following subchapters the stratigraphy, sedimentology and lithology of till sheets on land are examined in detail to determine the local and regional variability of till properties and hence the applicability of longer-ranged, onshore-offshore correlations.
4.2 STONY TILL PLAIN

4.2.1 Morphology

A till plain is defined (Goldthwait, 1989, p 274) as a nearly flat, or slightly rolling, and gently inclined plain with mostly thick cover, often with multiple till layers (varying compositions) and nonglacial beds, completely masking bedrock undulations. The "Stony Till Plain" is defined by irregular, hummocky topography or topography characterized by linear, en-echelon ridges (ribbed moraine) with large boulders (from <1 m up to 20 m in diameter) strewn on the surface. Ribbed moraine is best developed in southern Nova Scotia near Kejimikujik Lake (Map 1) and in the Chignecto Peninsula (Map 1, Stea et al., 1986). In land areas these ribbed moraine are occasionally found draped over drumlinoid landforms. Ribbed moraine have also been identified in the offshore regions using multibeam imaging (Figs 40, 51). In the offshore areas ribbed moraine are also draped over drumlins (Fig 51). Drift thicknesses in the stony till plain vary widely, generally increasing from 1-5 m in upland areas (>200 m elevation) to >7 m along the coast (Fig 60).

4.2.2 Stratigraphy and Lithology

Along the Eastern Shore study region the stony till plain consists of a clast-supported, sandy diamicton derived almost exclusively from local bedrock sources and characterized by a high percentage of angular, cobbles and boulders (Figs 14, 61). It has been termed the Beaver River Till from the type section near Yarmouth, Nova Scotia (Map 1, Grant, 1980b, Williams et al., 1985). For this study three textural and lithological "facies" of the Beaver River Till have been differentiated: ST/A, ST/B and ST/C. Facies "ST/A" is a olive-bluish grey, clast-supported, sandy diamicton associated with ground or ribbed moraine regions. It has a
Figure 60. Generalized cross-section through the "stony" till plain and drumlins along the Eastern Shore of Nova Scotia.
Figure 61.
homogenous, autochthonous composition with pebble percentage of the underlying or adjacent rock unit (in the sampling area mainly Meguma Group metasediments) averaging 86% and generally greater than 90%, whereas foreign or non-Meguma Group pebbles average 6% (Table 8, Fig 62) ST/A averages 64% gravel or clast weight percent with a matrix weight percent average of only 37% and a clast/matrix ratio of nearly 2 (Table 8, Fig 61) Only an average of 2% of the clasts are striated (Table 8) Facies ST/A covers most of the Eastern Shore study region (Map 1) Facies "ST/B" is similar in colour and texture to ST/A, but is found near drumlin fields and on top of drumlins It has inherited components of older tills, either directly as till inclusions or indirectly with higher erratic percentages Meguma metasedimentary (local) pebbles average 79% while foreign percentages average 19% Facies "ST/C" differs from the previous two facies with a coarser matrix texture and gravelly-sand and sand inclusions

4.2.3 Mushaboom Reference Section

A reference section for the stony till plain is a borrow pit described previously by Stea and Fowler (1979), located approximately 100 km east of Halifax, south of a contact between a large granitic pluton (Musquodoboit Batholith) and Meguma group metasediments (Keppie and Muecke, 1979, Fig 1, Section 61, Map 1) Several southward-trending Lawrencetown Till drumlins occur north of the section Bedrock outcrops nearby revealed parallel striae sets trending 175° (oldest) and 155° (youngest) (Map 1)
Table 8. Average values of lithic parameters for terrestrial tills in the study region.
Figure 62. Triangular plots of the ratio of percentages of Meguma metasedimentary pebbles, granitic (Meguma Zone granitic plutons) and "foreign" or erratic pebbles (from terranes outside the Meguma Zone). Sample locations given in Fig. 7 and data in Appendix 8. Beaver River Till facies ST/A and ST/B are lumped together as they show no significant differences in pebble percentages. The same is true for Lawrencetown Till facies A and B.
4.2.3.1 Stratigraphy and Lithology

At this site 14 m of bouldery, clast-matrix supported, sandy diamicton (Beaver River Till) overlies metagreywacke bedrock (Fig 63). The stony diamicton contains 55-75% metagreywacke clasts of the underlying Meguma Group Goldenville Formation and 21-45% clasts derived from the Musquodoboit Batholith, 100 m to the NW (Fig 63). The abundance of granitic clasts within the section implies southeastward clast dispersal from the Musquodoboit Batholith, rather than southward, parallel to the druin-forming flow phase which would not have crossed granitic rocks (Fig 63). The precise azimuth of flow is uncertain, because of variable and indistinct till fabrics (Graves and Finck, 1988). Clast/matrix weight ratios in this section vary from 0.9 to 2.0, compared with regional average values of less than 0.4 for the Hartlen and Lawrencetown Tills (Table 8, Fig 63). This stony diamicton is capped by a surface boulder layer dominated by granites. Large, Bluish, metagreywacke and smaller, white, megacrystic, granitoid boulders are found embedded within the section (Fig 63). The clasts are generally angular to subangular with no evidence of surface erosional markings such as striations. Pebbles do not display the distinctive, glacier flat-iron shapes like the metagreywacke-dominated Hartlen Till at West Lawrencetown and Wine Harbour.

4.3 STREAMLINED DRIFT-DRUMLINS

4.3.1 Morphology

The most prominent glacial landforms in the Eastern Shore terrestrial study area are drumlins. They have a characteristic shape defined by a steep, up-glacier facing (stoss) side.
Figure 63  Location, stratigraphy, lithology and clast/matrix weight percent ratios of till samples of the Mushaboom section (Fig 1, Section 61-Map 1)  The dashed region refers to Meguma metasediments whereas the dotted pattern to Meguma Zone (Musquodoboit Batholith) granitic lithologies. On the lithology profile plot the non-shaded portion (2%) is the non-Meguma Zone component of the clast fraction.
Sample site

Beaver River Till (ST/A)

Meguma outcrop

metagreywacke boulder

granite boulder

Clast/matrix ratio

Figure 63
and a gentler lee slope (Fig. 51). Nova Scotia has 4 major drumlin fields with thousands of individual drumlins (Map 1). The long axes of drumlins are parallel to the ice flow that formed them (Flint, 1971). V-shaped, "palimpsest" drumlins are present in Nova Scotia drumlin fields with axes parallel to both earlier and later ice flow phases (Stea and Brown, 1989; Stea, 1994). Drumlins are generally drift-cored and can exceed 30 m in total thickness at the stoss summit.

4.3.2 Stratigraphy and Lithology

Drumlin sections along the Eastern Shore reveal thick, predictable drift sequences which can be correlated regionally (Stea and Fowler, 1979). At the core of many drumlins is a grey, overconsolidated, matrix-supported, silty diamicton (Hartlen Till) overlain by a reddish, erratic-rich, matrix-supported, silty diamicton termed the Lawrencetown Till. The contact is sharp, often associated with striated, bullet-shaped boulder horizons or pavements (Grant, 1963; Nielsen, 1976). The red matrix colour of the Lawrencetown Till is very distinctive, entirely foreign to the Meguma Zone, which is comprised largely of grey and black slates, metawackes and white granites. In some sections the Beaver River Till (ST/B) overlies the Lawrencetown Till with an erosive contact (Fig. 64).

The Lawrencetown Till is divided into three textural facies (LT/A, LT/B and LT/C). Facies "LT/A" is a reddish, muddy diamicton with a high percentage of foreign (non-Meguma Zone) lithologies. It averages 20% gravel and 80% matrix regionally with a clast/matrix ratio of 0.3 (Table 8). The percentages of Meguma Group and foreign lithologies are nearly equal (47 and 49% respectively; Table 8). Facies "LT/C" is a brownish, sandy diamicton containing an average of 27% gravel and 73% matrix (sand+mud). It has generally higher percentages
Figure 64. (A) Three tills exposed in a drumlin section near Smith Point, Nova Scotia. The lowest unit is the Hartlen Till, the middle unit is the Lawrencetown Till and the upper unit the Beaver River Till (ST/B) (B) Beaver River Till (ST/B) overlying oxidized Hartlen Till at the Wine Harbour Section. Note the abundant boulders in the younger Beaver River Till.
Figure 64.
of Meguma pebbles (70%, Table) and less foreign or erratic lithologies (26%, Table 8). Facies "LT/B" is intermediate in lithic properties between these two "end-members" (Table 8). In many drumlin sections a tripartite diamicton sequence is recognized, with the Beaver River Till (ST/B) overlying the Lawrencetown and Hartlen Tills (Fig. 64). Grant (1963) originally interpreted these diamicton units as tills formed by separate glaciations, with contacts (including boulder pavements and thin waterlain units) representing interstadial intervals. Nielsen (1976) described this widespread three-till sequence as a single depositional package from one ice flow encompassing lodgement, and englacial and supraglacial melt-out tills (Nielsen, 1976). The author's present view is a compromise of these positions, with each till unit assigned to separate phases of ice flow from evolving ice centres and divides without intervening nonglacial intervals (Stea et al., 1992 a, c).

4.3.3 West Lawrencetown Reference Section

The West Lawrencetown section is in a wave-eroded drumlin located 3 km east of Cole Harbour near Halifax (Fig. 1, Section 56-Map 1), first described by Grant (1963) and later by Nielsen (1976). The drumlin has a v-shaped, lobate form with a main axis trending 150° and another axis trending 185° (Fig. 65). Between the two lobes is a circular depression, partially eroded by the sea with a base resting on a lower diamicton unit. The depression was originally interpreted as a kettle (Grant, 1963, p. 107). Bedrock outcrop on the western flank of the drumlin complex reveals wide, parallel grooves 10-40 cm in width and striae that trend 155°. The sense of flow is determined by nailhead striae (Prest, 1983). Another distinct set of slightly divergent, finer striae trending 180-190° cut across the older grooves inscribing only the north-facing slopes (cf. Lowell et al., 1990). Three sections were profile-sampled in order to assess local-scale lithic variations in the till sheets.
Figure 65  Stratigraphy, till fabric, and locations of detailed profiles (1, 2 and 3) at the West Lawrencetown reference section (Fig 1, Section 56, Map 1)  Arrows on the map delineate distinct lobes of the drumlin. Bedrock striae are shown on the upper map. Fabric 1 is located beside profile 1 (black square). Fabric 2 is located near profile 2. Till fabrics are represented by contoured lower-hemispheric, equal-area projections of the trend and plunge of the a-axes of 50 elongate stones (3 1) (modified after Stea, 1994). Black dots represent locations of stoss/lee (bullet) boulders with directional data presented in Stea (1994).
4.3.3.1 Stratigraphy and Lithology

The southern, tapered end of the drumlin has been truncated by the sea. Two exposures separated by the inter-lobe depression reveal two compositionally-distinct diamicton units termed the Hartlen and Lawrencetown Tills (Fig. 65). The Hartlen Till was originally designated by Stea and Fowler (1979) for the indurated "core" diamicton of drumlins along the Atlantic Coast whose type section is at Hartlen Point a few km east of Cow Bay (Section 55-Map 1). The Lawrencetown Till defined by Grant (1975) represents the reddish drumlin "mantle" till. The Lawrencetown Till type section is designated as Sandwich Point, south of Halifax (Section 53-Map 1, Grant in Williams et al., 1985).

The Hartlen Till is a grey, matrix-supported, silty diamicton. It is compacted or overconsolidated, forming steeper slopes (60 degrees) than the overlying Lawrencetown Till. Bullet-shaped, stoss-lee boulders embedded within the Hartlen Till and at the Hartlen-Lawrencetown Till contact (Fig. 65) trend parallel to the major drumlin axis (Stea, 1994). Fabrics in the Hartlen Till trend parallel to the main drumlin axis and local bedrock striae (Fig. 65). Pebble counts in the Hartlen Till average 59% metagreywacke, (derived from of the underlying Cambro-Ordovician Meguma Group), 23% whitish, megacrystic Meguma Zone (MZ) granitic pebbles and 17% erratic (non Meguma Group) pebbles (Fig. 66). Erratic lithologies include red and grey siliciclastic rocks, pink, alkali-feldspar granitoids, diorites and felsic and mafic volcanic rocks. This pebble assemblage is derived from the Cobequid Highlands (Stea and Mott, 1990; Stea, 1994, Table 2). The percentage of MZ granite and erratic pebbles increase to 40 and 20% respectively towards the top of the Hartlen Till in profile 3 (Fig. 66). Gravel percentages in the Hartlen Till vary from 19.4 to 32.5% (Figs 66, 67). Clast/matrix ratios in the Hartlen Till at this section vary from 0.2 to 0.4 (Appendix 9). Striae percentages vary from 4-13%.
Figure 66  (Top) Simplified geological map of central Nova Scotia showing the upland igneous and metamorphic complexes of the Avalon and Meguma Zones and the pebble assemblage source areas for the Hartlen and Lawrencetown Tills (Facies A) from Stea, (1994)  (Bottom) Stratigraphy and vertical variation in pebble lithology and texture (clast/matrix ratio) for profile 1 (see Fig 65 for location) Flow directions for till units are indicated by erratic boulder assemblages in the Lawrencetown Till (after Stea, 1994) The patterned region between sample points on the pebble graph represents the variability between duplicates counts on the same sample
A reddish, muddy, matrix-supported diamicton (Lawrencetown Till Facies A) overlies the Hartlen Till with a knife-sharp contact. It is distinguished from the Hartlen Till by its red colour and high erratic content (Figs. 66, 67). The Lawrencetown Till can be divided into an upper, sandy (LT/C) and a lower clay-rich facies (LT/A) with sand inclusions. It is thickest in the southern lobe section of the drumlin, whereas the Hartlen Till is thickest in the southeastern lobe of the drumlin (Fig. 65). LT/A exhibits a southward-trending till fabric, parallel to the youngest striae set on the adjacent bedrock outcrop and lobate drumlin axis.

In all profiles sampled, erratic pebble percentages increase dramatically in the Lawrencetown Till concomitant with the decline in MZ granite and Meguma Group metasediment percentages (Figs. 66, 67). The erratic fraction is dominated by distinctive hornblende-bearing, equigranular, salmon-pink granites with varying textures, followed by granodiorite-diorites, and mafic and felsic volcanics in order of abundance. Table 9 is a list of erratic lithologies visually identified by Dr. H.V. Donohoe (pers. comm., 1991) from a large sample (30) of sectioned boulders. The source of the plutonic and volcanic erratics in the Hartlen and Lawrencetown tills is likely to be the eastern Cobequid Highlands, 80 km north of the section (H.V. Donohoe, pers. comm., 1991, Fig. 66). The dominance of granites and diorites would preclude the Antigonish Highlands as a possible source (Table 2). Gravel weight percentages in LT/A vary from as low as 8% to 35% (Figs. 66, 67). The clast/matrix ratio varies from 0.1 to 0.5 (Appendix 9). Clast striae percentages average 12%. LT/C exhibits a marked increase in the percentages of local Meguma Group pebbles (av. 57%, Figs. 66, 67) and a sympathetic increase in clast/matrix ratio (0.3 - 0.4, Fig. 66). Striae percentages vary from 7-11% in this facies (Figs. 66, 67).
Table 9. Lithologies from visual identification of sectioned boulders from the West Lawrencetown Section.
Figure 67  Stratigraphy and vertical pebble lithology variations in profiles 2 and 3 (see Fig. 65 for location)
Figure 67.
PM-1 3½" x 4" PHOTOGRAPHIC MICROCOPY TARGET
NUS 1010a ANSI/ISO #2 EQUIVALENT

PRECISIONSM RESOLUTION TARGETS
4.3.4 Wine Harbour Reference Section

The Wine Harbour reference section is located 180 km northeast of Halifax along the Eastern Shore at Wine Harbour Bay (Fig 1, Section 64-Map 1) Two drumlins are cut transverse to 160°-trending long axes The section is located on the easternmost of two drumlins Drumlin long axes trends vary from 150° to 175° within the region of the sections and in the Country Harbour drumlin field (Map 1) Bedrock striations mapped locally also show three main trends, the oldest sets trending 150° to 160°, followed by minor sets of parallel striations with mean trends of 133° and 179°

4.3.4.1 Stratigraphy and Lithology

The Wine Harbour Section has been previously described by Stea and Fowler (1979) and Stea and Brown (1989) They recognized four main drumlin diamicton unit from base to top

1 A greyish-red to olive-grey, matrix-supported, silty diamicton with few stones (Unit 1)
2 Recumbently-folded laminated silt, sand and clay strata (Unit 2, Fig 68) which may underlie Unit 1, but the contact relationships are unclear
3 A brown, stony, silty diamicton (Unit 3) which comprises most of the drumlin
4 A reddish, sandy diamicton (Unit 4)

The section was revisited by the author in 1991 and 1992 Unit 3 was further subdivided into two units, a brown, matrix-supported, sandy-silt diamicton (Unit 3) which comprises much
of the drumlin and a sandy diamicton (Unit 4) with a higher clast/matrix ratio, which can be found on both flanks of the drumlin in gradational contact with Unit 3 (Fig 68). Separating Units 1 and 3 east of the drumlin core are lenticular pods or inclusions of silty-sand with deformed strata. The reddish diamicton is now termed Unit 5.

Unit 1 has the lowest percentage of local Meguma metagreywacke pebbles (57%). Local pebble percentages increase in Unit 3 to 66-72% (Fig 69). Unit 4 has the highest percentage of local Meguma Group pebbles (82%). Meguma Zone granites are only found in Unit 3 from a sample in the centre of the drumlin attaining 3% of the entire pebble fraction (Fig 69).

Erratic pebbles present in the Wine Harbour section fall into 6 main lithological groups (Fig 70):

1. Hard, maroon to red mudstone
2. Greenish and grey, coarse, feldspathic arenite
3. Red-orange-purple-grey porphyritic felsic and intermediate volcanic rocks
4. Reddish, altered, porphyritic granitoid rocks
5. Hard, green and black, finely laminated mudstone

Unit 1 has the highest erratic percentages in the section (43%), dominated by volcanic rocks and hard, red mudstones, typical rocks of the Antigonish Highlands (Fig 70, Table 2). The Fraser Brook member of the Keppoch Formation in the Antigonish Highlands is the probable source of red mudstones and volcanics (Murphy et al., 1991). Green, hard, finely-laminated...
Figure 68  Stratigraphy of the Wine Harbour section (Fig. 1, Section 64-Map 1). (A) Photo log of the eastern half of the drumlin at Wine Harbour Bay. (B) Close up of folded and boudinaged sand beds and silt and clay laminac (after Stea and Brown, 1989).

(C) Photograph of the eastern part of the drumlin exposure showing Units 1 (Hartlen Till) and 4 (Beaver River Till, see Fig. 64 B)  (D) Interpretation of the drumlin section.
Figure 69. Pebble lithology of selected samples throughout the Wine Harbour Section. Pie diagrams show the percentages of local Meguma Group pebbles (metagreywacke), and the percentages of "foreign" and Meguma Zone granites (SMB-South Mountain batholith-type granites).
Figure 70  (Top) Dispersal of pebble and boulder assemblages to sample points along the Eastern Shore study area (Legend on Fig. 66) (after Stea and Murphy, 1994). Areas of geology in black are appinite-diorite complexes. Source areas are defined by pebble-boulder assemblages, distinctive erratics (ie hornblendite, appinite) and boulder whole-rock geochemistry (Stea and Murphy 1994) (Bottom) Structural data from the Wine Harbour Section. Till fabric is represented by lower-hemispheric equal area projections of the trend and plunge of the a-axes of elongate stones (3:1) and simple rose diagrams depicting the long axis orientations of pebble-sided stones in till. Some structural data from the folded clay unit at the core of the drumlin (after Stea and Brown, 1989). Boulder long axis and upper surface striae orientation are depicted by striae symbols above boulders.
Boulder source areas

ANTIGONISH HIGHLANDS

PHASES 2+3

WINE HARBOUR SECTION

Figure 70.
siltstones occur in the Moose River Member of the Keppoch Formation. White, megacrystic monzogranite or granodioritic lithologies are probably derived from the Sherbrooke Pluton, 20 km to the northwest (Fig. 70). The flow direction of Unit 1 interpolated from this pebble assemblage, is between 150° and 160°. This is parallel with the drumlin axis, and the oldest striation set on adjacent rock outcrops. Pebble fabric in the unit, however, shows a southwest orientation, nearly perpendicular to the presumed ice flow based on provenance (Fig. 70).

Stea and Brown (1989) interpreted this paradox as contour deflection of the formative ice around the incipient drumlin form. This anomalous fabric orientation persists into the drumlin core itself. A more likely explanation for the discrepancy of provenance and boulder and pebble fabrics is fabric overprinting or reorientation by a late southwestward (Phase 4) ice flow (Fig. 4), defined regionally by striae, erratic transport and esker orientations (Stea et al., 1992a, Map 1).

Unit 3 has lower erratic percentages than Unit 1 (28-37%) but the same pebble types are represented. The flow direction of Unit 3, interpolated from this pebble assemblage, is also between 150° and 160°. Glaciotectonic deformation in laminated, fine-grained sediments at the base of the drumlin section indicate a southeastward flow (Stea and Brown, 1989), but pebble fabric and boulder orientations are largely southward-trending.

Units 4 and 5, the uppermost till units, have the lowest percentage of erratic lithologies (18-28%) probably reworked from the older units. Till fabric, boulder striae and orientation indicate a southwestward ice flow (Fig. 70). Unit 4 can be correlated to adjacent ground moraine regions overlain by Beaver River Till (ST/A). Bedrock striations underneath sections of ST/A and till-embedded boulders are oriented 170° to 180°. The red coloration of Unit 5 may be derived from flow across red-beds within Carboniferous basins to the north and northeast.
Units 4 and 5 formed during late southward and southwestward flow events mapped regionally (Ice Flow Phases 3 and 4, Fig. Map 1). The decline in foreign pebble percentages is consistent with reworking of the underlying erratic-rich till units (1 and 3).

