# Wisconsinan Glaciation of the Continental Shelf - Southeast Atlantic Canada

by

Lewis H. King and Gordon B. Fader

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# WISCONSINAN GLACIATION ON THE CONTINENTAL SHELF-

# SOUTHEAST ATLANTIC CANADA

Lewis H. King[1] and Gordon B. Fader

Geological Survey of Canada, Bedford Institute of Oceanography
P.O. Box 1006, Dartmouth, Nova Scotia, B2Y 4A2

## ABSTRACT

Through the interpretation of high resolution seismic reflection profiles of the Scotian Shelf, eastern Gulf of Maine, and the Grand Banks of Newfoundland, and the employment of a model for a marine ice shelf (Carey and Ahmad, 1961), we have developed a conceptual model for the deposition of glacial and glaciomarine deposits from these areas.

The Scotian Shelf Drift was derived from subglacial melt-out debris from a neutral to negatively buoyant active ice shelf in direct contact with the seabed. Emerald Silt formed from subglacial melt-out debris from a pinned but floating ice shelf. The debris is thought to have settled through a water column of variable thickness to form the conformable, rhythmically banded deposits which mimic a highly irregular substrate which is recognized over very broad areas. Horizontal migration of the ice-seabed contact (buoyancy line), induced by changes in ice thickness and changes in relative sea level, leads to the development of thick regional moraines interbedded with the glaciomarine deposits. Wedge-shaped till deposits (till tongues) are often formed at the distal side of the moraines through advance and subsequent retreat of the buoyancy line. The configuration of the seabed beneath an ice shelf is another important factor in the stratigraphic develop-

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[1]Present address: Marine Geological Consulting, 50 Swanton Drive, Dartmouth, Nova Scotia, B2W 2C5

ment of these marine deposits.

Type areas representative of the offshore Wisconsinan section were chosen on the basis of (1) what appeared to be the most complete seismostratigraphic sections representing both the Emerald Silt (glaciomarine) and Scotian Shelf Drift (glacial till) Formations, (2) their accessibility to sampling by piston corer, (3) the occurrence of unique structural and stratigraphic characteristics which were useful in the development of our conceptual models concerning glacial deposition in marine areas, and (4) their correlative relationships with the glacial geology of the entire shelf so that groundtruth information, for example radiocarbon dates, could be extrapolated within and beyond the type areas using continuous seismic reflection control.

We have used the Carey and Ahmad model as an aid in interpreting the seismostratigraphy of the surficial sediments on the continental shelf and integrated these studies with sample data to postulate the glacial history. In late Early Wisconsinan time the entire shelf was occupied by an ice sheet (the Scotian Shelf-Grand Banks advance). This wasted to an ice shelf which became buoyant in the deeper basins at about 46000 yBP. At this time a till blanket with lift-off moraines on its surface and associated glaciomarine sediment was deposited. During the period 46000 to 32000 yBP the buoyancy line oscillated intermittently, resulting in the development of till tongues at the periphery of the banks and outer edge of the inner shelf which were intercalated with glaciomarine sediments in the basins. At the western end of the Scotian Shelf the ice may have receded along the Bay of Fundy reentrant to deposit the Salmon River beds (38000 yBP) north of Yarmouth.

During the latter part of the Middle Wisconsinan and Late Wisconsinan (32000 to 16000 yBP) the buoyancy line receded to the coastal areas except in the eastern Gulf of Maine where the ice shelf again grounded to deposit till in the areas peripheral to Georges Basin, the southern flank of Browns Bank, and the Fundian Moraine of Sewell Ridge. The youngest moraine ( 17000 yBP) occurs on Truxton Swell. Subsequently the buoyancy line receded to the coastal areas of Maine and New Brunswick, and the resulting glaciomarine deposits emerged through glacioisostatic rebound. Evidence for the Late Wisconsinan sea level, which we date within the range of 15100 to 14465 yBP, and for the Late Wisconsinan-Holocene transgression is well expressed along the entire shelf.

# INTRODUCTION

In eastern Canada, during the Wisconsinan glaciation, the Laurentide ice sheet, and the local ice caps of the Appalachian Region advanced and retreated several times. According to Dreimanis (1975) and others, the general pattern of major glacial fluctuations in eastern Canada have led to a subdivision of the Wisconsinan glacial stage into three substages in the Great Lakes and Interior Plains regions. These are: Early Wisconsinan (approximately 120000 to 65000 yBP), Mid-Wisconsinan (65000 to 23000 yBP), and Late Wisconsinan (23000 to 10000 yBP). The chronology of events to 50000 yBP was based on radiocarbon dating, but timing of the earlier substage boundaries was established by correlation of the observed stadials and interstadials with the major fluctuations in the oceanographic record. The most detailed classification including several stadials and interstadials is best developed in the St. Lawrence-Lake Ontario-Lake Erie-Lake Huron region. The ice marginal positions can be correlated with levels of pro-glacial lakes which were controlled by the opening and closing of the lake outlets by glacial retreats and advances. In the northeast portion of the region a correlation with sea level changes is also possible. Dreimanis (1975) further suggested that east of the Great Lakes, the Mid-Wisconsinan retreats were either less pronounced or even absent from the stratigraphic records, and the main interstadial (St. Pierre) is assigned to the Early Wisconsinan.

According to Prest (1977) both interglacial and interstadial deposits are known from Cape Breton Island and mainland Nova Scotia, but neither has been recognized from Newfoundland nor Prince Edward Island. The stratigraphy

of eastern Canada pertains only to the last major glaciation with the exception of organic deposits that predate the last glaciation reported from the Magdalen Islands and from the Eastern Townships. In Nova Scotia, organic deposits buried beneath one or more tills are widespread, and are generally beyond the range of radiocarbon dating. Pollen analyses from the organic deposits provide evidence of a warm interglacial environment and are tentatively referred to the Sangamon. The organic deposits indicate a cool climate and are tentatively referred to the St. Pierre Interstade, although they may in part represent very late Sangamon or earliest Wisconsinan time. Grant (1981) suggested that the Sangamon is represented by a widespread elevated wave cut platform with associated littoral and fluvial gravels that sometimes grade to organic beds. He further suggests (pers. comm.) that this interglacial period spans a long interval between 125000 and 75000 yBP which accordingly compresses the Early Wisconsinan substage. The best known Mid- Wisconsinan deposit occurs along the Salmon River in southwestern Nova Scotia (Grant, 1976; Nielsen, 1974) and is of glaciomarine origin and dated at 38000 yBP, although this is somewhat in dispute, Grant 1980.

Glacial till is by far the most widespread of all the surficial deposits in Nova Scotia. This ubiquitous layer can be subdivided on the basis of colour, direction and mode of transport, and its relationship to underlying local source rocks. Grant (1963, 1972), on the basis of lithology recognized three distinctive till sheets on the Altantic slope of mainland Nova Scotia; the composition, thickness and texture of which vary according to differences in the erodability of the local source rock. Rock fragments within the till are generally of local origin, but a distinct proportion (up to 20%), are

derived from northern Nova Scotia and New Brunswick. Till colour was influenced strongly by a source of bright red sediment from the Minas Basin and Bay of Fundy which at times contributed to the formation of local hybrid tills, but sometimes was transported over distances in excess of 100km, maintaining its distinctive red colour. Nielsen (1976) emphasized the importance of the source and mode of transport to explain variations in lithology and colour and recognized distinct basal, englacial, and ablation phases. More recently, Stea and Fowler (1979, 1981), Stea (1982), and Stea and Grant (1982) used the above mentioned parameters together with geochemical data to produce a series of Quaternary lithostratigraphic maps covering much of mainland Nova Scotia.

The regional southeasterly trend of ice-flow features (glacial striae, drumlins, eskers, and end moraines) across Maine, New Brunswick, southern Nova Scotia, and the Scotian Shelf is well documented (Prest and Grant, 1969, and others) and is believed by Grant (1980) to be of Early Wisconsinan age. Subsequently, a redirection of glacial flow from southerly to westerly and northerly toward the Bay of Fundy, is considered to be the result of drawdown as a consequence of marine invasion in the Fundy basin. The southeasterly flow pattern contributed to a generally held concept of a regional radial flow of Laurentide ice across the Maritime Provinces. The absence of the southeasterly trends that might be expected from a major flow of Laurentide ice across the entire area of northern Nova Scotia, Cape Breton Island, Prince Edward Island and eastern New Brunswick was attributed by Goldthwait (1924) to erasure by later ice influenced by drawdown to the Gulf of St. Lawrence. This ice produced easterly flow patterns in Prince Edward Island

and for the most part northeasterly patterns in Cape Breton. Prest and Grant (1969) concluded that at the last glacial maximum the Maritime ice and the Laurentide ice were confluent, but the later was not as influential in the region as was formerly assumed. The Laurentian Channel effectively served to draw down the Laurentide ice and allow its greatest influence, especially erosion in Newfoundland. In the Maritime Provinces the Laurentide ice was not as pronounced as it was toward the continental interior, but was locally influenced by an Appalachian ice complex. During the Late Wisconsinan recessional phase there were several active centres throughout the Maritimes. In

the offshore area King and co-workers (King, 1970; MacLean and King, 1971; Drapeau and King, 1972; MacLean, Fader and King, 1977; Fader, King and MacLean, 1977; Fader, 1984 and Fader, King and Josenhans, in press) mapped the lithostratigraphy of the Scotian Shelf, eastern Gulf of Maine and Bay of Fundy, and the western Grand Banks using acoustical data to define the boundaries of major formations and then qualified them in terms of their gross sedimentary characteristics using bottom samples. They mapped the distribution of glacial till, the occurrence of a submarine moraine complex and associated glaciomarine deposits (King, 1969; King et al., 1972), and later formations associated with the decline of glaciation and an associated low sea level stand. They suggested that the end moraine complex was formed at the buoyancy line of an ice shelf. Evidence was not available to determine if it was formed during retreat of the ice from the edge of the continental shelf or if it represented a terminal position for the last advance on the shelf.

A study by Alam and Piper (1977) and Alam et al. (1984) of cores from the tops of seamounts close to the continental shelf provides evidence of the

earliest ice advances in the Atlantic region. The cores contained sequences of alternating clays and foram-nanno ooze representing glacial and interglacial events to the Pliocene. It was inferred that the Illinoian glaciation was most intense, and was associated with a strong influx of red sediment which indicates extensive erosion of the Gulf of St. Lawrence and Laurentian Channel.

The mapping methods for the earlier offshore work included the use of echograms which provided stratigraphic information on muddy bottoms and the lithologic character was interpreted from the echograms using an estimate of the degree of signal pulse stretching. Across the harder bottoms the airgun profiles provided the stratigraphic data but the resolution was not sufficiently good to resolve the details of the Pleistocene-Holocene section. Since 1975 a Huntec deep-tow high-resolution seismic system (DTS) (Hutchins et al., 1976) was employed which provides resolution to about 0.25m and in most areas provides stratigraphic information throughout the entire surficial section.

A number of the more interesting and critical areas of the shelf were reexamined on an opportunity basis using the DTS system and piston cores were collected in the glaciomarine formation using the high resolution seismic control for the selection of key sites. Radiocarbon dates at several horizons in each core and sediment analyses were correlated with the seismic stratigraphy. This approach has provided new insight into the nature of the glacial environment on the shelf, the relationship between the glaciomarine and glacial sediments, the depositional processes which led to the formation of these units, and the extent and age of glaciation on the shelf and its

relationship to the glacial history of the mainland. The glaciomarine section appears to be complete and represents a time interval from within the Early Wisconsinan to the last influence of floating ice on the shelf during Late Wisconsinan time.

Field work in support of this study involved several cruises: Hudson 75-009, 76-016, 77-011, 78-012, 79-011, and 80-010. The Huntec DTS track plots used in this interpretation are shown in Figures 1a and 1b.

## OFFSHORE SURFICIAL SUCCESSION

### REGIONAL SETTING

This discussion draws on a large number of studies on the bedrock and surficial geology of the Scotian Shelf and the Grand Banks of Newfoundland which were summarized in King (1980).

A diagrammatic section of the surficial succession on the Scotian Shelf from an area south of Halifax across Emerald Basin and Emerald Bank is shown in Figure 2 and a surficial map showing the areal distribution of the formations is shown in Figure 3. Bedrock underlying the surficial lithostratigraphic units consists of Cambro-Ordovician metasediments (Meguma Group) which continue offshore from mainland Nova Scotia and form the basement for a thick prism of Mesozoic-Cenozoic sediments filling the East Coast Geosyncline. The latter beds dip gently seaward and show a marked erosional unconformity with the surficial succession. The surface of the bedrock forms a submerged coastal plain which has subsequently been modified by glacial erosion. Its shape controls to a high degree the bathymetry of the present shelf. The surficial succession occurs as a thin blanket (approximately 50m thick) across the submerged coastal plain surface, and is comprised of five formations: Scotian Shelf Drift (glacial till), Emerald Silt (glaciomarine), Sambro Sand, LaHave Clay, and Sable Island Sand and Gravel.

The major factors dominating the Late Pleistocene and Holocene geological evolution of the Scotian Shelf and possibly that of the whole of the eastern Canadian shelf were: (1) the morphology of the bedrock surface which to a large degree was inherited from a former coastal plain environment, (2)

the advance and retreat of the continental ice sheet and its boundary relationships with the ocean, and (3) the low stand of the Late Wisconsinan sea level at about 120m and its subsequent transgression.

The Scotian Shelf is a typical example of a formerly glaciated shelf characterized by a well defined, rough inner shelf which is an offshore continuation of the land area, an area of longitudinal and transverse depressions with adjacent isolated banks forming the central shelf, and a chain of large, shallow banks with intervening saddles along the outer edge of the shelf. The morphology of such shelves is often thought to have resulted from glacial scour, but a study of seismic reflection data indicates that the major landforms are bedrock controlled and represent cuestas and mesas with adjacent lowland areas developed to varying degrees of maturity on the bedrock surface. These landforms are typical of a coastal plain environment developed by subaerial erosion across gently dipping strata of variable lithology. In general, they bear little relation to glacial landforms. Cuesta development is most prominent on the outer banks of the Scotian Shelf and the mesas are dispersed as isolated banks throughout the lowlands of the central shelf. The longitudinal troughs, particularly well developed along the Labrador Shelf, are formed on relatively soft Mesozoic- Cenozoic strata. The troughs are parallel to the contact with the more resistant rocks of the inner shelf (submerged Piedmont province).The troughs generally coalesce with transverse depressions forming lowlands or basins. The troughs appear to be analogous to the lowland areas of the emerged coastal plain south of New Jersey where the rivers flowing off the Piedmont at the fall zone change course and run for some distance subparallel to the inner margin of the

coastal plain before crossing it to the ocean.

The coastal plain landforms are of Late Tertiary age but in some areas inheritance from earlier Tertiary erosional cycles influenced the development of the surface. Submergence of the coastal plain along the Canadian margin is probably a result of tectonic subsidence caused by thermal contraction and sediment loading of the crust following the opening of the North Atlantic Ocean. Sedimentation during much of the Quaternary probably was slow because of the cover provided by the continental ice sheet. Consequently, the pre-Pleistocene topography was not completely masked by subsequent sedimentation and its impression is commonly evident in the present morphology of the eastern Canadian shelf. During times of glacial deposition, especially during recessional phases, sedimentation rates were high but probably not of sufficient duration to fill the depressions on the coastal plain surface.

The best evidence for modification of the coastal plain by glacial erosion is the degree of overdeepening that can be observed in the lowland areas which varies from areas of little modification to an extreme where the dominant landforms are obviously the result of glacial scour. The fiord-like features such as the Laurentian Channel and its tributary hanging valleys which incise the western Grand Banks of Newfoundland and originate in the fiords along the south coast of Newfoundland are examples.

The distribution (Fig. 3) of till and lag deposits of reworked debris signify that at least one ice sheet or ice shelf covered the greater part of the Scotian Shelf. Evidence for the last major ice advance on the shelf is clearly indicated by a submarine end moraine complex (Fig. 4) (King _et al._, 1972) and evidence from the present study indicates that the main complex was

fully developed in Middle Wisconsinan time at approximately 26000 yBP. The complex is more or less continuous over a distance of 840km, from the central part of the Gulf of Maine to the Laurentian Channel. East of the Laurentian Channel the moraine complex occurs south of the Burin Peninsula (Fader et al., in press). Several lobes which probably belong to the same system were described east of Newfoundland (Dale and Haworth, 1979) and on the Labrador Shelf (Grant, 1972; Van der Linden et al., 1976; Fillon and Harmes, 1982).

Following the period of glacial deposition the glacial and glaciomarine deposits were in part eroded and reworked to form the geological formations representing Late Wisconsinan and Holocene history (Figs. 2 and 3). During this period the shelf environment was influenced by isostatic adjustment of the crust and the glacioeustatic lowering of sea level. Apparently, the ice had receded from the shelf and isostatic adjustment was complete well before sea level reached the minimum still stand. The submarine terrace which was cut during the minimum sea level stand (115-120m) is unwarped and occurs at the same depth across the entire shelf. The present study has helped to define the age of the terrace, indicated in Figure 3 by the contact between the Sambro Sand and the Sable Island Sand and Gravel. Emerald Silt and glacial till in the peripheral areas of the basins were reworked to form the sublittoral Sambro Sand, while the bank areas were subjected to subaerial erosion. During the Late Wisconsinan-Holocene transgression all sediments between the old and present shoreline were modified in response to a high energy beach environment. As the transgression progressed it produced large areas of clean, well sorted, sands and gravels (Sable Island Sand and Gravel Formation) across the outer banks, leaving underlying truncated deposits of

Emerald Silt and glacial till. The winnowed fines were deposited in the adjacent basins to form the LaHave Clay. On the inner shelf between the end moraine complex and the present shoreline of mainland Nova Scotia the gradient was sufficiently steep to allow undercutting of the glacial material by the advancing sea. Consequently, the glacial and glaciomarine deposits of this area were almost completely removed and most of what remains is coarse gravel confined to depressions on the bedrock surface. This, to a large degree, isolated the offshore glacial and glaciomarine section from the mainland glacial deposits except along the ancient river channels.

The sedimentary distributions (Fig. 3) established during Late Pleistocene and Holocene times are still preserved at the present seabed indicating that much of the surface is relict except in shallow areas or areas of extreme exposure where deposits of Sable Island Sand and Gravel are still being reworked. The occurrence of relict iceberg furrows on the glacial till surface (King, 1980) demonstrates that indeed these surfaces are relict and have only been slightly modified (Fig. 5, see B).

## SEISMOSTRATIGRAPHY OF THE SURFICIAL FORMATIONS AND ASSOCIATED FEATURES

The surficial formations are described in terms of their seismic reflection characteristics as they appear on the Huntec DTS profiles. These characteristics include the presence or absence of coherent internal reflections, their amplitude, continuity, spacing, relief, structural form, and boundary relationships. Table I shows the relationships between the seismostratigraphy, lithostratigraphy and thickness of each of the formations together with a brief description. Figures 6a to 6h describe the distribution

and stratigraphy of the surficial succession from the Gulf of Maine to the Northeast Newfoundland Shelf. Column sections interpreted from the Huntec DTS profiles, were chosen at piston core locations and in areas where significant relationships between formations exist. This format consolidates the many line kilometres of data on the seismic profiles. Only those formations and associated features essential to the development of the glacial model are described in detail.