4.5 CORRELATION OF EASTERN SHORE TILL UNITS

A correlation diagram of Eastern Shore sections described here and on the Surficial Map of Nova Scotia (Map 1) is presented (Fig. 71). Compacted, drumlin-base and core diamicton units with low clast/matrix ratios and moderate to high erratic contents are all correlated with the Hartlen Till at the type and reference sections at Hartlen Point (Map 1) and West Lawrencetown. At several sections, waterlain units and boulder horizons (pavements) separate the Hartlen Till from later till units. The Lawrencetown Till comprises all the reddish to brown, erratic-rich diamicton facies (LT/A/B/C) which overlie the Hartlen Till. They are distinguished from the Hartlen Till by stratigraphic position, less consolidation, higher erratic content, sand-silt interbeds and inclusions (Table 8). Drumulins throughout the Eastern Shore region display varying proportions of the Lawrencetown Till facies (Stea and Fowler, 1979) but all three facies are generally present within individual drumulins and drumlin fields. The stony, locally-derived surface till (Beaver River Till) at Mushaboom, Smith Point (4 km east of Section 63, Map 1) and Holland Harbour (Section 65, Map 1) is the next youngest till unit. It is lithologically and texturally distinct from the drumlin Lawrencetown and Hartlen Tills (Table 8). In areas remote from drumlin fields the Beaver River Till is compositionally uniform, consisting of 90% or greater local or underlying bedrock. It is also distinguished from the drumlin tills by angular to subangular clast-shapes and few striated or bullet-shaped clasts (Table 8).
Figure 71   Correlation of major sections along the Eastern Shore of Nova Scotia: 1. West Lawrencetown  2 Smith Point east (1)  3 Smith Point west (Fig 64)  4 Wine Harbour  5 Holland Harbour (Section 65-Map 1)  6 Mushaboom

1  HT = Hartlen Till
2  LT/A/C = Lawrencetown Till facies A and C
3  ST/A/B = Beaver River Till facies A and B
ST/A  }  Beaver River Till (end member (A) and inherited (B) facies)
   loose, slaty, locally-derived, clast-matrix-supported, sandy diamictite

ST/B  }  Lawrencetown Till (end member (A) and hybrid facies (C))
   moderately compacted, polymictic, matrix-supported, reddish-brown to brown, sandy-silty diamictite

LT/C  }  moderately compacted, polymictic, grey-brown, silty diamictite

HT  }  Hartlen Till
   compact, polymictic, grey-brown, silty diamictite
4.6 AGE

The till sequences along the Eastern Shore are interpreted to span the entire Wisconsinan (70-11 ka). A correlative of the Hartlen Till in central Nova Scotia overlies forest beds believed to encompass the Sangamon Interglacial interval (Mott and Grant, 1985; Sections 117, 120-Map 1) On the mainland of Nova Scotia, organic horizons have not been found within or between till layers, indicating continuous ice cover throughout the last glacial period (Stea et al., 1992c) Mid-Wisconsinan interstadial inter-till fossiliferous and organic sediments have only been found in northern Cape Breton Island (Grant, 1989).

4.7 INTERPRETATION

Fabric, striation and provenance data strongly imply that the drumlin diamictons (Hartlen and Lawrencetown Tills) were formed by successive, divergent ice flows from differing ice centres. These drumlin diamictons are interpreted as subglacial, primary till based on over-consolidation, abundant striated bullet-boulders, and strong, unidirectional till fabrics (Dreimanis, 1989). Drumlins are largely relict from an early southeastward ice flow (Ice Flow Phase 1, Fig 4) During Ice Flow Phases 1a and 1b, southeastward ice flows from an Appalachian source glacier produced a fine-grained, matrix-dominated till (Hartlen Till) as ground moraine and drumlins (Fig 4). Some of the fine-grained matrix material may have been derived from interglacial residuum, and perhaps pre-existing glacio-lacustrine deposits (Fig 72).

Some drumlins are palimpsest forms, moulded from proto-drumlins composed of the Hartlen Till during a southward flow that formed the Lawrencetown Till (Ice Flow Phase 2; Stea,
The southward ice flow brought the "flood" of red, fine-grained material from the red beds in the Maritimes Basin north of the Cobequid Highlands (Boehner et al., 1986) to the Atlantic Coast. As the ice sheet crossed the Cobequid Highlands, red, fine-grained material, and highland clasts were shunted to an englacial position by upward vectors of flow and regelation (Boulton, 1975) or shearing (Sugden and John, 1976, Fig. 72).

The autochthonous, Beaver River Till (ST/A/B/C) represents the last stages of ice formation during the Late Wisconsinan Late, southeastward ice flow events (Ice Flow Phase 3) formed the Beaver River Till and associated stony till plain. Striae associated with Phase 3 are hard to differentiate from relict, Phase 1b southeastward patterns (Fig. 4). Coker et al. (1988) and MacEachern and Stea (1986) defined southeastward (140° to 160°) dispersal trains from known ore bodies of gold and related elements within the Beaver River Till. Phase 3 flow appears to have shifted to the south in the vicinity of the Wine Harbour section.

Ice Flow Phase 4 produced southwestward-trending striae, eskers and pebble-dispersal trains (Turner and Stea, 1987, Stea et al., 1989) in northeastern Nova Scotia (Map 1). This flow produced thin, discontinuous, stony tills (Unit 5-Wine Harbour) and a few constructional till-landforms.

Diamictons termed the Beaver River Till are interpreted to be subglacial till based on these characteristics:

1. Local derivation and ubiquitous areal extent (MacEachern and Stea, 1986)

2. Rapid transition from one dominant lithology to another across bedrock.
contacts (Peltoniemi, 1985; Dreimanis, 1989). "Renewal" distances are generally less than 1 km (Graves and Finck, 1988).


A subglacial hypothesis is not supported by the lack of striated clasts and clast angularity. Stea et al. (1989) suggested that the contradictory properties of the Beaver River Till are an indicative of proximity to an ice divide, well documented over Nova Scotia (MacNeill, 1951; Hickox, 1962; Stea, 1984; Grant, 1989; Graves and Finck, 1988; Stea et al., 1992c). It may have formed underneath a thermally-variable glacier, with cold-based and warm-based zones (Hughes, in Denton and Hughes, 1981). This divide was short-lived, so the lack of significant transportation and abrasion may be a function of time within the traction zone or "maturity" or in the sense of Dreimanis and Vagners (1971). Basal velocities in the ice divide environment are low, hence transport in the tractional zone is limited. The Beaver River Till can be termed an "ice divide till" (Fig. 72).

Drumlins are rarely overlain by the lithologically homogenous, Beaver River Till, but rather "swim" within the stony till plain (Grant, 1963). A possible explanation for this phenomenon may be increased basal ice velocities due to sole lubrication (meltwater trapped in a low permeability clay till) and substrate deformation (Boulton and Jones, 1979; Clark, 1992; Hart and Boulton, 1992). Relatively high ice velocities will decrease mixing rates and maintain debris in transport (Clark, 1987). Lawrencetown Till facies C, is a result of "overprinting" by the ice from Scotian Ice Divide (Ice Flow Phase 3-Fig. 4). It is produced as fast-moving ice as it deforms the earlier drumlin tills (Fig. 72).
Figure 72  Schematic cross sections showing the genesis and source glaciers for the Eastern Shore till units

1 During Phase 1a and 1b, southward flow from Appalachian source glaciers (Fig. 4) produced a fine-grained, matrix-dominated till (Hartlen Till) as ground moraine and molded into drumlins. Some of the fine-grained matrix material may have been derived from interglacial residuum, and perhaps pre-existing glacio-lacustrine deposits.

2 Ice Flow Phase 2 stemming from the Escuminac ice Centre (Fig. 4), produced a reddish, fine-grained, matrix-dominated till (Lawrencetown Till) that mantles drumlins. This till is largely derived from englacial debris, produced when the glacier crossed the Cobequid Highlands to the north. The Lawrencetown Till also contains inherited components of the Hartlen Till, derived from local erosion.

3 During Ice Flow Phase 3 an ice divide developed over Nova Scotia itself, and a distinctive, sandy, clast-dominated till was produced. This "ice divide till" is characterized by local transport, and lack of abrasion. Drumlin tills were "overprinted" with local rocks creating Lawrencetown Till facies C. Ice Flow Phase 4 was produced from reactivated glaciers remnant from the Scotian Ice Divide and produced thin, discontinuous, tills similar in texture to the Beaver River Till.
CHAPTER 5: CORRELATION OF MARINE AND TERRESTRIAL GLACIAL EVENTS ALONG THE EASTERN SHORE

5.1 INTRODUCTION

In the last chapter till-sheets were correlated and then linked to regional ice flow phases. All these ice flow events extended past the present day shoreline of Nova Scotia (Fig 4; Map 1). In this next chapter the author will attempt to correlate the terrestrial till sheets with offshore tills and morainal landforms. In this manner these former ice sheets can be reconstructed with both flowlines and margins. A complicating factor in this correlation is the modification of seabed landforms and deposits between the modern shoreline and 70 m BSL by the post-glacial transgression.

5.2 LAND-SEA CORRELATION BASED ON ICE FLOW PATTERNS

Four ice flow phases can be traced to the present shoreline (Fig. 4). What evidence exists for ice movement beyond the present coastline, considering that the marine transgression may have removed former morainal margins? Erratics are present in almost all of the offshore diamicton samples (Figs. 6; 73). Dioritic, granitic and volcanic erratics are distributed in two distinct dispersal fans; southeastward (Phase 1b) and southward (Phase 2) (Figs. 4; 73). The glaciers that formed these fans must have traversed the inner shelf, as most erratics have been transported at least that far.
Figure 73. Erratic pebble percentages in marine and terrestrial samples (Fig. 6). Triangles represent the relative abundances of each pebble class. The source area geology is simplified after Keppie and Muecke (1979). The maximum extent of distinctive erratic assemblages is shown by arrows from lines connected to the upland source regions. Southeastward transport is indicated by the extent of North Mountain basalt pebbles (Ice Flow Phase 1). Dioritic, volcanic and granitic pebble assemblages down-ice from the Antigonish and Cobequid Highland source areas indicate southward ice flow (Ice Flow Phase 2).
It is generally true that end moraines are oriented normal to the direction of ice flow except at lobate margins (Flint, 1971). The prominent northeastward trend of the Scotian Shelf End Moraine Complex (Fig. 3) is compatible with two flow events, Ice Flow Phase 1b, and Ice Flow Phase 3 (Fig. 4). Ice Flow Phase 1b is a regional glaciation that crossed the Bay of Fundy and probably extended to the shelf edge based on erratic transport (Grant and King, 1984). Ice Flow Phase 3 is the most likely candidate, but if buoyancy is a major factor in moraine formation then the flow orientation may not solely define the trends of moraines. Southward ice flow during Phase 2 may have been restricted to fast-moving ice streams (cf. Grant, 1963) that disintegrated rapidly, leaving no depositional record. Long-distance transport and the restricted areal extent of the Lawrencetown Till is compatible with fast ice flow and ice streams (Denton and Hughes, 1981).

The symmetrical and asymmetrical moraines of the Morainal Zone are oriented NW-SE, indicating a change in ice flow directions from the prominent NE-SW trend of the Scotian Shelf End Moraine Complex (Fig. 16; Map 2). Southwestward-trending channels are found at the distal end of the Morainal Zone. Ice Flow Phase 4, a southwestward flow which postdates Phase 3, is normal to this morainal complex (Map 1; Figs. 3, 4) and can be traced offshore. The Morainal Zone is interpreted as a marginal depositional zone of this ice flow, centred in northeast Nova Scotia. The NW-SE portion of the Country Harbour Moraine may have also formed during Ice Flow Phase 4.

Patterns of ice flow have been interpreted from the orientation of deeply-incised, "tunnel" valleys on the outer Scotian Shelf near Sable Island Bank (Boyd et al., 1989; Loncarevic et al., 1992; Fig. 74). The origin of these valleys is controversial. Previous authors have suggested glacial erosion of pre-existing river valleys (King et al., 1974) and subglacial
Figure 74. Digital bathymetry shadowgram of the Scotian Shelf southeast of Cape Breton Island (after Loncarevic et al., 1992). Some tunnel valleys are indicated by white lines. (A) Sable Island Bank and morainal complex. (B) Banquereau (C) Note the lack of tunnel valley incisions in the presumed offshore continuation of the Scotian Ice Divide (Map 1; Fig. 4).
Figure 74.
meltwater erosion (Boyd et al., 1983, Loncarevic et al., 1992) Cross-cutting tunnel valleys suggest several glacial configurations (Loncarevic et al., 1992, Fig 74)

The E-W orientation of Sable Island Bank and southward-trending tunnel valleys imply a southward-flowing ice sheet (McLaren, 1988) These flow patterns are consistent with Ice Flow Phase and an offshore continuation of the Scotian Ice Divide (Fig 4) Northwest-trending tunnel valleys on the shelf west of Cape Breton Island are parallel to northwestern ice flow patterns on Cape Breton (Grant, 1989) The undissected part of the shelf (C-Fig 74) between the incised zones may be the location of the Scotian ice Divide The northern edge of Sable Island bank may correspond with recessional margins of the Late Wisconsinan Escuminac Ice Centre and Scotian Ice Divide

5.3 STRATIGRAPHIC CORRELATION

Three seismic units can be discerned within the Scotian Shelf End Moraine Complex (Fig 17) Three incoherent acoustic units have also been noted within the nearshore basins of the inner shelf by Hall (1985), who correlated these units with three major till units on land near Halifax Piper et al. (1986) defined and sampled two massive acoustic facies on the inner shelf off Mahone Bay termed the "Late and "Early" tills They correlated the shore-proximal "Late" till with the Beaver River Till and the distal "Early" till with the drumlin tills The tripartite acoustic divisions within the Eastern Shore Moraine (Fig 17) may be correlated with the three-till land sequences, the uppermost till in the moraine sequences being equivalent to the Beaver River Till "facies B" (ST/B) (Figs. 64, 71) The significant percentages of erratics within the Scotian Shelf End Moraine Complex samples (Tables 4, 8) suggests erosion of underlying Lawrencetown and Hartlen Tills.
These moraines may be palimpsest, sites of readvances or stillstands during several phases of ice flow. Earlier accumulations of till would serve as pinning points for later readvances (Powell, 1984).

Grant and King (1984) correlated both the Hartlen and Lawrencetown Tills with the Scotian Shelf Drift (Table 1) and proposed a shelf-wide, extensive southeastward glaciation in the Early Wisconsinan described as the "great red flood". They did not recognize the independent, southward flow (Phase 2) first elucidated by Chalmers (1895) then by Goldthwait (1924) and later by Prest et al. (1972). According to their model, after the major Early Wisconsinan glaciation, Late Wisconsinan ice extended slightly past the present coastline, not as far as the Scotian Shelf End Moraines. I interpret the Scotian Shelf Drift as a composite of three glacial diamictons, formed from different ice centres spanning the Wisconsinan age.

5.4 MORPHOLOGIC CORRELATION

The offshore Morainal Zone and the Stony Till Plain are correlative based on the presence of ribbed moraine (Fig. 41) and coarse, bouldery, surface deposits (Fig. 36). The Stony Till Plain, which occupies most of the Eastern Shore study region, is expected to have an offshore counterpart. Ribbed moraines on land have not been widely recognized, but may be more common than realized. Ridges with amplitudes of less than 5 m, readily visible on the Hunttec and the multibeam bathymetric records offshore (Figs. 40, 41) may not be apparent on land due to tree cover. After deglaciation, land moraines were deflated by wind and water. Offshore moraines were spared subareal erosion below the transgression.
Drumlins are evident in the multibeam bathymetric image in the region south of Peggy's Cove (Loncarevic et al., 1994, Fig 51). Below -65 m, they are similar in size and shape to land drumlins of the Lunenburg Drumlin Field (Fig 51, Map 1). Ribbed moraines are draped on the drumlins, indicating that they formed later. These offshore drumlins and ribbed moraines indicate that drumlin and till-plain-forming Ice Flow Phases (1-3) extended well offshore.

5.5 LITHIC CORRELATION

The lithic composition of the inner shelf diamictons is uniform (Fig 75). Offshore diamictons are generally olive-grey with some samples exhibiting a brownish hue (Appendix 3). The coloration reflects derivation from greenschist facies Meguma Group metasediments (Stea and Fowler, 1979). Other than isolated inclusions, no reddish matrix samples comparable in colour to the Lawrencetown Till were obtained (Appendix 3). The colour of offshore diamictons implies a correlation with the Beaver River Till.

The pebble lithology of offshore diamictons is also compatible with the Beaver River Till. Offshore sediments consist of 75-98% autochthonous (local bedrock) components with a mean value of 86% (Tables 4, 8). Local rocks consist mainly of Meguma Group metawacke, except for samples from the Country Harbour Moraine and adjacent till-tongues, which are underlain by a wedge of gently-dipping Mesozoic or Cenozoic rocks (visible in the sleevegun records) and contain mostly brown marlstones (King and MacLean, 1974). Mean Meguma pebble percentages in the Beaver River Till facies vary from 77-86% (Table 7). The ground moraine facies of the Beaver River Till (ST/A) contains an average Meguma pebble percentage of 86%. A box and whisker plot of foreign/local clast percentage ratios...
Figure 75. Contour plot of Meguma Group metasedimentary pebble percentages in samples from the study area. Note the uniformity of the inner shelf diamictons and the lobes of erratic-rich till associated with drumlin fields onshore (Map 1). Also note the down-ice uptake of Mesozoic/Cenozoic brownish marlstone clasts in the marine diamictons south of the Meguma-Mesozoic contact (see also Fig. 2, indicating glacial erosion of the substrate, and supporting a primary or secondary till-origin for these diamictons.
graphically and statistically demonstrates the similarities between the Beaver River Till (ST/A/B/C) and the marine diamictons (Fig. 76). Offshore sediments in all mapped terrain zones of the inner shelf have significantly different lithological ratios than the drumlin Lawrencetown and Hartlen Tills and are statistically indistinguishable from the Beaver River Till (Tables 4, 8, Fig. 76).

The lack of a reddish, muddy component and low foreign content precludes correlation of the offshore diamictons with the drumlin Lawrencetown and Hartlen Tills. Down-ice dilution of foreign components cannot explain lithological differences for these reasons.

1. Lawrencetown Till samples on land show a negligible down-ice uptake of underlying bedrock (P. Finck, pers. comm., 1994). Regional mapping of clast lithology over the South Mountain batholith revealed a considerable "skip zone" with little uptake of local rocks in the Lawrencetown Till (Finck and Stea, 1990, Finck et al., 1994 a,b,c,d,e). "Renewal distances" can exceed 30 km. A regression curve of Lawrencetown Till samples over the South Mountain batholith in southern Nova Scotia (Fig. 2) show negligible uptake of local rocks (Fig. 77). The predicted foreign clast percentage in the Lawrencetown Till at the Scotian End Moraine complex is between 33 and 12% (Fig. 77). This is a conservative estimate because only surface Lawrencetown Tills were used in this analysis and these are largely the overprinted LT/C type.

2. Lack of red matrix components in the offshore diamictons. The red colour persists in land sections 50 to 100 km south of the Carboniferous source regions and out to the edge of the shelf where anomalous red mud has been
Figure 76. Notched Box and whisker plots of foreign/local clast percentage ratios plotted for terrestrial till facies and offshore diamictons. The box and whisker plot is an effective way to demonstrate summary statistics graphically. The notched plot (McGill et al., 1978) corresponds to the confidence interval of the median, whereas the width of the box is proportional to the square root of the number of observations in the data set. If the two notches on the boxes do not overlap the median values of the data distributions do not significantly overlap at a 95% confidence level. The dashed lines show the overlap boundaries of the drumlin tills.

Sedimentary Units: Onshore Tills
1. Hartlen Till
2. Lawrencetown Till facies A
3. Lawrencetown Till facies B
4. Lawrencetown Till facies C
5. Beaver River Till facies A
6. Beaver River Till facies B
7. Beaver River Till facies C

Offshore diamictons and gravels
8. Emerald Silt
9. Morainal Zone (Scotian Shelf Drift)
10. Scotian Shelf End Morainal Complex (Scotian Shelf Drift)
11. Sable Island Sand and Gravel
12. Unit B
13. Unit C
Figure 77. Regression plot of erratic percentages in Lawrencetown Till vs. distance down-ice from the Carboniferous basin/Meguma Group bedrock contact in southern Nova Scotia. Data from a regional till survey over the South Mountain Batholith (Graves and Finck, 1987a, b; Finck et al., 1993, 1994a,b,c,d,e). The regression curve of Lawrencetown Till samples over the South Mountain batholith in southern Nova Scotia (Fig. 2) shows slow uptake of local rocks, an indication of englacial transport. The predicted foreign clast percentage in the Lawrencetown Till at the Scotian End Moraine complex is between 33 and 12% with a median of 20%. This is a conservative estimate because only "overprinted" surface Lawrencetown Tills were used in this analysis.
MEGUMA GROUP BEDROCK

Scotian Shelf End Moraines

ONSHORE

OFFSHORE

Figure 77.
attributed to glacial transport (Hill, 1981, Mosher et al., 1989)

3 Homogeneity of composition throughout the inner shelf. Erratic dilution cannot be discerned from proximal to distal areas of the inner shelf. In the Truncation Zone less competent pebbles may have been preferentially destroyed by the transgression, but a significant proportion of exotic clasts are competent lithologies. Piper et al. (1986) found that the petrology of nearshore gravels were indistinguishable from parent tills onshore.

The stratigraphic position of the Hartlen Till would preclude correlation with seabed-outcropping marine diamictons because it is invariably overlain by the Lawrencetown Till and sometimes the Beaver River Till in land sections. The small (<10%) but significant percentages of non-Meguma lithologies in most morainal diamicton samples of the inner shelf, indicate that an underlying or nearby source of Lawrencetown or Hartlen Till has been eroded. "End member" ground moraine facies of the Beaver River Tills (ST/A), removed from drumlin fields, contain virtually no foreign lithologies, but hybrid facies (ST/B), contain erratic percentages of greater than 10% (Table 8, Fig. 62). Erratics within the "hybrid" Beaver River Till can be attributed to the proximity of Lawrencetown Till drumlins or subcropping Lawrencetown Till. It is probable that the Lawrencetown Till extends out at least as far as the Scotian Shelf End Moraine Complex. Significant grain size and density differences between the Beaver River Till and the Lawrencetown and Hartlen Tills would account for acoustic impedance boundaries as observed in the sleevegun profiles of the Eastern Shore Moraine (Fig. 17).
Two factors which make the correlation between marine and terrestrial tills using grain size parameters difficult are

1. Incorporation of marine muds (especially at grounding-line) by glacial erosion

2. Sampling difficulties and loss of fines upon IKU retrieval. This does not apply to matrix-rich moraine samples (Appendix 3) which were retrieved without loss of fines, however gravel percentages in Sable Island Sand and Gravel samples are anomalously high.

A box and whisker plot of the clast/matrix weight percentage ratios for the entire data set is shown on Figure 78. The Beaver River Till is clearly distinct from other onshore tills. The Sable Island Sand and Gravel (Truncation Zone) and Morainal Zone samples are statistically distinct from the other marine sediments. The Truncation Zone lag gravels are clast-rich because fines have been removed by wave action and sample retrieval. The Morainal Zone "lump" or "matrix" samples (Appendixes, 3, 4) should be representative of the texture of the true morainal material. Textural overlap with the Beaver River Till is apparent.

Detailed grain size analysis verifies the textural similarity between the Beaver River Till and the Morainal Zone samples and suggests similarity between the Beaver River Till and the Scotian End Moraine surface diamicton samples. The Beaver River Till and offshore morainal samples overlap in the fine sand and silt size grades (3-5 phi, Fig. 14) but are deficient in clay (>8 phi, Fig. 14) unlike the clay-rich drumlin tills.
Figure 78  Notched box and whisker plots of clast/matrix weight ratios plotted for terrestrial till facies and offshore diamictons  See Figure 76 for box legend

Sedimentary Units  Onshore Tills

1  Hartlen Till
2  Lawrencetown Till facies A
3  Lawrencetown Till facies B
4  Lawrencetown Till facies C
5  Beaver River Till facies A (ST/A)
6  Beaver River Till facies B (ST/B)

Offshore diamictons and gravels

7  Emerald Silt
8  Morainal Zone (Scotian Shelf Drift)
9  Scotian Shelf End Morainal Complex (Scotian Shelf Drift)
10  Sable Island Sand and Gravel
11  Unit B
12  Unit C
Figure 78.
The Scotian Shelf End Moraine samples have lower clast contents than the Beaver River Till and are comparable with the drumlin tills, probably as a result of incorporation of Emerald Silt at or behind the grounding line (Figs 14, 78).

The highest percentages of striated clasts occur in the Scotian Shelf Drift and the lowest in the Beaver River Till (Tables 4, 8, Fig 79). Assuming correlation of these two units, this discrepancy can be explained by differing basal processes of till formation between the Scotian Ice Divide and its margin (Fig 4). Outward from the Scotian Ice Divide (Fig 4) basal sliding velocities increased. Clast abrasion, comminution and matrix formation are enhanced with higher ice velocities, hence the decrease in clast modes and increases in clast-surface striae (Figs 72, 78). The change in till properties from divide to margin could be summarized as follows:

Striated clast percentages increase—>
Clast weight percentages decrease—>

Ice divide—> Stony Till Plain —> Morainal Zone —> SSEMC

5.6 ICE DYNAMICS AND THICKNESSES

Glacier ice physics can be used to elucidate the processes of formation of the Scotian Shelf End Moraines and their relationship to the Scotian Ice Divide. Ice Flow Phase 3 overtopped the 300 m high Cobequid and Antigonish Highlands in northern Nova Scotia (Stea and Finck, 1984, Map 1, Fig 4). Ice thickness over the highlands was sufficient to allow for basal-melting and basal slip over the substrate and the production of erosional and depositional landforms (Denton and Hughes, 1981). An initial ice thickness value of 300 m may be assumed based on the temperature distribution of modern glaciers (Paterson, 1981).
Figure 79. Scattergram of clast weight percentages vs striated clast percentages for terrestrial till units and offshore diamictons.

Offshore diamictons

1. Scotian Shelf End Morainal Complex (Scotian Shelf Drift)
2. Emerald Silt
3. Unit B
4. Morainal Zone (Scotian Shelf Drift)

Onshore tills

5. Lawrencetown Till (LT/C)
6. Lawrencetown Till (LT/A/B)
7. Hartlen Till
8. Beaver River Till (ST/A/B/C)
1. Hypothetical trend line of increasing abrasion away from the Scotian Ice Divide to the offshore.
height of the Scotian Ice Divide can then be calculated from this initial starting point, using a simple ice model based on equations of ice motion for a perfectly plastic ice sheet (Paterson, 1981)

\[ H = \left( \frac{2\tau_o L}{\rho g} \right)^{1/2} \]  

Here \( \rho \) is ice density (0.91 g/cm^3, assumed constant), and \( g \) is acceleration due to gravity (981 cm/sec^2). \( \tau_o \) is the yield or shear stress, \( L \) is the distance to the ice margin, \( H \) is the ice thickness. The author modified an iterative scheme by Schilling and Hollin (1981) to model Nova Scotia ice caps with variable bedrock topography based on equation (1) to include variable basal shear stresses (Appendix 8). The ice thickness equation for this iterative scheme is

\[ H_{i+1} = H_i + \frac{\Delta X}{\rho g t_i} \]  

where \( H_i = \) initial ice elevation (including bedrock elevation)

\( \Delta X = \) iterative step length

\( t_i = \) initial ice thickness

In this model steady-state equilibrium is assumed and isostatic adjustments are ignored, because for this discussion ice thickness is the primary concern.

The ice profile was constructed in a north to south transect across the former Scotian Ice Divide (Ice Flow Phase 3, Fig. 80). The model starts with an initial (minimum) ice thickness of 300 m where Ice Flow Phase 3 overtopped the Cobequid Highlands at 300 m elevation. The Cobequid Highlands have a rough, bedrock-dominated surface topography. Basal shear
Figure 80. Calculated profiles of the Ice Flow Phase 3-Scotian Ice Divide and an estimate of the thickness of the earlier Phase 2 ice cap (Escuminac Ice Centre). The solid line in the Phase 3 glacier reconstruction represents iterated ice sheet profile calculations from a program modified from Schilling and Hollin (1981). Dotted glacier profiles are interpolated using the previous ice slope calculations. (B) Bedrock areas with basal shear stress set at 0.8 bar. (D) Deformable beds with basal shear stresses set at 0.2 bar. Ice Flow Phase 2 ice thickness estimated from isostatic rebound at Scotian End Moraines (maximum thickness).
Figure 80.
stresses in this environment would be relatively high. A minimum value of 80 kPa (0.8 bar) is assumed (Schilling and Hollin, 1981). The Minas Basin region south of the Cobequid Highlands is a till plain. Boulton and Jones (1979) calculated shear stresses of 20 kPa (0.2 bar) for till substrates. The South Mountain region (like the Cobequid Highlands) is a bedrock-dominated environment with a thin veneer of bouldery till. A basal shear stress value of 80 kPa was also used in this region. With these model constraints, the height at the centre of the Scotian Ice Divide is calculated to be 978 m (Fig. 80).

Another estimate of ice thickness can be derived from the depth relationships of the Scotian Shelf End Moraines. With the exception of the Country Harbour Moraine, the morainal ridges of the Scotian Shelf End-Moraine Complex are found in water depths of 140-160 m. King et al. (1972) originally suggested that the consistent depth of most of the morainal ridges implies that ice buoyancy may be a formational factor. The depth of water required for flotation is governed by this relationship from Reeh (1968):

\[ h(w) = (\rho_w / \rho_i) h_i \quad (3) \]

\( h(w) = \text{water depth}, \ h_i = \text{ice thickness}, \ \rho_i = \text{density ice,} \ \rho_w = \text{density water} \)

If ice advances into water approximately 1.1 times its height, then the ice mass will become buoyant. As it does, basal shear stresses tend towards zero and the ice mass spreads out and thins. This thinning will result in retreat of the grounding line until an equilibrium is attained between ice production through surface accumulation and lateral advection and ice wastage by bottom melting and calving (Thomas, 1979). Powell (1991, p. 78) suggested that a maximum or critical water depth exists for calving glaciers, independent of ice thickness, beyond which the front becomes unstable. At this water depth calving rates greatly exceed
ice flux to the terminus (Powell, 1991). The grounding line of unconfined glaciers along the polar East Antarctic glaciated shelf are commonly found at 200-300 m water depth (Schilling and Hollin, 1981). Similarly, the maximum depth of the grounding line in temperate Alaskan tidewater glaciers was found to be around 300 m (Brown et al., 1982). The modal maximum height of modern icebergs in the Labrador Shelf of eastern Canada is 300 m (Hotzel and Miller, 1983, Josenhans and Fader, 1989).

During moraine formation at about 16 ka (Gipp and Piper, 1989), RSL lowering in Barbados was 115 m (Fairbanks, 1989). Assuming gravitational and hydro-isostatic effects to be minimal, much of the present day continental shelf would have shoaled to water depths that would not support ice with an assumed "critical" lift-off thickness of 300 m. The formation of grounding-line moraines on the Scotian Shelf implies that isostatic depression compensated for the eustatic decrease in sea-level. It is assumed that isostatic depression of the crust under the moraines resulted from a major ice flow to the shelf edge (Phase 2, Fig 4). The amount of crustal depression can be calculated from the difference between the present day moraine depths and the depth of moraine formation during eustatic minima with a simple equation similar to that used by Morner (1971) and Belknap et al. (1987) using present day sea-level as datum (Fig 81)

\[ I = B + S \]  \hspace{1cm} (4)

\[ B = E + 300 \text{ m} \]  \hspace{1cm} (5)

where

\[ I = \text{isostatic uplift at the outermost moraines} \]

\[ E = \text{eustatic sea-level depression (115 m)} \]

\[ S = \text{present day average Scotian Shelf Moraine depths} \]

\((-150)\)
Figure 81  Schematic diagram illustrating the concepts of isostatic recovery at the Scotian Shelf End Moraine Complex

I = isostatic uplift at the outermost moraines
E = eustatic sea-level depression (115 m)
S = present day average Scotian Shelf Moraine depths (-150)
B = critical buoyancy depth at eustatic minima (below present sea level) = 415 m
B = critical buoyancy depth at eustatic minima (below present sea level) = 415 m

Therefore \( I = 265 \text{ m} \)

If isostatic equilibrium is established then a thickness \( h \) of ice will depress bedrock by (Weertman, 1961)

\[
h \left( \frac{\rho_i}{\rho_r} \right) = I \quad (6)
\]

using \( I = 265 \text{ m} \),

\[ h = 795 \]

Using an approximate 1/3 ratio of ice density to rock density (Hughes, 1981) an ice thickness of 795 m is produced for the Scotian Moraines. This is a crude estimate, but a reasonable one, because factors affecting the model such as greater ice volumes, ice-water attraction, disequilibrium etc would tend to both increase and decrease the estimate of \( I \). The Scotian Shelf End Moraines are generally found in water depths of 140-160 m. The Country Harbour Moraine crest is a notable exception in only 85 m water depth. This may reflect thicker ice over the site.

It is important to remember that these values represent ice thicknesses overlying the Scotian Shelf End Moraines from the previous larger ice mass (Escuminac Ice Centre; Ice Flow Phase 2, Figs 4, 80) that extended out to the shelf edge. I am proposing that rebound-induced stabilization of the ice front during rapid recession of phase 2 ice may have been responsible for the formation of the moraines.
5.7 CORRELATION OF MARINE AND TERRESTRIAL DEGLACIATION RECORDS

Seismic sequences within the sedimentary infill of the Basin Zone and valleys of the Truncation Zone record deglaciation and sea-level change. The sedimentary record of Core 91018-53 within the Halifax subzone spans the time from 13 ka to 2 ka. On land, lake core records span the same length of time (Mott et al., 1986, Ogden, 1987) and can be used for correlation.

Most lake cores in Nova Scotia show successions of pollen zones and lithostratigraphy that indicate climatic fluctuations (Livingstone and Livingstone, 1958, Mott et al., 1986, Ogden, 1987). At the base of most lake cores is inorganic sand or mud, sometimes rhythmically-laminated. In southern Nova Scotia, this is succeeded by black or brown organic mud accumulation (gyttja) which has been radiocarbon-dated between 13 and 12 ka (Jetté and Mott, 1989, Stea and Mott, 1990). The pollen successions in lake gyttja during this time invariably record a change from tundra to spruce forests indicating climatic amelioration (Railton, 1972, Ogden, 1987, Jetté and Mott, 1989). Around 10.8 ka an abrupt and pronounced climatic deterioration occurred that strongly affected the landscape and its vegetation cover. Lake cores reveal a sedimentological "oscillation" that marks the Younger Dryas Chronozone (11,000-10,000 ka, Mangerud et al., 1974). This oscillation is defined by a distinct inorganic clay layer that separates gyttja near the base of the cores (Mott et al., 1986) from gyttja that constitute the rest of the cores. Changes in pollen assemblages in these lakes, imply a warmer-colder-warmer climatic cycle. In southern Nova Scotia, the clay layer may have formed by solifluction of the adjacent devegetated slopes during the reversion to cold climatic conditions (Mott and Stea, 1990, 1994) or runoff from ephemeral or permanent
snowpacks (Stea and Mott, 1992) The remaining record in the lake cores indicates an initial rapid warming and climate fluctuations during the Holocene (Ogden, 1987) In northern mainland Nova Scotia, stratigraphic sections reveal diamictons, clay and sand overlying peats that predate the distinctive inorganic and clay layer in the lake sediment records Radiocarbon ages from wood and peat increments at the top of the peat layer cluster around 10.8 ka (Mott and Stea, 1994) Stea and Mott (1989) interpret the overlying sediments as largely glacigenic, implying a renewed glaciation in this region

The offshore seismic sequences and events can be correlated to the onshore lake stratigraphy The Yankee Bank Formation is correlative with the Younger Dryas clay layer in the lakes and the glacigenic sediments that overlie peat in land sections On land and sea a zone of increased inorganic sedimentation is bracketed by organic strata with biota indicating a warming trend The inorganic sedimentation events were initiated at around 10.8 ka and represent climatic cooling and snowfield/glacier buildup on land and increased sea-ice/icebergs in the marine realm The land and ocean events appear synchronous, in spite of the postulated 400 year lag time of oceanic $^{14}C$ equilibration (Mangerud, 1972) Isolation from the ocean may have decreased the mixing time within these nearshore basins

5.8 SUMMARY

In this chapter, a direct land to sea correlation of glacial deposits and landforms was made These are the salient conclusions

1 Ice Flow Phases 1 and 2 extended to at least as far as the Scotian Shelf End Moraine Complex because erratic
dispersal trains, associated with each flow phase, can be traced to the SSEMC moraines (Fig 82)

2 The Scotian Shelf End Moraines represent a margin of Ice Flow Phase 3 based on direct correlation between Beaver River Till formed under the Scotian Ice Divide and the morainal diamictons (Fig 82) The SSEMC was probably a recessional margin that originally formed when Ice Flow Phase 2 retreated from the outer shelf. Rebound-induced stabilization of the ice front at the moraines may have been partly responsible for the buildup of the Scotian Ice Divide. Flow from the Scotian Ice Divide may also have briefly extended past the morainal complex as an ephemeral ice shelf during readvances or surges that formed till-tongues.

3 The Morainal Zone represents the margin of Ice Flow Phase 4, the last, till-forming, widespread glacier advance in northeastern Nova Scotia (Fig 4) Sequence 2 (ca 12.9) in the Basin Zone, probably formed during this event.

4 A climatic event ca 10.8 ka to 10 ka that is recorded in land sections and lake cores is also recorded in the offshore basins by the Yankee Bank Formation (Sequence 5)
Figure 82  Schematic cross section through the Scotian Ice Divide to the outer shelf showing correlation of till units and major depositional zones (A) Scotian Shelf End Moraine Complex (B) Drumlín fields (C) Stony Till Plain (D) Bedrock and residuum
Scotian Ice Divide
Phase 3

offshore

onshore

Beaver River Till
Lawrencetown Till
Hartlen Till
Residuum

Sea Level
Glacier flow line
Erosion and entrainment

Figure 82
CHAPTER 6: DISCUSSION AND CONCLUSIONS: WISCONSINAN GLACIAL AND SEA LEVEL HISTORY OF EASTERN CANADA AND THE CONTINENTAL SHELF

6.1 INTRODUCTION

The processes, correlations and sea-level history established for the inner Nova Scotian shelf are crucial in the interpretation of glacial events for the entire Eastern Canadian-Appalachian region. Previous Wisconsinan glaciation models discussed in the introduction, are simplistic, and not in accord with the land and sea evidence. The author will present a new empirical model of ice advances and retreats and relative sea-level changes for Eastern Canada incorporating the inner shelf study area concepts. Some of these ideas are speculative, but should serve as a basis for further study and a development of ideas on the glaciation of the onshore-offshore region. The new compilation of Wisconsinan flow paths for Eastern Canada (Fig. 83) which includes offshore margins is a major step forward from the previous ice flow phase concept (Fig. 4). L.H. King (pers. comm., 1994) has also proposed a similar marine-land ice flow model based on the author's ice flow phase concept.

6.2 ICE FLOW PHASES 1A AND 1B

The oldest observed ice flow patterns in Maritime Canada are eastward and southeastward (Ice Flow Phases 1a, 1b). Eastward flow patterns (Phase 1a) prevalent along the Northumberland Strait region (Map 1; Fig. 83), were originally interpreted by Chalmers (1895) to be the Appalachian-derived Northumberland Glacier. This flow phase
Figure 83. Evolution (advance and retreat) of ice divides and domes over Maritime Canada during the Wisconsinan (70-10 ka). (A) Ice Flow Phase 1a in Atlantic Canada and margin. (B) Phase 1b (C) Ice Flow Phase 2 from the Escuminac Ice Centre and Divide on the Magdalen Plateau. A retreat margin over Sable Island, Western, Emerald Banks (after Boyd in McLaren, 1988). (D) Advance and retreat of Phase 3 ice (Scotian Ice Divide) (E) Ice Flow Phase 4. Ice flow data from Chalmers, 1895; MacNeill, 1951; Prest and Grant, 1969; Rampton et al., 1984; Pronk et al., 1989; Rappol et al., 1989; Stea et al., 1992a).

LEGEND FOR ICE FLOW MAPS

- Ice flow line
- Ice flow line (later)
- Ice margin
- Ice shelf
- Ice rise
- End moraines

G Gaspereau Ice Centre/Divide
E Escuminac Ice Centre/Divide
C Chignecto Glacier
Figure 83.
Figure 83.
Phase 4
ca 14-12.7 ka

Figure 83.
may represent a separate, older phase of glaciation. The ice configuration later changed to a centre in New Brunswick (Gaspéreau Ice Centre; Rampton et al., 1984) which produced southeastward striae flow patterns on the southern New Brunswick highlands (Caledonia Phase; Rampton et al., 1984) and distinctive till units in Nova Scotia (Hartlen Till-East Milford Till; Stea et al., 1992a, c). The East Milford Till has been assigned an Early Wisconsinan age because it postdates the last interglacial interval and is overlain by recessional sediments and Late Wisconsinan tills. Land and offshore deep sea oceanic records show a major glaciation at this time (Clark et al., 1993). The Hartlen Till is correlated with the middle of three major reflections seen in the Scotian Shelf end moraines and the lowest wedge-shaped, massive reflections on the Scotian Slope interpolated to be Early Wisconsinan by Mosher et al. (1989). Evidence of the passage of Laurentide ice across New Brunswick is equivocal (Rampton et al., 1984; Pronk et al., 1989; Rappol, 1989). Shield anorthosite boulders in western Prince Edward Island (Prest and Nielsen, 1987) and in the Gulf of St. Lawrence (Loring and Nota, 1973; Stea, 1991) indicate that Laurentide ice may have been confluent with ice from the Gaspereau Ice Centre (Fig. 83).

Ice Flow Phases 1a and 1b left a strong imprint on the Nova Scotia landscape. These flow patterns can be traced throughout the Maritimes region (Fig. 83) and across the inner shelf. They represent the most "vigourous" ice flow events characterized by extensive erosion and deposition of thick, overconsolidated, matrix-rich tills (cf. Grant, 1977, 1989). The margin of these events are placed at the shelf\slope break at water depths of approximately 300-600 m, the maximum depth of relict iceberg furrowing (Mosher et al., 1989; Dodds and Fader, 1986).
6.3 ICE FLOW PHASE 2

Late Wisconsinan ice flow trends across much of northern Nova Scotia are southward and southwestward. Along the Nova Scotia - New Brunswick border region, ice flow recorded by striae is largely southwestward (Prest, 1973), (Stea and Finck, 1984, Rampton et al., 1984; Foisy and Prichonnet, 1991). Northward ice flow has been reported in the Magdalen Islands (Parent et al., 1990). These flow trends have been attributed to an ice dome or divide near Prince Edward Island, termed the Escuminac Ice Centre (Rampton et al., 1984; Fig. 84). Northward flow features in the Baie des Chaleurs region of New Brunswick have been attributed to the development of the Escuminac Ice Centre (Pronk et al., 1989). H. Josenhans (pers. comm., 1990) interprets an incoherent seismic unit that pinches out in the Launenian Channel as a till formed by ice from the Magdalen plateau region. Flow from a centre in the region of Prince Edward Island was likely responsible for the "flood" of red mud (cf. Grant and King, 1984) from the vast areas of Carboniferous red-beds in northern Nova Scotia onto igneous and metamorphic terrains to the south. Anomalous red muddy diamictons have been described in some offshore cores adjacent to the Lunenburg (Piper et al., 1986) and Chezzetcook drumlin fields (Forbes et al., 1988) but were not encountered in grab samples in this study. The Sable Island and Western Bank regions are characterized by high percentages of red-stained (hematite) detrital quartz grains with a possible Carboniferous red bed source (Cok, 1970). Cok (1970) also describes a sizeable "skip zone" with little or no red-stained quartz across the inner shelf, which corresponds to the extent of Ice Flow Phase 3 (Fig. 83). Amos and Miller (1990) reported abundant red siltstone lithic fragments in the cores of southwest Sable Island Bank. Piston cores in continental slope regions to the south and west of the Sable Island and Western banks have a characteristic, consistent stratigraphy consisting of
an olive-grey mud, overlying a red-brown mud (Hill, 1981; Mosher et al., 1989). Conolly et al. (1967) also described a similar stratigraphy in cores on the slope off the Laurentian Channel. Hill (1981) determined a Carboniferous source region for the red mud based on reworked Carboniferous palynomorphs. According to Hill (1981) the red unit was deposited from suspension fallout from the ice sheet that formed the Lawrencetown Till. The olive-grey unit is Holocene in age and derived from adjacent banks.

The "red flood" of Ice Flow Phase 2 crossed the inner shelf to the outer banks and slope, in 300 m water depths (Dodds and Fader, 1986; Mosher et al., 1989). The timing of this flow event is between 18 and 21 ka (Mosher et al., 1989; Baltzer et al., 1994). The inferred maximum extent of Phase 2 is shown in Fig 83. This is comparable to the "maximum model" and consistent with the ice thickness estimates on the inner shelf (1 km) obtained earlier (Fig. 80). Ice outflow from the Escuminac Ice Divide and Centre during Phase 2 may have been in the form of ice streams or ice "currents" (Grant, 1963). These rapidly flowing ice streams would have converged into inter-bank channels, drawn by rapid calving of the ice margin in deeper water.

Late Ice Flow Phase 2 ice flow indicators along the northern part of the study region appear to veer to the southeast parallel to the earlier Phase 1b ice flow (Fig. 83). The east-west orientation of Sable Island and Banquereau and north-south trending tunnel valleys imply a major southward-flowing ice sheet (Fig. 74; Boyd et al., 1988b). The southeastward flow trajectory in the Wine Harbour area may be due to a localized ice stream that merged with southward-trending streams from an extensive, former ice divide across the Magdalen Plateau and possibly, Cape Breton Island. This divide (Escuminac
Ice Divide) may have formed during the Middle to Late Wisconsinan (Stea et al., 1992a) when Laurentide Ice was removed from the Laurentian Channel, cutting off Magdalen Plateau ice from a Laurentide source (Occhietti, 1989). Alternatively, (Phase 2) southeastward ice flows may represent later, radial flow from a remnant centre (Fig. 83).

6.3.1 Retreat of Phase 2 Ice

Retreat of the Phase 2 ice margin was hastened by calving at the marine shelf/slope break margin. Calving was a primary feedback mechanism for initiation of the retreat of the northern hemisphere glaciers (Denton and Hughes, 1981). Mosher et al. (1989) reported Late Wisconsinan iceberg furrowing at 300 m adjacent to the margin. Rapid retreat of Ice Flow Phase 2 ice from the shelf edge margin and from the Laurentian Channel may have contributed to the formation of Heinrich events in the deep ocean dated 18.9-21.4 ka (Heinrich, 1988; Broecker et al., 1992; Bond et al., 1992; Grousset et al., 1993). Ice streams in the Laurentian Channel emanating from a Newfoundland Ice Cap (Grant, 1989) and the Escuminac Ice Divide (Fig. 83) eroded large quantities of limestone from the Anticosti and Magdalen Basins and also carried a substantial quantity of Appalachian terrane debris. Based on the sand mineralogy of continental rise sediments off southeast Canada, Piper et al. (1994) suggested the development of Appalachian ice caps during marine oxygen isotope stages 4 and 2. A Baffin Bay source has been postulated for Heinrich layers, based on petrography and geochemistry (Grousset et al., 1993), but the technique cannot rule out iceberg debris derived from the complex bedrock terranes of the eastern Canadian margin. Individual Heinrich layers (H11 and H12) appear to thicken towards the Appalachian margin (Grousset et al., 1993). The ice margin may have initially retreated as far as the inner shelf. Boyd (in McLaren, 1988) and King (1994)
proposed that Sable Island Bank was the site of a significant ice marginal stand. According to these authors, a glacier grounded near sea level produced an outwash-delta-dominated moraine. Boyd (in McLaren, 1988) correlated this margin across Western and Emerald Bank as far as Georges Bank.