Scotian Shelf Drift

At the base of the surficial succession overlying the bedrock surface is the Scotian Shelf Drift Formation (glacial till). It appears on the Huntec DTS profiles as a uniform dense grey pattern of incoherent reflections (Fig. 5). Where the till is thick (i.e. 60m) and penetration with the DTS is limited, small airgun seismic reflection systems define the till thickness to bedrock.

The regional distribution of till at the seabed is shown in Figure 3. With the exception of the inner shelf and shallow bank areas, where the Late Pleistocene-Holocene marine transgression eroded most deposits, till occurs in the subsurface as a continuous blanket of relatively uniform thickness (i.e. 10 to 15m). The thickest deposits ( 100m) occur in the moraines on the flanks of the basins. Numerous wedge-shaped bodies of till situated on the distal side of the moraines, are also found on the flanks of the basins and depressions of the shelf. These features are referred to as "till tongues" (Fig. 5). They are usually found at the stratigraphic transition zones between the Scotian Shelf Drift and the Emerald Silt Formations. Figure 7a

shows the known distribution of till tongues from the Gulf of Maine to Northeast Newfoundland Shelf. The column sections of Figures 6b to 6h also serve to indicate variations in till tongue distribution and thickness.

The surface of the till exhibits several variations in morphological character. Its surface is conformable to the underlying bedrock surface, where the till occurs as a continuous sheet of ground moraine. Along the Scotian Shelf Moraine Complex which occurs 30 to 40km offshore (Fig. 4), the till exists as a belt of low ridges, which are up to 120m in height, and are parallel to the present coastline. In contrast to this large scale morphology is the occurrence of relict iceberg furrows on the surface of the till formed by grounding icebergs. These furrows have average depths of 2-3m and sometimes extend to over 200m in width. Figure 5 shows typical iceberg furrows developed across the surface of the Fundian Moraine. The surface of the till, as seen on the DTS profiles, in many of the broad basins and isolated depressions of the shelf displays a unique hummocky morphology. In most areas these hummocks are buried beneath thick sediments; however, at a few localities such hummocks have been found exposed at the seabed. Sidescan sonograms revealed their true morphology to be sub-parallel ridges (Fig. 8) which we refer to as "lift-off moraines". These features will be described in detail later.

Emerald Silt Formation

The Emerald Silt Formation is subdivided into three seismic facies: (A) high-amplitude, continuous, coherent reflections, (B) medium- to low-amplitude, continuous, coherent reflections, (C) discontinuous coherent re-

flections (also referred to as the transitional facies). The following acoustic characteristics are common to facies A and B. Continuous coherent reflections are smooth, parallel and closely spaced at an average distance of 0.3m. Throughout the vertical section individual reflections may vary in intensity from weak to strong; however, in the horizontal direction individual reflections maintain uniform intensity over large distances despite changes in their depth of occurrence within the section. Point source reflections (hyperbolae) can occur within all the seismic facies, but are largely confined to facies B.

The Emerald Silt Formation varies from a few metres to over 100m in thickness. The thickest deposits occur within the basinal areas of the central Scotian Shelf, outer Gulf of Maine and in isolated depressions on the eastern Scotian Shelf and Grand Banks of Newfoundland. Isolated erosional remnants of Emerald Silt are occasionally found on the inner Scotian Shelf and bank areas and some are indicated in Figures 6b to 6h.

Seismic facies A occurs mainly at the base of the Emerald Silt and the characteristic high-amplitude continuous coherent reflections in some places appear as a grouping of reflections as discrete bands on the profiles. The facies attain a thickness of up to 80m. A dominant characteristic of this facies is the conformability of the reflections which parallel the surface morphology of the underlying till. Reflections with dips of 5° occur over the steepest till surfaces, but most reflections range in dip from 0° to 2°. Facies A in some places is characteristic of younger sections (B in Fig. 7b) but this is not a common occurrence.

Seismic facies B consists of medium- to low-amplitude continuous coher-

ent reflections (Fig. 5). It overlies facies A in the basins and depressions of the shelf and reaches a maximum thickness of 40m. The structural style of the facies changes from a conformable character in the lower part of the section to a more ponded style in the upper part of the section with associated onlap on the basin flanks. An unconformity of wide regional extent (in some places disconformable) occurs at the base of facies B. The unconformity is best developed on the flanks of the basins and appears on the DTS profiles as truncations of continuous coherent reflections (Fig. 9). It is often well developed locally around bedrock or till topographic highs and is sometimes associated with a moat (Fig. 9) surrounding positive features. Point source reflections appear to occur more frequently within facies B. Sections where they occur range in thickness up to 40m and are largely confined to the southern Gulf of Maine and the western Scotian Shelf. Seismic facies C of the Emerald Silt Formation consists of discontinuous coherent reflections transitional between Emerald Silt facies A and the Scotian Shelf Drift (Figs. 5 and 10). The coherent aspect of the reflections is of limited extent in both the vertical and horizontal dimensions and reflections cannot be correlated throughout the deposits. This facies is of limited extent and is best developed in the southern Gulf of Maine where it attains a maximum thickness of 20m (Fig. 10).

Till Tongues

Till tongues are wedge-shaped bodies of till interbedded with continuous coherent reflections of the Emerald Silt Formation (Figs. 5 section C and 7b and 7c). At the thin or feather edge of the tongues, continuous coherent

reflections of the Emerald Silt, splay, and continue under and over the feather edge of the tongues. Beneath the till tongues and in a more proximal position the Emerald Silt section often terminates abruptly against the till (root), but in some areas may continue beneath the moraine complex. Figure 13 at the 24.5km mark shows the termination of a 40m section of Emerald Silt beneath a till tongue on the Scotian Shelf. The tongues vary in length from 100m to over 25km and the thickness of the till is highly variable ranging from a few metres to over 40m. The rate of thickening can vary from a maximum thickness of 40m over a distance of 700m, to 8m over a distance of 3000m.

The distribution of till tongues from the Gulf of Maine across the Scotian Shelf to the Northeast Newfoundland Shelf is indicated in Figures 6b to 6h and 7a. Where seismic coverage allows, such as in the Gulf of Maine, correlation of till tongues between adjacent ship's tracks has been attempted. In other areas the tongues have only been identified where they occur along individual profiles. Also, in Figures 6b-6h the stratigraphic position of the till deposits is indicated numerically. These stratigraphic designations only apply along individual profiles and correlation across the shelf is not implied. Most till tongues are confined to the flanks of the depressions on the shelf. Along the inner shelf, till tongues are generally confined to the distal, seaward side of the moraine complex.

A complex stratigraphic section that includes eight till tongues occurs in the northeastern corner of Emerald Basin and extends on the northwestern area of Middle Bank. Figure 11 is a composite airgun and Huntec DTS interpretation from north to south across the area. The cross-section is supple-

mented by DTS profiles (Figs. 12 and 13) parallel to and normal to the section. They illustrate details of the stratigraphic relationships. These eight till tongues are sequentially numbered upwards through the section beginning with number 2 above the base of the section. Till number 1 is a thin till which overlies the bedrock surface as ground moraine. In order to develop a three dimensional model of the distribution of the till tongues in this study area, a plan view diagram showing the extent of each till tongue is provided (Fig. 14). Proximal to the feather edge of the till tongues near the Country Harbour Moraine, till tongues 2 to 4 appear to grow out of, or are rooted in the moraine at depth (Figs. 14 and 11). In contrast, till tongues 5 to 9 have been eroded at their rooted ends and the intervening Emerald Silt between tongues has been exposed. This provided an opportunity for samples to be obtained throughout the section.

The relationship between the till tongues and the Emerald Silt differ between the top and bottom surfaces of the tongues. The bottom surfaces are generally smooth with little or no relief and they are parallel to the underlying Emerald Silt reflections. Where steeper dips and rapid changes in structure occur within the Emerald Silt (Fig. 7b), the till generally conforms with the underlying deposits. One exception is shown in Figure 13 from the 10-16km mark where till tongue number 7 shows irregularities with relief up to 10m at its base indicating an erosional contact.

In the vertical direction the till tongues terminate abruptly and the contact between the Emerald Silt and the till is clearly defined. However, in the horizontal direction along the upper tongue surfaces, the transition is gradational. Individual Emerald Silt reflections become discontinuous and

less defined, and eventually terminate within the till.

Lift-off Moraines

As discussed earlier, lift-off moraines are parallel ridges of till which occur on the surface of the basal till in the basin areas. They are illustrated by sidescan data in Figures 8 and 15. These moraines commonly occur in fields or groups, the distribution of which is shown in Figure 16a. The ridges vary in height from a few metres to 20m, in width from 20 to 150m with the average at 40 to 50m, and are spaced from 30 to 400m apart, but on average about 40m. Figure 16b shows several DTS profiles collected over lift-off moraines from the Gulf of Maine, Scotian Shelf and the Grand Banks of Newfoundland. They show variations in internal structure and complex relationships with the adjacent and overlying Emerald Silt. Lift-off moraines generally occur in association with facies A of the Emerald Silt, and it is sometimes difficult to define where the upper surface of the moraine ends and the overlying conformable glaciomarine sediments begin. The flanks of the ridges are very steep and are generally defined by an abrupt termination of Emerald silt reflections which occur between the ridges. Figure 16b illustrates how discrete bands of Emerald Silt can be correlated between ridges across an entire field of lift- off moraines, and suggests contemporaneous deposition of the till and glaciomarine sediment. If the glaciomarine sediment had been deposited later, it would mimic the shape of the ridge. In contrast, the till ridges sometimes occur at the seabed without any associated glaciomarine sediment.

## LITHOSTRATIGRAPHY AND CHRONOLOGY OF THE GLACIAL SURFICIAL FORMATIONS

Previous studies (King, 1970 and 1980) utilized information from the seismostratigraphy of the surficial formations including distribution, boundaries, structure and thickness, combined with lithologic information from samples, to establish the lithostratigraphic classification which forms the basis for Figures 2 and 3. Data from the piston cores collected for the present study supplement the previous descriptions for the Scotian Shelf Drift and the various facies of the Emerald Silt Formation. The results are summarized in Table I and discussed in detail below.

### Piston Core Analysis

A total of 25 cores were obtained with a split piston, Benthos corer to represent the various glacial, glaciomarine, and postglacial units in the eastern Gulf of Maine, Scotian Shelf, Laurentian Channel, and Grand Banks of Newfoundland. Their locations are indicated in Table II and on the seismic column section maps, Figure 6b to 6h. Core sites were chosen along high resolution seismic reflection profiles which had been previously collected and interpreted, with the objective to obtain maximum stratigraphic coverage during sampling. In the process of sampling, the DTS system was again employed to define precise site locations with respect to stratigraphy. This technique was of particular value in areas where the section was exposed at the seabed by virtue of an unconformity. The use of standard navigational controls alone could have resulted in the collection of samples with stratigraphic errors as great as 15 to 20m.

The cores were split, x-rayed, subsampled for grain size analysis,

radiocarbon dating and micropaleontological analysis, and described in terms of lithology and structure. The radiocarbon dates were based on total organic carbon in the sediment after removal of carbonate carbon. For the paleontological analysis only those parameters are reported which are considered most significant in defining glacial and postglacial environments (Vilks and Rashid, 1976), that is, total abundance of foraminifera, percentage of Elphidium excavatum and planktonic foraminifera. Foraminiferal abundance is reported as the number of tests per 60ml of wet sediment. Sand fractions remaining from the paleontological analysis were examined microscopically. Each core was classified according to stratigraphic formation and seismic facies, and correlations were made with seismic events where possible. Unfortunately, colour was generally not noted because of the time lapse between sampling and examination; however, most of the features enhanced by colour in core descriptions, especially sedimentary structure, are preserved in the x-radiograms so that this detailed information was not lost.

An index for the core description is shown in Figures 17a and 17aa and the detailed logs are shown in Figures 17b to 17z. They are grouped according to area and stratigraphic position. The following discussion of the various analyses generalize and consolidate the detailed descriptions of the individual cores in the diagrams.

## Gulf of Maine Suite

The Gulf of Maine suite of samples is comprised of five cores (76-016-2 to 76-016-6; Fig. 17) and represent four major stratigraphic units: Emerald Silt, facies A, B and C, and LaHave Clay. Facies A occurs near the base of

core 76-016-5 and only includes the uppermost 2m of this unit. Facies B occurs in cores 76-016-2, 3, 4 and 5, facies C in core 76-016-6, and LaHave Clay in cores 76-016-2 and 3.

Lithology

Emerald Silt, facies A core 76-016-5 is characterized by weakly rhythmically banded, gravelly muds. The rhythmic bands arise from alternating bands of silt and clay, both of which vary from 1 to 3cm in thickness, but the variations between the thickness of clay and silt for any given couplet can be wide. The boundaries between the silt and clay layers are distinct and undisturbed and the degree of contrast between the layers appears to be a function of contrast in textural composition. The gravel content is widely dispersed and the clast diameters range from 0.1 to 2cm. The clasts do not appear to disturb the regularity of the boundaries between the clay and silt bands and they appear to be uniformly dispersed through both members of the couplets; although, the gravel content may vary over broader intervals involving several couplets.

The average number of couplets or complete cycles/m is approximately 25 for this section of core 76-016-5. Because of the weak nature of the banding it is difficult to obtain an accurate count and the number probably represents a minimum value. Their is no indication of bioturbation which if present would also tend to decrease the number of cycles/m.

Grain size analysis indicates an average composition of 3% gravel, 10% sand, 40% silt and 47% clay with a slight increase in sand content upward through the section. When plotted on a scatter diagram of mean grain size

versus sorting (Fig. 20), these sediments fall within the upper part of the Emerald Silt grouping. Individual clay and silt members of the rhythmic bands were not sampled so that the analyses also represent an average of these variations.

The sand fraction is dominated by quartz and feldspar with minor quantities of lithic grains and foraminiferal tests. A relatively high percentage of the quartz grains show a glossy, secondary growth of silica, thus many are subangular to subrounded in shape.

In Emerald Silt, facies B (cores 76-016-2 and 76-016-5) rhythmic banding is very weak and is barely visible on the split core face but can be seen in x-radiographs. Otherwise, the band thickness and distribution of gravel is similar to facies A, but the clasts are slightly smaller (0.1 to 1.0cm in diameter). The average number of cycles/m in the cores is approximately 23 but ranges as high as 30 cycles/m in cores 76-016-4 and 5. The average textural composition is 3% gravel, 13% sand, 47% silt, and 37% clay, which is slightly coarser than facies A and the composite cumulative curves and frequency distributions (Fig. 19) show a biomodal distribution. This is interpreted as arising from bulk sediment analysis of sections of rhythmically banded silt and clay layers so that the analyses represents an average of these variations. Quartz and feldspar are common in the sand fraction and significant amounts of quartz show secondary growth. The biogenic fraction consists of foraminiferal tests and variable amounts of ostracods. Shells are sparsely dispersed throughout much of core 76-016-3. Pyritized worm tubes are common near the top of the section in cores 76-016-2 and 3, and in the lower half of core 76-016-4.

Emerald Silt, facies C was only sampled in one core, 76-016-6. It is characterized by strongly laminated, sometimes steeply dipping beds grading to rhythmically banded wispy beds. The average textural composition is 3% gravel, 7% sand, 53% silt, and 37% clay, which is similar to facies B. The sand fraction is also similar. These samples when plotted on the scatter diagram (Fig. 20), tend to group together in the Emerald Silt facies A & B field; however, a few samples are more widely displaced.

The LaHave Clay is a wispy-laminated, sandy silt; a peripheral facies of the typical LaHave Clay of LaHave and Emerald Basins which is normally a silty clay. The average textural analysis is 1% gravel, 27% sand, 55% silt, and 17% clay. Fragile, woody organic fragments are often present in the sand fraction.

Samples of till from the Gulf of Maine were collected with bottom grab samplers, and represent only the surface of the till at the seabed. Winnowing of the till surface by bottom currents and the effects of iceberg furrowing are post depositional processes that have modified the upper till section. The till is a dark greyish to reddish brown, poorly sorted sediment, containing angular fragments of pebble to boulder sized material. The average range for the median diameter of the till is 0.18 to 1.5mm, and the average textural analysis, based on approximately 30 samples, is 38% gravel, 41% sand, 11% silt and 10% clay (Fig. 18). When plotted on the scatter diagram (Fig. 20) these till samples fall within the Scotian Shelf Drift field together with till samples from the Western Banks of Newfoundland. Striations and polished surfaces occur on many of the larger clasts. The gravel fraction is generally angular to subangular, but some fragments are well rounded.

The gravel material from the Gulf of Maine consists of fragments of sandstone, mudstone, metamorphic rocks, granite and basalt.

## Micropaleontology

Emerald Silt, facies A recorded a low abundance of foraminifera ($10^2$) per sample, dominated by Elphidium excavatum. In facies B the abundances increased upwards in the section from $10^2$ to $10^4$ and are accompanied by an increase in diversity dominated by Elphidium excavatum. Facies C also falls in the intermediate range of abundances. LaHave Clay shows high abundance ($10^5$ to $10^6$) and high diversity of forams. The introduction of planktonic forams, and a sudden decrease in Elphidium excavatum is also evident. Ostracods are present in variable amounts throughout. The distributional trends of foraminifera from the Gulf of Maine compare favorably with the average distribution between formations for the entire suite of cores (Fig. 21).

## Correlation of Cores and Radiocarbon Dates

The correlation of piston cores from the Gulf of Maine is mainly based on seismostratigraphic control by (1) tracing specific reflections which can be identified as lithologic events in the cores, (2) tracing characteristic seismostratigraphic units such as the high amplitude reflection unit (facies A), and (3) by identifying on the seismic records the formational changes such as the contact between LaHave Clay and Emerald Silt, facies B, recognized in the cores on the basis of lithologic and biogenic changes. Using the correlations of Figure 22a it is possible to compile a composite of the pis-

ton cores, and core 76-016-3 was chosen as the control core. The control core was selected from the suite of cores because it contained the most complete stratigraphic section, and was used as a reference to which the other cores were seismically correlated. It was also important to establish the composite section of piston cores at a site where the section was most complete as indicated by the seismic profiles. The type acoustic section is normally chosen at the control core site but in this case a more representative seismic section was chosen 1.1km south of the control core location (Fig. 23a, 1.1km south of core 76-016-3 position). Figure 23a shows that approximately half the section was sampled.

In Figure 24a radiocarbon dates (Table III) are plotted against their stratigraphic position at the composite section (Fig. 23a). Range bars indicate the analytical error on the age determinations and an estimated stratigraphic error resulting from core positioning and seismostratigraphic correlations. All data are included in the plot except those which did not give a finite result. With the exception of samples 2801 and 2755 all points fall close to the line of best fit. Although it is unfortunate that material other than total organic carbon was not available to supplement the age determinations, it is promising that the dates show a systematic relationship to the stratigraphy, and the straight-line relationship for samples of glaciomarine origin (ages 17000 yBP) provides some justification for extrapolation through the lower, unsampled portion of the section. It seems reasonable to accept the results as a basis for regional correlation across the study area and for establishing at least a relative chronostratigraphy for the offshore acknowledging the possibility that the ages are likely to be old

rather than too young because of the presence of "dead" (non-radioactive) carbon.

In Figure 24a significant stratigraphic events are plotted against the chronology for the Gulf of Maine section. The Emerald Silt, facies A deposition dates from 38000 to 26500 yBP, facies B from 26500 to 13500 yBP, and LaHave Clay from 13500 yBP to the present. Both facies of Emerald Silt show an average sedimentation rate of approximately 1.8m/1000 y. The sedimentation rate for LaHave Clay decreases to approximately 0.3m/1000 y. The bulk of the Fundian Moraine between its till tongue and top, appears to have been deposited between 32000 and 26500 yBP.

Emerald Basin Suite

Fifteen cores constitute the Emerald Basin suite: 79-011-1 and 2; 79-011-6 to 12; 79-011-2V to 4V (V indicates vibrocores) and 82-003-4, 6 and 7. They represent four major stratigraphic units: Scotian Shelf Drift, Emerald Silt, facies A and B, and LaHave Clay. Scotian Shelf Drift occurs in core 79-011-8 which sampled a till tongue associated with the Country Harbour moraine. Facies A occurs in cores 79-011-2, 2V, 3V, 4V, 6, 7, 9, 10 and 82-003-7; facies B in cores 79-011-1, 11, 12 and 82-003-4, 6 and 7; and LaHave Clay in cores 79-011-1 and 82-003-4. With the exception of core 79-011-12 which is located to the south in the deeper part of the basin, all samples were obtained in the northeast end of the basin and northwest flank of Middle Bank. The object of this sample program was to obtain the most complete Wisconsinan section possible. This area was unique in that much of the section was exposed at the seabed along a well developed seabed unconformity.

Lithology

Scotian Shelf Drift from till tongue 7 is a silty clay with a minor gravel content, and is structureless except for several weak bands indicated on the x-radiographic log. Sub-samples of the core are uniform in composition and average approximately 1% gravel, 7% sand, 43% silt and 49% clay (Fig. 18); similar to the composition of Emerald Silt, facies A. On the scatter diagram (Fig. 20), these samples plot closely together in the Emerald Silt field, unlike till samples from the Gulf of Maine and the Grand Banks of Newfoundland sampled by bottom grab.

Emerald Silt, facies A is characterized by strong rhythmic banding, more distinct than those of the equivalent facies in the Gulf of Maine. The bands range in thickness from 1 to 3cm and alternate from clay to silt in each band. In freshly cut cores the bands alternate in colour from 5y 3/1 to 5y 5/1 (Munsell soil colour) respectively. The average number of couplets/m ranges from a minimum of 32 to a maximum of 52. The rather wide range results from the difficulty in identifying some individual cycles. In general, the muds are slightly finer than those from the Gulf of Maine, and have less associated gravel. The average textural composition is 1% gravel, 7% sand, 38% silt and 54% clay (Fig. 18). The samples of Emerald Silt facies A on the scatter diagram (Fig. 20) plot in the same grouping as those from facies B. Quartz grains with secondary overgrowths are common in the sand fraction. This appears to be a common feature of Emerald Silt, facies A formation and appears related to environment of deposition and not source.

Emerald Silt, facies B is characterized by a wide range in structure,

from weak to strong rhythmic banding and weak to strong laminations. The average number of couplets/m is approximately 32 for the section represented in core 79-011-11, Table IV. Bioturbation is common and tends to mask primary structures. Size analysis yields an average distribution of 15% sand, 45% silt and 40% clay. Gravel occurs in minor quantities throughout the facies. A wide variety in grain sizes is characteristic of facies B and some samples contain up to 40% sand. The sand fraction is dominated by quartz and feldspar with minor amounts of lithic fragments, foraminiferal tests and ostracods. Disseminated pyrite, organic fragments, and shells and shell fragments are common throughout the facies.

The LaHave Clay is a weakly to strongly laminated sediment. The average textural analysis yields 5% sand, 45% silt and 50% clay. Gravel is virtually absent from the sediment. The sand size sediment is dominated by quartz and feldspar together with foraminiferal tests, minor pyrite and woody organic fragments. Near flanks of the basin the sand content increases to 35%. LaHave Clay is heavily bioturbated, and shells and shell fragments are common.

## Micropaleontology

Emerald Silt, facies A shows a low abundance of foraminifera (10 to $10^2$) dominated by *Elphidium excavatum*, which is similar to the equivalent facies in the Gulf of Maine. In facies B the abundance increases upwards from $10^2$ to $10^3$. *Elphidium excavatum* decreases within facies B towards the top of the formation from 90% to 50%. The LaHave Clay shows high abundance of foraminifera ($10^4$), the presence of planktonic foraminifera and a

dramatic lowering in the percentage of Elphidium excavatum to between 5 and 10%.

## Correlation of Cores and Radiocarbon Dates

Figure 22b shows a correlation of piston cores from Emerald Basin. All cores were included with the exception of cores 79-011-6, 10, 11 and 12 where seismic control was lacking. Based on these correlations, a composite section of piston cores was compiled (Fig. 23b) where the seismic records indicated the most complete section. Stratigraphic positions of the samples where radiocarbon dating had been carried out (Table III) were obtained from this diagram and plotted against the dates (Fig. 24b). After the line of best fit was established, data from cores 6, 10 and 11 were positioned on the diagram to establish the stratigraphic position of these cores which previously could not be seismically correlated. With the exception of samples 8539 and 8540 the plot shows little scatter indicating that the radiocarbon dates are reliable, at least in a relative manner with respect to the stratigraphy. If the dates were strongly influenced by the presence of "dead" (non-radioactive) carbon one would not expect a systematic plot over such a wide stratigraphic range. In Figure 24b, where the significant stratigraphic events are shown against the chronostratigraphic plot, it appears that Emerald Silt, facies A deposits occurred from 46000 to 32000 yBP, facies B from 32000 to 14500 yBP and LaHave Clay from 14500 yBP to present. Emerald Silt, facies A shows an average sedimentation rate of 3.6m/1000y, facies B; 1.1m/1000y and LaHave Clay; 1.1m/1000y.

## Laurentian Channel Suite

The Laurentian Channel suite of samples is composed of 3 piston cores, only one of which is presented in this study. The remaining cores were collected in an area of intense iceberg furrowing which tended to disturb the stratigraphy. Core 73-003-351 penetrated Emerald Silt, facies B and LaHave Clay.

## Lithology

Emerald Silt, facies B is characterized by weakly, rhythmically banded sandy mud with occasional gravel. The gravel is widely dispersed throughout the facies with several local zones of concentration. An average grain size distribution for facies B is 5% gravel, 20% sand, 35% silt and 40% clay. A slight increase in clay content together with a decrease in sand occurs towards the bottom of this facies. There is little evidence for bioturbation within facies B. The sand fraction is dominated by quartz and feldspar with minor amounts of lithic grains and occasional pyrite fragments and concretions. The biogenic component consisted of foraminiferal and ostracod tests. Many of the quartz grains showed secondary silica overgrowth.

The LaHave Clay formation is a bioturbated, sandy, silty clay typical of the type section found in Emerald Basin. The average textural analysis shows 6% sand, 32% silt and 61% clay with less than 1% gravel. Pyrite in the form of worm tube and crack fillings is common throughout.

## Micropaleontology

The Emerald Silt, facies B shows a lower abundance of foraminifera than

the LaHave Clay, and is dominated by Elphidium excavatum. The total abundance of foraminifera increases upwards in facies B from $10^3$ to $10^4$. LaHave Clay shows an increase in planktonic foraminifera from less than 5% to over 40% towards the upper part of the section, and a sharp decrease to 20% in Elphidium excavatum from 80% within the underlying facies B. Ostracods occur throughout the section in variable amounts.

## Correlation of Cores and Radiocarbon Dates

Figure 24c is a plot of the radiocarbon dates (Table III) against the stratigraphic section at the core location shown in Figure 23c. Range bars in this case indicate only the analytical error for the age determination. The line of best fit was extrapolated downward to the top of the basal till. These data show that the Emerald Silt, facies A was deposited from 40500 to 31500 yBP, facies B from 31500 to 23000 yBP and LaHave Clay from 23000 yBP to the present. Both facies A and B show an average sedimentation rate of approximately 0.7m/1000y, and LaHave Clay shows a decrease to 0.5m/1000y. The Laurentian Moraine, a major feature of the Laurentian Channel area, was deposited at approximately 35000 yBP.

## Grand Banks of Newfoundland Suite

The Grand Banks suite consists of a vibrocore, 75-009-21V, collected from Downing Basin in the central Grand Banks area, and 3 piston cores, 78-012-132, 78-012-223 and 78-012-321 collected in outer Placentia Bay southwest of the Avalon Peninsula. These samples represent three stratigraphic formations, Emerald Silt, facies A and B, and LaHave Clay. Numerous grab samples

of glacial till were collected along the shelf south of Newfoundland and in the Avalon Channel. Emerald Silt, facies A was only sampled in core 75-009-21V. Facies B occurs in cores 78-012-321 and 78-012-223, and LaHave Clay in cores 78-012-132 and 78-012-223. As a result of a major unconformity within the section at Placentia Bay, core 78-012-223 sampled LaHave Clay and the lower part of facies B.

Lithology

The samples of till show a wide variation in particle size distribution. Samples from the western area contain an average of 22% gravel, 53% sand, 15% silt and 10% clay (Fig. 18). Samples from the Burin Moraine, western Placentia Bay, averaged 38% gravel, 37% sand, 20% silt and 5% clay. The former area is underlain by Pennsylvanian sandstone while the latter is flanked by metamorphic and volcanic rocks which probably accounts for the wide variations in textural composition.

Emerald Silt, facies A from Downing Basin is a rhythmically to wispily banded sediment ranging in composition from a muddy sand to a sandy mud. Gravel is virtually absent from core 75-009-21V. The average number of couplets/m is approximately 40, and there is no evidence for bioturbation. The textural composition averages 50% sand, 35% silt and 15% clay, with an increase in the sand content to 70% towards the bottom of the section. Minor gravel occurs near the base of the section.

Emerald Silt, facies B is a weakly rhythmically banded sediment which ranges from a gravelly to sandy mud. Core 78-012-223 collected near the Burin Moraine shows an average composition of 20% gravel, 15% sand, 30% silt

and 35% clay, close to that of the nearby till. Core 78-012-321 from the eastern side of Placentia Bay is composed of 1% gravel, 6% sand, 47% silt and 46% clay. The sand fraction is again dominated by quartz and feldspar, with a small percentage of the quartz grains showing secondary growth. The biogenic component consists of foraminiferal tests. Shells were common in the core from the eastern side of Placentia Bay but were absent from the western core.

The LaHave Clay is a strongly to weakly laminated sandy to silty clay. The average textural analysis is 3% sand, 47% silt and 50% clay with the clay content increasing towards the bottom of the section. This increase in coarseness at the top results from modern reworking. The sand fraction consists of quartz and feldspar fragments with foraminiferal tests. In the upper part of the section the foraminiferal tests account for more than 50% of the sand fraction. Lithic fragments increase in abundance towards the bottom of the section, and zones of pyritized worm tubes and crack fillings are common throughout. Zones of bioturbation are evident and shells are common throughout the section.

Micropaleontology

Emerald Silt, facies A from Downing Basin shows a moderate abundance ($10^3$) of foraminiferal tests dominated by *Elphidium excavatum*. Planktonic foraminifera are absent from the bottom of the core but increase to over 10% at the top. Facies B from Placentia Bay shows a more or less constant foraminiferal total abundance of $10^2$ throughout the core with *Elphidium excavatum* dominating. LaHave Clay shows an upward increase in total

abundance from $10^2$ to $10^3$ accompanied by a decrease in Elphidium excavatum from 85% to 30%, and an overall absence of planktonics.

## Correlation of Cores and Radiocarbon Dates

Figure 24d is a plot of radiocarbon dates (Table III) against the stratigraphy of the type locality shown in Figure 23d for Placentia Bay. This dates the deposition of facies A from 29500 to 28000 yBP. The facies is represented acoustically in Figure 24d as a 2m section overlying basal till. The Burin Moraine which occurs to the west, interbedded with facies A, Emerald Silt, was formed during this time interval. Facies B dates from 28000 to 15500 yBP, and LaHave Clay from 15500 yBP to the present. Both facies A and B and the LaHave Clay show a sedimentation rate of approximately 1.3m/1000y.

## DEPOSITIONAL MODEL OF GLACIAL AND GLACIOMARINE FACIES AND THE ORIGIN OF TILL TONGUES, LIFT-OFF MORAINES, AND REGIONAL, SUBGLACIAL, ICE SHELF MORAINES.

Through the interpretation of high resolution seismic reflection profiles of glaciomarine deposits on the Scotian Shelf, Grand Banks, and eastern Gulf of Maine, and the employment of a model for a marine ice shelf (Carey and Ahmad, 1961), we have developed a conceptual model regarding the mechanism of deposition of glacial and glaciomarine sediments. The most informative profiles are those that exhibit boundary relationships between the various sedimentary facies and glacial features; specifically, (1) in the area of till tongues which occur on the distal side of the prominent moraines on the inner Scotian Shelf (King *et al.*, 1972), (2) in the area of till occurrences, including till tongues, on the periphery of the outer banks, 50 to 75km to the south of the moraine complex across Emerald Basin, and at the same seismostratigraphic horizon within the Emerald Silt as the till tongues, and (3) in the area of lift-off moraines, which are of common occurrence on the surface of many of the till deposits. Our studies of the seismic reflection profiles from these areas and the compilation of maps (Fig. 3) to define sediment distribution and depositional geometry not only serve as a source of ideas for depositional mechanisms and thoughts concerning the origin of these particular features, but they enable us to discuss and evaluate the relevance of the Carey-Ahmad model (1961) to the offshore Wisconsinan environment of southeast Atlantic Canada.

### SUMMARY OF CAREY-AHMAD (1961) MODEL

Carey and Ahmad (1961) recognized the following environments of sedimentation in relation to a glacier that extends to sea (Fig. 25).

A. Terrestrial - where the base of the glacier is above sea level

B. Grounded shelf - where the base of the glacier is below sea level but not floating

C. Floating shelf - where the glacier is floating, the buoyancy line defines the boundary with the grounded shelf

D. Inner iceberg zone - from the ice-barrier to the limit of winter pack ice

E. Outer iceberg zone - beyond the limit of pack ice but within limit of icebergs

Whether a glacier is wet-base or dry-base is of importance since wet-base glaciers have a base at melting temperatures which extends long distances back from the line of buoyancy, whereas a dry-base glacier has a base below melting temperatures landward from the buoyancy line. The wet- base glacier differs from the dry-base glacier in that there is little surface indication of where the floating shelf touches down and becomes grounded except for the presence of tidal cracking, within the ice. Buoyancy exerted by the sea reduces friction at the base of the ice shelf and causes thinning of the ice. On the other hand, the dry-base glacier experiences a sharp break in slope at the buoyancy line increasing its thickness by a factor of four to five times in a landward direction. Their model indicates that melting is the most significant factor contributing to deposition from a marine glacier.

Carey and Ahmad (1961) developed conceptual ideas regarding the sediment types characteristic of each of the zones in both the wet-base and dry-base models by applying their model to the Permian glacial deposits of Tasmania. Wet-base glaciers produce great thicknesses of unfossiliferous tills in the grounded shelf zone. Some of this sediment is transported to the floating shelf zone by dozing action to a foreset slope, forming beds of sand and mud interdigitating with the till. These give rise seaward to banded marine muds and silts transported by meltwater and turbidity currents. In contrast, the dry-base glacier, lacking meltwater, turbidity currents and foreset till flows, produces few tills, but yields marine pebbly mudstones and erratics all of which are dropped through water. Reading and Walker (1966) applied the Carey-Ahmad model to the Eocambrian glacial deposits of Norway, Edwards, in Reading (1978) later elaborated on the facies associations found in the marine glacial environment.

## TILL TONGUES AND THEIR ASSOCIATED FACIES

The occurrence, morphology, structure, and seismostratigraphic relationships of till tongues with their associated facies were discussed in a previous section, and the present discussion is mainly concerned with the genesis of the features and an improved knowledge of the depositional mechanism of glacial and glaciomarine sediments.

The stratigraphic relationships of the till tongue shown in Figure 5, where till overlies glaciomarine sediment, gives the impression of an advance and retreat of the ice mass that deposited the till. However, closer inspection shows no evidence for a forward thrusting motion during a glacial

advance; otherwise, the Emerald Silt underlying the tongue would show some indication of glaciotectonic deformation. If the ice depositing the massive till were to be in direct contact with the seabed, a condition which we think to be essential for the formation of glacial till, and the sediment were to be derived through subglacial melt-out, then the configuration of the till tongue could for the most part be formed by migration of the point of contact between the ice and the seabed. Horizontal migration of the ice- seabed contact would be controlled by a combination of several factors, for example changes in water depth due to variations in the topography of the seabed, by changes in the size and draft of the ice mass itself, influenced by climatic conditions, and by changes in isostatic and eustatic sea level. During times of advance of the buoyancy line the ice would make contact with the seabed by vertical motion, touching down upon, rather than plowing the seabed. Till would be deposited behind and glaciomarine sediment in front of the lift-off point (buoyancy line), and during a complete cycle of transgression and recession of the lift-off point, a wedge of till between underlying and overlying layers of glaciomarine sediment would be formed. These processes are represented diagrammatically in Figure 26 which shows the sequential development of a moraine and associated till tongue applying the concept that melt-out debris from grounded ice is deposited through accretion at the seabed as glacial till. In the floating position the melt-out debris is dispersed in the water column and deposited from suspension as glaciomarine sediment.

Till tongues are typical features of the distal side of the moraine complex along its entire length from the Gulf of Maine to Placentia Bay, Newfoundland. We have also observed them on published seismic reflection

profiles from Lake Michigan (Lineback et al., 1974) and MacLean and Josenhans (pers. comm.) have noted them on seismic profiles from the Baffin and Labrador Shelves. They are probably of common occurrence in glaciomarine and glaciolacustrine deposits where short-lived advances or glacial surges have occurred.