6.4 ICE FLOW PHASE 3

Flow from the Scotian Ice Divide was northwestward into the Bay of Fundy and southeastward over the Eastern Shore, veering southward east of Country Harbour (Fig. 83). It was also funnelled northward into Georges Bay and the Cape Breton Channel from Cape Breton and the mainland. At an early stage of development, the Scotian Ice divide merged with glaciers in the Baie des Chaleurs in New Brunswick to flow northward through the Shediac Channel into the Laurentian Channel (Fig. 83). An ice dome off Cape Breton Island proposed by Grant (1977) was probably part of this divide which can be traced eastward through Chedabucto Bay onto the continental shelf. Offshore tunnel valleys, oriented northeastward towards the Laurentian Channel and southward towards Sable Island (Boyd et al., 1988b), converge in an undissected zone, probably the offshore location of the Scotian Ice Divide (Figs. 74, 83).

Three hypotheses can be invoked to explain the formation of the Scotian Ice Divide; two based on autocyclic, self-regulating mechanisms, and one based on external factors. They are:

1. Ice retreated rapidly, through calving, into the marine channels bordering Nova Scotia including the Gulf of Maine-Bay of Fundy (Prest
and Grant, 1969), and Cape Breton Channels. In this manner Nova Scotia
ice was cut off from the Escuminac and Gaspereau ice sources (Fig. 83).

2. Stabilization of the retreating Phase 2 ice margin at the Scotian Shelf
End Moraine complex due to crustal rebound and marginal sediment
accumulations. At Sable Island, crustal depression was minimal, and
outwash deltas formerly graded to a sea level of 75-80 m bsl (McLaren,
1988).

3. Change in climatic factors, favouring snow accretion on the Nova Scotia
mainland (MacNeill and Purdy, 1951).

Simple drawdown (Hypothesis 1) may not completely explain an ice sheet whose flow-
lines can be traced across the northern mainland highland areas exceeding 400 m. A
combination of all 3 formative mechanisms is likely. Snow accretion may be favoured if
jet stream position and seasonal storm tracks were shifted southward around the larger
Laurentide ice sheet (Shinn and Barron, 1989).

The Scotian Ice Divide may have extended as far as Sable Island Bank and the series of
banks which line the outer shelf. The southern margin of the Scotian Ice Divide may have
been a short-lived ice shelf, with ice rises briefly acting as centres of outflow during
recession (Gipp, 1989; 1994). Ice later settled at the Scotian End Moraine Complex, the
site of former ice retreat margins from earlier advances (Phases 1a, 1b) and Banquereau
Bank (Fig. 83). These margins are nearly equidistant from the inferred divide (Fig. 83).
The Scotian Shelf End Moraine Complex formed at approximately 15.5 ka in the northeast
(Gipp, 1989; Gipp and Piper, 1990), but ages in the southern portion are not well constrained.

6.4.1 Retreat of Phase 3 Ice

Calving retreat of the Scotian Shelf and Banquereau ice margin was initiated when sea levels rose again following the outer shelf lowstand at 17-15 ka (Amos and Miller, 1990; King, 1994). By 14 ka the ice margin was close to the present day coastline over much of Nova Scotia (Piper et al., 1986; Stea and Wightman, 1987; Stea and Mott, 1989; 1990; Piper, 1991; Gipp, 1994), but it stubbornly remained there for at least 1500 years (Stea and Mott, 1989; Gipp, 1994). The lack of raised features along the Atlantic Coast of Nova Scotia indicates that an ice carapace prevented shoreline formation during isostatic recovery. This was not the case on the Gulf of Maine and Bay of Fundy, where vigorous ice stream drawdown cleared ice from these regions (Mayewski et al., 1981; Belknap et al., 1991). Raised beaches, deltas and marine deposits dated from 16 to 12.6 ka are found along the Bay of Fundy (Grant, 1980a, b, 1989; Scott and Medioli, 1980a; Stea et al., 1986; 1987; Honig and Scott, 1987; Stea and Wightman, 1987). In northern Nova Scotia deglaciation was delayed by as much as 2000 years until the region around Georges Bay was cleared of ice by landward propagation of a calving bay in the Cape Breton Channel.

The rapid calving-style retreat of the Phase 3 ice margin can be correlated with the youngest of the Heinrich events (13.4-15.0 ka) in the deep North Atlantic ocean (Grousset et al., 1993). Sequence 2 of the inner shelf Basin Zone records much of the iceberg activity in the form of point-source diffractions (boulders and boulder dumps). Bacchus (1993) and Gipp (1994) also report extensive iceberg activity in the Scotian Shelf and Gulf
of Maine at this time. From 15 to 13 ka much of the ice was removed from the Gulf of Maine-Bay of Fundy system and the Laurentian Channel (Piper, 1991).

6.5 ICE FLOW PHASE 4

During this phase ice caps remnant from the Scotian Ice Divide briefly reactivated (Fig. 83). These small ice caps or glaciers formed over Southern Nova Scotia, (South Mountain Ice Cap-MacNeill, 1951), the Chignecto Peninsula (Chignecto Glacier-Chalmers, 1895) and Antigonish Highlands in northern Nova Scotia (Chedabucto Bay Glacier Complex-Stea et al., 1989; Figs. 4, 83). During Phase 4, ice was strongly funnelled westward into the Bay of Fundy and southwestward to the Atlantic coast and beyond. Rapid advance of Phase 4 ice offshore formed the Morainal Zone. Southwestward-trending channels which cut the Phase 3 ice margins and are initiated at the Morainal Zone (Fig. 16; Map 2) may have served as meltwater conduits from this glacier. Truncation of Emerald Silt in water depths between 140 and 160 m may have been synchronous with the Phase 4 margin.

The Country Harbour End Moraine is a complex, palimpsest feature. The southwest-trending ridge may have initially formed during Ice Flow Phase 3. Phase 4 ice eroded and reshaped the NW-SE segment, probably at the same time as the formation of the Morainal Zone. The complex-interdigitating succession of till-tongues and thick Emerald Silt section are a result of ice advance and retreat from the successive Phase 3 and 4 ice margins.

Ice Flow Phase 4 on land can be traced to re-advance margins in the Annapolis Valley
(Hickox, 1962) and in southern Nova Scotia near Chester (Graves and Finck, 1988; Map 1; Figs. 4, 83). In the Annapolis Valley this margin is marked by outwash deltas. Marine limit beyond the margin is +40 m whereas inside the margin it is <25 m (Stea et al., 1987; Fig. 5). The time of formation is bracketed between dates on the oldest marine limit indicator south of the margin (ca. 14 ka, Grant, 1989) and the youngest, north of the margin (ca. 12.4 ka; Stea et al., 1987). This event may correspond with the erosional event that formed the Sequence 2 boundary in the Basin Zone dated at approximately 13.5 to 12.7 ka. Ice Flow Phase 4 is correlative with the Robinson's Head advance in Newfoundland dated at 12.6 (Grant, 1989).

6.5.1 Retreat of Phase 4 Ice and the Sea Level Lowstand

After 12.6 ka, the climate warmed considerably (Mott and Stea, 1994) resulting in ice retreat to small, remnant terrestrial centres (Map 1). Trees migrated into southern Nova Scotia after 12 ka (Mott and Stea, 1994). Sea-level fall stranded Phase 4 glaciers into isolated pockets. Ablation by downwasting ensued, leaving tracts of hummocky ground moraine (Map 1) throughout the marginal regions. Phase 4 ice may have disappeared by 11 ka (Stea and Mott, 1989; Gipp, 1994). Sea levels dropped to their lowest point at -65-70 m around 11.6 ka. A shoreline formed during a short period of crustal stability (Transition Subzone). Sea-level rose rapidly after 11.6 ka and a large area of the inner shelf was transformed by erosion into a low-relief surface, barren of depositional features (Platform Subzone). The rate of relative sea-level rise slowed markedly between 11 and 9 ka, a result of forebulge passage, and depositional features were again established (Estuarine Subzone).
6.6 THE YOUNGER DRYAS EVENT

Around 11,000 yr B.P. an abrupt and pronounced climatic deterioration occurred that strongly affected the terrestrial landscape, its vegetation cover and the nature of sedimentation and fauna of adjacent oceanic basins. A wide variety of sediments covered organic sediments at many sites in the Maritime Provinces. Lake cores on land and the inner shelf basins reveal a sedimentological "oscillation" that marks the Younger Dryas (11,000-10,000 ka; Mott et al., 1986). Changes in pollen and foraminiferal assemblages in these depositional areas imply a warm-cool-warm cycle. Stea and Mott (1989) proposed that small glaciers developed from pre-existing ice or reformed during this time on land. King (1994) proposed a major glacial advance terminating at Sable Island.

Using a general circulation model, Rind et al. (1986) predicted accentuated cooling in northern Europe and northeastern North America during the Younger Dryas with increased onshore wind flow, storm activity and precipitation. Shinn and Barron's (1989) climate model of the last glacial maximum also evokes a northward shift of storm tracks along the margin of the Laurentide ice mass and net increase in precipitation along the margin. According to their theory, storms over the region reached a maximum frequency when the retreating Laurentide and Appalachian ice masses were somewhat removed from the region. The increase in storminess, coupled with a North Atlantic meltwater-induced cooling (Broecker, et al., 1985) would explain a local buildup of ice during the Younger Dryas.
APPENDIX I: Mapping methods-marine geophysical data acquisition.

Seismic data for the study were collected during Cruise 91018 (CSS Dawson; Fader et al. 1993) on the eastern inner Scotian Shelf (Fig. 1). The Huntec Deep Tow boomer system was used to generate analogue seismic data with vertical resolution of less than 1 metre (c.f. Hutchins et al. 1976). Operating frequencies ranged from 500 Hz to 2 KHz with a power output of 4-6 kV. Two recorders were employed, an internal transducer mounted underneath the towed sound source and an external array. The deep tow system is towed at 10-75m water to minimize the effects of surface swell and wave noise. Seismic reflection data with less resolution but more penetration were obtained using a 40 cubic inch Haliburton sleevegun system operating at 5-500 Hz. Bathymetric information was obtained using data produced by a hull-mounted 3.5 KHz acoustic profiler, from Canadian Hydrographic Service charts, and from the depth of the first multiple below the seabed in the Huntec records (Sylwester, 1983). Seafloor topography/reflectivity were obtained using the Bedford Institute of Oceanography-designed sidescan operating at 70 KHz with a range of 1 kilometer and Klein dual frequency sidescan sonar systems at 100 and 500 KHz with a range of 100 to 400 m.

Additional data were obtained from seismic records of a previous CSS Dawson cruise in 1987 (Forbes et al. 1991). The Nova Scotia Research Foundation Corporation (NSRFC) deep-tow seismic system was deployed on this cruise and included a nine element streamer and a 30-tip sparker source operating at 200j.

14.25 kHz echograms collected by the Canadian Hydrographic Service in the 1960’s for charting purposes were also used in the mapping of the offshore. A region from Sheet
Harbour to Halifax was surveyed at a track spacing of 0.93 km (King et al. 1972).

Late in this thesis study, high resolution swath bathymetry was made available to the author. A region of the inner shelf off Halifax was surveyed using the Simrad EM-100 multibeam sounder. It provides a digital image of the sea floor. The coverage is 1.7 X water depth across track (80° angle) with 32 narrow beams, each about 2.5°. It operates at 95 KHz and has a depth range of 10 to 600m. Spatial resolution of a single beam is 5 metres (Costello et al. 1993).

Offshore digital terrain data obtained from this system was gridded using GRASS GIS software (R. Courtney pers comm, 1992). The digitized bathymetric 14 kHz data was gridded and a terrain model produced using AUTOMINER software. Digital elevation data was also obtained from the Lunenburg drumlin field through the Land Survey and Registration Service (Prince Edward Island). This was also gridded and plotted using GRASS GIS software.
APPENDIX 2: Marine and terrestrial sampling methods and sample preparation.

Most samples of the seafloor were obtained with a large (0.3 ton capacity IKU) wire-line grab (Fig). These grab samples were described carefully and subsampled on deck using a shovel and five gallon pails. If the IKU was full sampling proceeded from the top down. Five gallon pail samples were collected from the surface lag, from the zone in the middle of the sediment block, and from the bottom. The preservation of stratigraphy within the filled grabs suggest minimal modification of the sediment package by the sampler. At other sites, however, the grab did not "bite" into the substrate. These grabs brought back a surface lag of cobble and boulders and occasionally some matrix. Other, coarser-grained samples were thoroughly mixed when the sampler was discharged on the deck.

Shallow cores in coarser, hard sediments were taken with a Brooke Ocean vibrocorer capable of 3m penetration. In softer sediments (muds) a modified benthos piston corer was employed with a penetration capability of 20m. Cores were split, photographed, described and subsampled for grain size and biostratigraphic analysis using foraminifera (Costello, 1994).

2-5 kg till samples were obtained from road cuts, sea cliffs and borrow pits. Sixty-five samples of till pebble separates from a previous survey in 1977 (Stea and Fowler, 1979) were utilized to augment the pebble count database. The reference section at West Lawrencetown was systematically profile-sampled at regular intervals. One sample per metre of depth was obtained.
A 1-2 kilogram aliquot of sample was removed from the sample bags of terrestrial and marine samples. This portion was dried overnight and then weighed with the container (aluminum pan). If the sample flowed freely then it was dry sieved through 63mm, 4mm and 2mm pans. The clast fractions (>2mm) were weighed and the matrix fraction discarded. The >4mm <30 mm fraction was retained for pebble lithological analysis. Lumpy samples (clay-rich) were wet sieved through 2mm screens after disaggregation and dispersal in a calgon solution. The clast fraction was dried overnight and then processed like the previous dry samples.
## APPENDIX 3: Marine samples locations and field notes.

### Vibro-Core data

<table>
<thead>
<tr>
<th>SAMPLE #</th>
<th>TYPE</th>
<th>DAY/TIME (GMT)</th>
<th>LATITUDE/LONGITUDE</th>
<th>DEPTH (MTRS)</th>
<th>LOCATION</th>
<th>NOTES</th>
</tr>
</thead>
<tbody>
<tr>
<td>001</td>
<td>VIBRO</td>
<td>1551913</td>
<td>44 38 65N 63 33 36W</td>
<td>21 9</td>
<td>HALIFAX HARBOUR</td>
<td>CATCHER SAMPLE IN A BUCKET (125 ML). ANN MILLER HAS TAKEN THIS SAMPLE TO WASH FOR FORAMS. NO CUTTER SAMPLE. SEISMIC TIME FROM NAVICULA 89009 DATA</td>
</tr>
<tr>
<td>003</td>
<td>VIBRO</td>
<td>1561615</td>
<td>44 33 62N 63 32 17W</td>
<td>31 0</td>
<td>HALIFAX HARBOUR</td>
<td>1 BUCKET WITH CUTTER SAMPLE. ANN MILLER HAS THE SAMPLE. THIS IS THE BOT VIBROCORE NOT THE AGC VIBROCORE</td>
</tr>
<tr>
<td>004</td>
<td>VIBRO</td>
<td>1561722</td>
<td>44 33 25N 63 32 07W</td>
<td>24 0</td>
<td>HALIFAX HARBOUR</td>
<td>STOPPED VIBROCORING WITH 30CM TO GO BECAUSE OF THE LACK OF PENETRATION THROUGH SEDIMENTS OVER A FEW MINUTES. FINE SAND VERY DRY. GREEN-BLACK WELL-SORTED FEW SHELL FRAGMENTS</td>
</tr>
<tr>
<td>018</td>
<td>VIBRO</td>
<td>1591637</td>
<td>44 15 60N 64 10 43W</td>
<td>56 0</td>
<td>SOUTH CROSS ISLAND</td>
<td>LATE TILL. PIPERS YOUNG MORaine SOUTH OF CROSS ISLAND TILL MOUND. LINER JAMMED IN THE BARREL BUT THERE ONLY WAS A CATCHER SAMPLE. FINE GRAY GRAVEL AND A COBBLE IN THE CATCHER</td>
</tr>
<tr>
<td>022</td>
<td>VIBRO</td>
<td>1601438</td>
<td>44 15 74N 64 09 69W</td>
<td>65 0</td>
<td></td>
<td>GOUDES IN THE CORE BARRELL, BOULDERS % SILTY SAND. GRAY SULPHIDE-H2S SMELL IN CORE. BOTTOM GRAVELLY SAND (COARSE) WITH FEW PEBBLES IN CORE. CATCHER SAMPLE STORED IN A 1 GAL BUCKET. SAMPLE CUT INTO 2 SECTIONS A-B = 152CM, B-C = 66CM. SECTION A B SLUMPED (COMPRESSED) 20CM WHEN HELD UPRIGHT. MAIN SECTION COARSE SAND, GRANULE, GREY (SLATE FRAGMENTS) NOT WELL-SORTED. ANGULAR ROCKS CORE ON DISTAL (LATE TILL) MOUND. TOTAL RECOVERY 2.2M INCLUDING CATCHER</td>
</tr>
<tr>
<td>014</td>
<td>VIBRO</td>
<td>1621710</td>
<td>44 25 09N 63 27 01W</td>
<td>82 3</td>
<td>EAST SAMBRO LEDGES</td>
<td>OLIVE GREY STONY DIAMICTON. ANGULAR SLATE FRAGMENTS. SILTY CLAYEY MATRIX WITH GREY MOTTLING. SEISMIC UNIT (C) SURFACE TOP CRUDELY LAMINATED MASSIVE ACOUSTIC UNIT. TILL-GlACIAL MARINE PROXIMAL ?? CATCHER IMPLODED. CUTTER DAMAGED BY BOULDER. CORE CUTTER SAMPLE BAGGED IN A BUCKET.</td>
</tr>
<tr>
<td>036</td>
<td>VIBRO</td>
<td>1621920</td>
<td>44 23 36N 63 29 13W</td>
<td>80 5</td>
<td>OLD SACKVILLE RIVER VALLEY</td>
<td>OLIVE GREY SILTY CLAY DIAMICTON. FEW STONES COBBLING ON SURFACE. SHELL FRAGMENTS MUCH MORE CLAY. THEN 034 SOME SAND HORIZONS-INCLUSIONS. SHELL FRAGMENTS ON SURFACE. SEISMIC UNIT (C) OUTCROPS ON SURFACE. APPARENT LAMINATION. CORE CUTTER SAMPLE BAGGED IN BUCKET</td>
</tr>
</tbody>
</table>
Appendix 3

038 VIBRO 1631200 44 23 65N 73.0 DELTA ON FLANK OLD SACKVILLE RIVER

KEPT core as 1 piece because of coarse nature of the sample. Sample would be damaged by cutting. No catcher or cutter samples. Possible Delta 7 mound with reflectors dipping down slope terminating on flat horizontal reflectors at base. Top 92cm olive grey coarse sand, granules cobbles on top, shell fragments throughout. Sharp contact with lower 80cm C = olive grey silty fine sand with a few pebbles-massive no shells.

039 VIBRO 1631330 44 47 95N 64.0 NORTH EAST OF SAMPLE 038

Catcher sample + 15cm sample bagged in a bucket under unit (C) or (B) outcrops on bedrock channel flank unit (C) incised in channel and filled by younger sediments (Holocene) top 15cm sandy, grey green, cobbles, shell fragments core catcher grey diamicton with reddish tinge, silty, metagraywacke clasts (angular-subrounded), abundant shell fragments core did not sample upper unit but sampled lower unit (C). Penetrated 2 metres without sampling unit.

040 VIBRO 1631425 44 26 24N 80.0 SACKVILLE RIVER SOUTHERN TRIBUTARY

OLIVE GREY SILTY SAND (SOME CLAY), DIAMICTON COBBLEY SURFACE, NO VISIBLE SHELLS-FILL OR GLACIAL MARINE, METAGREYWACKE CLASTS (SUBANGULAR-SUBROUNDED). 1 GRANITE CLAST

041 VIBRO 1631620 44 30 48N 60.0 DUNCANS COVE

NO RECOVERY.

053 BENTHOS PISTON 1641630 44 17 72N 168.0 HALIFAX MORaine

LAHAVE CLAY APPROX 6M THICK, TOP 22 FT OLIVE-GREY SILTY CLAY BOTTOM 5 FT GREY-BLACK, REDUCED MUD, H2S ODOR, CUTTER SLIGHTLY DAMAGED ATTEMPT TO CORE COARSE UNIT AT BASE OF LAHAVE CLAY, STRATIGRAPHY-LAHAVE CLAY OVER FACIES A, THIS COULD BE THE YOUNGER DRYAS EVENT (FACIES A) BENEATH LAHAVE CLAY, NO TRIGGER WEIGHT-DANIEL TOO LONG TO CLEAR THE HAIL.

066 VIBRO 1671630 44 37 28N 64.0 SOUTH OF MUSHABOom

OLIVE GREY GREEN SILTY SAND WITH COBBLE LAG, BRITTLE STAR ON SURFACE, GREY SILT AND CLAY, NO STONES AT 20CM, DEPTH 3 (1 GAL) PAILS 1=TOP 10CM 2= BOTTOM 20CM, 1 CORE CUTTER.

067 VIBRO 1671900 44 31 88N 62 27 13W 137.0 SOUTH OF SHEET HARBOUR

STRATIGRAPHY OF CORE UNIT-2 OLIVE-GREY SANDY MATRIX, UNIT-1 GREEN AND BROWN SILTY MUD FEW GRANULES (METAWACKE)

078 VIBRO 1681654 4°00 28N 61 30 38W 80.0 SOUTH OF COUNTRY HARBOUR

SEISMIC OUTCROP OR OLDER UNIT (C) OLIVE-GREY SAND (20CM OVERLYING BLACK ORGANIC CLAY (4CM), CORE EXTRUDED INTO 2 (1 GAL)
BUCKETS 1 IS LABELLED 78 TOP 20CM AND THE OTHER IS LABELLED 78 BOTTOM 4CM. CORE MOTOR HOUSING FRAME BROKEN WHILE HITTING THE SIDE OF THE SHIP. ONLY A FEW WELDS LEFT. ONLY BOLTS HOLDING THE FRAME ON THE VIBROCORE.

CUTTER CONTAINED OLDER UNIT (C). VENEERED BY SABLE ISLAND SAND. UNIT 3 = COBBLE-PEBBLE LAG METAGRAYWACKE QUARTZ VEIN GRANITE. UNIT 2 = OLIVE-GREY COARSE TO MEDIUM SAND. UNIT 1 = GREY-BLACK ORGANIC CLAY SLOWED DOWN THE DESCENT OF THE VIBROCORE TO THE BOTTOM AND GOT BETTER RECOVERY/APPROXIMATE PENETRATION RATIO AS A RESULT.

ONLAP OF THICK SABLE "LAND SAND? WEDGE ONTO DOME STRUCTURE. OLIVE-GREY SAND WITH BLACK SANDY BANDS APPROXIMATE PENETRATION OF 2.1 METRES (RECORDER) BUT ONLY 78CM. RECOVERY MUCH OF THE SAMPLE MAY HAVE WASHED THROUGH THE CATCHER BAG.

SEISMIC WEDGE OF SAND NORTH OF DOME. OLIVE-GREY SAND. CORE SOCK DID NOT HOLD SAMPLE. EXTRUDED FROM THE BOTTOM ONLY 40CM RETAINED. BAGGED IN A BUCKET.