Additional information concerning the position of the subglacial surface of the ice with respect to the seabed and its control on sedimentation can be inferred from an inspection of the boundaries between the Emerald Silt and glacial till. Close examination of many seismic profiles show that the Emerald Silt beds often grade laterally, interpenetrating the till over a large range of distances. For example, in Figure 5 the transition from Emerald Silt to glacial till takes place over a distance of approximately 1.5km. That is from continuous coherent reflections in the silt, to an intermediate zone of discontinuous coherent reflections, ultimately to a zone of incoherent reflections in the massive glacial till. Along the upper surface of the till tongue (Fig. 5, section A) some onlap occurs, but in addition several very thin beds of silt terminate in the till where the transition occurs within several metres. Thus, the length of the transition zone can be highly variable, ranging from several metres to tens of kilometres. On the other hand, the vertical boundaries between the silt and till are relatively sharp ranging from the resolution limit of the system (approximately 0.25m) to several metres. The nature and variability in scale of these vertical and lateral boundary relationships have been observed in numerous other occurrences at the distal side of the end-moraine complex and some are illustrated in Figures 7b and 7c, 11, 12, 13, 27 and 29. On the basis of these

interpretations the position of the subglacial surface of the ice with respect to the seabed is critical in defining the sharp vertical contacts between the till and glaciomarine sediment. The sharp contact in the vertical section between the till and the glaciomarine sediments is the result of introduction or removal of the water column between the ice and the seabed. If the separation between ice and seabed is very small, the glaciomarine sediments present may be influenced by tidal pumping and develop facies C Emerald Silt on a very small scale.

In one instance a disturbance of the Emerald Silt underlying a till tongue was noted (Fig. 13) apparently arising from the partial erosion of a thin layer of the silt, and the depressions were subsequently infilled with glacial till. The depressions do not appear to be buried iceberg furrows; otherwise, their occurrence would extend beyond the edge of the till tongue. They may have been associated with irregularities at the base of an ice shelf as it grew and settled on the seabed. Once again, there seems to be no evidence for lateral motion of the ice with respect to the seabed. The fact that buried iceberg furrows were not observed beyond the feather edge of the till tongues, and indeed within the Emerald Silt Formation in general, constitutes evidence that the Emerald Silt was deposited from an ice shelf and not from calving bergs.

In some instances the glaciomarine unit grades laterally from well defined coherent reflections to a zone of mixed coherency (facies C) and back to continuous reflections; the massive till development being absent. For example, in Figure 10 north of the Fundian Moraine, such a transitional facies occurs as a 20 to 30m thick deposit, suggesting that the conditions

which gave rise to the facies change were stable over a substantial period of time. Depth appears to be the controlling factor. The transitional facies (C) occurs in slightly shallower water.

An inspection of the seismic profiles also shows differences in thickness of Emerald Silt and glacial till for a given depositional time interval. For example, in Figure 5, section C at the feather edge of the till tongue, a stratigraphic interval of 3m in the Emerald Silt is represented by an increasing thickness of till towards the core of the moraine. The equivalent thickness of till reaches a maximum of about 40m. The same relationship is also apparent in many other areas. For example in Figure 12 and 13 till bodies are much thicker than their equivalent Emerald Silt. At this site we attempted to quantify the difference in sedimentation rates in till tongues and the adjacent glaciomarine deposits. This was accomplished by tracing the stratigraphic control established for the composite section of piston cores in Figure 23b along the seismic section of Figure 13 to a more proximal area where several till tongues occur in the section and greatly expand its thickness. Sedimentation rates at this site are approximately 9m/1000y for the Emerald Silt and approximately 21m/1000y for the glacial till of the till tongues. A rate of 33m/1000y was estimated at the thickest section of till tongue number 7, but this is a short term rate and appears to be greater than the normal rate of till depositon. A sedimentation rate for the till in the moraine complex is estimated at 4 to 10m/1000y as opposed to an average rate of 1 to 2m/1000y for Emerald Silt, facies A in the basins and further afield from the marginal zone.

Till tongues do not appear to form as a single catastrophic event as

would be the case for a tongue of flowtill. The interdigital relationship between the glaciomarine and till units indicate a progressive development over a period of time (Fig. 26), and the proximal increase in sedimentation rate is also in part responsible for the overall morphology of the till tongue.

In some of the more atypical types of till tongues with irregular bases (Figs. 7b, 7c and 27) the ice conformed to the shape of the seabed at the time of lift-off and maintained its characteristic shape throughout the period when the glaciomarine unit was being deposited. This suggests on a local scale that between the time of lift-off and subsequent grounding, the ice shelf remained pinned and did not experience major changes in subglacial morphology. Further evidence that the ice shelf was pinned arises from regional considerations discussed later.

## MULTIPLE TILL SECTIONS ON THE OUTER SHELF AND THEIR SIGNIFICANCE

Seaward of the inner shelf moraine, till covers the bedrock surface of the shelf to and probably beyond the southern extent of the DTS coverage. On the inner periphery of the outer banks and around the isolated banks within Emerald Basin, a second till (Fig. 27) overlies the continuous blanket and is separated by Emerald Silt of variable thickness. In most respects the relationship of this till to the Emerald Silt is similar to that of the till tongues on the moraine proper. It occurs at the same stratigraphic horizon in the Emerald Silt as does a till tongue 60km to the north on the Sambro Moraine (Cross-section, Fig. 6c). The silt thins towards the shallower central parts of the banks where the tills often coalesce. There is no evidence

for glaciotectonics, and the previously described intimate boundary relationships between the Emerald Silt and till, are maintained. This suggests that the buoyancy of the ice, water depth, and basal melting are important factors in the depositional environment as previously suggested for the till tongues.

Although icebergs and drift ice from a calving front could provide a mechanism for transport of glacial debris across Emerald Basin to the outer banks, such a mechanism could not be utilized to explain the distribution and nature of the contacts between the massive till and Emerald Silt such as have been discussed under till tongues. The consistent and complete stratigraphic succession of the glaciomarine unit, the transitional interbedded contacts between the till and the glaciomarine unit, as well as the wedge-shaped morphology of the till body could not be explained by such a random process as ice rafting. Furthermore, there is no indication of extensive ice-scour phenomena produced by grounded bergs. On the other hand, the concept of subglacial deposition from a pinned ice shelf extending across the entire Scotian Shelf, at times grounded on the banks and floating in the basins, could explain the continuity of glacial deposits observed on both the inner and outer shelves and account for the conformable nature of the lowermost section (facies A) of the Emerald Silt.

The presence of the foraminiferal and ostracod assemblages observed in the piston cores of Emerald Silt (Figs. 17b to 17z, and Fig. 21) introduces certain restrictions to the ice shelf concept and must be taken into consideration. The foraminiferal assemblage grades from a rich and diversified fauna at the top of facies B to a much poorer fauna at the base of facies A,

dominated by Elphidium excavatum. This is indicative of a trend towards colder, less saline, and more shallow water conditions (Vilks and Rashid, 1976). In the underlying facies A, the practically monospecific assemblage of Elphidium excavatum gradually becomes even more sparse at its base. The relative abundance of foraminifera to ostracods fluctuates throughout the column but no pattern was observed.

From studies of Antarctic ice shelves it is often thought that, because of the attenuation of light by the overlying layers of ice, biological activity is restricted. In order to explain the fossil assemblage of a diamicton core sampled through the Ross Ice Shelf (Brady and Martin, 1979) postulated open marine water conditions, at least on a seasonal basis for the time of deposition. On the other hand, Arnaud (1975) believed that a typical Antarctic biota exists under any Antarctic ice shelf because currents are surely present which could carry food. Lipps et al. (1977) conclude that the question of whether or not life can exist far from the open sea beneath the permanent ice shelves of Antarctica is unresolved because of the lack of critical evidence.

Taking into consideration the possible constraints imparted by the faunal data, we suggest the following possible sequence of events leading to the formation of multiple till sections on the outer banks. The occurrence of the lowermost till blanket at or close to the shelf edge shows that ice was probably grounded across the entire shelf and was probably laid down as melt-out debris from the basal zone of an ice sheet. Thinning of the ice sheet during its recessional phase led to the development of a regional grounded ice shelf and the beginning of the major depositional phase. Further thin-

ning resulted in lift-off in the deeper basins accompanied by deposition of Emerald Silt conformable with the underlying till, and in places, bedrock surfaces. Development of a sparse benthic community was initiated at this time. Large areas of the ice shelf remained grounded on the banks, but we have not observed evidence such as glaciotectonics or scouring and removal of section to suggest that the grounded ice was dry- based and formed domes or local active ice sheets at this time. However, this does remain a possibility especially on the larger banks, but our present state of knowledge and seismic coverage is too incomplete to resolve this question at the present time. The floating ice shelf of the basin areas retreated to the basin perimeters with open water conditions in the centre of the basins allowing the development of an abundant faunal assemblage in the glaciomarine sediment and a concomitant deterioration of the conformable style of sedimentation in the upper part of the Emerald Silt.

Further evidence in support of this general framework is dicussed in the following section on depositional environment, and we will also return to the discussion of a regional ice shelf and its control on the development and distribution of "regional, subglacial, ice-shelf moraines" using evidence from a broader area of the continental shelf.

## DEPOSITIONAL ENVIRONMENT OF THE GLACIAL AND GLACIOMARINE SEDIMENT

In our interpretation of the depositional environment and assessment of the Carey-Ahmad model, the interface between the glacial and glaciomarine sediments is thought to have formed in the zone where the wet-base ice shelf became buoyant. The character of the interface varies from large scale in-

terfingering of till (till tongues) and Emerald Silt, to gradational changes expressed as an interpenetration of units, to an abrupt relationship where the contemporaneous units abut and show no evidence of onlap. The rhythmically banded glaciomarine unit (Emerald Silt) may extend for hundreds of kilometres distally from the marginal zone where the glacial and glaciomarine units interface.

From our interpretation of the seismic profiles in the marginal zone (area of the buoyancy line) we infer that the massive till, represented by the seismostratigraphic unit characterized by incoherent reflections, was derived from subglacial melt-out debris from a neutral to negatively buoyant active ice shelf in direct contact with the seabed. Anderson et al. (1980b) from a study of Antarctic sediments involving deposits from grounded ice also reasoned that it is difficult to envision sedimentation of marine glacial till from floating ice even in totally quiescent water.

Seismic interpretation of profiles from the marginal zone and distal to this critical zone, suggests that the glaciomarine unit, represented by the seismostratigraphic unit characterized by repetative continuous coherent reflections with occasional point source reflections (dropstones), was also derived from subglacial melt-out debris. The debris is thought to have settled to the seabed from a floating active ice shelf through a water column of variable thickness to form highly conformable banded deposits which mimic a highly irregular substrate over very broad areas.

A transitional unit representing the gradational areas between the till and glaciomarine units of the marginal zone is characterized by discontinuous coherent reflections. It is inferred to have been laid down by subglacial

melt-out from an active ice shelf in close proximity and periodically in direct contact with the seabed.

Within the marginal zone Carey and Ahmad (1961) postulated the formation of a foreset deposit comprised of till, slumped till or flowtill, interbedded with meltwater outwash silts, and laminar, rhythmic silts and clays. Edwards in Reading (1978) described such a marginal facies association as consisting of random laminae with dropstones, subaqueous outwash, and basal till. We tend to place less emphasis than Carey and Ahmad, and Edwards on sediment dozing action by the ice, slumping and the formation of flowtills and turbidity currents to explain rhythmic bedding in the glaciomarine unit. Although these processes may be important on a local scale we do not recognize large scale glaciotectonics. The conformable character of the banded Emerald Silt over very broad areas of varying relief precludes a turbidite origin.

Sediment Source

In our model of deposition of glacial and glaciomarine deposits, melt-out debris is assumed to have originated from the basal and englacial debris zones of an active ice shelf. This material was presumably entrained through scouring processes by a thick proximal ice sheet moving across present terrestrial and shallow marine areas, and possibly some was derived locally on the bank areas of the shelf especially if ice domes did exist. Although the amount of debris carried by an active ice shelf has been generally assumed to be small as indicated by observations of Antarctic ice by Odell (1952) and Warnke (1970), more recent observations by Anderson _et al_. (1980a) on sediment-laden icebergs suggest that the former observations may have underesti-

mated the quantity of transported debris. Anderson et al. (1980a) recorded basal debris zones up to 15m in thickness on several icebergs calved from the Antarctic ice shelf. Sedimentation rates measured in the Weddell Sea range from 0.01m/1000y on the shelf to 0.02 to 0.07m/1000y for the outer shelf and upper continental slope (Orheim and Elver Løi, 1981).

Modelling studies of the Brunt and Ross ice shelves by Drewry and Cooper (1981) suggest that ice shelves are of major importance for sedimentation on the continental shelf, especially in the grounding-line zone. Due to relatively thin basal debris zones, their model indicates that sedimentation is highly sensitive to melting rate. Strong melting close to the grounding line for instance, may remove all debris from the ice within the first few tens of kilometres of the ice shelf. The effect of basal freezing prior to complete release of sediments is to shift the area of deposition farther along the flow line of the ice shelf. They predicted sedimentation rates as high as 1m/y at the grounding line tapering to zero within a distance of 30 to 50km.

The Antarctic studies indicate a wide variation in the amounts of debris entrained in an ice shelf, some of which far exceed the amounts in the Scotian Shelf ice shelf required to explain till and glaciomarine thicknesses which we describe. The ice shelf covering the Scotian Shelf proper was probably present over a period of approximately 14000 yBP during deposition of facies A (Fig. 24b) before it began to disintegrate during facies B deposition. As previously stated, we estimate average rates of till deposition during the formation of the moraines of 4 to 10m/1000y and short term rates of 20 to 30m/1000y in some of the till tongues. For the Emerald Silt, facies A we recognized a pronounced sedimentational gradient from 1 to 2m/1000y in

the basin to 9 to 10m/1000y in the marginal zone, over a distance of approximately 30km.

The steep sedimentational gradient for the marginal zone suggested by Drewry and Cooper (1981) would lead to a disproportionate distribution of till versus glaciomarine sediment for the Scotian Shelf. Of most significance is the fact that such gradients can be demonstrated beneath both modern and ancient ice shelves. The above observations also provide credibility to the assumption that a subglacial melt-out source from a floating ice shelf can produce sufficient debris to reasonably account for the glaciomarine section on the Scotian Shelf, given sufficient time.

Origin of Rhythmic Bands

Rhythmic bands are best developed and preserved in facies A of the Emerald Silt and thus their discussion will be limited to this facies. The sedimentological periodicity we describe is similar to descriptions of cores for example, from Kaipokok Bay, Labrador (Kontopoulus and Piper, 1982), to the glaciomarine Presumpscot Formation (Bloom, 1963), to the Permian section of Tasmania (Carey and Ahmad, 1961), to accounts of early Paleozoic and Precambrian sections described by Reading (1978), and the Proterozoic Toby Conglomerate of British Columbia (Aalto, 1971). The widespread occurrence and similarity between deposits of all ages suggests that relatively few processes lead to their formation.

From our earlier description we know that the banding is related to the alternate layering of silt and clay (Figs. 17b to z), and is well defined on the x-radiograms. Typically, the individual bands of silt and clay range

from 1 to 3cm thickness with some fluctuation in thickness between the silt and clay for any given couplet. All boundary relationships are sharp and it is quite common to observe pyrite concretions at the top surface of the silt layers. At times the silt layers are laminated and it becomes more difficult to define distinct couplets and obtain a precise count of the number of cycles/m of core.

The above and earlier descriptions appear to be characteristic for periodic deposition from suspension under extreme quiescence. Each couplet appears to represent a short period of sediment influx and fall out, with settling times sufficiently slow to allow sorting of the silt and clay fractions. The uniform seismic character of high amplitude, regularly spaced reflections of facies A throughout the basins, the maintenance of their conformability across irregularities on the substrate, the uniformity of depositional style observed in cores from different areas of the basins, and the lack of evidence for any current activity especially at the sediment water interface, suggest that the periodic influx of sediment originated from a blanket source directly above the depositional interface. All the above characteristics are compatible with the model of deposition by subglacial melt-out from a floating ice shelf, but the cyclical style remains to be explained.

We suspect that the periodicity may be related to short term fluctuations in the ice shelves associated with such deposits, and one likely mechanism is repeated freezing and thawing at the subglacial surface of a floating shelf. Carey and Ahmad (1961) recognized both frozen base and melting ice shelves in their models and Drewry and Cooper (1981) suggested how basal

freezing prior to complete release of sediment from the subglacial surface could control sedimentation rates around grounding-line zones. A delicate balance between freezing and thawing over broad areas of the subglacial surface could give rise to cyclical variations in the release of sediment from the ice shelf. Also, during the freezing stage, time would be provided for regeneration of the sediment load entrained in an active ice shelf. Fallout from a plume generated from underflows at the bouyancy line (Mackiewicz, 1983) may be an alternate explanation; however, this would not explain the widespread distribution of the coarse clasts which are commonly associated with the glaciomarine sediment. Possibly the gravel is a bimodal component introduced directly above from the subglacial surface.

A quantitative estimate of the cyclical period was obtained from a count of the total number of cycles in the composite section of piston cores using the data of Figure 23b and Table IV. Because of the subjective element in identifying some individual cycles, we express the total count as a range between extreme maximum and minimum totals, or 1664 to 2704 cycles for the entire section. Facies A represents approximately 14000 years of sedimentation (Fig. 24b) which gives a cyclical period of 5 to 8 years. A period of this order may be related to sunspot activity.

Origin of Facies A, B and C in Emerald Silt

From the previous discussions of till tongues and the origin of rhythmic bands, much has already been said about the origin of facies A. The interface between the glacial and glaciomarine sections is restricted to this facies, both are subglacial in origin, both are derived from subglacial melt-

out, and the basic difference in mode of origin is their respective formation from grounded and floating ice shelves. It is also important to note that facies A is time transgressive following the retreat of the buoyancy line to the near-shore. For example, most of our observations were made on the Emerald Basin section which was deposited between 46000 and 32000 yBP while younger sections occur in the Gulf of Maine associated with the Fundian Moraine covering the respective period 38500 to 26000 yBP, and the Truxton Moraine approximately 17000 yBP.

We have also suggested that the cyclical sedimentation pattern was formed from a blanket, subglacial, sediment source controlled by repetative freezing and thawing. The depositional gradient across the marginal zone could possibly arise from lateral variations in melt-out rate from a blanket source. On the other hand, the gradient could be influenced to some degree through proximity to local subaqueous sources at the buoyance line; however, we do not envisage a dominating influence by subaqueous outwash. The main reason for differences in sedimentation rate between the glaciomarine sediment and the till may be the fact that the latter was deposited in greater confinement between the ice and the sediment interface and less material was lost by dispersion through plumes in a water column.

In facies B of the Emerald Silt the evidence for deposition by floating ice is still strong as indicated by the presence of dropstones and smaller clasts, and the presence of a cold water but more diversified and abundant fauna. In contrast with facies A, Emerald Silt, the degree of bioturbation increases upward through the section in sympathy with the changes in faunal assemblage. A stronger degree of lamination in the silt bands also develops

indicating a stronger influence of bottom currents and possibly waves. Barrie and Piper (1982) argue for waves not currents to explain similar unconformities from Makkovik Bay, Labrador. Current influence is also indicated by the seismostratigraphic relationships evident in the basins. This is expressed as a trend towards thinning of the section on the local topographic highs and in some areas the development of a pronounced basinal unconformity (Fig. 9). Most of these changes appear to suggest more open water conditions and transformation from a floating ice shelf environment to one dominated by drifting ice; that is, a change from glacial to proglacial conditions. These conditions extended over a period of approximately 16000 y for the main area of the Scotian Shelf.