OLIVE-GREY SAND LITHOTHAMNION ENCRUSTED COBBLE ON TOP. BOTTOM IN COBBLY SAND. SEISMIC CHANNEL FILLED WITH LAMINATE ON TOP FILL. SOCK IN CORE CATCHER PREVENTED SOME LOSS. 1 GALLON BUCKET WITH CORE CATCHER MATERIAL.

SEISMIC-CHANNELS IN BEDROCK FILLED WITH TRANSGRESSIVE ON TOP FACIES WITH UNIT (E). MYSTERY IN BASE OF CHANNELS OLIVE-GREY SILTY SAND, COBBLE PINK COVERED CLASTS ON SURFACE (LITHOTHAMNION), LAG, WELL SORTED, BLACK REDUCED ZONES MOTTLING (SULPHIDE) FEW GRANULES 2 SUBSAMPLES (CUTTER-CATCHER).


CORE HIT STONE IN UPPER UNIT(3) AND CONTINUED TO PENETRATE WITHOUT FURTHER RETENTION. THE RESULTANT CORE WAS AN OLIVE-GREY SAND (UNIT 3). THE TOP AN APPARENT TRANSGRESSIVE SAND. THE OUTSIDE OF THE BARREL WAS COVERED WITH A
## Appendix 3

<table>
<thead>
<tr>
<th>SAMPLE NUMBER</th>
<th>TYPE OF JULIAN DAY/TIME</th>
<th>LATITUDE (MTRS)</th>
<th>DEPTH (MTRS)</th>
<th>GEOGRAPHIC LOCATION</th>
<th>NOTES</th>
</tr>
</thead>
<tbody>
<tr>
<td>002</td>
<td>IKU 1561448</td>
<td>44 33 62N</td>
<td>31 0</td>
<td>HALIFAX HARBOUR</td>
<td>GREENISH GREY COMPACTED MUDDY SEDIMENT WITH QUOHOG SHELLS IN SEDIMENT HIGHLY DISTURBED SO NO PUSH CORE TAKEN SAMPLE STORED IN 1 (5 GAL) BUCKET SMALL SAMPLE</td>
</tr>
<tr>
<td>005</td>
<td>IKU 1581231</td>
<td>44 19 50N</td>
<td>20 00</td>
<td>LUNENBURG HARBOUR</td>
<td>POCKED BEDROCK SURFACE ON SIDE SCAN HALIFAX SLATE COBBLES</td>
</tr>
<tr>
<td>006</td>
<td>IKU 1581247</td>
<td>44 19 46N</td>
<td>22 00</td>
<td>LUNENBURG HARBOUR, NORTH CROSS ISLAND</td>
<td>BEDROCK SURFACE POSSIBLY HALIFAX SLATE 3 (5 GAL) BUCKETS TAKEN IN AREA OF BEDROCK ON THE SIDESCAN WITH UNUSUAL PITS AT SEABED POSSIBLY OLD MINING PITS BY SEAGOLD</td>
</tr>
<tr>
<td>007</td>
<td>IKU 1581326</td>
<td>44 17 08N</td>
<td>62 0</td>
<td>LUNENBURG</td>
<td>1ST ATTEMPT JAWS OPEN BROUGHT BACK ROCKS IN THE JAWS</td>
</tr>
<tr>
<td>008</td>
<td>IKU 1581334</td>
<td>44 17 08N</td>
<td>62 0</td>
<td>LUNENBURG</td>
<td>NO RECOVERY</td>
</tr>
<tr>
<td>009</td>
<td>IKU 1581345</td>
<td>44 16 95N</td>
<td>62 0</td>
<td>LUNENBURG</td>
<td>MUDDY SANDY GRAVEL, MOSTLY SLATE SAMPLE STORED IN 5 (5 GAL) BUCKETS PANNED SAMPLE HEAVIES MAGNETITE GARNET, ILMenITE NO GOLD SUBSAMPLES - GRAVEL FINES</td>
</tr>
<tr>
<td>010</td>
<td>IKU 1581409</td>
<td>44 16 12N</td>
<td>65 0</td>
<td>LUNENBURG</td>
<td>SAMPLE STORED IN 4 (5 GAL) BUCKETS 2 PHOTOS TAKEN</td>
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<tr>
<td>011</td>
<td>IKU 1581441</td>
<td>44 15 23N</td>
<td>67 0</td>
<td>LUNENBURG</td>
<td>SAMPLE ROCKS/MACROFAUNA SAMPLE STORED IN 1 (5 GAL) BUCKET</td>
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<tr>
<td>012</td>
<td>IKU 1581608</td>
<td>44 09 89N</td>
<td>106 0</td>
<td>LUNENBURG</td>
<td>NO RECOVERY</td>
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<td>013</td>
<td>IKU 1581618</td>
<td>44 09 89N</td>
<td>106 0</td>
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<td>NO RECOVERY</td>
</tr>
<tr>
<td>014</td>
<td>IKU 1581624</td>
<td>44 09 90N</td>
<td>106 0</td>
<td>LUNENBURG</td>
<td>SAMPLER JAMMED OPEN WITH LARGE BOULDER 1/3 FULL PANNED SUBSAMPLE, NO GOLD BOULDERS COBBLES SAND SILT AND MUD STARFISH (BRITTLE) GREENISH BROWN COLOR AREA OF EXPOSED BEDROCK WITH SAND AND GRAVEL IN RIPPLES BETWEEN RIDGES</td>
</tr>
<tr>
<td>015</td>
<td>IKU 1581825</td>
<td>44 01 22N</td>
<td>165 0</td>
<td>PENNANT PT</td>
<td>3 PUSH CORES (AREA E) 40CM EACH GREY MUD NO STONES (LAHAVE CLAY) FEW SHELLS MORaine TOP SAMPLE (PENNANT PT) MAY HAVE MISSED TOPOGRAPHIC HIGH MISSED TILL GOT LAHAVE CLAY ALL MUD NO STONES SAMPLE STORED IN 3 (5 GAL BUCKETS) LABELLED - SURFACE LAG JUST BENEATH SURFACE, BOTTOM</td>
</tr>
<tr>
<td>016</td>
<td>IKU 1591236</td>
<td>43 55 27N</td>
<td>165 0</td>
<td>LUNENBURG</td>
<td>ICEBERG FURROWS ON SEABED GOOD TILL-SOUTH SHORE MORaine SLATE FACETTED-STRIATED SOME GRANITE CLASTS GREY BROWN MUDDY MATRIX TOP</td>
</tr>
</tbody>
</table>
Appendix 3

SAMPLE PANNED: 2 SUBSAMPLES 1= PEBBLE 2= GRANULE SAMPLE STORED IN 5 (5 GAL BUCKETS) 1 TOP, 2 MIDDLE, 2 BOTTOM PUSH CORE WAS STARTED IN THE MIDDLE OF GRAB AVOIDED SURFACE LAG PUSH CORE LABELLED AS 91-18-015 SUBSAMPLES LABELLED AS IKU-16 PEBBLE, IKU-16 GRANULE

017 IKU 1591424 44 05 16N 64 02 71W 82 3 LUNENBURG (SEISMIC) SABLE ISL SAND ON REWORKED TILL, CLASTS ANGULAR AND ROUNDED SLATE GRANITE 5-10 %, SANDY MATRIX, COARSE GRANULES, GREY QTZ VEIN IN FLOAT 2 SUBSAMPLES IKU-17 GRAVEL GRANULE, IKU-17 GRANULE, SAMPLE STORED IN 2 (5 GAL) BUCKETS 1 BUCKET RETAINED FOR PANNING

019 IKU 1591746 44 15 60N 64 10 43W 56 0 SOUTH CROSS ISLAND JUST RECOVERED ROUNDED BOULDERS OF METAGRAYWACKE 1-2 FT IN DIAMETER 10-12 BOULDERS IN 4 TRIES

020 IKU 1591838 44 14 85N 64 09 40W 53 0 SOUTH CROSS ISLAND (SEISMIC) BOULDERY SURFACE INTERMIXED WITH GRAVEL TILL OR BEDROCK COVER MAYBE LATE TILL 1 BOULDER RETAINED BOULDER IN FIRST 2 ATTEMPTS, WATER IN THIRD

021 IKU 1591928 44 12 96N 64 16 75W 40 0 LUNENBURG GRAVEL RIPPLES IN BEDROCK FIRST ATTEMPT WAS WATER SECOND ATTEMPT BOULDERS AND SOME GRAVEL SAMPLE STORED IN 1 (5 GAL) BUCKET

023 IKU 1601650 44 19 33N 64 15 01W 20 0 DIRECTLY OFF OVENS PARK 8 ATTEMPTS- ONLY ONE HAS SMALL SAMPLE MUDDY GRAVEL- OTHERS HAVE BOULDERS IN JAWS UNUSUAL BOULDERS OF BRIDGEPATCH CONGLOMERATE IN ONE ATTEMPT MUDDY GRAVEL WASHED DOWN 3 BAGGED SUBSAMPLES FROM PANNING 1= GRAVEL SAMPLE, 2= GRANULE SAMPLE 3= HEAVIES SAMPLE 1 BUCKET SAMPLE OF BRIDGEPATCH CONGLOMERATE

024 IKU 1601850 44 18 19N 64 00 06W 70 0 LUNENBURG MATRIX-SILTY-SAND, SOME CLAY GRAVEL- PEBBLES GREY BROWN LARGE CLASTS SOME STRIATED HALIFAX SLATE TOP OF INTERPRETED DRUMLIN (SEE CONTACT) DRUMLIN EXPOSURE 3 (5 GAL) BUCKETS OF LARGE CLASTS 1 BAG OF MATRIX

025 IKU 1601915 44 17 33N 64 00 96W 71 0 LUNENBURG DRUMLIN EXPOSURE TOP OF DRUMLIN 1 (5 GAL) BUCKET SAMPLE GOOD TILL, LOTS OF MATRIX GREY-BROWN, SILTY-SAND MATRIX

026 IKU 1611148 44 35 95N 63 21 36W 18 0 SOUTH OF LAWRENCE TOWN SAMPLING BOULDER ARMOR SOUTH OF LAWRENCE TOWN BEACH 2 (5 GAL) BUCKETS OF GRAVEL ROCKS

027 IKU 1611240 44 29 68N 63 19 65W 62 0 SOUTH OF THREE FATHOM HARBOUR HUMMOCKY TILL SURFACE BOULDERY ROUNDED AND ANGULAR CLASTS 5 AND Y MATRIX 3 (5 GAL) BUCKETS + 1 BAG OF MATRIX STORED IN A (1 GAL) BUCKET

028 IKU 1611432 44 12 75N 63 17 83W 164 0 HALIFAX MORaine BROWN-GREY SILTY TILL, STRIATED CLASTS SURFACE LAG BOULDERY, SOME EDGE ROUNDED INCLUSIONS OF UNOXYDIZED GREY-BLACK SILTY TILL (DRY) WITHIN BASE OF GRAB THIS WAS SAVED AS
Appendix 3

A bag sample, Halifax Moraine continuation, mound of till (seismic). 1 push core labelled 91-18-028 and 4 (5 gal) buckets. Push core was beneath surface lag. Buckets labelled as 1= top, 2= middle and 2= bottom.

029 IKU 1611640 44°01'13"N 63°13'50"W 810 HALIFAX HARBOR North Sambro Bank: till mound or moraine sample. Grey-brown stony till; silty-clay-sand matrix clasts; metawacke, pink granite, inclusions of red sediment. Few shells edge rounding. Sample stored in 5 (5 gal) buckets. 1 top, 2 middle and 2 bottom. 1 push core labelled 91-18-029. Refer to 3 5kHz record at 161/1640.

031 IKU 1621415 44°29'45"N 63°19'51"W 640 OLD SACKVILLE RIVER VALLEY Surface till sheet transgressed; sandy matrix, slate-metawacke clasts, edge rounded, looks like Beaver River till. Local lithology: large boulders, cobble, 2 (5 gal) buckets. Large boulders, pebbles, some matrix recovered put in bucket.

032 IKU 1621440 44°28'41"N 63°19'57"W 820 EAST OF OLD SACKVILLE RIVER VALLEY Surface till sheet or outwash. Mounds? gravel surface. (Sidescan) grey-brown sandy matrix. Metawacke clasts rounded to subangular. Beaver River till: 7 large boulders. Same as previous till: 3 (5 gal) buckets. 2 boulders-pebble lags, 1 bucket of matrix till.

033 IKU 1621650 44°27'52"N 63°23'49"W 640 SOUTH OF DARTMOUTH Till sheet: hummocky surface. (Sidescan) gravel-boulders. Grey sandy matrix (silt < 20%) unwashed, clasts subrounded to subangular. Metagraywacke > 90%, granite < 5% shell fragments, pelecypod. Beaver River till: 3 (5 gal) buckets. 2 for top lags and 1 for bottom till. 1 subsample of matrix till for grain size labelled as 91-18-33.


044 IKU 1631825 44°29'73"N 63°23'73"W 550 TILL SHEET, SOUTH CONRODS HEAD Cobble-few boulders: seismic-sidescan cobble lag subrounded to subangular. Metagraywacke, < 2% granitoid (south MT). Matrix-coarse sand, olivewrey. Some matrix retained in grab. (Beaver River till?) Sample stored in 2 (5 gal) buckets.

047 IKU 1631904 44°29'30"N 63°24'31"W 540 INNER SHELF, SOUTH OF HALIFAX Sample stored in 2 (5 gal) buckets. Labelled matrix and lag black sidescan cobble surface, a few boulders. diamicton-till, cobble-boulder lag. Olive grey silty.
Appendix 3

SAND TILL UNDERNEATH LAG SURFACE PROTECTS AN UNDERLYING TILL MATRIX WHICH IS NOT WASHED > 90% METAGRAYWACKE CLASTS; SANDY MATRIX

SAMPLE STORED IN 2 (5 GAL) BUCKETS LABELLED MATRIX AND LAG TILL-SECTION (SEISMIC INTERP) OLIVE- GREY SANDY MATRIX; UNWASHED-SUBROUNDED-SUBANGUUR CLASTS; LAG SURFACE OF METAGRAYWACKE 1 LARGE METASOMATIZED ? GRANITE BOULDER

SAMPLE STORED IN 4 (5 GAL) BUCKETS LABELLED AS 1=TOP, 1=MIDDLE, 1=BOTTOM AND 1 LABELLED RED LAYER TILL MOUND ADJACENT TO POSSIBLE LIFTOFF? MORaine OLIVE GREY SAND/ SILTY DIAmICTON-TILL STRIATED FACETTED METAGRAYWACKE CLASTS AND GRANITE CLASTS ANOMALOUS RED GREY LAYER IN TILL (SAMPLE)

TILL MOUND SURROUNDED BY GLACIO MARINE EMERALD SILT OLIVE GREY SILTY TILL MORaine GREEN SAND AND RED CLAY INCLUSIONS SHELL FRAGMENTS METAGRAYWACKE CLASTS, SUBROUNDED TO SUBANGUUR, SURFACE COBBLE LAG-TILL DOMINATED BY PEBBLE-SIZED CLASTS 1 PUSH CORE FROM AREA (E) SAMPLE STORED IN 4 (5 GAL) BUCKETS LABELLED 1=TOP, 1=MIDDLE AND 2 FROM THE BOTTOM

SURFACE LAG SILTY-SANDY DIAmICTON-TILL? WITH SUBROUNDED TO SUBANGUUR CLASTS-METAGRAYWACKE (>90 %) RED SILTSTONE, VOLCANIC, METAMORPHIC CLASTS <5%, GRADES INTO MASSIVE GREYISH CLAY-SILTY-CLAY FEW STONES 1 PUSH CORE OF BOTTOM SILTY-CLAY SEDIMENTS POSSIBLE EMERALD SILT (A) (SEE SEISMIC) GRAB WELL STRATIFIED, NOT MIXED IN SITU 1 SUBSAMPLE OF LITHOLOGY SAMPLE STORED IN 5 (5 GAL) BUCKETS 1 FROM TOP, 2 FROM MIDDLE AND 2 FROM THE BOTTOM

OLIVE GREY SANDY-SILTY DIAmICTON (TILL) SAND INCLUSIONS, COBBLE-BOULDER LAG SURFACE METAGRAYWACKE CLASTS >90 % SUBROUNDED TO SUBANGUUR, 1 PINK GRANITOID CLAST ON SURFACE (SYENITE) SAMPLE STORED IN 4 (5 GAL) BUCKETS LABELLED 1=TOP, 1=MIDDLE AND 2 FROM BOTTOM 1 SUBSAMPLE OF ERRATIC GRANITE IN A BAG

TOP OF MORaine-ICEBERG FURROWS? OLIVE-GREY SILTY SANDY MATRIX (DIAmICTON) METAGRAYWACKE >90 % GRANITE, MUD CLASTS (CARB 7) <5 %) SANDY INCLUSIONS BURROWS INSIDE SAND 2 BUCKETS FROM THE TOP AND 2 FROM THE BOTTOM BASE OF CLASTS-METAGRAYWACKE GRANITE, MUD CLASTS 1 GREY GREEN MUDSTONE

TOP OF EASTERN SHORE MORaine SURFACE LAG-BOULDER-COBBLE LAYER
Appendix 3

METAGRAYWACKE ANGULAR 10 SUBROUNDED, CLASTS (> 90%) SOME FACETTED AND STRIATED BECOMING CLAYISH AND LESS STONY BENEATH THE SURFACE MATRIX OLIVE-GREY SILTY-CLAY SAND RED-BROWN CLAYEY LAYERS- INCLUSIONS 1 GRANITE CLAST SOME CLASTS VERY ANGULAR AND SOME WITH EDGE ROUNGING SAMPLE STORED IN 5 (5 GAL) BUCKETS 2 FROM TOP, 2 FROM BOTTOM AND 1 OF CLASTS

EMERALD SILT (ERODED) COBBLE- BOULDER SURFACE LAG METAGRAY- WACKE >90% SILTY-CLAY SAND MATRIX BELOW OLIVE GREY SILTY DIAMICTON COBBLE-PEBBLE SIZED CLASTS Boulders ENCRUSTED WITH POLYPS- LARGE WORM RECOVERED 1 BAG OF PURPLE SLIME OF UNKNOWN ORIGIN SAMPLE STORED IN 3 (5 GAL) BUCKETS 1 FROM THE TOP AND 2 FROM THE BOTTOM

HUMMOCKY MOUNDS OF TILL BOULDERY LAG IN IKU, METAGRAYWACKE GRANITE SOME MATRIX (OLIVE GREY SAND), VERY BOULDERY SAMPLE STORED IN 2 (5 GAL) BUCKETS 1 OF BAG BOULDERS AND 1 OF MATRIX

SAND BEDFORMS OVERLYING ARMOUR UNCONFORMING ESTUARINE/GlacIOmarine SEDIMENTS (SEISMIC) ARMOUR OF COBBLES BOULDERS, DISCOIDAL CLASTS (SHINGLE) OVERLYING OLIVE GREY BLACK SILTY SAND MATRIX, TILL? (DIAMICTON) SAMPLE STORED IN 3 (5 GAL) BUCKETS 1 LABELLED LAG 1 LABELLED MIXED AND 1 LABELLED MATRIX

SAND BODY (WHITE ON SIDESCAN) COBBLE LAG-DISCOID CLASTS (SHINGLE SHAPE) METAGRAYWACKE, GRANITE LITHOTHAMNION ENCRUSTATION ON CLASTS, MATRIX, SAND 2 (5 GAL) BUCKETS

BEACH GRAVEL (SEISMIC) DISCOID CLASTS METAWACKE-GRANITE SAND MATRIX WAS WELL SORTED GREY-GREEN IN COLOR (BEACH) MISSED ON 1ST ATTEMPT, OVERSHOT, 2 (5 GAL) BUCKETS

UNIT-C OUTCROPS ON SURFACE BETWEEN 2 DEEP VALLEYS FILLED WITH BEACH (UNIT-B) AND ESTUARINE SEDIMENT (UNIT-G) (SEISMIC) OLIVE-GREY STONEY SANDY DIAMICTON (TILL) METAGRAYWACKE CLASTS 1 GRANITE CLAST 3 (5 GAL) BUCKETS OF MIXED SAMPLE (LAG + SUBSURFACE) (SMALL VOLUME IN GRAB)

TILL MOUNDS (SEISMIC) OLIVE-GREY STONEY DIAMICTON (TILL) SUBANGULAR TO SUBROUNDED CLASTS, MOSTLY SUBROUNDED SANDY MATRIX METAGRAYWACKE CLASTS >90% SAMPLE WELL WASHED DURING RECOVERY 4 (5 GAL) BUCKETS 1= TOP, 3= MATRIX
| 068 | IKU 1681127 | 45 05 78N | MOUTH OF COUNTRY HARBOUR | OLIVE-GREY MEDIUM GRAINED SAND WITH ABUNDANT SAND SIZED SHELL FRAGMENTS SAMPLE ALSO CONTAINS ABUNDANT SAND DOLLARS 4 (5 GALL) BUCKETS |
| 069 | IKU 1681147 | 45 05 46N | MOUTH OF COUNTRY HARBOUR | DESCRIPTION- OLIVE-GREY MEDIUM SAND WITH SAND SIZED SHELL FRAGMENTS NICE SMELL TO THE SAMPLE!! PROBABLY A HIGH CONCENTRATION OF HEAVIES IN SAMPLE SAND DOLLARS AND OYOHOG SHELL 4 (5 GALL) BUCKETS |
| 070 | IKU 1681202 | 45 05 11N | MOUTH OF COUNTRY HARBOUR | ON THE DOME GRAVEL LAG OVER MATRIX MUDDY SAND LITHOTHAMNION ON ALL ROCKS 2 (5 GALL) BUCKETS OF MATRIX ONLY 4 (5 GALL) BUCKETS OF FULL SAMPLE THIS IS A SPECIAL SAMPLE FOR ASSAY SAMPLE ACCUMULATED IN 4 GRABS AT THIS SITE |
| 071 | IKU 1681226 | 45 05 12N | MOUTH OF COUNTRY HARBOUR | MUDDY GRAVEL CLASTS ARE COVERED WITH LITHOTHAMNION SLATE-METAGRAYWACKE >90% 1 (5 GALL) BUCKET PANNED SAMPLE, NO VISIBLE GOLD LOTS OF HEAVIES (GARNET) |
| 072 | IKU 1681236 | 45 04 81N | MOUTH OF COUNTRY HARBOUR | OLIVE-GREY MEDIUM SAND WITH GRAVEL CLASTS SLATE METAGRAYWACKE >90%-GRANITE-QUARTZ VEIN (NO VG) 1 (5 GALL) BUCKET PANNED SAMPLE NO VISIBLE GOLD BUT LOTS OF HEAVIES |
| 073 | IKU 1681242 | 45 04 60N | MOUTH OF COUNTRY HARBOUR | OLIVE-GREY MEDIUM SAND (NO GRAVEL) SAND DOLLARS 2 (5 GALL) BUCKETS |
| 074 | IKU 1681329 | 45 04 52N | MOUTH OF COUNTRY HARBOUR | OLIVE-GREY MEDIUM SAND FEW COBBLES SAND DOLLARS (SAND WEDGE SEISMIC) PANNED SAMPLE - ABUNDANT GARNET, A FEW COBBLES SILT-SIZED 3 (5 GALL) BUCKETS SAMPLER WAS 1/4 FULL |
| 075 | IKU 1681349 | 45 03 70N | SOUTH OF COUNTRY HARBOUR | GRAVEL AND COBBLES SANDY MATRIX OLIVE GREY DISCOID CLASTS, SLATE-(DOME) METAGRAYWACKE > 90% SOME GRANITE, REWORKED TILL (SEISMIC) SABLE ISLAND SAND-GRANULE 2 (5 GALL) BUCKETS SAMPLER WAS 1/8 FULL |
| 076 | IKU 1681437 | 45 00 43N | SOUTH OF COUNTRY HARBOUR | (SEISMIC) SABLE ISLAND SAND-GRANULE BROAD CHANNEL COBBLE LAG METAGRAYWACKE (DOME) SUBANGULAR TO SUBROUNDED MOSTLY SUBROUNDED, MATRIX OLIVE-GREY (5%) SAND MEDIUM, NOT AS MUCH ROUNDED AS PREVIOUS SAMPLE SAMPLE THOROUGHLY WASHED ON RETRIEVAL 1 (5 GALL) BUCKET SAMPLER WAS 1/4 FULL |
| 077 | IKU 1681630 | 44 59 61N | SOUTH OF COUNTRY HARBOUR | SEISMIC RIPPLE ZONE BETWEEN 2 ROCK KNOBS OLIVE-GREY STONY DIAMICTON-METAGRAYWACKE (DOME) CLASTS > 90% SUBANGULAR TO SUBROUNDED (WIDE RANGE) SILTY-CLAY SAND MATRIX SOME BLACK ORGANIC CLAY IN MATRIX 3 (5 GALL) BUCKETS OF MATRIX SAMPLE LAG LEFT |
Appendix 3

082  IKU 1691250  44 50 05N  82 0  SOUTH OF COUNTRY HARBOUR
   61 31 71W
BEHIND SAMPLER WAS 1/2 FULL
ON TOP OF BRASS NUT MOUND,
TRANSRESSED COUNTRY HBR
MORAINE, LAG GRAVEL OVER TILL
LAG-COBBLE BOULDER SUBROUNDED, TILL;
OLIVE-GREY SANDY DIAMICTON,
SUBROUNDED TO ANGULAR CLASTS.
METAGRAYWACKE > 95% 2 (5 GAL)
BUCKETS SAMPLER WAS 1/4 FULL

083  IKU 1691320  44 50 18N  82 0  SOUTH OF COUNTRY HARBOUR
   61 31 55W
ONSIDE OF MOUND (BRASS NUT) COB SUBANGULAR TO
UNIT-3, SUBROUNDED, METAGRAYWACKE
AND QUARTZ VEIN OVERLYING
GREY-BROWN SANDY DIAMICTON, 20CM,
UNIT-2 OVERLYING GREY CLAY WITH
SHELLS, FEW STONES UNIT-1 (EMERALD
SILT) 3 (5 GAL) BUCKETS, 1=TOP, 1=MIDDLE
AND 1= BOTTOM SAMPLER WAS 3/4 FULL

084  IKU 1691409  44 40 25N  82 0  SOUTH OF COUNTRY HARBOUR
   61 24 11W
MORAINE MOUND (TRANSRESSED)
COBBLE-BOULDER LAG QUARTZ VEIN
IN BOULDER (MORAINE) METAGRAYWACKE
>95% TILL UNDERNEATH OLIVE-GREY
SANDY DIAMICTON METAGRAYWACKE
SUBROUNDED TO SUBANGULAR.
METAGRAYWACKE >95% 1 (5 GAL) BUCKET
OF LAG DEPOSIT AND 1 (5 GAL) BUCKET OF
TILL BENEATH (BOTTOM) SAMPLER WAS 1/2
FULL

085  IKU 1691515  44 42 46N  109 0  SOUTH OF COUNTRY HARBOUR
   61 21 71W
SEISMIC MOUND MORAINE OVERLAIN BY
VENEE OF EMERALD SILT SAMPLE 3
UNITS. TOP (MORAINE) TOP UNIT-
COBBLE-BOULDER LAG METAGRAYWACKE,
MIDDLE UNIT- (10 40CM)OLIVE GREY-SANDY
DIAMICTON UNIT 3+( 40CM) GREY-BROWN
(REDDISH TINGE) CLAY WITH FEW STONES.
3 (5 GAL) BUCKETS 1= TOP LAG, 1= MIDDLE
SAMPLE, 1= BOTTOM SAMPLER WAS 3/4
FULL

086  IKU 1691650  44 40 08N  107 0  COUNTRY HARBOUR
   61 20 95W
SURFACE OUTCROPPING TILL TONGUE
OVERLAIN BY EMERALD SILT, THEN
TRUNCATED BY MARINE EROSION ?
COBBLE-BOULDER LAG METAGRAYWACKE
SUBROUNDED TO SUBANGULAR OVERLAYER
OLIVE-GREEN-GREY SANDY DIAMICTON, >
95% METAGRAYWACKE CLASTS A THIN
CLAY LAYER NOTED AT BASE PROBABLY
OVERLIES EMERALD SILT PANNED SAMPLE.
CLASTS MOSTLY SUBANGULAR QUARTZ
VEIN, NO VG (VISIBLE GOLD) ILMANTITE
GARNET 2 (5 GAL) BUCKETS, 1= LAG
(SURFACE), 1= BOTTOM, SAMPLER WAS 1/2
FULL

089  IKU 1691920  44 44 62N  106 0  COUNTRY HARBOUR
   61 45 35W
MORAINE
BOTTOM-UNIT-1 GREY-SANDY CLAY-SILT/
CLAY MASSIVE, SHELL BEARING,
HEAVY > 6CM MIDDLE- UNIT 2 OLIVE GREY
SILTY SANDY DIAMICTON 2-6CM TOP- UNIT
3 BOULDER COBBLE LAG, SUBANGULAR TO
SUBROUNDED METAGRAYWACKE > 95%.
PUSH CORE WAS TAKEN UNDERNEATH THE
SURFACE LAG SAMPLE STORED IN 3
BUCKETS, 1 (2 1/2 GAL) BUCKET LABELLED
TOP 1 (2 1/2 GAL) BUCKET LABELLED
MIDDLE 1 (5 GAL) BUCKET LABELLED
Appendix 3

091
IKU 1701248
45 02 53N
61 18 24W
73 0
SOUTH OF TOR BAY

BOTTOM APPROX 70CM FROM TOP TO BOTTOM SAMPLER WAS FULL TO BRIMMING

BEDROCK RIDGE LAG DEPOSITS (SEISMIC) COBBLE-BOULDER LAG- WELL ROUNDED COBBLES-BOULDERS MIXED WITH SUBANGULAR COBBLES-PEBBLES, OVERLYING GREY DIAMICTON, METAGRAYWACKE >95% SOME RED INCLUSIONS IN MATRIX DERIVED LAWRENCTOWN TILL 1 (25 GAL) BUCKET LABELLED BOTTOM CONTAINS UNDERLYING TILL MATRIX AND 2 (1 GAL) BUCKETS LABELLED TOP CONTAINING TRANSGRESSIVE LAG + 1 BAG OF MATRIX FOR GRAIN SIZE SAMPLER WAS 1/2 FULL

092
IKU 1701340
45 01 77N
61 18 71W
93 0
SOUTH OF TOR BAY

ERODED UNIT (C) TILL IN BEDROCK CHANNEL (SEISMIC) COBBLE BOULDER LAG SLATE-ANGULAR (>80%) ROUNDED GRANITE (CANADIAN PLUTON?) UNDERLYING GREY MD BROWN SILTY CLAY DIAMICTON FEW STONES NOTE REDDISH TINGE ARM WAS BENT NEEDED 2 ATTEMPTS 2 (1 GAL) BUCKETS LABELLED TOP LAG AND 1 (2 1/2 GAL) BUCKET OF UNDERLYING TILL LABELLED BOTTOM + 1 BAG OF GRANITE (GLAST LITHOLOGY)

093
IKU 1701414
44 57 07N
61 21 64W
112 0
SOUTH OF NEW HARBOUR RIVER

ERODED EMERALD SILT (A) AT LOW SEA LEVEL POSITION (SEE HUNTEC, SIDESCAN) PEBBLE-LAG (5CM) BROWNISH-GREY DIAMICTON METAGRAYWACKE >90% (UNIT-3) OVERLYING GREYISH-GREEN SANDY SILT DIAMICTON (UNIT-2) OVERLYING (15-40CM) CLAY WITH REDDISH INCLUSIONS (SHELL FRAGMENTS) (UNIT 1) 4 (1 GAL) BUCKETS 2= TOP 1=MIDDLE AND 1= BOTTOM PUSH CORES (BLOW BY) 4-10CM OF SEDIMENT (COMRESS) SAMPLER WAS FULL

094
IKU 1701633
45 03 86N
61 48 87W
11 00
WINE HEAD

WINE HEAD SAND SAMPLE FIRST ATTEMPT HAD KELP AND MUSSELS WASHED AND REWORKED GRAVEL AND COBBLES WITH ORGANIC SILTY SAND (SECOND KELP) 1 (5 GAL) BUCKET + 3 BAGS SAMPLER WAS 1/2 FULL

095
IKU 1701648
45 03 85N
61 48 60W
18 00
WINE HEAD

BROWN COBBLY-SAND OXIDIZED OVER BLACK ORGANIC SILTY SAND WITH PEBBLES (ESTUARINE) 3 BAGS LABELLED TOP BOTTOM AND SURFACE GRAVEL SAMPLER WAS 1/2 FULL
Appendix 4

MARINE SEDIMENT DATA
91018-DAWSON CRUISE

SAMPLE TYPES, ASSOCIATED LANDFORMS AND LITHOLOGIC AND SHAPE OBSERVATIONS

1 = Sample number
2 = Sample type (0=full, 1=surface lag, 2=matrix, 3=top, 4=middle, 5=bottom.)
3 = Sediment Class (G-Gravel, GS-Gravelly sand, D-Diamicton, S-Sand, M-Mud)
4 = Landform (TD-Transgressive deposit, UC- Unit C, UB- Unit B, RM-Moraines (Morainal Zone), SM-Scotian Shelf End Moraines.
5 = Duplicate status (D-Duplicate if true .T)
## Appendix 4

<table>
<thead>
<tr>
<th>1</th>
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<th>3</th>
<th>4</th>
<th>5</th>
<th>Comments</th>
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<tr>
<td>006</td>
<td>0</td>
<td>G</td>
<td>TD</td>
<td>F</td>
<td>Rounded to well rounded. Foreigns include North Mountain basal, volcanics, pink, k-spar granitoids.</td>
</tr>
<tr>
<td>009</td>
<td>1</td>
<td>G3</td>
<td>TD</td>
<td>F</td>
<td>Subangular to subrounded.</td>
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<td>GS</td>
<td>TD</td>
<td>F</td>
<td></td>
</tr>
<tr>
<td>010</td>
<td>0</td>
<td>D</td>
<td>UC</td>
<td>F</td>
<td>Subangular to subrounded. Foreign include North Mountain basalt, volcanic fragments and pink k-spar granitoids.</td>
</tr>
<tr>
<td>011</td>
<td>0</td>
<td>D</td>
<td>UC</td>
<td>F</td>
<td>Angular to subrounded. 27 Fragments of an indurated diamict. Foreign include coarse, feldpathic sandstone, pink granitoid.</td>
</tr>
<tr>
<td>014</td>
<td>0</td>
<td>G</td>
<td>UB</td>
<td>F</td>
<td>Subangular to subrounded. Foreigns include amygdaloidal basalt, red and grey clastic sedimentary rocks, volcanics?</td>
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<tr>
<td>015</td>
<td>3</td>
<td>M</td>
<td>ES</td>
<td>F</td>
<td>No granule or pebble-sized clasts in sample.</td>
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<tr>
<td>015</td>
<td>5</td>
<td>M</td>
<td>ES</td>
<td>F</td>
<td></td>
</tr>
<tr>
<td>016</td>
<td>3</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to subrounded. Foreigns include foliated granitoids (gneiss), volcanic, red and grey clastics, pink k-spar granitoid.</td>
</tr>
<tr>
<td>016</td>
<td>4</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to subrounded mostly subangular. Metapelitic lithologies with abundance of siliceous, greenish (Epidote?) alteration and disseminated sulphides. Foreign include mafic granitoids, porphyritic felsic volcanics, and red metasiltstone. Presence of an indurated grey diamict, possible older tills?</td>
</tr>
<tr>
<td>016</td>
<td>5</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to subrounded. Foreigns include mafic and felsic volcanics, grey and red clastic sedimentary rocks, pink, k-spar granitoids.</td>
</tr>
<tr>
<td>016</td>
<td>5</td>
<td>D</td>
<td>SM</td>
<td>T</td>
<td>Angular to subrounded. Foreign include mafic and felsic volcanics, red and grey felsic volcanics, pink K-spar granitoids.</td>
</tr>
<tr>
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<td>UB</td>
<td>F</td>
<td>Subangular to subrounded. North Mountain Basalt Volcanic, red and grey sedimentary</td>
</tr>
<tr>
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<td>TD</td>
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<td>Subangular to subrounded. Foreign include pink k-spar granitoids.</td>
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<tr>
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<td>RM</td>
<td>F</td>
<td>Subangular to subrounded. Foreigns include foliated granitoids, gneiss?, grey sandstone.</td>
</tr>
<tr>
<td>025</td>
<td>0</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Subangular to subrounded. Foreigns include granitoid, Mwolcanics, mostly unrecognizable.</td>
</tr>
<tr>
<td>027</td>
<td>2</td>
<td>D</td>
<td>UB</td>
<td>F</td>
<td>Subangular to subrounded. Foreign lithologies include foliated mafic igneous rocks, granitoids.</td>
</tr>
<tr>
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<td>Comments</td>
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</tr>
<tr>
<td>028</td>
<td>3</td>
<td>D</td>
<td>UB</td>
<td>F</td>
<td>Subangular to subrounded mostly subangular. Meguma metapelitic clasts mineralized-chalcopyrite?</td>
</tr>
<tr>
<td>028</td>
<td>5</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subrounded to subangular.</td>
</tr>
<tr>
<td>029</td>
<td>3</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Mostly subrounded to subangular, no rounded clasts; two or more directions indicated on some clasts; One mineralized Meguma clast (sulphides -grey metallic-arsenopyrite-galena?)</td>
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<tr>
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<td>4</td>
<td>D</td>
<td>RM</td>
<td>T</td>
<td>Subangular to subrounded. Foreigns include grey and red clastic sedimentary rocks, pink, k-spar granitoids, diorite, and felsic volcanics.</td>
</tr>
<tr>
<td>029</td>
<td>4</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Subangular to subrounded most of the larger clasts (&gt;8.00) m are striated. Quartz vein fragments mineralized with pyrite. Foreign include 1 amygdaloidal basalt fragment, 3 k-spar granitoids and aphanitic granitoids.</td>
</tr>
<tr>
<td>029</td>
<td>5</td>
<td>D</td>
<td>RM</td>
<td>T</td>
<td>Subrounded to subangular. Foreign include volcanic, pink k-spar granitoids, limestone (micrite) with brachiopod shells-Windsor Group?, red mudstone, grey sedimentary.</td>
</tr>
<tr>
<td>029</td>
<td>5</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Subangular to subrounded. Foreigns include pink kspar granitoids, diorite, red and grey clastic sedimentary rocks.</td>
</tr>
<tr>
<td>031</td>
<td>0</td>
<td>D</td>
<td>UC</td>
<td>F</td>
<td>Rounded to subrounded mostly rounded. No surface markings. Many stained with iron oxide. Foreign include volcanics.</td>
</tr>
<tr>
<td>032</td>
<td>1</td>
<td>D</td>
<td>UB</td>
<td>F</td>
<td>Rounded to subrounded. No striated clasts. Foreigns include sandstones, volcanic, granitoids.</td>
</tr>
<tr>
<td>032</td>
<td>2</td>
<td>D</td>
<td>UB</td>
<td>F</td>
<td>Subrounded; bullet-nose clasts modified by abrasion, edge rounding, foreigns include foliated granitoids with pink k-spars.</td>
</tr>
<tr>
<td>033</td>
<td>3</td>
<td>D</td>
<td>UB</td>
<td>F</td>
<td>Subangular to subrounded, mostly subrounded. Foreign clasts include red and grey clastic sedimentary, pink, granitoids and felsic volcanic.</td>
</tr>
<tr>
<td>042</td>
<td>0</td>
<td>G</td>
<td>UC</td>
<td>F</td>
<td>Rounded to well rounded clasts some discoid. No striations. Foreign components consist of greyish diotitic granitoids with porphyritic textures, volcanics and sandstones.</td>
</tr>
<tr>
<td>044</td>
<td>0</td>
<td>D</td>
<td>UC</td>
<td>F</td>
<td>Subrounded to rounded. No striations. Some bullet clasts, apparently reworked. Several greenish (silicified) metapelites. Surface foreigns include sandstones and granitoids.</td>
</tr>
<tr>
<td></td>
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<td>---</td>
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<td>-----------</td>
</tr>
<tr>
<td>048</td>
<td>1</td>
<td>D</td>
<td>UC</td>
<td>F</td>
<td>Subangular to subrounded, mostly subrounded. A mineralized, subangular and striated clast has flecks of chalcopyrite. Also appears silicified.</td>
</tr>
<tr>
<td>048</td>
<td>2</td>
<td>D</td>
<td>UC</td>
<td>F</td>
<td>Subangular to rounded, mostly subrounded, surface feature obliterated by algal growth; a mineralized clast seen, may be chalcopyrite.</td>
</tr>
<tr>
<td>049</td>
<td>3</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Angular to rounded; felsic volcanics, pink granitoids, red-grey ss, diorite-gabbro.</td>
</tr>
<tr>
<td>049</td>
<td>4</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Meguma clasts largely greenish metapelite few slate, Foreigns include porphyritic felsic volcanics, grey sandstone, pink kspar granitoids.</td>
</tr>
<tr>
<td>049</td>
<td>5</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Pebbles more rounded than 50 Bottom; subrounded mainly, a few subangular; Foreigns include porphyritic aphanitic microgranitoid, mafic igneous rocks, red ss, pink, k-spar granitoid-Meguma mostly grey metapelite</td>
</tr>
<tr>
<td>050</td>
<td>3</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Roundness sorting value high angular to rounded two distinct populations apparent, well rounded and glacial clasts; felsic prophyritic volcanic, ss, granitoids</td>
</tr>
<tr>
<td>050</td>
<td>4</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Sample contains subangular to subrounded clasts one rounded mineralized clast disseminated shiny in greenish metapelite note*5=bottom 4=middle 3=top 2=matrix 1=lag 0=full codes for iktupe.</td>
</tr>
<tr>
<td>051</td>
<td>3</td>
<td>D</td>
<td>ES</td>
<td>F</td>
<td>Angular to rounded; wide range of grain shapes in sample 51. Foreign include volcanic with phenocrysts, grey and red clastic sedimentary rocks, pink k-spar granitoids.</td>
</tr>
<tr>
<td>051</td>
<td>4</td>
<td>D</td>
<td>ES</td>
<td>F</td>
<td>Subangular to subrounded. Meguma mostly black metapsammite-pelite with a few schistose fragments. Foreign include pink, k-spar granitoids, red and grey clastic sedimentary rocks.</td>
</tr>
<tr>
<td>051</td>
<td>5</td>
<td>M</td>
<td>ES</td>
<td>F</td>
<td>Subangular to rounded. Only a few rounded. Foreign include an amygdaloidal basalt fragment, red sandstone and conglomerate and phenocrystic volcanics.</td>
</tr>
<tr>
<td>052</td>
<td>3</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Angular to rounded; wide range of particle shapes. Meguma mostly grey metapelite and metapsammite. Foreigns include mafic igneous (diorite); pink K-spar granitoids and volcanics.</td>
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## Appendix 4

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<td>D</td>
<td>SM</td>
<td>F</td>
<td>Angular to subrounded. Foreigns include diorites, porphyry, k-spar pink granitoids, red mudstone, sandstone, rhyolites.</td>
</tr>
<tr>
<td>052</td>
<td>5</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Angular to rounded. Meguma lithologies consist mainly of greenish metapelites and metapsammites. Foreigns consist of sandstone, volcanic porphyry, pink k-spar granitoids.</td>
</tr>
<tr>
<td>055</td>
<td>3</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Angular to subrounded. Meguma clasts have disseminated sulphides, veinlets of quartz. Foreign include grey and red mudstones and felsic volcanics.</td>
</tr>
<tr>
<td>055</td>
<td>5</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Angular to rounded. Meguma lithologies consist primarily of grey metapelite. Foreigns include porphyritic volcanics, grey and red clastic sedimentary rocks, pink, k-spar, phaneritic and aphanitic granitoids.</td>
</tr>
<tr>
<td>055</td>
<td>5</td>
<td>D</td>
<td>SM</td>
<td>T</td>
<td>Angular to subrounded, mostly subrounded. Foreign include red and grey clastic sedimentary rocks, greenish volcanics and pink granitoids.</td>
</tr>
<tr>
<td>056</td>
<td>3</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to rounded, mostly subrounded. Meguma consists of grey and greenish metapelite and metapsammite. Foreigns include grey and red clastic sedimentary rocks, volcanic porphyry, and pink k-spar granitoids.</td>
</tr>
<tr>
<td>056</td>
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<td>SM</td>
<td>F</td>
<td>Angular to rounded. Foreigns include gneiss and pink k-spar granitoids, volcanics and red sandstone.</td>
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<tr>
<td>057</td>
<td>3</td>
<td>D</td>
<td>ES</td>
<td>F</td>
<td>Subangular to subrounded. Some well rounded, discoid clasts. Foreign include red mudstone, granitoids, volcanics.</td>
</tr>
<tr>
<td>057</td>
<td>4</td>
<td>M</td>
<td>ES</td>
<td>F</td>
<td>Subangular to subrounded. Foreigns include granitoid clasts and grey clastic sedimentary rocks.</td>
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<tr>
<td>059</td>
<td>2</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Subangular to subrounded. Meguma lithologies consist of grey and green metapelite speckled with sulphides. Foreigns may be volcanic, origin doubtful.</td>
</tr>
<tr>
<td>060</td>
<td>1</td>
<td>G</td>
<td>UC</td>
<td>F</td>
<td>Subrounded to well rounded. No striated clasts.</td>
</tr>
<tr>
<td>061</td>
<td>0</td>
<td>G</td>
<td>TD</td>
<td>F</td>
<td>Subrounded to rounded. Foreign include volcanic?, grey sandstone.</td>
</tr>
<tr>
<td>063</td>
<td>0</td>
<td>D</td>
<td>UB</td>
<td>F</td>
<td>Subangular to well rounded. South Mountain type large granitoids.</td>
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<tr>
<td>064</td>
<td>2</td>
<td>D</td>
<td>RM</td>
<td>F</td>
<td>Subrounded to rounded. Foreign include phenocrystic volcanic and red clastic sedimentary rocks.</td>
</tr>
</tbody>
</table>
### Appendix 4