Facies C, Emerald Silt is regarded as a transitional facies between facies A and glacial till. As indicated earlier it is limited in areal extent to the marginal zone and is generally recognized on the basis of seismic interpretation alone. Only one core was obtained from the facies, core 76-016-6 from the Gulf of Maine. In this core the beds are either strongly laminar or wispy and are often inclined, which may be related to current influenced deposition. We interpret the facies as subglacial which formed during intermittent contact between the ice and seabed, accompanied by strong currents which were probably generated by tidal pumping. Our example occurs late in the history of events on the shelf as the glacial environment receded landward. This possibly accounts for the increase in abundance of foraminifera in core 76-016-6, well above the normal abundance generally found in facies A to which it is most closely related. The occasional occurrence of sand layers in several of the cores could be accounted for by local

environments similar to those suggested for facies C.

LIFT-OFF MORAINES

A discussion of the occurrence and distribution of discrete hummocks at the surface of blanket till deposits was introduced earlier and the features were referred to as lift-off moraines (Figs. 8, 15, 16a and b). Their distribution is shown in Figures 6b to h and 16a and b, and the descriptive details on morphology and scale are discussed under seismostratigraphy. The features are generally buried by Emerald Silt which mimics their morphology, but they are sometimes exposed at the seabed where the sidescan sonograms (Figs. 8 and 15) reveal their true morphology as ridges. A recent sidescan mosaic and seismic survey in the Strait of Belle Isle shows a detailed depositional pattern of lift-off moraines exposed at the seabed (pers. comm. J.Y. Guigné and E.L. King). Their trends, shape, and size are similar to occurrences on the adjacent Newfoundland shore mapped by Grant (1970).

The seismostratigraphic relationships between the till cores of the ridges and the adjacent Emerald Silt show that the Emerald Silt and till are contemporaneous and over very short distances the silt beds grade into and interfinger with the till at the flanks of the ridges. The internal structure of the ridges is generally incoherent and well contrasted against the Emerald Silt, but in some cases the ridges are moderately well stratified and some of the individual reflections match and are continuous with major reflections in the Emerald Silt (Fig. 16b). Unfortunately, there is no sample data to aid our interpretation and provide better information on the sediment texture of the ridges.

The similarities between lift-off and De Geer moraines is striking. However, the latter are defined as annular ridges while the former appear to form simultaneously in groups or fields and probably require a much longer period for their complete evolution. Sugden and John (1976) described De Geer moraines as a succession of discrete, delicate, narrow ridges ranging from short and straight to long and undulating, and sometimes with cross-ribs. The ridges are seldom more than 15m high, and they may be regularly spaced up to 300m apart. They are occasionally steep-sided and are made of variable till with a capping of sub-angular and sub-rounded boulders and are best developed in broad open depressions. Lenses of sand and other stratified water-lain deposits may occur in the ridges, and varved sediments occasionally lie in the intervening depressions, supporting the idea that De Geer moraines developed beneath ice which was grounded in deep water.

Our seismic control over areas of buried lift-off moraines is too sparse to provide a spatial description of the features, but two very detailed surveys from the South Fladen area and the Norwegian Trench of the North Sea provide good examples of the true morphology of the moraines (M. Hoveland and A. Judd, pers. comm.). By correlation between adjacent closely spaced seismic lines they demonstrate the existence of linear furrows and ridges. Their ridges correspond to the features we describe in seismic section as lift-off moraines. Also, the morphological patterns which they mapped are similar in appearance and scale to our descriptions.

Origin

Our opinion regarding the general depositional environment is that the

lift-off moraines are subglacial, recessional features for the following reasons; (1) they always occur at the top surface of the till deposits, (2) fields of the lift-off moraines are remarkably well preserved over thousands of square kilometres and could not have survived regionally during a glacial advance, (3) the ridges were deposited contemporaneously with glaciomarine beds between the ridges and are generally overlain conformably by later glaciomarine beds (Fig. 16b). This typical succession is thought to develop as a grounded ice shelf lifts off the seabed. If the conformable succession of glaciomarine beds above the moraines is thick, the implication is that the ice shelf remained pinned for an extended period after lift-off. If the succession is thin, the shelf probably disintegrated shortly after lift-off and the moraines probably formed closer to the ice-frontal position. The latter category would be more readily observed on land because of the surface morphological expression. Also, the near-surface moraines are probably of more common occurrence on the emerged land areas having formed at a later stage of the recession when the ice shelves were less extensive and the ice-frontal positions were receding at a faster rate. Conversely, the more deeply buried moraines are probably more common in the deeper basins of the continental shelf.

Ideas regarding the mechanism of formation of the lift-off moraines can also be obtained from an interpretation of the seismic profiles. The intimate and highly contrasting sharp relationships between the till and Emerald Silt between the ridges is repetative over large areas and most likely were controlled by very abrupt changes at the interface between the subglacial surface and the seabed. We again suggest that such changes were most readily

achieved through a condition whereby the proximity of the base of the ice shelf to the seabed was the critical factor in determining whether till or Emerald Silt was deposited from the subglacial melt-out debris.

The depositional pattern of the lift-off moraines leads one to think in terms of structural control to form long subparallel ridges. It is difficult however, to conceive of a pushing mechanism whereby the till ridges could develop without some evidence for deformation in the glaciomarine sediments between the ridges. A more plausible explanation may be achieved by imparting a structural control to the ridges through mechanisms involving features in the overlying ice, for example fractures at the base of the ice or other structural features such as keels.

In discussing basal melting and the creation of subglacial melt-out tills, Sugden and John (1976) wrote the following" ..... a special circumstance is the controlled provision of geothermal heat (Mickelson, 1971). This heat is largely responsible for the basal melting process where there is little ice movement and it may cause the very gradual sedimentation of 'ideal' melt-out tills. On the other hand, a new set of special circumstances is encountered which might disturb the till structure. Where a glacier is stagnant, the geothermal heat flux can cause melting upwards through the basal ice layers. As long as the meltwater created can be expelled laterally, the annual layers of melt-out till, as they accumulate, will retain more or less the fabric characteristics of the dirty parent ice. However, this is a delicate equilibrium situation which is all too easily disturbed. If meltwater is not evacuated efficiently, the melt-out till can become saturated and lateral flowage will occur under the pressure of overlying ice. If late-

ral flowage is impeded, then injection features may well be created in the till mass (Hartshorn and Ashley, 1972). If the water accumulates and cannot escape the ice surface will 'lift-off' from a small part of the till surface. This will create a basal cavity, and further particles released from the ice will be dropped onto the till surface, perhaps with the creation of a new fabric (Marcusson, 1973)".

In light of the above discussion it seems possible that fractures in the ice could be created as a result of the basal melting processes and the ice fractures could control the depositional pattern of the till ridges. For example, in a situation where a receding ice shelf is approaching a state of lift-off, till would be accreted at its point of contact with the seabed, and it is assumed that the meltwater would be evacuated through the process of lateral seepage. If for some reason, such as a local increase in the rate of basal melting, the capacity of the discharge system were to be exceeded, then the resulting accumulation of water would cause the ice surface to lift off from a portion of the underlying till surface. Stress fields would be induced within the ice leading to structural failure. Meltwater could escape through the fractures allowing the ice to remain in contact with the seabed and result in till deposition to form linear morainic ridges. Presumably stoping could occur along these crevasses and induce an even greater rate of till deposition. Pinching processes along the fractures could be active and contribute to the ridge forming process. In areas between the fractures the flowage of meltwater could be impeded and allow for the creation of basal cavities in which stratified sediment could be deposited.

Another cause for the formation of multiple, long, straight fractures in

ice is from tidal movements in the ice along the buoyancy line of ice shelves (Carey and Ahmad, 1961). This could allow a time transgressive development of till ridges as lift-off progressed around the periphery of broad depressions during a general recession. Also, seismic activity accompanying local fracturing and faulting of the bedrock caused by glacio-isostatic adjustments during recession (Morner, 1978), may have created fractures in the ice. An important consideration in favor of multiple ice fracturing is that they allow for the simultaneous development of fields of moraines as opposed to an annular development at individual ice fronts.

## REGIONAL, SUBGLACIAL, ICE SHELF MORAINES

The distribution of till in the Laurentian Channel is similar to that of Emerald Basin in that the lowermost part of the till section blankets the bedrock surface while the upper part is less continuous and only occurs along the edge of the channel and over bedrock high areas in the central part of the channel. These relationships result in a thickening of the till in the central part of the channel which we refer to as the Laurentian Moraine, and interpret it as a regional, subglacial, ice shelf moraine. Emerald Silt, contemporaneous with the upper section of till, occurs in the deeper parts of the channel and is interbedded with till at the edge of the moraine and along the edge of the channel at a water depth of approximately 400m. The relationships between the till and Emerald Silt are illustrated in Figure 28, and are similar to those found along the periphery of the outer and central banks on the Scotian Shelf.

Similar occurrences of regional, subglacial, ice shelf moraines and

associated glaciomarine deposits are found on the central Grand Banks in Halibut and Haddock Channels, Whale Deep, and Downing Basin. Figure 6f to h shows that they are all seaward of the inner shelf moraine off Burin Peninsula. Some of these occurrences are outliers and have an unconformity at their surface (Fig. 29). In most other areas of the Grand Banks the entire glacial and glaciomarine section was removed by the Late Wisconsinan- Holocene transgression and only thin sands cover the Tertiary bedrock surface.

The evidence seems to indicate that relatively thick and widespread moraines can be formed beneath ice shelves. The water depth in which these occur requires that the ice shelves were of the order of 200 to 500m thickness and formed on a regional scale during the recessional phase of an ice sheet which extended to the shelf edge. Moraine formation appears to have been controlled to a high degree by basal melting, water depth (which was a function of topographic configuration and glacio-isostatic conditions), and ice thickness. The fact that these moraines occur on a regional scale and within a given basin occur at approximately the same stratigraphic level, suggests that the ice shelves were of relatively uniform thickness over large areas.

Our past thoughts (King et al., 1972) regarding the origin of the principal moraine complex along the inner shelf were that it formed near the margin of a thick ice sheet, possibly along a buoyancy line associated with water depth, and that the till was derived from basal debris accumulating at a more or less stationary ice frontal position. Glaciomarine deposits along both the distal and proximal sides of the moraine were thought to have been laid down later in an onlap relationship with the till. Examination of the

radiocarbon dates and the newly acquired high resolution seismic data, suggests that this is not the case; rather, some of the glaciomarine deposits at the proximal side are contemporaneous with the distal side deposits which are interbedded with the till. In other words, it appears that at least some of these moraines were formed by grounded ice with a floating ice margin at both the distal and proximal sides. On the proximal side the glaciomarine sediments grade to till much more abruptly, and the interbedded relationship generally is not conspicuous except in areas where the slope of the till surface is moderate. The slope of the moraine surface generally is much steeper on the proximal side and the boundary between the till and glaciomarine sediment did not migrate laterally over large distances to produce till tongues and prominent interbedded and gradational relationships as observed on the distal side. This is illustrated in Figure 5 at the 10 to 11km mark along the section where the 50m section of Emerald Silt terminates abruptly against the till on the proximal side of the moraine. Age determination on cores from both sides of the moraine indicate that the distal and proximal sections of Emerald Silt are contemporaneous, and that the densely stratified band through both sections is approximately 26000 years old.

From the foregoing discussion it is suggested that the main Scotian Shelf moraine complex is also a regional, subglacial, ice shelf moraine, its general location having been largely determined by water depth with local control imparted by topographic highs at the incipient stage of formation. As the feature grew in height it was in a sense selfperpetuating, prolonging its period of growth through continued direct contact with the subglacial surface of the ice. As suggested earlier the rate of deposition of the till

was faster than the adjacent glaciomarine sediment, and this would also tend to extend the duration of contact between the ice and the seabed and further enhance growth of the moraine.

If the Scotian Shelf end-moraine complex is indeed a regional, subglacial, ice shelf moraine, the idea constitutes a significant step towards better integration of the glacial history of the offshore and onshore areas. It enables a reasonable correlation of Middle Wisconsinan outlier deposits of glaciomarine origin exposed in coastal sections with the Emerald Silt section offshore, providing a better understanding of the land deposits which in the past have been interpreted as representing interstadial events. Emerald Silt of an equivalent age to the moraines occurs along deeper water reentrants landward of the moraine (Figs. 6c, d and e) bringing it in much closer proximity to the glaciomarine outliers on land. Earlier ideas regarding the moraine as a deposit formed at the frontal position of an ice sheet obstructed attempts to make such correlations.

Other implications of the regional, subglacial, ice shelf moraine concept are that it provides the opportunity to (1) significantly refine our thoughts regarding the nature of marine ice margins, (2) recognize their influence on the distribution and stratigraphic style of the marine deposits which appear to have both the horizontal and vertical components of their stratigraphy well expressed and developed, and (3) recognize their control on the timing and extent of glacio-isostatic adjustments along the continental margin. The significance of these statements will be clarified and elaborated upon in the following sections. It should also be mentioned that regional, subglacial, ice shelf moraine is a descriptive designation, em-

ployed to help embrace and consolidate several concepts; it is not intended as a permanent name.

## GEOLOGICAL HISTORY

In this section we discuss the sequential development of the glacial history of the continental shelf, using our conceptual model to help describe the ice and sedimentational environments. The chronological framework is based on the correlation of the radiocarbon dates with the succession of significant events in Figures 24a to d and summarized in Table V. We will also attempt to integrate the offshore with chronostratigraphy on land, and discuss the major events on land.

### MODEL APPLICATION TO GLACIAL HISTORY

Figure 30 is a diagrammatic representation of the inferred environmental history of the Scotian Shelf at five critical stages of its development. The section is oriented north-south across the type area in the Country Harbour Moraine, Emerald Basin area and extends from the terrestrial zone to the edge of the continental shelf. The major physiographic divisions are indicated at the top of the diagram and are controlled by the shape of the bedrock surface. The small depression in the bedrock surface at the boundary of the inner shelf and Emerald Basin immediately north of the moraine has particular significance in that it represents the westernmost extension of a large reentrant along the south coast of Cape Breton Island and approaches to Chedabucto Bay. Events depicted in the small depression at the various stages of evolution indicate on a small scale the depositional environment in the reentrants north of the moraine. Figure 30, Stage 1 covers a period prior to approximately 50000 to 46000 yBP as indicated by extrapolation of the radiocarbon dates of Figure 24b to the blanket till deposits at the base of

the section. Based on correlations with the general chronostratigraphy on land, as well as with the North Atlantic stratigraphy (Fig. 34), Stage 1 may have begun as early as 70000 yBP. Our seismostratigraphic interpretation indicates that there are no previous glacial deposits older than the Middle and Late Wisconsinan recessional deposits which we have described and which occur as a complete, undisturbed section over many thousands of square kilometres. It seems reasonable to assume that earlier glaciations would have deposited similar recessional successions on the shelf, but the fact that they are not in evidence, except for the possible existence of undiscovered outliers, strongly suggests that such deposits were essentially removed by erosion during the last major glacial advance. Therefore, it is suggested that during Stage I the shelf was covered by a dry-base continental ice sheet which probably extended to the shelf edge. This event up to the time of recession appears to have been dominantly erosional except for the possible deposition of a thin layer of basal till at the base of the section overlying bedrock, and for subglacial deposition at the margin of the ice sheet, at the continental slope. The presence of a thick ice sheet on the shelf was undoubtedly accompanied by extensive isostatic depression of the crust; however, this isostatic influence is not incorporated into the diagrammatic presentation because of the lack of information on its nature and extent. Increasing water depths because of the isostatic influence would have initiated a trend towards thinning of the ice sheet and this effect together with the beginning of glacial recession would have led to the development of wet-based glaciation and lift-off in the deeper areas of the shelf.

The early recessional stage, Stage 2, probably occurred at about 50000 to 45000 yBP (Fig. 24b), and represents the period of deposition of the uppermost part of the blanket till, the development of the lift-off moraines at the top surface of this early till, and the very early glaciomarine sediment between the morainic ridges. The presence of the lift-off moraines and our suggested mechanism for their development leads us to think that the till represents the early depositional phase of the type area and that it is not an earlier till of a previous glaciation of the shelf, although it may include some basal till deposited during the advance. Further evidence comes from the fact that the conformable beds of glaciomarine sediment immediately overlying the lift-off moraines can be traced on the continuous seismic profiles to the distal side of the moraine complex where they are interbedded with the lowermost till of the moraine. At this time of early recessional deposition in the basins of the Scotian Shelf, deposition of glaciomarine sediment probably continued beyond the shelf edge, but the ice remained grounded on the bank areas. Limited biological activity in the basins was initiated during or immediately following this stage.

Stage 3 represents the period between 45000 and 32000 yBP (Fig. 24b) during deposition of the Emerald Silt, facies A section. It should be appreciated that facies A is time transgressive and that the present discussion only pertains to the basins of the Scotian Shelf. During the early part of this stage the recession reached a maximum. The Scotian Shelf moraine complex was beginning to develop as a major feature and remained in contact with the ice shelf in accordance with our model throughout the period. The greater part of Stage 3 was characterized by minor surges and retreats of the

buoyancy line. This oscillation created till tongues along the main moraine, as well as at the periphery of the outer banks. Glaciomarine deposition was continuous in the basins throughout this stage and was characterized by the conformable style and cyclical pattern of sedimentation. Sedimentation rates ranged from 1m/1000y for the Emerald Silt to 4m/1000y for the till with respective rates as high as 10 and 30m/1000y over short periods. The faunal abundance remained at a low but more or less constant level. It appears that the ice shelf maintained its integrity throughout the period, grounded as wet-base ice and depositing till over the banks for at least part of this stage, and as a floating, active ice shelf across the basins. The larger banks may have sustained local ice domes with dry-base ice for periods during this stage accompanied by local erosion on the banks and drawdown to the basins; and this would have contributed to sediment supply in the basins and beyond the continental shelf edge. Landward of the main moraine, the blanket till, lift-off moraines, and Emerald Silt, facies A were being deposited in the Chedabucto reentrant south of Cape Breton Island. The formation of a floating ice shelf in this area was probably enhanced by deeper water conditions associated with isostatic depression by the adjacent ice sheet on land, as well as its proximity to the deeper water of the Laurentian Channel. The northern extent of the deposits in the reentrant is not fully known, but during the latter part of Stage 3 and earlier part of Stage 4 they may have extended on land across some of the low lying areas of Cape Breton Island around the Bras D'Or Lakes, and possibly as far east as the southern area of the Burin Peninsula in Newfoundland. They may be correlatives of isolated glaciomarine and shell bearing till deposits in these areas. These land

deposits include shell bearing tills within multiple till sections in the River Inhabitants area (Grant, 1975), and marine beds interbedded with till on the Burin Peninsula (Tucker and McCann, 1978). The northern extent of Emerald Silt on the inner shelf between the moraine and mainland Nova Scotia is also uncertain because most of this section was eroded during a later transgression. It is assumed that north of the present mainland coastline the ice shelf thickened to sheet proportions and melt-out deposits graded to normal basal till. Farther to the west in the Gulf of Maine-Bay of Fundy reentrant, lift-off probably extended well to the north. Its glaciomarine deposits underlying the Fundian Moraine may correlate with the 38000 yBP glaciomarine deposits (Salmon River section, Grant, 1976; Neilson, 1974) on land. We feel that the dominant aspect of the deposit is glaciomarine and not interglacial and suggest the evidence for warmer conditions was short lived within the reentrant during the glaciomarine interval.