<table>
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<td>065</td>
<td>D</td>
<td>ST</td>
<td>F</td>
<td>Subangular to subangular. Till sample taken on land while in port at Sheet Harbour. One clast appears to be a transitional bullet-nose form. Foreign includes red sandstone.</td>
<td></td>
</tr>
<tr>
<td>070</td>
<td>070</td>
<td>G</td>
<td>TD</td>
<td>F</td>
<td>Subangular to rounded.</td>
<td></td>
</tr>
<tr>
<td>070</td>
<td>070</td>
<td>G</td>
<td>TD</td>
<td>F</td>
<td>Subrounded to well rounded. Meguma lithologies with sulphides. Foreign include granitoids, sandstone.</td>
<td></td>
</tr>
<tr>
<td>071</td>
<td>071</td>
<td>G</td>
<td>TD</td>
<td>F</td>
<td>Subrounded to well rounded. Greenish metapelites and spotted slates.</td>
<td></td>
</tr>
<tr>
<td>072</td>
<td>072</td>
<td>S</td>
<td>TD</td>
<td>F</td>
<td>Subangular to rounded. Meguma lithologies include schistose rocks, mostly metapelites, some hornfels.</td>
<td></td>
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<tr>
<td>075</td>
<td>075</td>
<td>S</td>
<td>TD</td>
<td>F</td>
<td>Subrounded to well rounded. Meguma lithologies mostly metapelite, a few schistose, muscovite-biotite bearing rocks (higher meta grade?)</td>
<td></td>
</tr>
<tr>
<td>076</td>
<td>076</td>
<td>S</td>
<td>TD</td>
<td>F</td>
<td>Subangular to subrounded. Meguma metapelites.</td>
<td></td>
</tr>
<tr>
<td>077</td>
<td>077</td>
<td>D</td>
<td>UC</td>
<td>F</td>
<td>Subangular to rounded.</td>
<td></td>
</tr>
<tr>
<td>082</td>
<td>082</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to subrounded. One clast appeared to be a glacially-reworked beach? pebble; well rounded with a basal facet and striations. Foreigns include banded volcanic rocks, and porphyritic granitoid.</td>
<td></td>
</tr>
<tr>
<td>083</td>
<td>083</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subrounded to rounded. Foreigns include volcanic, granitoids, sandstone.</td>
<td></td>
</tr>
<tr>
<td>083</td>
<td>083</td>
<td>4</td>
<td>SM</td>
<td>F</td>
<td>Subangular to rounded with many rounded clasts. Wide variety of Meguma lithologies including grey, greenish metapelites, metapsammites and buff metawackes. Foreigns include granitoids and volcanics.</td>
<td></td>
</tr>
<tr>
<td>083</td>
<td>083</td>
<td>5</td>
<td>SM</td>
<td>F</td>
<td>Subangular to well rounded. Meguma lithologies also include staurolite-andalusite? schistose fragments. Foreigns include pink, K-spar granitoid, red sandstone.</td>
<td></td>
</tr>
<tr>
<td>084</td>
<td>084</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to rounded, mostly rounded. Evidence of reworking of glacial clasts. Foreign clasts include a buff marlstone of unknown age (Tertiary?) as well as sedimentary and volcanic species.</td>
<td></td>
</tr>
</tbody>
</table>
## Appendix 4

<table>
<thead>
<tr>
<th>1</th>
<th>2</th>
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<tbody>
<tr>
<td>084</td>
<td>2</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to rounded with many rounded clasts. Wide variety of Meguma lithologies including grey, greenish metapelites, metapsammites, and buff metawackes. Foreigns include granitoids and volcanics.</td>
</tr>
<tr>
<td>085</td>
<td>3</td>
<td>D</td>
<td>ES</td>
<td>F</td>
<td>Rounded to subrounded. Abundance of soft, buff-coloured marlstones Tertiary?.</td>
</tr>
<tr>
<td>085</td>
<td>5</td>
<td>M</td>
<td>ES</td>
<td>F</td>
<td>Subangular to rounded. Many fragments of brownish marlstone. One fossiliferous limestone.</td>
</tr>
<tr>
<td>086</td>
<td>3</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to subrounded, but no striated. This in unusual because many diamictons with this angularity have striated clasts. Foreigns include small granitoids.</td>
</tr>
<tr>
<td>086</td>
<td>5</td>
<td>D</td>
<td>SM</td>
<td>F</td>
<td>Subangular to subrounded. Foreigns include two fossiliferous limestones (corals?, crinoid stems, brachiopods). Others include granitoids, volcanics, red and grey sedimentary rocks.</td>
</tr>
<tr>
<td>089</td>
<td>3</td>
<td>D</td>
<td>ES</td>
<td>F</td>
<td>Subangular to subrounded. Granitoid and volcanic clasts in foreign component.</td>
</tr>
<tr>
<td>089</td>
<td>5</td>
<td>M</td>
<td>ES</td>
<td>F</td>
<td>Angular to well rounded, an extreme mix of grain shapes. Foreigns include red mudstone.</td>
</tr>
<tr>
<td>091</td>
<td>3</td>
<td>D</td>
<td>UB</td>
<td>F</td>
<td>Angular to subrounded mostly subangular.</td>
</tr>
<tr>
<td>091</td>
<td>5</td>
<td>D</td>
<td>UB</td>
<td>F</td>
<td>Angular to subrounded mostly subangular. Meguma lithologies consist of coarse metapelite. Foreign include volcanic, pink k-spar granitoid.</td>
</tr>
<tr>
<td>092</td>
<td>3</td>
<td>D</td>
<td>ES</td>
<td>F</td>
<td>Subangular to rounded. Foreign include conglomerate, sandstone, pink, k-spar granitoid, volcanics?</td>
</tr>
<tr>
<td>092</td>
<td>5</td>
<td>D</td>
<td>ES</td>
<td>F</td>
<td>Subangular to subrounded. No striations in spite of the angular nature of the Sediments. Foreigns include red mudstone, grey sandstone and pink granitoids.</td>
</tr>
<tr>
<td>093</td>
<td>5</td>
<td>M</td>
<td>ES</td>
<td>F</td>
<td>Subangular to rounded. One porphyritic felsic volcanic clast.</td>
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<tr>
<td>094</td>
<td>0</td>
<td>G</td>
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<td>F</td>
<td>Subrounded to rounded. None striated. Metapelites with greenschist to biotite grade metamorphism. Foreigns include red and grey elastic sedimentary rocks, felsic, porphyritic volcanics, mafic igneous rocks.</td>
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**APPENDIX 5:** Terrestrial till data: locations and till types.

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<tr>
<th>Year</th>
<th>Sample</th>
<th>Subsample</th>
<th>Till type</th>
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APPENDIX 6: Terrestrial till data 1977-1992

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<td>subangular shapes. Meguma- mostly psammites. Foreigns- quartz, grey sandstone, diorite, red sandstone.</td>
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<td>35</td>
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<td>subangular to subrounded. Foreigns- greenish sandstone, hard red sandstone, pinkish granitoid, felsic volcanic</td>
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</table>
mostly subrounded shapes. Meguma- slates and metagreywackes. Foreigns- hard green siltstone (banded), diorites, syenitic granites, felsic volcanics

subangular to subrounded. meguma mostly metawackes and psammites. foreigns- grey sandstone, hard brown siltstone, volcanic silicic volcanic.

subrounded to subangular shapes. Meguma- pelites (striated) and psammites. Foreigns- grey micaceous sandstone, purple quartz fragments, diorite.

subangular shapes. Meguma- mostly psammites.

subrounded shapes. Foreigns- red sandstone, syenites, volcanics.

subangular shapes. Meguma- many schistose. Foreigns- sandstones.

subangular to subrounded shapes. foreigns- grey sandstones, mudstone, volcanics.

subangular shapes. Meguma- mostly psammites. Foreigns- quartz and feldspar fragments.

subrounded-subangular shapes. Meguma- mostly psammites. Foreigns- quartz, syenites, grey sandstones, diorite.

subangular shapes. Foreigns- grey sandstone, syenite, quartz.

subrounded- subangular shapes. Foreigns- red and brown sandstone, volcanics, syenites, quartz and feldspar, diorite.

subangular shapes. foreigns- quartz and feldspar fragments.

subangular. Meguma- mostly psammites. Foreigns- quartz.

subangular shapes. mostly psammites. Foreigns-grey sandstones, syenites, red sandstones, diorites.

subangular shapes. Foreigns- till fragments, grey sandstone, diorite, quartz, syenite.

subrounded- subangular shapes. Meguma- mostly pelites. Foreigns- red sandstone, grey sandstone, diorites, volcanics, syenites, quartz.
subangular to subrounded shapes. Foreigns- hard red siltstone, dioritic granites, volcanic ?.

subrounded- subrounded shapes. Meguma-mostly psammmites. Foreigns- grey sandstones, quartz, volcanics.

subrounded to subangular. Foreigns- grey sandstones, red sandstones, volcanics, diorite, syenite.

subangular to subrounded shapes. Foreigns- hard red purplish siltstone, granitoid, diorite.


subangular shapes. Meguma- mostly psammmites. Foreigns- grey sandstone, syenites.

subangular to subrounded shapes. Foreigns- red sandstones, grey sandstones, syenites, quartz, more subangular than subrounded. Foreigns- pink syenogranite, grey sandstone, volcanic subangular shapes. Meguma- psammmites and pelites. Foreigns- sandstones, diorites, syenites.

subrounded- subangular shapes. Foreigns- reddish sandstone, diorites, grey sandstone.

subangular shapes foreigns- reddish sandstones, syenites, volcanics, quartz and feldspar fragments, diorites.

subrounded to subangular shapes. Foreigns- red sandstones, grey sandstones, syenites, diorites, quartz fragments.

more subrounded than subangular shapes. Meguma-some schistose. Foreigns- red sandstones, grey sandstones, green-grey sandstones, syenites, diorites, volcanics, quartz and feldspar fragments, till fragment.
<table>
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| 91 | 22 | 4  | 280| 42 | 6  | 52 | 6   | subangular to subrounded shapes. Meguma-mostly psammites. Foreigns- red and grey sandstone, diorite, syenite, volcanics, quartz fragments, micaceous sandstone, gneissic ?.
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APPENDIX 7: Grain size data

SAMPLE: Sample numbers (see Appendixes 3-7)
TT: Till type
TOT: Total sample weight (dry)
PHI1: Weight of -1 PHI fraction (Gravel)
GR%: Gravel percentage
MA%: Matrix percentage
PWT: PAN WEIGHT
RAT: GRAVEL/MATRIX RATIO
Terrestrial samples

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Appendix 8

C ABLAT.FOR: SCHILLING & HOLLIN: SURFACE PROFILES FOR VALLEY GLACIERS
C
C PROGRAM SOURCE: THE LAST GREAT ICE SHEETS, DENTON G.H. & HUGHEST J. EDs,
C WILEY 1981 pp. 212-3
C
C TAU IS THE BASAL SHEAR STRESS
C RHO IS DENSITY OF THE ICE
C GR IS THE GRAVITATIONAL CONSTANT
C S IS THE "SHAPE: FACTOR"
C G IS THE BEDROCK ELEVATION
C H IS THE SURFACE ELEVATION OF THE ICE
C T IS THE ICE THICKNESS
C NOTE: LENGTHS CONVERTED TO METRES IN THE PROGRAM
C ELEVATION AT SNOUT NOT ENTERED.
C
C MODIFIED BY: D.W.RIGBY, NSDME DEC 87 FOR MICROSOFT FORTRAN V3.0
C
C SUBROUTINES GETFIL/SAVFIL CALLED FROM ULIB.LIB AVAILABLE AT NSDME
C
C FTAU -FIRST SHEAR FACTOR TA(3) 3 SUBSEQUENT SHEAR FACTORS
C
C ---------------------------------------------
C PROGRAM ABLAT
C
REAL G(15),H(15),T(15),POS(15),FT(15),HS(15),TS(15)
REAL DELX,S,TAU,FTAU,TA(3)
INTEGER ITA(3)
LOGICAL CTAU,CERR,IERR
CHARACTER INFIL*12,OUTFIL*12,OTIT*30,ORD(3)*6,INP*8,ANS*1

DATA ORD/' First','Second',' Third'/

WRITE (*,*) ' WELCOME TO SCHILLING'S GLACIER SURFACE PROGRAM'
WRITE (*,*) (' INSTRUCTIONS? Y/N <Ret> =N >> "\")')
READ (*,*(A1)) ANS
IF (ANS.EQ.'Y'.OR. ANS.EQ.'y') THEN
WRITE (*,3)
3 FORMAT (""" THE PROGRAM RECONSTRUCTS THE GLACIER SURFACE"",/  
*   "",/  
* ' INPUT FILE:',/  
* ' Max 99 ground elevations in feet col 1-10 (Real #s)="/  
* ' Entered one per record with last record as 999999="/  
* "/,  
* ' PROMPTED FOR="/  
* ' Output title - max 30 chars="/  
* ' Horizontal step distance from ice snout="/  
* ' Initial ice thickness in metres="/  
* ' Shaping factor="/  
* ' Initial basal shear stress="/  
* ' Additional 3 stress factors after defined steps="/  
* "/,  
* ' OUTPUT FILE:'/,  
* ' Prints/plots on printer ground and reconstructed="/  
* ' ice surface.""")
ENDIF
C
OPEN INPUT/OUTPUT FILES
WRITE (*,('/" Enter data filename > > ")')
CALL GETFIL (1,INFIL)
WRITE (*,('/" Enter output filename > > ")')
CALL SAVFIL (3,OUTFIL)

WRITE (*, ('" Enter output title; <Ret>=" * * >> ")')
READ (*,'(A20)') OTIT
4 WRITE (*,5)
5 FORMAT (""" Enter Horizontal step distance (metres): <Ret> =5000. > >  
*"")
READ (*,'(A8)') INP
IF (INP.EQ. ') THEN
   DELX = 5000.
   WRITE (*,'(1X,F8.3)') DELX
ELSE
   CALL CTOR (INP,DELX,CERR)
   IF (CERR) THEN
      WRITE (*,'(" Please enter a number")')
      GOTO 4
   ELSE

   END
WRITE (*,'(1X,F8.3)') DELX
ENDIF
ENDIF

C

6 WRITE (*,7)
7 FORMAT (' Enter Initial ice thickness (metres): < Ret > =200. > > '')
READ (*,'(A8)') INP
IF (INP.EQ.' ') THEN
T(1) = 200.
WRITE (*,'(1X,F8.3)') T(1)
ELSE
CALL CTOR (INP,T(1),CERR)
IF (CERR) THEN
WRITE (*,'('' Please enter a number'')')
GOTO 6
ELSE
WRITE (*,'(F8.3)') T(1)
ENDIF
ENDIF

C

8 WRITE (*,'(/' Enter "Shape" Factor: < Ret > =1.000 > '))'
READ (*,'(A8)') INP
IF (INP.EQ.' ') THEN
S = 1.000
WRITE (*,'(F8.3)') S
ELSE
CALL CTOR (INP,S,CERR)
IF (CERR) THEN
WRITE (*,'('' Please enter a number'')')
GOTO 8
ELSE
WRITE (*,'(F8.3)') S
ENDIF
ENDIF

C

9 WRITE (*,'(/' Enter "Basal Shear Factor": < Ret > =0.8 > '))'
READ (*,'(A8)') INP
IF (INP.EQ.' ') THEN
TAU = 0.800
WRITE (*,'(F8.3)') TAU
ELSE
CALL CTOR (INP,TAU,CERR)
IF (CERR) THEN
    WRITE (*,'(" Please enter a number")')
    GOTO 9
ELSE
    WRITE (*, '(F8.3)') TAU
ENDIF
ENDIF
FTAU = TAU

CTAU = .FALSE.
WRITE (*,10)
10 FORMAT (' Change shear factor at 3 user defined steps? Y/N <Ret> =N
* => >;'>)
READ (*,'(A1)') ANS
IF (ANS.EQ.'Y'.OR.ANS.EQ.'y') THEN
    CTAU = .TRUE.
DO 20 I =1,3
13 WRITE (*,'(/" Enter step > > ")')
READ (*,'(A8)') INP
CALL CTOI (INP,ITA(I),IERR)
IF (IERR) THEN
    WRITE (*,'(" Please enter an integer number")/')
    GOTO 13
ELSE
    WRITE (*, '(F8.3)') TAU(I)
ENDIF
20 CONTINUE
WRITE (*,'("")')
ENDIF

C
RHO = .91
GR = 981.
I = 1
SUM = 0.
TU = TAU*1.E06
C = TU/(RHO*GR*100.*S)
WRITE (3, '(1X,A30/)') OTIT
WRITE (3, 44)
44 FORMAT (11X,'VERTICAL MEASUREMENTS IN FEET',/)  
WRITE (3,22)  
22 FORMAT (/,IX,'DISTANCE (KM.)',3X,'GROUND LEVEL',3X,'ICE SURFACE'  
*,3X,'THICKNESS',/)  
C
READ (1,11) FT(1)
C  NOTE CONVERSION TO METRES
G(1) = FT(1)*.3048
H(1) = G(1) + T(1)
TS(1) = T(1)*3.2808
HS(1) = H(1)*3.2808
POS(1) = DELX/1000.
SUM = SUM + T(1)
WRITE (3,33) POS(1),FT(1),HS(1),TS(1)
I = 2

C
30 READ (1,11) FT(I)
11 FORMAT (F10.2)
C  NOTE CONVERSION TO METRES
IF (FT(I).GT.900000.) GOTO 99
C  CHANGE BASAL STRESS FACTOR BY STEPS (CTAU .TRUE.)
IF (CTAU) THEN
  IF (I.GT.ITA(1)) TAU = TA(1)
  IF (I.GT.ITA(2)) TAU = TA(2)
  IF (I.GT.ITA(3)) TAU = TA(3)
  TU = TAU*1.E06
ENDIF
C
C = TU/(RHO*GR*100.*S)
G(I) = FT(I)*.3048
H(I) = H(I-1) +(C/T(I-1))*DELX
T(I) = H(I) -G(I)
C
THE FOLLOWING CARDS STOP PROGRAM AT NUNATAKS
IF (T(I).LE.25.) GOTO 89
SUM = SUM + T(I)
HS(I) = H(I) * 3.2808
TS(I) = T(I) * 3.2808
POS(I) = (I) * DELX/1000.
WRITE (3,33) POS(I), FT(I), HS(I), TS(I)
I = I+1
GOTO 30

C
89 WRITE (3,41)
41 FORMAT (/1X,'NUNATAK OR MAXIMUM HEIGHT')
99 XI = I-1
AVE = SUM/XI
AVEFT = AVE*3.2808
WRITE (3,47) DELX, T(I), S, FTAU
47 FORMAT (/'' HORIZONTAL STEP DISTANCE = ',F6.0, /1X,'INITIAL ICE THICKNESS = ',F5.0, /1X,'SHAPE FACTOR = ',F5.3, /1X,'INITIAL BASAL SHEAR STRESS = ',F5.3)

C
DO 50 I = 1,3
WRITE (3, 25) ITA(I), TA(I)
25 FORMAT (IX,'SHEAR FACTOR AFTER STEP ',I3,' = ',F5.3)
50 CONTINUE

C
WRITE (3,55) AVE, AVEFT
55 FORMAT (///,I1X,'AVERAGE ICE THICKNESS IN METRES: ',F7.2, /1X,'AVERAGE ICE THICKNESS IN FEET: ',F7.2, ///)

C
CALL PLOT(FT,HS)

C
WRITE (*,'(/" OUTPUT IN FILE: ",A12/)') OUTFIL
CLOSE (1,STATUS = 'KEEP')
CLOSE (3,STATUS = 'KEEP')
STOP
END

C
SUBROUTINE PLOT(G,H)

C
WRITTEN BY RALPH STEA 1982 FOR FORTRAN IV ON CYBER 17C

C
MODIFIED BY DOUG RIGBY DEC 87 FOR MICROSOFT FORTRAN V3.0

C
DIMENSION STATEMENTS CHANGED TO REAL
C MULTIPLE INITIALIZATION OF IG,IH OMITTED
C MATRIX C CHANGED FROM INTEGER HOLLORITH TO CHARACTER
C DOUBLE CHARACTER QUOTES CHANGED TO SINGLE QUOTES
C SPACING CHANGED TO 8 BY 11 PAPER FOR 70 RECORDS
C
C CUMULATIVE HORIZONTAL DISTANCE (POS) REPLACED WITH J
C COUNTER
C TO PLOT STEPS ONE PER COLUMN.
C
REAL G(15),H(15)
CHARACTER C(35,15)
INTEGER IG(15),IH(15)
C
DO 10 J=1,15
IG(J) = IFIX(G(J))*0.025
IH(J) = IFIX(H(J))*0.025
10 CONTINUE
D0 987 J = 1,15
DO 987 I = 1,35
987 C(I,J) = ' ' 
DO 11 J=1,15
IF(IH(J).EQ.0)GOTO11
IF(IG(J).EQ.0)IG(J) = 1
KX=IH(J)
LX=IG(J)
KX = 36-KX
LX=36-LX
C(LX,J) = '#'
C(KX,J) = '*' 
11 CONTINUE
WRITE(3,33)((C(I,J),J = 1,15),I = 1,35)
33 FORMAT(1X,' ',2X,'T',15(A1,' '),7X,72('_'))
END
C
SUBROUTINE CTOR (CVAR,RVAR,CERR)
C
C INTEGER TO REAL CONVERSION AND ALPHA ERROR TRAP
C GETS CVAR: CHARACTER*8
C CONVERTS CVAR TO INTEGER AND/OR REAL
C RETURNS REAL AS RVAR
C IF ALPHA CHARACTER, RETURNS "CERR" AS .TRUE.
INTEGER IVAR
REAL RVAR
CHARACTER CVAR*8
LOGICAL CERR

C
CERR = .FALSE.
C IF INTEGER, CONVERT TO REAL; BUT FIRST FLUSH SCREEN BUFFER
WRITE (*,'(" ")')
READ (CVAR,*,ERR=50) IVAR
C IF ERROR GOTO LABEL 50
RVAR = FLOAT(IVAR)
RETURN
50 READ (CVAR,'(F8.3),ERR=60) RVAR
RETURN
60 CERR = .TRUE.
RETURN
END

C
SUBROUTINE CTOI (CVAR,IVAR,IERR)
C
C Written by D.W. Rigby NSDME, Feb 5/88
C Character to integer conversion and alpha error trap
C Gets CVAR: Character*8
C Converts CVAR to integer
C Returns integer as IVAR
C If real or alpha character, returns IERR as .true.
C
C D.W Rigby, NSDME Feb 87
C
INTEGER IVAR
CHARACTER CVAR*8
LOGICAL IERR
C
IERR = .FALSE.
C Flush screen buffer
WRITE (*,'(" ")')
READ (CVAR,*,ERR=50) IVAR
RETURN
C If error go to label 50
50 IERR = .TRUE.
   RETURN
END
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PRECISION™ RESOLUTION TARGETS
LEGEND

GEOLOGICAL UNITS

- VALLEY FILL
- SADDLE FILL
- BASIN FILL (LAHAVE CLAY)
- ERODED BASIN FILL (OUTCROPPING EMERALD SILT)
- ASYMMETRICAL MORAINES
- SYMMETRICAL MORAINES
- HUMMOCKY MORAINES
- TILL-TONGUE MORAINES
- BEDROCK

SCALE (km):
PM-1 3½"x4" PHOTOGRAPHIC MICROCOPY TARGET  
NBS 1010a ANSI/ISO #2 EQUIVALENT

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PRECISION® RESOLUTION TARGETS
### ADDITIONAL SOURCES OF GEOLOGICAL INFORMATION

- **AGI Report on the geology of the Antigonish area.**
- **AGI Report on the geology of the Canso area.**
- **AGI Report on the geology of the Stellarton area.**
- **AGI Report on the geology of the Digby area.**
- **AGI Report on the geology of the Yarmouth area.**
- **AGI Report on the geology of the Cape Breton area.**
- **AGI Report on the geology of the Richmond area.**
- **AGI Report on the geology of the Pictou area.**
- **AGI Report on the geology of the Kings County area.**
- **AGI Report on the geology of the Annapolis area.**
- **AGI Report on the geology of the Prince Edward Island area.**

### LIST OF REFERENCE SECTIONS

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<tr>
<th>Number</th>
<th>Title</th>
<th>Page Numbers</th>
<th>Reference</th>
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</thead>
<tbody>
<tr>
<td>1</td>
<td>Prehistorical Case</td>
<td>87-89</td>
<td>Gifford (1977)</td>
</tr>
<tr>
<td>2</td>
<td>Redhead</td>
<td>90-96</td>
<td>Gifford (1977)</td>
</tr>
<tr>
<td>3</td>
<td>Labrador River</td>
<td>97-102</td>
<td>Gifford (1977)</td>
</tr>
<tr>
<td>4</td>
<td>South Coast</td>
<td>103-108</td>
<td>Gifford (1977)</td>
</tr>
</tbody>
</table>


These stones are parts of the area that used to be a beach, now covered by debris and cut banks. The stones are part of a series of levelled debris that were cut by erosion and deposited by wave action. The stones are part of a series of levelled debris that were cut by erosion and deposited by wave action.