Stage 4 represents a general retreat of the ice shelf from the central shelf to the present land areas, except for one short advance, and covers the approximate period of 32000 to 16000 yBP. It is represented by the Emerald Silt, facies B. Early in this period open water conditions developed on the shelf proper, bottom currents exercised a greater influence in the distribution of sediments and were responsible for the development of a fairly widespread unconformity by approximately 30000 yBP (Fig. 9). An opportunity to examine recently collected seismic profiles (A. Jenkins, pers. comm.) shows this unconformity continues north of the Country Harbour Moraine and along the ancient Country Harbour River Channel to Country Harbour. In several areas, including the north flank of the Country Harbour Moraine, the uncon-

formity is overlain by a thin layer of glacial till which provides the evidence for the return of a grounded ice shelf to parts of the inner and central shelf for a short interval of Stage 4. It is in turn overlain by Emerald Silt, facies B. In facies B, the cyclical pattern of sedimentation gradually deteriorated as a result of increased bioturbation, bottom currents and decreasing continuous floating ice cover. The faunal assemblage progressively increased in diversity and abundance, and planktonic assemblages began to appear. Sedimentation rates in Emerald Basin decreased to approximately 1.1m/1000y. In general the environment of the shelf proper evolved to one with a proglacial aspect, and the evidence for a strong ice influence suggests that the ice did not retreat far inland from the present coastline. Piper (pers. comm.) has recently acquired data from the Lunenburg coastal area to suggest that a morainal system occurs just offshore which appears to be of Late Wisconsinan age (latter part of Stage 4 or Stage 5).

In the Gulf of Maine and Bay of Fundy region events during the first half of Stage 4 appear to be significantly different. A grounded ice shelf returned to this area by 31000 yBP to deposit the Fundian Moraine (31000 to 26500 yBP) and tills to the south of this feature around the periphery of Georges Basin and the northern flank of Northeast Channel off the Browns Bank area. Lift-off then advanced north to Truxton Swell, in the area of column section 12, Figure 6b. Here, till is deposited on beds of Emerald Silt dating about 17000 yBP. No information is available in the immediate area to refine a date for deposition of this till other than to suggest that it probably lies between the 17000 yBP maximum date, and dates on the raised glaciomarine deposits associated with the Presumpscot Formation in the coastal

areas of Maine which date at about 13500 yBP (Stuiver and Borns, 1975), and raised marine ice marginal deposits in central and southern New Brunswick of approximately 13000 yBP (Gadd, 1973). The occurrence of Emerald Silt and lift-off moraines in the area of column section 3, Figure 6, possibly overlies the younger till, but samples were not obtained for dating purposes. The surface of this till has also been modified by iceberg furrows.

During the evolution of the glacial history of the continental shelf to an including Stage 4, the glacio-isostatic response of the crust can only be discussed in general terms because of the lack of direct evidence. At the time when the Wisconsinan ice sheet occupied the entire shelf (Stage 1) isostatic depression presumably reached a maximum, but by the time of Stages 2 and 3 (dominated by the presence of the ice shelf), isostatic rebound was probably well advanced. During the stage of final retreat (Stage 4), possibly as early as 30000 yBP, the crust had more or less fully recovered except for the influence of the remaining ice sheet on the adjacent land areas. A well defined submarine terrace at a depth of 115 to 120m on the Scotian Shelf and 100m on the Grand Banks occurs across large areas. This correlates approximately in time and magnitude with the Late Wisconsinan maximum lowering of eustatic sea level (King, 1970, 1980). Its uniform depth of occurrence over such a broad area strongly suggests that glacio-isostatic rebound had fully recovered and stabilized before the low sea level stand was recorded; otherwise, tilting and differential warping would possibly be in evidence. Exceptions to the consistent depth of the terrace occur near shore south of Cape Breton Island, MacLean _et al._, 1977, nearshore south of Lunenburg (Piper _et al._, 1974) in the eastern Gulf of Maine and approaches to the Bay of Fundy

where ice appears to have delayed rebound until a later date (Fader et al., 1977), and in western Placentia Bay near the Burin Peninsula (Fader et al., 1984).

Stage 5 (16000 to 10000 yBP) covers the period of lowest sea level, its subsequent transgression, and the disappearance of ice influence in the offshore. Evidence for the low sea level stand and subsequent transgression is widespread. The trace of the terrace is indicated by the boundary between the Sambro Sand and Sable Island Sand and Gravel Formations (Fig. 3). This boundary is defined by a textural change between the formations, from sublittoral sands and gravel of the Sambro Sand with a mud content up to about 10 per cent, to clean, well sorted sands and well rounded gravels of the basal transgressive Sable Island Sand and Gravel. In many areas the boundary appears as a knickpoint on the echograms and seismic profiles, and it coincides with a truncation of Emerald Silt beds around the periphery of the banks. For example, this unconformity is illustrated on the seismic profiles (Figs. 12, 13 and 29) across the type area for the Wisconsinan section on the Scotian Shelf, and in fact the erosional surface was responsible for the exposure of the early part of the section. Another example of the terrace is shown in Figure 31 where it is cut into Tertiary bedrock and overlain by up to 6m of Sable Island Sand and Gravel.

As Figure 3 shows, large areas of the shelf were exposed subaerially and there is some evidence to suggest that the geotechnical properties of the sediment were modified during subaerial exposure (Table VI), possibly as a result of desiccation and to some degree removal of overlying sediment by erosion. Piston cores in Emerald Silt above the old shoreline penetrate the

Emerald Silt for about 1 to 1.5m, but below the shoreline and in more or less the same stratigraphic section a penetration of approximately 10 to 15m can be achieved (Fig. 32).

The terrace indicates the position of the lowest sea level and beginning of the Late Wisconsinan-Holocene transgression during which time all sediments between the old and present shoreline were modified in response to a high energy beach environment. As the transgression progressed across the outer banks it produced large areas of clean, well sorted sands and gravels (Sable Island Sand and Gravel Formation). The fines were deposited in the basins as LaHave Clay. On the inner shelf, bedrock is exposed or nearly exposed at the seabed, but these areas were originally covered with till and glaciomarine deposits. During the transgression these deposits were undercut and eroded by the advancing sea and essentially all that remains is coarse gravel confined to depressions on the bedrock surface, and a few thin outliers of till and Emerald Silt as indicated on Figure 6 by the column sections and on the geological cross-sections. The slope of the inner shelf was sufficiently steep to allow undercutting and slumping to dominate and destroy the continuity between the glacial and glaciomarine deposits of the offshore with the glacial deposits on land.

The age of the submarine terrace can be bracketed by our data on the age of the uppermost truncated beds of Emerald Silt, facies B, and the age of the overlying basal beds of LaHave Clay (Fig. 13 at the 1km mark). Within these constraints we suggest an age between 15100 and 14465 yBP. Our previous estimates of the age of the terrace were based on sea level curves of Milliman and Emery (1968) which showed the lowest stand of 125m at 15000 yBP.

This is in good agreement with the above age of 15100 to 14465 yBP at 115 to 120m.

On the Grand Banks, the Late Wisconsinan-Holocene transgression was effective in removing the Pleistocene deposits on a regional scale leaving only outliers in a few small basins and channels (Figs. 6f to h, 29 and 33). Consequently, the glacial history is fragmented, but at least it can be inferred from their scattered positions on the Banks that the entire area was glaciated. One of the more complete sections is in Placentia Bay and approaches, and these deposits are typical of those observed on the Scotian Shelf, including sections of till, glaciomarine sediment, a regional moraine, lift-off moraines, till tongues, and blanketing Holocene deposits. Data from Figures 23d and 24d and Table V indicate that the Placentia Bay deposits and events are generally younger than those of the Scotian Shelf, and correlate better with late Middle Wisconsinan events which include the Fundian Moraine and associated deposits in the eastern Gulf of Maine. This is probably because of its close proximity to the land, and we assume that the outliers farther offshore are older and comparable to the deposits of the central Scotian Shelf.

One of the last vestiges of ice influence on the offshore sediments is revealed by the occurrence of iceberg furrows. They are formed by the scouring action of grounded icebergs moving across the seabed under the influence of winds and currents. Both modern and relict iceberg furrows are of common occurrence along the entire eastern Canadian shelf and have been documented in papers by Harris and Jollymore (1974), Van der Linden _et al_. (1976), King (1976), Fader and King (1981) and Lewis and Barrie (1981). Their dimensions

are highly variable but typically they are approximately 2 to 5m deep, 30 to 40m wide, and can be many kilometres in length. It is difficult to distinguish between relict and modern furrows because they both can display a fresh well preserved appearance. Through regional considerations concerning the distribution of ice conclusions can be drawn regarding their age. For example, most iceberg furrows on the western Scotian Shelf (Fig. 6) must be relict because icebergs have been absent since the continental ice sheet withdrew. Almost all furrows on the Scotian Shelf occur on glacial till. Farther to the east, for example south of Newfoundland, furrows occur on the younger LaHave Clay (Fader et al., 1984). These were probably formed by calving bergs from the late ice (12500 yBP) of southwest Newfoundland described by Brooke (1974). The major local sources for icebergs in the Gulf of St. Lawrence had probably disappeared by 12000 yBP. Across the Grand Banks east of Newfoundland the seabed has been affected by iceberg scour throughout the Pleistocene and to the present, and below depths of 115 to 120m furrows covering the complete spectrum of ages may occur (Fader and King, 1981). Above the 115 to 120m terrace the furrows post-date the Late Wisconsinan-Holocene transgression. Any earlier furrows having been removed by erosion.

Unfortunately, our seismic control does not cover the areas where glacial and glaciomarine deposits would have been deposited in the offshore by the late ice of Newfoundland, but presumably such deposits do occur in the approaches to the fjords and in the nearshore areas around the northern and eastern areas of the Gulf of St. Lawrence.

## REGIONAL STRATIGRAPHIC INTERPRETATION

Figure 34 shows a correlation between our synthesis of glacial fluctuations of the Scotian Shelf and Grand Banks with the post-Sangamon chronostratigraphic framework developed by Grant (1977, 1980), Prest and Grant (1969) and Dreimanis et al. (1981) for the land areas of the Atlantic Provinces; Alam, Piper and Cooke (1984) for areas on the continental slope and top of nearby seamounts; Dreimanis (1975) for southern Ontario and southwest Quebec; and Ruddiman and McIntyre (1981) for the west North Atlantic oceanographic record.

The best section of pre-Sangamon stratigraphy occurs on the Fogo Seamounts, close to the continental shelf and west of the Grand Banks (Alam and Piper, 1977; Alam et al., 1984), and according to Prest (1977) other sections are sparse throughout the rest of the area. Cores from the seamounts contain sequences of alternating clays and foram nanno ooze representing glacial and warmer periods respectively, back to the mid- Pleistocene. An early Illinoian glaciation with a source from across Newfoundland and the Grand Banks appears to have been the most severe. This was followed by a strong late Illinoian advance accompanied by a major influx of red sediment indicating significant erosion of the Carboniferous redbeds of the Gulf of St. Lawrence and Laurentian Channel. Alam and Piper (1977) suggested that the Wisconsinan glacial stages were much weaker and did not extend to the shelf edge, whereas our data show that at least one Early to Middle Wisconsinan advance reached the shelf edge. It appears that the Fogo Seamount data provides the best information on the relative magnitudes of the various Pleistocene stades for southeast Atlantic Canada, but the shelf data help calibrate the seamount section and provides more definitive data on the Middle and Late Wisconsinan

section.

In addition the seamount data provide useful evidence for interpreting the chronology of the erosional history of the area. The Laurentian Channel and its tributaries are the most prominent erosional landforms and the seamount data indicate a Late Illinoian age for a major erosional phase. The morphological evidence (Loring and Nota, 1973, and King and MacLean, 1976) suggests that the channels were formed by ice streams along a former drainage pattern. Overdeepened rock basins along the main channel or trough are associated with the confluence of hanging valley, tributary systems with heads in the deep fjords along the south coast of Newfoundland. Some additional erosion apparently occurred during the Early to Middle Wisconsinan advance because they were sufficiently strong to ground and remove any former glacial and glaciomarine deposits. The only remaining deposits are recessional which formed during the retreat of the Wisconsinan ice.

The section on the continental shelf (Figs. 30 and 34) begins with a continuous blanket of till overlying bedrock with lift-off moraines at the surface of the till. This section is best developed in the basin areas, and we believe that it represents the base of a continuous section of recessional glacial and glaciomarine deposits arising from subglacial melt-out from beneath an ice shelf. Remnants of early glacial deposits may occur in some local depressions, but we are unable to identify them as such on the basis of our present knowledge. We believe that the recessional event was preceded by the advance of an ice sheet which extended to the edge of the continental shelf. Extrapolation of our earliest radiocarbon date, 41000 yBP (Fig. 24b), which is near the base of the section would suggest that the recession was

well advanced by approximately 50000 yBP, and the ice sheet is presumed to have wasted to an ice shelf configuration at that time. It is also probable that the ice sheet postdated the St. Pierre interstade because this is thought by some to have been a strong event across southern Ontario and Quebec, and is often associated with organic deposits. The St. Pierre interstade at approximately 75000 yBP is also defined by the oceanographic record and can be recognized in the data of Alam, Piper and Cooke (Unit B6, 1984), so it seems reasonable to assume that an ice sheet would not have been sustained on the continental shelf during this interstade. Therefore, we postulate that the Scotian Shelf - Grand Banks advance followed the St. Pierre interstade and correlates with the Guildwood stade of southern Ontario. Ice cover was continuous in Quebec following St. Pierre time, and the Guildwood stade and Scotian Shelf - Grand Banks advance appear to be marginal fluctuations from this body of ice to the north. An earlier Wisconsinan stade is suggested by Grant in Dreimanis _et al_. (1981) and designated the Fundy and Cabot stades, as well as by Alam, Piper and Cooke (1984) as their B7 stade, but evidence for this event on the continental shelf, if indeed it did exist, appears to have been removed by erosion. Grant (pers. comm.), has later suggested that the last interglacial includes the St. Pierre and that the Early Wisconsinan between St. Pierre and Sangamon time was not represented by a major ice sheet in the Atlantic Provinces. Rather, the period was to some degree a cooler extension to the Sangamon interglacial and includes all the organic deposits of the Maritime Provinces. The North Atlantic oxygen isotope data (Ruddiman and McIntyre, 1981) show intermediate values during this period which could corroborate such an alternate hypothesis.

At approximately 50000 yBP the Scotian Shelf - Grand Banks ice sheet had degenerated to ice shelf thicknesses, and was beginning to lift off in the basin areas. As discussed earlier this condition persisted throughout the greater part of the Middle Wisconsinan with minor fluctuations at the periphery of the banks between the grounded and floating portions of the ice shelf. We refer to this major event as the Scotian Shelf - Grand Banks recession. It was a gradual recession, at time approaching a stillstand, with minor oscillations throughout the period from approximately 45000 to 32000 yBP. This condition was probably maintained by a thick ice sheet over much of the adjacent land areas, terminating somewhere on the inner continental shelf. The recession on the continental shelf appears to be compatible with the prolonged Port Talbot and Plum Point interstades of southern Ontario which were marginal to an area of continuous glaciation across Quebec. The recession also correlates rather well with the warmer units B2 and B4 of Alam, Piper and Cooke (1984) and with the decrease in global ice volume predicted by the western North Atlantic oceanographic record (Ruddiman and McIntyre, 1981). Also, Sirkin and Stuckenrath (1980) reported a warm interval indicated by sediments near the moraines of Long Island and Block Island which contain peat and oyster beds and range in age from 43800 to 21750 yBP.

At times during this recession floating ice shelf conditions, possibly accompanied by open water conditions, may have extended along major reentrants to several areas of the mainland where Middle Wisconsinan, glaciomarine sediments and shell bearing tills are known to occur. Grant (1980) established the Clare interstade on the basis of these deposits.

In summary, the stratigraphy for the late Early Wisconsinan and Middle Wisconsinan on the continental shelf comprises the Scotian Shelf - Grand Banks advance representing a major ice sheet from the north carrying debris off Nova Scotia and Newfoundland to the continental shelf, followed by a long period of gradual recession (Scotian Shelf - Grand Banks recession) grading into late Middle and Late Wisconsinan time.

This interpretation of the stratigraphy is somewhat in contrast with that suggested by Grant (in Dreimanis et al., 1981) and Grant (1977) for the early part of the Middle Wisconsinan. He postulated the Isle Royale and Burin stades of Cape Breton and southern Newfoundland as the major events of this period, and on the basis of striae and other evidence indicated a direction of northern flow for the ice, suggesting ice domes on the continental shelf as a source for these advances. We see no evidence for the continental shelf acting as a primary source area for the northerly flow, and suggest as a possible explanation that the northerly flow was caused by drawdown of some shelf and Cape Breton ice towards the Gulf of St. Lawrence and Laurentian Channel during the early stages of recession (Stage 1, Fig. 30). Areas of the seabed between the postulated domes and Cape Breton Island are underlain by Stage 3 and subsequent deposits and would have been disrupted and eroded by a later northerly movement of ice from the offshore.

Tucker and McCann (1980) in their study of Quaternary events on the Burin Peninsula and Islands of St. Pierre and Miquelon concluded as follows from the available stratigraphic evidence; (1) the earliest event was pre-Wisconsinan or perhaps Early Wisconsinan in age and is represented by indurated, weathered till deposited from an all encompassing glaciation of New-

foundland-centered ice from the north, (2) Newfoundland-centered ice again advanced in Early Wisconsinan time, depositing the lower unit of the multiple till section, (3) marine overlap occurred depositing fossiliferous sands and silts which are correlated with the Salmon River deposits of Nova Scotia (38000 yBP) on the basis of their fossil content, (4) partial glaciation by ice from an offshore source deposited the upper unit of the multiple till section, (5) limited glaciation by Late Wisconsinan, Newfoundland-centered ice. Their major glacial advance (event 2) probably corresponds to the Scotian Shelf - Grand Banks advance and appears to be compatible in magnitude and direction, but we believe that an early Middle Wisconsinan date is more likely. Their analysis provides further documentation on the northerly pattern of flow for a glacial advance (event 4) and dates the event as late Middle Wisconsinan (post Clare interstade) whereas Grant (in Dreimanis et al., 1981) suggested a pre-Clare date and correlated it with the Burin stade. According to our evidence the late Middle Wisconsinan date weakens even more the argument for an offshore source to explain the northerly direction of ice flow, because at that time the grounded ice shelf was beginning to lift off the banks and complete its withdrawal from the shelf. It could, however, still be explained by drawdown of local ice towards Fortune Bay.

Based on our present knowledge of the distribution and age of glacial till on the shelf, it appears that grounded portions of the ice shelf in late Middle Wisconsinan time (approximately 30000 yBP) had withdrawn to the coastal areas except in the eastern Gulf of Maine. Here, there appears to have been a strong resurgence to form the Fundian Moraine and associated deposits of 30000 to 26000 yBP. This resurgence may have been a forerunner of the

Late Wisconsinan advance of approximately 20000 yBP recorded in the Cape Cod - Long Island area and southern Ontario, but not expressed elsewhere in the offshore except in the coastal areas. North of the Fundian Moraine the age of the youngest till (Truxton Swell) lies between approximately 17000 and 13000 yBP. Deposition of the ice related sediment in Emerald Basin was continuous (Emerald Silt, facies B, proglacial) and was present in the offshore section until approximately 15000 yBP. At this time a faunal discontinuity occurs where older benthic foraminiferal assemblages dominated by Elphidium excavatum change to a more diverse present-day assemblage (Vilks, 1980), and this break coincides approximately with the boundary between the Emerald Silt and overlying LaHave Clay. This boundary also shows time transgressive aspects as indicated in Table 5 for the various regions.

The offshore Late Wisconsinan events appear to be compatible with estimates on the magnitude of the Scotia Stade on the mainland. Prest and Grant (1969) envisaged ice flow from local independent upland sources, and thought that the upland sources were remnants of a former more extensive ice sheet which became isolated mainly as a result of incursions by the sea. They further concluded that the Late Wisconsinan ice sheet was not as active in the Maritimes as it was toward the continental interior. Grant (1977) further elaborated on limits of Late Wisconsinan ice. Late Wisconsinan glaciers springing from numerous local ice caps extended to and only slightly beyond the present coast and landward of the Scotian Shelf Moraine complex. He stated that the Late Wisconsinan ice failed to reach the outer Fundy coast at all points, apart from a possible ice shelf in the Gulf of Maine. We corroborate this view in finding evidence for a grounded ice shelf in the

central and eastern portion of the Gulf of Maine, but no evidence for thicker ice. He further suggested the possibility of an ice shelf in the Gulf of St. Lawrence; we intuitatively agree on the basis of our present knowledge in other reentrants, but our data base does not extend into the Gulf. A seismostratigraphic investigation of the Gulf of St. Lawrence would elucidate many of the problems associated with the nature and extent of ice cover in the Atlantic Provinces during Late Wisconsinan time and possibly qualify much of our thinking on earlier Wisconsinan problems as well. It is of interest to note that Prest (1957) described glaciomarine sediments intercalated with waterlain till on the Magdalen Islands of uncertain age.

In the Long Island - Cape Cod area the evidence for a strong Late Wisconsinan advance is well documented (Schafer, 1961; Kaye, 1964; Schafer and Hartshorn, 1965; Borns, 1973; Sirkin, 1976; Oldale, 1976 and Sirkin and Stuckenrath, 1980). The ice advanced through southern New England approximately 20000 yBP and may have been close to its maximum position as late as 15300 yBP or possibly earlier. In coastal and eastern Maine the glacial margin is characterized by recessional marine deposits (Borns, 1973 and Smith 1981) between 13500 and 12500 yBP. These deposits include the Presumpscot glaciomarine sediments (probably equivalent to Emerald Silt) and numerous small moraines which could include lift-off moraines. According to Smith (1981), marine silt and clay commonly overlie till but locally are interbedded with deposits of waterlain till. Small end moraines are overlain and underlain by the Presumpscot Formation, and distal deposits of large end moraines intertongue with the marine sediments.

The glacial style of the Maine deposits is surprisingly similar to what

we have described for the basins of the continental shelf and the eastern Gulf of Maine, except for a minor advance at 12700 yBP which formed the Pineo Ridge Moraines. This readvance may have been caused by an increase in calving rate along coastal Maine rather than by a general climatic change (Borns, 1973 and Borns and Hughes, 1977). Similarly, the Kennebunk readvance appears to have been one of many local fluctuations on the ice front during general recession (Smith, 1981). In general, the glacial deposits of coastal Maine appear to be part of a long time-transgressive, recessional succession beginning on the continental shelf off Nova Scotia at about 50000 yBP and ending in Maine at about 12500 yBP.

The boundary between the Maine recessional domain and the terrestrial, terminal deposits of southern New England is not well established. We had initially thought (King et al., 1972) that the Scotian Shelf Moraine complex correlated with the Cape Cod moraines, but we now realize that they differ widely in mode of origin and that the Scotian Shelf Moraine more closely resembles some of the Maine deposits except for the difference in age. By applying our depositional model to the piston core and seismic data of Tucholke and Hollister (1973) from Stellwagen Basin, we would suggest that their youngest till lies at the base of the section below that sampled by the piston core, and probably represents deposition from the Cape Cod Bay and Great South Channel lobes of the Late Wisconsinan advance. There seems to be no evidence from the seismic profiles to suggest that ice grounded subsequently in the basin; although, an ice shelf may have grounded on the adjacent banks when the glaciomarine section represented by their piston core was being deposited. Their oldest date of 18900 yBP in glaciomarine sediment

appears to be old in relation to the best known dates for the Cape Cod Bay and Great South Channel advances. Offshore Cape Cod is indeed a critical area for investigating the problem of how a typical terrestrial glacial environment interfaces with a glaciomarine environment over what appears to be a relatively short distance.

Like southern New England, southern Ontario also experienced a strong Late Wisconsinan glacial advance, but as we have indicated this advance was weak in the marine areas and adjacent land areas of the Atlantic Provinces. Furthermore, this trend appears to continue north on the Labrador Shelf. Cores from a basin in Cartwright Saddle (Vilks and Mudie, 1978) indicate the presence of a sedge-tundra environment near the basin as early as 21000 yBP, and they suggest that the glacial limit of Late Wisconsinan time may have been close to the shore leaving some of the headlands and islands exposed where vegetation could have established. If their core was in typical glaciomarine sediment it is possible that the pollen was transported by the open ocean and deposited beneath an ice shelf, and that the ice retreat was not at such an advanced stage as they suggest; nevertheless, the presence of an ice shelf would suggest that at least a recessional trend was well established at 21000 yBP. Seismostratigraphic studies on the Labrador Shelf by Grant (1972), Van der Linden _et al_. (1976), Fillon (1976) outline the distribution of glacial sediments, and MacLean and Josenhans (pers. comm.) have recognized till tongues and a glaciomarine succession similar to what we have described on the Scotian Shelf. The chronology of these deposits is not yet known; however, the results of Vilks and Mudie (1978) seem to suggest that most of the till deposits are older than Late Wisconsinan.

This general trend of strong Early to Middle Wisconsinan events tapering to weaker events for the Late Wisconsinan which appears to be characteristic for the Atlantic Provinces and the offshore, is similar to what Andrews et al. (1972), Miller et al. (1977) and Andrews and Barry (1978) have shown for parts of the east Canadian Arctic where the Early Wisconsinan advance was stronger than subsequent advances.

Ives (1978) in discussing the extent of the Laurentide Ice Sheet in eastern and northern North America reviewed the evolution of ideas which have led to the development of two opposing schools of thought regarding the terminal position of Wisconsinan ice. He designated the extreme views as "the maximum Wisconsinan viewpoint" supported by proponents advocating an ice sheet extending to terminal positions on the continental shelf as emphasized by Denton and Hughes (1981), as opposed to "the minimum Wisconsinan viewpoint" adopted by those supporting a more limited ice cover. To some degree the extreme views are a function of terminology and the rather vague meaning associated with the expression, "last ice", as well as a lack of knowledge of the nature, age and extent of glacial deposits on the continental shelf. We think that our synthesis provides a base for resolving at least some of the inherent difficulties of the debate and that it reveals a strong element of truth in both viewpoints. We consider the "last ice" to include the Scotian Shelf - Grand Banks advance of Early to Middle Wisconsinan age and feel that this major event which terminated at the shelf edge would satisfy the arguments in favor of the maximum Wisconsinan viewpoint. On the other hand, we interpret the Late Wisconsinan advance as a resurgance related to a long general recession of the Scotian Shelf - Grand Banks advance and this could

satisfy the minimum viewpoint, recognizing of course that the Late Wisconsinan event strengthened in the New England and central parts of the continent.

Differences of opinion are also aggravated by the difficulty of precisely defining ice margins, and more emphasis should be placed on a distinction between the terminus of ice sheets and adjacent ice shelves when mapping ice margins. This paper emphasizes the fact that ice margins when dominated by ice shelves result in broad regional synchronous deposits, not characteristic of sharp terminal events. Major glaciotectonic features and the limits of strong isostatic effects are probably more useful in defining ice sheet limits.

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## Figure Captions

Figure 1a and b

Index maps for the study area showing the regional distribution of Huntec deep-towed seismic reflection profiles used in this study. Additional airgun seismic reflection, echogram and sample control for the interpretation of the surficial formations (Fig. 3) is not indicated.

Figure 2

Diagrammatic cross-section of the Scotian Shelf through Emerald Basin showing the distribution and relationships among surficial formations, bedrock, and the last low sea level position (King, 1980).

Figure 3

Distribution of surficial formations from the Bay of Fundy, Gulf of Maine, Scotian Shelf, Laurentian Channel and the western Grand Banks of Newfoundland.

Figure 4

The Scotian Shelf end-moraine complex and the southern boundaries of known till and significant gravel occurrences (areas with more than 10% gravel) on the Scotian Shelf and Georges Bank area (King _et al_., 1972). The moraines are now interpreted as regional, subglacial, ice shelf moraines. See text for explanation of their formation.

Figure 5

Huntec DTS (A), sidescan sonogram (B), enlarged sections of the Huntec DTS profile (C and D), and geological interpretation (E) across the Fundian Moraine, Gulf of Maine. Facies A, Emerald Silt Formation is

interbedded with till of the moraine and extends under the moraine for a distance of 8km, where it pinches out against the bedrock surface. Near the top of the moraine, facies A grades to facies C, Emerald Silt with an acoustic character of continuous-discontinuous coherent reflections over a distance of 1.5km. The surface of the till of the moraine is covered with iceberg furrows up to 5m in depth. The wedge shaped body of till interpenetrating facies A Emerald Silt, is referred to as a till tongue. The positions of the piston cores are indicated and the type section for the Gulf of Maine area is located 1.1km south of core 76-016-3. The morphology and position of the bedrock surface was interpreted from an adjacent airgun seismic reflection profile.

Figure 6a

Legend for the column section maps Figures 6b-6h from the Gulf of Maine to the Grand Banks of Newfoundland. The column sections are interpretations of Huntec DTS profiles chosen at locations to best represent the regional surficial stratigraphy as well as critical areas of geological significance. The numbers at the top of the column sections are referenced along the Huntec DTS survey tracks. Where piston cores were collected at column section sites, the core numbers appear directly below the column section. The bedrock is not subdivided for this presentation.

Figure 6b

Column section map for the eastern Gulf of Maine area. The Fundian Moraine forms Sewell Ridge on the northern flank of Georges Basin, cross-section A-B. The Truxton Moraine occurs 40km to the north of

Sewell Ridge on Truxton Swell. Cross-section A-B at its northern end, and cross-section C-D near D, both illustrate the distal till tongue development of the Truxton Moraine and its relationship with the surficial formations. Lift-off moraines are widespread in the basinal areas of the Gulf of Maine and outcrop at the entrance to the Bay of Fundy near column section 2. The thickest deposits of till occur in the Fundian Moraine and at the approaches to the Bay of Fundy. Facies C, Emerald Silt occurs south of Truxton Swell, column section 16 and cross-section A-B. Note the thin surficial cover over Browns Bank, and in the nearshore zone off southwest Nova Scotia, column sections 11, 12, 20, 21 and 22.

Figure 6c

Column section map of the western half of the Scotian Shelf. Study area A, the type section for the Scotian Shelf is shown in Figure 6d. Lift-off moraines occur across Roseway, LaHave and Emerald basins. Cross-section E-F-G, which extends continuously from the inner shelf across Emerald Basin to the northern edge of Western Bank, shows till tongues of the Sambro and Halifax Moraines interbedded with Emerald Silt facies A at the same seismostratigraphic horizon as the till tongues on the northern edge of Western Bank. The thickest deposits of till occur in these moraines. Note the presence of thin basal transgressive sand and gravel deposits (Sable Island Sand and Gravel Formation) above 100m water depth, column sections 27, 31, 37, 40 and 34. Large dropstones, recognized acoustically, occur within facies A Emerald Silt only in LaHave Basin, section 32 and 33. Note the regional occurrence of till

tongues which are found on the inner shelf, rim the basins of the central shelf and occur on the northern edges of the outer banks.

Figure 6d

Column section map of the type section area Scotian Shelf, shown as area A (Fig. 6c). In this diagram all the column sections are seismostratigraphic interpretations at core locations. The till deposits are all interbedded with facies A Emerald Silt Formation. With the exception of section 61, which shows relief of up to 4m on the lower till tongue surface, all other till tongues across the Scotian Shelf exhibit nearly flat and smooth, lower and upper surfaces.

Figure 6e

Column section map for the inner part of the eastern Scotian Shelf. Note the presence of lift-off moraines in the basins and till tongues on the basin flanks.

Figure 6f

Column section map for the western Grand Banks of Newfoundland and the Laurentian Channel. Column section 76 is from the distal side of the Laurentian Moraine, a prominent topographic feature 30m in height, which projects across the floor of the Laurentian Channel. The till of the moraine is interbedded at the same seismostratigraphic horizon with the Emerald Silt as is the till on the flank of the channel, section 74. The relationship is also shown in Fig. 28. Placentia Bay represents one of the few areas on the Grand Banks of Newfoundland where all five surficial formations occur. Most glacial sediments were removed from the bank areas during the Late Pleistocene-Holocene transgression. Iceberg

furrows occur on LaHave Clay, section 77, north of Burgeo Bank and were probably formed by late glacial ice originating from the fiords of the south coast of Newfoundland.

Figure 6g

Column section map for the eastern Grand Banks of Newfoundland. The thickest surficial sediments occur in the Avalon Channel, the Downing Basin area and north of Conception Bay. Over most of the Grand Banks of Newfoundland in depths shallower than 100m the sediments are thin, generally less than 10m, cross-section J-K.

Figure 6h

Column section map for the northern Grand Banks - south Northeast Newfoundland Shelf area. The thickest glacial deposits occur in Conception Bay, off the mouth of Trinity Bay and on the outer continental shelf. Till is continuous across the shelf from Trinity Bay to the shelf edge, and its surface is covered with iceberg furrows.

Figure 7a

Known distribution of till tongues from the Gulf of Maine, Scotian Shelf and Grand Banks of Newfoundland based on an interpretation of Huntec high resolution seismic reflection profiles. The arrows point in the direction of the thinning feather edge of the till tongues interbedded with the glaciomarine sediment.

Figure 7b and 7c

A suite of till tongues as they appear on Huntec DTS profiles. See text for detailed description.

Figure 8

Sidescan sonogram A, Huntec DTS profile B, and interpretation of Huntec DTS profile C, across an area of lift-off moraines from the northwestern Grand Banks of Newfoundland. The sidescan sonogram A shows that the features are linear ridges and not isolated mounds or hummocks. The lift-off moraines range in height from 8-10m and are spaced from 0.1 to 0.25km apart. Note the consistant character of the Emerald Silt reflections throughout the field of lift-off moraines.

Figure 9

Huntec DTS profile from Emerald Basin, Scotian Shelf, showing the stratigraphic relationships among bedrock, glacial till, Emerald Silt facies A and B and LaHave Clay. A basinal unconformity occurs near the top of facies A Emerald Silt and is expressed at the 1.5km mark as a truncation of continuous coherent reflections. In many areas it becomes disconformable. Note the occurrence of a moat surrounding the bedrock high. These features are widespread and are believed to be formed by a combination of lower sedimentation rates and erosion associated with bottom currents. Note the interpreted seismic facies are continued to the left on the diagram.

Figure 10

Huntec DTS profile of facies C Emerald Silt formation from the Gulf of Maine south of Truxton Swell. Facies C is called the transitional facies as it displays both the acoustic character of till and Emerald Silt. It consists of areas of incoherent reflections mixed with zones of discontinuous coherent reflections. Note that in slightly deeper water to the south and north along the profile, at the 1 and 5km marks

the facies is an equivalent of facies A Emerald Silt. The facies was formed during intermittent contact between the ice and seabed.

Figure 11

Interpretation of airgun seismic and Huntec DTS profiles north to south, across northeastern Emerald Basin (Fig. 6d). Till tongues 2-9 are interbedded with facies A Emerald Silt. Till tongue 4 does not occur along this profile and the position indicated only represents its stratigraphic horizon. This part of the County Harbour Moraine (Middle Bank) is composed entirely of glacial materials and is not located on a bedrock high. Note the continuous deposits of Emerald Silt above till tongue 3, which extend beneath the entire moraine across the section. This area of the shelf contains the thickest section of glacial sediments and the broadest development of till tongues across the entire Scotian Shelf and Grand Banks of Newfoundland.

Figure 12

A Huntec DTS profile along Figure 11 from the 14 to 40km marks (Fig. 6d). A wide variation in till tongue shapes occurs across this section. The upper and lower surfaces of the till tongues are generally smooth and flat with the exception of till tongue 7 from the 8-10km mark, which exhibits 2m irregularities on its upper surface. These relationships indicate contemporaneous deposition of till and silt.

Figure 13

Huntec DTS profile normal to the Huntec DTS profile in Figure 12 (Fig. 6d) to help develop a three dimensional picture of study area A. Note the well developed unconformity across the Emerald Silt formation as

evidenced by truncation of reflections at the seabed and the presence of till tongues 7, 5, 2 and 4. Till tongue 7 is the only one sampled in this study, (Core 79-011-8). At the 24.5km mark along the profile a 40m section of Emerald Silt abruptly terminates against a deposit of till - Scotian Shelf Drift.

Figure 14

Map of study area A, northeast Emerald Basin showing the distribution of each of 8 till tongues. The hachured areas are zones where till tongues have been eroded at the seabed. The remaining till tongues are rooted at depth to the main moraine complex (see cross-section Figure 11). Till tongues 2-6 are rooted or eroded in the north with the feather edge thinning to the south. Till tongues 7-9 are rooted or eroded in the east at the edge of Middle Bank and the till tongues thin in a westerly direction.

Figure 15

A 70kHz sidescan sonogram across an area of iceberg furrowed lift-off moraines from the northwestern Grand Banks of Newfoundland. The occurrence of the lift-off moraines was confirmed by Huntec DTS data. Note the large iceberg furrows in the bottom sonogram cutting across the field of lift-off moraines.

Figure 16a

Map of the known distribution of lift-off moraines from the Gulf of Maine, Scotian Shelf and the Grand Banks of Newfoundland. The lift-off moraines were interpreted from Huntec DTS data and a small amount of sidescan information. Exposed lift-off moraines at the seabed are limi-

ted in their area of occurrence to the Gulf of Maine and the northwest Grand Banks of Newfoundland.