At the mouth of Mahone Bay, a delta exposure reveals a slope at the edge of the beach, which is cut by erosion and deposited by wave action. The stones are part of a series of levelled debris that were cut by erosion and deposited by wave action.

next line

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<table>
<thead>
<tr>
<th>Number/Locality</th>
<th>Stratigraphic Description</th>
<th>Date (yr B.P.)</th>
<th>Principal reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>30 West Side</td>
<td>Basal in top 6 cm organo-poor peat with wood fragments, facial grass; red, silt: fine ash with wood fragments and charcoal; grasses, shrub, tree, charred. Similar to sample 12 dated 1680 BC and 2240 BC: 1680 BC detrital (GSC 258) organic-rich.</td>
<td>1680 BC</td>
<td>Hall, 892, Grant and King, 884</td>
</tr>
<tr>
<td>34 Big Brook</td>
<td>Base in top 1 m organo-poor peat with wood fragments and charcoal; grasses, shrub, tree, charred. Similar to sample 12 dated 1680 BC and 2240 BC: 1680 BC detrital (GSC 258) organic-rich.</td>
<td>1680 BC</td>
<td>Hall, 892, Grant and King, 884</td>
</tr>
<tr>
<td>62 Whycomah</td>
<td>Base in top 1 m organo-poor peat with wood fragments and charcoal; grasses, shrub, tree, charred. Similar to sample 12 dated 1680 BC and 2240 BC: 1680 BC detrital (GSC 258) organic-rich.</td>
<td>1680 BC</td>
<td>Hall, 892, Grant and King, 884</td>
</tr>
<tr>
<td>63 Ribbons</td>
<td>Base in top 1 m organo-poor peat with wood fragments and charcoal; grasses, shrub, tree, charred. Similar to sample 12 dated 1680 BC and 2240 BC: 1680 BC detrital (GSC 258) organic-rich.</td>
<td>1680 BC</td>
<td>Hall, 892, Grant and King, 884</td>
</tr>
<tr>
<td>64 Hines</td>
<td>Wood and peat horizon by context and soil.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>65 Gates Port</td>
<td>Base in top 1 m organo-poor peat with wood fragments and charcoal; grasses, shrub, tree, charred. Similar to sample 12 dated 1680 BC and 2240 BC: 1680 BC detrital (GSC 258) organic-rich.</td>
<td>1680 BC</td>
<td>Hall, 892, Grant and King, 884</td>
</tr>
<tr>
<td>Number/Location</td>
<td>Stratigraphic Description</td>
<td>Dates (yr BP)</td>
<td>Principal Reference</td>
</tr>
<tr>
<td>----------------</td>
<td>--------------------------</td>
<td>--------------</td>
<td>---------------------</td>
</tr>
<tr>
<td>17 Lorraine Conc</td>
<td>Basal breccia</td>
<td>17 K 200 1990</td>
<td>Stea and Stock 1990</td>
</tr>
<tr>
<td>18 How Brook Conc</td>
<td>Basal breccia</td>
<td>17 K 200 1990</td>
<td>Stea and Stock 1990</td>
</tr>
<tr>
<td>28 Shepsquasha</td>
<td>Basal breccia</td>
<td>17 K 200 1990</td>
<td>Stea and Stock 1990</td>
</tr>
<tr>
<td>27 Theobald Beach</td>
<td>Breccia breccia</td>
<td>17 K 200 1990</td>
<td>Stea and Stock 1990</td>
</tr>
<tr>
<td>35 Miller Creek Quarry</td>
<td>Breccia breccia</td>
<td>17 K 200 1990</td>
<td>Stea and Stock 1990</td>
</tr>
</tbody>
</table>

**Note:** The dates are given in years before present (BP), and the principal references are cited for each location.
### Table: Geologic Setting and Ages

<table>
<thead>
<tr>
<th>Number/Narrative</th>
<th>Stratigraphy Description</th>
<th>Name or Ref.</th>
<th>Principle Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>13 Elbow Lake</td>
<td>Base to top 25 m exposed, clay-</td>
<td>13Elbow Lake</td>
<td>Myers and Stea, 1986</td>
</tr>
<tr>
<td></td>
<td>dominated, wedge intrudes on clay.</td>
<td></td>
<td>(Ref. 13)</td>
</tr>
<tr>
<td>10 West Lake</td>
<td>Base to top 25 m, exposed, sandy, impervious sandstone with</td>
<td>10 West Lake</td>
<td>Myers and Stea, 1986</td>
</tr>
<tr>
<td></td>
<td>wedge intrudes on clay.</td>
<td></td>
<td>(Ref. 10)</td>
</tr>
<tr>
<td>14 Twenty 20</td>
<td>Base to top 25 m, exposed, sandy,</td>
<td>14 Twenty 20</td>
<td>Myers and Stea, 1986</td>
</tr>
<tr>
<td></td>
<td>wedge intrudes on clay.</td>
<td></td>
<td>(Ref. 14)</td>
</tr>
<tr>
<td>8 New Glasgow</td>
<td>Base to top 25 m, exposed, clay-</td>
<td>8 New Glasgow</td>
<td>Myers and Stea, 1986</td>
</tr>
<tr>
<td></td>
<td>dominated, wedge intrudes on clay.</td>
<td></td>
<td>(Ref. 8)</td>
</tr>
<tr>
<td>7 Esk Lake</td>
<td>Base to top 25 m, exposed, sandy,</td>
<td>7 Esk Lake</td>
<td>Myers and Stea, 1986</td>
</tr>
<tr>
<td></td>
<td>wedge intrudes on clay.</td>
<td></td>
<td>(Ref. 7)</td>
</tr>
</tbody>
</table>

### Figure: Shale Superior Divides

- **Stringerite/Shale**: The formations are identified based on their characteristic features and stratigraphic positions.

**Principle reference**: Myers and Stea, 1986; Stea et al., 1987; and other cited sources.
Neuport: The Neuport Till is a post-glacial till with a mainly fine gravel content, enriched with debris of till. The till is composed of clasts in the 2-3 cm range, with a few larger clasts in the 3-4 cm range. It is characterized by a brownish-grey coloration, with a fine-grained matrix. In some areas, it shows a lamination structure.

Dective Lake: The Dective Lake Till is a post-glacial till with a mainly fine gravel content, enriched with debris of till. The till is composed of clasts in the 2-3 cm range, with a few larger clasts in the 3-4 cm range. It is characterized by a brownish-grey coloration, with a fine-grained matrix. In some areas, it shows a lamination structure.

Glenora: The Glenora Till is a post-glacial till with a mainly fine gravel content, enriched with debris of till. The till is composed of clasts in the 2-3 cm range, with a few larger clasts in the 3-4 cm range. It is characterized by a brownish-grey coloration, with a fine-grained matrix. In some areas, it shows a lamination structure.

Port Elgin: The Port Elgin Till is a post-glacial till with a mainly fine gravel content, enriched with debris of till. The till is composed of clasts in the 2-3 cm range, with a few larger clasts in the 3-4 cm range. It is characterized by a brownish-grey coloration, with a fine-grained matrix. In some areas, it shows a lamination structure.

Tanami: The Tanami Till is a post-glacial till with a mainly fine gravel content, enriched with debris of till. The till is composed of clasts in the 2-3 cm range, with a few larger clasts in the 3-4 cm range. It is characterized by a brownish-grey coloration, with a fine-grained matrix. In some areas, it shows a lamination structure.
SURFICIAL GEOLOGY OF THE PROVINCE OF NOVA SCOTIA

MAP 92-3

Compiled by R R Stea, H Conley and Y Brown

SCALE
1:500,000 or 7.99 miles to 1 inch

NOVA SCOTIA DEPARTMENT OF NATURAL RESOURCES
MINES AND ENERGY BRANCHES

Honourable C. W. McNeil M.D
Minister
Honourable John McCann
Deputy Minister

HALIFAX, NOVA SCOTIA
1992

The funds for preparation and publication of this map were provided through the Canada — Nova Scotia Mineral Development Agreement: 1984-1989

Map reviews: R. C. Butter and D. B. MacDowell, Nova Scotia Department of Natural Resources; R. N. W. Dilworth and R. J. Fulton, Geological Survey of Canada


The Surficial Geology of the Province of Nova Scotia is available in digital form through Land Registration and Information Service, Amherst, NS

For information on errors or omissions in the Surficial Geology, contact the Nova Scotia Department of Natural Resources.
Figure 3 shows the Late Wisconsin. The map indicates the movement of the glacier, which advanced and retreated multiple times. The ice flow phases are marked on the map. The legend explains the symbols used for different features, such as ice advance and recession, lake and bog locations, and raised beaches.

ICE FLOW PHASE 3
- Peat buried by diamicton
- Flow paths
- Ice center or divide early
- Ice center or divide late
- Ice lens or divide
- Lake or bog not recording oscillation
- Lake or bog recording oscillation
- Raised beach ca. 14,000 yr B.P.

ICE FLOW PHASE 4
- Peat buried by diamicton
- Peat buried by waterlain sediment
- Raised beach ca. 14,000 yr B.P.
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For notification of errors or omissions to the Surficial Geology contact the Nova Scotia Department of Natural Resources.
LATE GLACIAL SITES
(14,000-10,000 yr B.P.)

- Peat buried by diamicton
- Peat buried by waterlain sediment
- Lakes, bogs not recording oscillation
- Lakes, bogs recording oscillation
- Ice marginal delta ca 14,000-12,600 yr B.P.
- Raised beach ca. 14,000 yr B.P.

**SYMBOLS**

- Dune
- Aerial
- Raised marine terrace
- Ice marginal delta
- Raised beach
- Lakes, bogs not recording oscillation
- Lakes, bogs recording oscillation
- Peat buried by diamicton
- Peat buried by waterlain sediment
- Ice marginal delta
- Raised beach
- fence

Location of stratigraphic sections of special interest is indicated on information sheet at the back of this report.
EXPLANATION

This map depicts the distribution and nature of Quaternary glacial and other surface deposits in Nova Scotia. Outlined on the map is the glacial history of Nova Scotia. Surface deposits form the present landscape for most sites in Nova Scotia. The data on this map are used in environmental studies, resource exploration, agricultural surveys, and more.

Figure 1 is a summary of the stratigraphic and temporal relationships of Nova Scotia Quaternary deposits. Quaternary lithostatigraphic units in Nova Scotia are arranged by region and area from the lowest or earliest units to the highest or latest units. The stratigraphic sequence (most recent) is also shown on this figure (see also Fig. 2). Ice flow phases are succeeded by till sheets by comparing regional ice flow directions with till provenance and fabric. Normal stratigraphic units are shown on this figure and described in more detail on the back of the map. Relevant sections used to construct this diagram are marked on the bottom of the diagram. Refer to the map on the reverse side for section locations and descriptions.

Columns on Figure 1 refer to lithostatigraphic units described in the legend (e.g. gravel units are 1, gravel—glacial drift, etc.). Several sections on the diagram imply ice-free conditions prevailing, indicated by soil formation. Arrows indicate the stratigraphic age assignments in the compilation. Data from borehole units are shown on the diagram in their stratigraphic and temporal context. Rectangular units containing yellow till have arrow pointing to their preglacial source area. Roman numerals correspond to pollen stratigraphic units (see list on left or back of sheet for descriptions).

Intensive erosion and deposition (Bridgewater Marine Complex) may precede the earliest glaciation. A pre-Emmon stratigraphic unit (18,000-30,000 yr B.P.) is indicated by dates on till in the southwest area of Nova Scotia. The (2) stratigraphic unit is represented by marine and terrestrial deposits underlying till of the last glacial period. The Westernmost Till is in the northwest area of Nova Scotia and has a warm-water fauna indicating present-day conditions. During this time, sea level was up to 5 m higher than at present. This higher sea level is a consequence where marine ice was not encroaching beneath glacial deposits. At the time, forested communities changed from hardwood-dominated (Unit 1) during the warm part of the interglacial to spruce forest (Unit 2) in the cooler phase.

During certain ice ages, glaciers developed as the Wisconsinan. In northern Cape Breton Island, ice-free conditions may have persisted until 50,000 yr B.P., with a period of warm interglacial conditions. Continued ice cover through most of the Wisconsinan stage is indicated by the names of the units on the stratigraphic chart (see also Fig. 2). Arrows indicate the direction of ice flows for each unit.

The earliest Wisconsinan ice flow was in Nova Scotia were eastward (Phase 1a, Fig. 2) from southwest (Phase 1b, Fig. 2) to northeast (Phase 1c, Fig. 2). Several late-stage glacial deposits were erosional features formed by ice flow. In fact, the eastward ice flow may represent a separate, earlier phase of glaciation. The Harriets, East Maitland, Richmond and Riverhead Tills (Fig. 1) are fluvial in these areas. The East Maitland Tills contain blocky of glacial till from the Colchester Highlands and boulders from the North Mountain which were transported southeastward and are not mingled in the Atlantic Coast, up to 20 km south of the coastline. This phase may represent glaciers with an Appalachian or Laurentide source.

The second major ice flow trend (Phase 2a, Fig. 2) was southeastward and southwestward from the Escuminac Ice Center in the Prince Edward Island region. This event is succeeded by southwestward-trending ice covering all of the land south of the region. Nova Scotia and New Brunswick. The Lakehead-Scituate and Eastville Tills (Fig. 1) were formed during this phase. Material from the crustal fault in southern mainland Nova Scotia and Colchester Basin, which was transported southeastward and can be traced to the Atlantic Coast, up to 20 km south of the coastline. This phase may represent glaciers with an Appalachian or Laurentide source.

During the next ice flow event (Phase 3a, Fig. 2) glaciers from the Atlantic Coastal-Pennsylvanian Province were transported northward onto the North Mountain. The Harris and Souris River Tills were formed during this ice flow phase. Emphasis from the Colchester Highlands can be found throughout the New Brunswick—Prince Edward Island region. Northeastward-trending streams can be traced across the northern margin of Nova Scotia (Fig. 2). This well-documented northwest ice flow was clearly in response to the development of an ice divide at eastern Nova Scotia (Fig. 3). The flow was northwestward and southwestward from this divide across the east of the Nova Scotia mainland. This divide may have formed as a result of marine incursion into the Bay of Fundy or climatic event.

This map is not to scale and may not be a true representation of the glacial history of Nova Scotia. The map is a summary of the stratigraphic and temporal relationships of Nova Scotia Quaternary deposits. Quaternary lithostatigraphic units in Nova Scotia are arranged by region and area from the lowest or earliest units to the highest or latest units. The stratigraphic sequence (most recent) is also shown on this figure (see also Fig. 2). Ice flow phases are succeeded by till sheets by comparing regional ice flow directions with till provenance and fabric. Normal stratigraphic units are shown on this figure and described in more detail on the back of the map. Relevant sections used to construct this diagram are marked on the bottom of the diagram. Refer to the map on the reverse side for section locations and descriptions.
### Glaciomarine Deposits

#### Description
- Formed by marine ice streams
- Deposited in a ponded body of water

#### Environmental Significance
- Source of aggregate materials
- Excellent aquifer in some areas
- Acidic and impeded cultivation
- Groundwater source

#### Topography
- In coastal lowlands
- Drainage and slope inclines

#### Legend
- Coastline
- Wild area
- Glacier
- Aquifer
- Agricultural area
- Drainage
- Meltwater

#### Thickness
- Generally 45 m

### Glaciallake Deposits

#### Description
- Deposited in a ponded body of water
- Formed by coastline retreat

#### Environmental Significance
- Deposited in a ponded body of water
- Source of aggregate materials
- Excellent aquifer in some areas
- Acidic and impeded cultivation
- Groundwater source

#### Topography
- In coastal lowlands
- Drainage and slope inclines

#### Legend
- Coastline
- Wild area
- Glacier
- Aquifer
- Agricultural area
- Drainage
- Meltwater

#### Thickness
- Generally 45 m

### Glacial Till

#### Description
- Silty till, drumlins
- Granite, greywacke, slate
- Dewed facies

#### Environmental Significance
- A complex mixture of glacial material
- Deposited in a ponded body of water
- Source of aggregate materials
- Excellent aquifer in some areas
- Acidic and impeded cultivation
- Groundwater source

#### Topography
- In coastal lowlands
- Drainage and slope inclines

#### Legend
- Coastline
- Wild area
- Glacier
- Aquifer
- Agricultural area
- Drainage
- Meltwater

#### Thickness
- Generally 45 m
During the last ice flow phase (Phase 3), boulders were transported northward onto the North Mountain. The Hants and Basin Rivers Terrace were formed during this phase. As glaciers withdrew and upland areas of New Brunswick and New Brunswick, the ice sheets continued winning the Panhandle and advancing northwards. The Coast Mountains and Cape Breton formed in the Prince Edward Island region. This phase resulted from the development of an ice divide in southern Nova Scotia (Fig. 2). The southerly ice flow was deflected by topography to the north, and as a result, the only glacier to form was a small glacier in the Fundy coastal plain. This glacier reached the coastline near Cape Blomidon, Nova Scotia, and deposited till and moraines. During this phase, residents of the Bay of Fundy had to abandon their homes and find new homes in areas farther inland.

The final phase of ice flow occurred during the deglaciation. Glacial advances and retreats caused sea levels to fluctuate. During Phase 4, the coast was eroded, and sea levels rose, causing the coastline to retreat and leave behind a series of sea cliffs. As the climate warmed, glaciers began to recede, leaving behind a series of moraines. As the climate continued to warm, the coastline retreated further inland, leaving behind a series of sea cliffs. These cliffs were later eroded by the action of waves and the sea, leaving behind a series of sea cliffs.

As the climate continued to warm, sea levels rose, and the coastline retreated further inland. This process continued until the climate stabilized, and the coastline reached its current position. During this phase, the coastline was eroded by the action of waves and the sea, leaving behind a series of sea cliffs. These cliffs were later eroded by the action of waves and the sea, leaving behind a series of sea cliffs.

By 10,000 years ago, the climate had stabilized, and the coastline had retreated. The sea level had stabilized, and the coastline had reached its current position. The coastline was eroded by the action of waves and the sea, leaving behind a series of sea cliffs. These cliffs were later eroded by the action of waves and the sea, leaving behind a series of sea cliffs.

The coastline was eroded by the action of waves and the sea, leaving behind a series of sea cliffs. These cliffs were later eroded by the action of waves and the sea, leaving behind a series of sea cliffs.
Figure 5: Evolution of the terraced shoreline deltas on the north shore of White Basin. For each stage (Stage I, 15,000 to 11,500 years B.P.; Stage II, 11,500 to 10,000 years B.P.; Stage III, 10,000 to 8500 years B.P.; Stage IV, 8500 to 7500 years B.P.; Stage V, 7500 to 5000 years B.P.); the contact between the foreland and the delta plain represents the former sea level. For each stage, the delta front is indicated by a dashed line. The delta plain is shown as a continuous line.

Figure 6: Diagrammatic representation of the evolution of the delta plain. The contact between the foreland and the delta plain represents the former sea level. The delta plain is shown as a continuous line. The delta front is indicated by a dashed line. For each stage, the delta plain is shown as a continuous line. The contact between the foreland and the delta plain represents the former sea level. For each stage, the delta plain is shown as a continuous line. The delta front is indicated by a dashed line.

Figure 7: Evolution of the terraced shoreline deltas on the north shore of White Basin. For each stage (Stage I, 15,000 to 11,500 years B.P.; Stage II, 11,500 to 10,000 years B.P.; Stage III, 10,000 to 8500 years B.P.; Stage IV, 8500 to 7500 years B.P.; Stage V, 7500 to 5000 years B.P.); the contact between the foreland and the delta plain represents the former sea level. For each stage, the delta front is indicated by a dashed line. The delta plain is shown as a continuous line. The contact between the foreland and the delta plain represents the former sea level. The delta plain is shown as a continuous line. The delta front is indicated by a dashed line. For each stage, the delta plain is shown as a continuous line. The contact between the foreland and the delta plain represents the former sea level.
**PRE LAST GLACIATION**

**LAST GLACIATION**
(WISCONSINIAN)

**POS**
<table>
<thead>
<tr>
<th><strong>Elastic State</strong></th>
<th><strong>Deposits</strong></th>
<th><strong>Characteristics</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Marine Deposits</strong></td>
<td>Gravel, sand, clay locally varved by post-glacial marine submergence</td>
<td>Formerly coastal; submergence through wave action, longshore currents, and tides</td>
</tr>
<tr>
<td><strong>Clastic State</strong></td>
<td><strong>Glacioklastite Deposits</strong></td>
<td>Gravel, sand, and breccia with sand, silt, and clay; pebbles of various sizes</td>
</tr>
<tr>
<td></td>
<td><strong>Glaciomarine Deposits</strong></td>
<td>Gravel and sand, brecciated deposits, sand, and clay, locally varved by ice water</td>
</tr>
<tr>
<td></td>
<td><strong>Outwash Fans, Deltas, and Valley Train Deposits</strong></td>
<td>Gravel and sand, brecciated deposits, sand, silt, and clay, locally varved by ice water</td>
</tr>
<tr>
<td><strong>Ground Moraine and Streamlined Drift</strong></td>
<td><strong>Hummocky Ground Moraine Deposits</strong></td>
<td>Till comprising gravel, sand, and mud; organic gravel, silt, and clay; sandy and loamy substrates</td>
</tr>
<tr>
<td><strong>Story Till Plan and Drawings</strong></td>
<td><strong>Gravel and Sand Deposits</strong></td>
<td>Gravel and sand, locally varved by ice water</td>
</tr>
<tr>
<td><strong>Salty Till Plan and Drawings</strong></td>
<td><strong>Glacial Lakes, Spillways, and Ice Margins Deposits</strong></td>
<td>Gravel and sand, locally varved by ice water</td>
</tr>
<tr>
<td><strong>Peck</strong></td>
<td><strong>Feodown</strong></td>
<td>Fragments, root communality of angiosperm flowers and trees, formed under ice</td>
</tr>
<tr>
<td><strong>Bedrock</strong></td>
<td><strong>Glacial State</strong></td>
<td>Material derived from bedrock; locally varved by ice water</td>
</tr>
</tbody>
</table>