Figure 16b

Huntec DTS profiles across lift-off moraines showing a wide variety of shapes, sizes and relationships with the adjacent and overlying sediments.

Figure 17a

Index for core logs.

Figure 17aa

X-radiographic positives showing rhythmically banded (a); rhythmically banded, wispy (b); and laminated (c) examples of structure from piston cores. Note the whispy aspect of (b) is best illustrated in the central part of the core.

Figure 17b-17z

Core logs, used in this study.

Figure 18

Composite cumulative curve envelopes and frequency distribution curves of samples of Scotian Shelf Drift (till) from Gulf of Maine (A); till tongue No. 7, core 79-011-08 from Emerald Basin (B); and Western Grand Banks of Newfoundland (C). The higher percentage of sand sized sediments in the Grand Banks samples is the result of glacial erosion of sedimentary bedrock offshore Newfoundland. The samples (B) from the till tongue in Emerald Basin plot similar to those of Emerald Silt, Fig. 19. However, the till tongues appear structureless both on X-radiographic positives and in their seismic reflection characteristics as com-

pared to the rhythmic banding of the Emerald Silt formation. Note change of scale in (A).

Figure 19

Composite cumulative curve envelopes and frequency distribution curves of samples of Emerald Silt; Emerald Silt facies B from Gulf of Maine (A); Emerald Silt facies B from Emerald Basin (B); and Emerald Silt facies A from Emerald Basin (C). Note the increased content of both sand and gravel in the Gulf of Maine samples compared to those of Emerald Basin. Note change of scale in (A).

Figure 20

Scatter diagram of mean grain size (mϕ) vs. sorting (oϕ) for representative Scotian Shelf sediment samples. The samples of till(Scotian Shelf Drift) are very poorly sorted. The samples from the till tongue in Emerald Basin fall within the Emerald Silt field, together with both Emerald Silt facies A and B samples from the same area. Several samples of the Sable Island Sand and Gravel formation are plotted for comparison and demonstrate their origin through reworking of the till during the late Pleistocene-Holocene transgression.

Figure 21

Distributional trends of foraminifera in Emerald Silt facies A and B and LaHave Clay.

Figure 22a

Correlation diagram for cores in the Gulf of Maine. Cores 2, 3, 4 and 6 were correlated on the basis of seismic reflection characteristics while core 5 was additionally influenced by age relationships. The cores are

positioned relative to water depth.

Figure 22b

Correlation diagram for cores from the Emerald Basin area of the Scotian Shelf. The cores were all correlated on the basis of seismostratigraphy. The largest unsampled section is 20m in thickness and occurs between core 8p and 2V.

Figure 23a

Huntec DTS profile at the type section in the Gulf of Maine. The sections sampled by the cores are bracketed. Emerald Silt facies A consists of medium- high intensity continuous coherent reflections. Emerald Silt facies B consists of medium intensity continuous coherent reflections with occasional dropstones. In this area LaHave Clay is coarser and has an acoustic character similar to facies B Emerald Silt. The till consists of incoherent reflections and the till tongue is interbedded with facies A Emerald Silt.

Figure 23b

Huntec DTS type section profile from Emerald Basin, Scotian Shelf. The core intervals have been shifted to a composite section away from the zone of till tongues. The positions of core 11p and 10p are less accurate as local unconformities and large distances made correlation difficult.

Figure 23c

Huntec DTS type section profile from the Laurentian Channel at the proximal side of the Laurentian Moraine. Facies B, Emerald Silt is characterized by weaker than normal continuous coherent reflections.

Figure 23d

Huntec DTS type section profile from Placentia Bay, western Grand Banks of Newfoundland. Emerald Silt facies A is very thin, approximately 2m in thickness and the LaHave Clay exhibits medium intensity continuous coherent reflections. A major unconformity occurs in core 223, hence the separation in sections sampled by the core.

Figure 24a

Sedimentation curve based on $C^{14}$ ages and seismostratigraphic analyses of Huntec DTS data for the Gulf of Maine. Range bars represent C*14* dating and seismostratigraphic accuracies. Plotted along the line of best fit are major events for the Gulf of Maine together with seismic facies. Extrapolated to the base of the section, initial ice lift-off occurred at 38000 yBP. Sedimentation rates remained constant until approximately 17000 yBP, prior to deposition of the LaHave Clay Formation when the rate decreased.

Figure 24b

Sedimentation curve based on $C^{14}$ ages and seismostratigraphic analyses of Huntec DTS data for Emerald Basin, Scotian Shelf. Initial ice lift-off occurred at approximately 46000 yBP. Till tongues 2-9 were developed during deposition of Emerald Silt facies A. Till tongue 4, the main tongue of the Scotian Shelf moraine complex, was deposited at 42000 yBP. Near the top of Emerald Silt facies A the sedimentation rate decreases and remains more or less constant until 8000 yBP.

Figure 24c

Sedimentation curve based on $C^{14}$ ages and seismostratigraphic

analyses of Huntec DTS data for the Laurentian Channel. Initial ice lift-off began approximately 40500 yBP and the Laurentian Moraine (Fig. 28) was formed at 36000 yBP. The sedimentation rate decreased during mid- Emerald Silt facies B deposition.

Figure 24d

Sedimentation curve for Placentia Bay, Newfoundland, based on $C^{14}$ ages and seismostratigraphic analyses of Huntec DTS data. Initial ice lift-off occurred at 29500 yBP. By 27000 yBP the Burin Moraine had developed and facies B was deposited. The sedimentation rate for Emerald Silt and LaHave Clay is similar.

Figure 25

Profile sections through dry-base (A) and wet-base (B) glaciers (Carey and Ahmad, 1961).

Figure 26

Diagrammatic model for the development of till tongues. Stage 1: ice sheet over bedrock, Stage 2: development of basal till, Stage 3a: floating ice shelf grounded at buoyancy line with deposition of glaciomarine sediment beneath ice shelf. Stage 3b: migration of buoyancy line seaward, Stage 3c retreat of buoyancy line and deposition of glaciomarine sediment over till tongue, Stage 4: total ice shelf development, no grounding.

Figure 27

Huntec DTS profiles from the inner shelf, Sambro Moraine area, across Emerald Basin to the northern edge of Sable Island Bank. These profiles form part of interpreted cross-section A-B on column section map 6c.

The till tongue of the Sambro Moraine is interbedded with facies A Emerald Silt at the same stratigraphic horizon as are till tongues on the northern edge of Sable Island Bank, across a distance of 75km.

Figure 28

Cross-section from the northern flank to the bottom of the Laurentian Channel, interpreted from Huntec DTS data. Area A shows a thin narrow till tongue on the flank of the channel interbedded with facies A Emerald Silt. Area B shows the Laurentian Moraine, a large 30m thick till tongue, which occurs on the floor of the Laurentian Channel, also interbedded with facies A Emerald Silt at the same seismostratigraphic horizon, across a distance of 15km.

Figure 29

Huntec DTS profiles from Middle Bank (Scotian Shelf) and Haddock Channel (Grand Banks of Newfoundland). In the upper profile, outliers of Emerald Silt, with a major unconformity developed on their surface attest to the widespread occurrence of glacial sediment on the bank areas of the shelf and their subsequent erosion through the Late Pleistocene-Holocene transgression. Haddock Channel shows at least three till deposits separated by Emerald Silt which have been eroded on the flank of the channel. Such deposits of limited extent occur only in depressions on the shelf. In most areas of the Grand Banks bedrock occurs at or near the seabed because the glacial section was removed by erosion.

Figure 30

Diagrammatic representation of the inferred environmental history of the Scotian Shelf at five critical stages of its development. See text for

explanation.

Figure 31

Huntec DTS profile across the Late Pleistocene terrace on the Scotian Shelf north of Banquereau. A major unconformity occurs across bedrock and glacial sediments from the 0.5 to the 3km marks. Overlying the terrace is a wedge shaped deposit of Sable Island Sand and Gravel. Section collected by D. Scott (Dalhousie University).

Figure 32

A plot of core length vs. water depth for cores from Emerald Basin, Middle Bank area (Fig. 6d). The cores were collected with the same core head weight and equal free fall distances. The plot shows a dramatic decrease in core penetration at approximately 115m. This is interpreted as representing the depth of the Late Pleistocene low sea level stand. See Table VI - geotechnical properties.

Figure 33

Airgun seismic and Huntec DTS profiles and interpretation from the Grand Banks of Newfoundland. Numerous channels occur at the bedrock surface as seen on the airgun data, but because of a 10m bubble pulse their acoustic character cannot be defined. The Huntec DTS profile indicates that the channels are infilled with till and glaciomarine sediment. Note the Late Pleistocene unconformity at the seabed across the glacial sediments. See also Figure 29. The absence of glacial deposits from the 17-24km mark is typical for most of the Grand Banks.

Figure 34

A time distance diagram and lithostratigraphy for the Scotian Shelf and

the Grand Banks of Newfoundland. This syntheses is correlated with the post-Sangamon stratigraphic framework developed by (1) Grant (1977, 1980), Prest and Grant (1979), and Grant in Dreimanis _et al_. (1981), (2) Alam, Piper and Cooke (1981), (3) Dreimanis _et al_. (1981) and Dreimanis (1975), (4) Ruddiman and McIntyre (1981). The time distance diagram is only in part representative of the Bay of Fundy and Gulf of Maine region and requires qualification which is included in the text.

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TABLE I - TABLE OF QUATERNARY FORMATIONS

| AGE | FORMATION | LITHOSTRATIGRAPHY | THICKNESS | SEISMOSTRATIGRAPHY |
|---|---|---|---|---|
| HOLOCENE | LaHave Clay | Greyish brown, soft, silty, clay grading to clayey silt, confined mainly to basins and depressions of shelf. Derived by winnowing of glacial sediments on banks and transported to basins. Time equivalent of Sable Island Sand and Gravel and Sambro Sand on banks | 0-70m | Generally transparent without reflections. Some weak continuous coherent reflections in base of section becoming stronger in nearshore sandy facies |
| | Sable Island Sand and Gravel | Fine to coarse, well-sorted sand grading to sub-rounded to rounded gravels. Unconformably overlies Emerald Silt and Scotian Shelf Drift, and derived from these deposits through reworking during Holocene transgression above 120m present depth. Time equivalent of LaHave Clay in basins | 0-50m generally veneer | Highly reflective seabed. Generally closely spaced continuous coherent reflections if deposit is of sufficient thickness to resolve |
| | Sambro Sand | Silty sand grading locally to gravelly sand and well-sorted sand. Deposited sublittorally with respect to the Pleistocene shoreline below 120m present depth. Time equivalent to basal LaHave Clay and upper Emerald Silt, facies B | 0-20m generally veneer | Similar to Sable Island Sand and Gravel |
| PLEISTOCENE | Emerald Silt, facies C | Not well sampled | 0-100m | Discontinuous coherent reflections; transitional between facies A Emerald Silt and glacial till |
| | Emerald Silt, facies B | Dark greyish brown, poorly sorted clayey and sandy silt with some gravel. Poorly developed rhythmic banding; proglacial in origin | 0-40m | Medium to low amplitude continuous coherent reflections, and to some degree a ponded sedimentational style |
| | Emerald Silt, facies A | Dark greyish brown, poorly sorted clayey and sandy silt, some gravel. Well developed rhythmic banding; subglacial in origin. Time equivalent to parts of Scotian Shelf Drift | 0-100m | High amplitude continuous coherent reflections, highly conformable to substrate irregularities |
| | Scotian Shelf Drift | Very dark greyish brown, cohesive glacial till comprised of poorly sorted sandy clay and silt with variable gravel | 0-100m | Incoherent reflections, sometimes with scattered point source reflections |

TABLE II - CORE DATA

| CORE # | WATER DEPTH(m) | LATITUDE | LONGITUDE | CORE LENGTH(m) | GEOGRAPHIC AREA |
|---|---|---|---|---|---|
| 73-003-351 | 454 | 46°34.2'N | 58°24.2'W | 9.9 | Laurentian Channel |
| 75-009-21v | 139 | 46°50.1'N | 50°44.8'W | 4.3 | Downing Basin |
| 76-016-2 | 198 | 43°03.25'N | 67°03.0'W | 4.3 | Gulf of Maine |
| 76-016-3 | 268 | 42°40.6'N | 67°06.0'W | 8.2 | Gulf of Maine |
| 76-016-4 | 241 | 42°41.9'N | 67°06.2'W | 3.1 | Gulf of Maine |
| 76-016-5 | 238 | 42°48.2'N | 67°05.8'W | 8.4 | Gulf of Maine |
| 76-016-6 | 194 | 43°00.0'N | 67°04.1'W | 4.6 | Gulf of Maine |
| 78-012-132 | 263 | 46°50.3'N | 54°43.7'W | 9.6 | Placentia Bay |
| 78-012-223 | 227 | 46°52.8'N | 54°51.8'W | 2.4 | Placentia Bay |
| 78-012-321 | 181 | 46°45.6'N | 54°36.2'W | 2.3 | Placentia Bay |
| 79-011-1 | 142 | 44°33.0'N | 61°25.2'W | 8.7 | Northeast Emerald Basin |
| 79-011-2 | 135 | 44°41.9'N | 61°25.0'W | 3.1 | Northeast Emerald Basin |
| 79-011-6 | 97 | 44°44.4'N | 61°13.9'W | 1.7 | Northwest Middle Bank |
| 79-011-7 | 91 | 44°43.5'N | 61°12.6'W | 1.2 | Northwest Middle Bank |
| 79-011-8 | 111 | 44°43.4'N | 61°18.0'W | 1.9 | Northwest Middle Bank |
| 79-011-9 | 111 | 44°43.9'N | 61°17.75'W | 1.9 | Northwest Middle Bank |
| 79-011-10 | 133 | 44°42.0'N | 61°46.5'W | 6.5 | Northeast Emerald Basin |
| 79-011-11 | 150 | 44°39.28'N | 61°46.86'W | 6.4 | Northeast Emerald Basin |
| 79-011-12 | 159 | 43°47.8'N | 62°36.3'W | 6.2 | Southwest Emerald Basin |
| 79-011-2v | 124 | 44°44.1'N | 61°24.5'W | 3.1 | Northeast Emerald Basin |
| 79-011-3v | 120 | 44°45.0'N | 61°24.2'W | 3.2 | Northeast Emerald Basin |
| 79-011-4v | 117 | 44°46.25'N | 61°25.4'W | 1.7 | Northwest Middle Bank |
| 82-003-4s | 135 | 44°38.48'N | 61°25.12'W | 2.8 | Northeast Emerald Basin |
| 82-003-6s | 135 | 44°40.12'N | 61°26.24'W | 3.4 | Northeast Emerald Basin |
| 82-003-7s | 135 | 44°41.24'N | 61°25.06'W | 3.4 | Northeast Emerald Basin |

TABLE III - AGE DETERMINATION

| CORE # | INTERVAL (cm) | FORMATION AND FACIES | AGE X (1000 yBP) | LABORATORY # |
|---|---|---|---|---|
| 73-003-351 | 90-125 | LaHave Clay | 16.17 ± 0.52 | GX-6695 |
| | 230-265 | LaHave Clay | 17.245 ± 0.45 | GX-6696 |
| | 620-655 | Emerald Silt (Facies B) | 27.15 + 2.26 -1.77 | GX-6697 |
| | 880-915 | Emerald Silt (Facies B) | 30.26 + 2.5 - 1.9 | GX-6698 |
| 75-009-21v | 160-190 | Emerald Silt (Facies A) | 22.015 + 1.35 - 1.15 | GX-8803 |
| 76-016-2 | 25-55 | LaHave Clay | 6.09 ± 0.17 | GSC-2947 |
| | 130-160 | LaHave Clay | 11.1 ± 0.15 | GSC-2944 |
| | 224-249 | Emerald Silt (Facies B) | 17.0 ± 0.9 | GSC-2709 |
| 76-016-3 | 170-195 | Emerald Silt (Facies B) | 15.3 ± 0.39 | GSC-2697 |
| | 550-575 | Emerald Silt | 19.0 | GSC-2711 |
| | 733-783 | Emerald Silt (Facies B) | 21.6 ± 0.69 | GSC-2735 |
| 76-016-4 | 40-80 | Emerald Silt (Facies B) | 19.0 ± 0.53 | GSC-2939 |
| | 275-300 | Emerald Silt (Facies B) | 18.0 ± 0.99 | GSC-2755 |
| 76-016-5 | 130-165 | Emerald Silt (Facies B) | 22.6 ± 1.08 | GSC 2967 |
| | 207-232 | Emerald Silt (Facies B) | 18.0 | GSC-2770 |
| | 430-475 | Emerald Silt (Facies B) | 24.8 ± 1.13 | GSC-2962 |
| | 519-544 | Emerald Silt (Facies B) | 18.0 | GSC-2789 |
| | 735-760 | Emerald Silt (Facies A) | 26.6 ± 1.6 | GSC-2715 |
| 76-016-6 | 125-150 | Emerald Silt (Facies C) | 17.6 ± 0.62 | GSC-2810 |
| | 415-440 | Emerald Silt (Facies C) | 17.4 ± 0.67 | GSC-2801 |
| 78-012-132 | 30-85 | LaHave Clay | 7.49 ± 0.08 | GSC-2933 |
| | 810-850 | LaHave Clay | 14.1 ± 0.2 | GSC-2926 |

| | | | | |
|---|---|---|---|---|
| 78-012-223 | 15-50 | LaHave Clay | 14.8 ± 0.33 | GSC-2874 |
| | 160-195 | Emerald Silt (Facies B) | 22.2 ± 1.45 | GSC-2866 |
| 78-012-321 | 56-95 | Emerald Silt (Facies B) | 22.7 ± 0.73 | GSC-2999 |
| | 170-210 | Emerald Silt (Facies B) | 21.8 ± 0.76 | GSC-2890 |
| 79-011-1 | 15-55 | LaHave Clay | 6.48 ± 0.07 | GSC-3255 |
| | 324-359 | LaHave Clay | 11.6 ± 0.12 | GSC-3258 |
| | 650-680 | Emerald Silt (Facies B) | 15.1 ± 0.23 | GSC-3260 |
| | 807-837 | Emerald Silt (Facies B) | 16.4 ± 0.18 | GSC-3272 |
| 79-011-2 | 10-40 | Emerald Silt (Facies A) | 34.0 | GSC-3263 |
| | 155-190 | Emerald Silt (Facies A) | 34.3 ± 0.96 | GSC-3264 |
| | 262-295 | Emerald Silt (Facies A) | 38.0 | GSC-3265 |
| 79-011-6 | 115-155 | Emerald Silt (Facies A) | 32.2 ± 1.03 | GSC-3095 |
| 79-011-7 | 15-55 | Emerald Silt (Facies A) | 41.8 ± 1.79 | GSC-2979 |
| 79-011-8 | 21-61 | Scotian Shelf Drift | 37.8 ± 1.9 | GSC-3271 |
| | 124-165 | Scotian Shelf Drift | 32.18 + 3.4 - 2.2 | GX-8541 |
| 79-011-9 | 25-65 | Emerald Silt (Facies A) | 33.0 | GSC-3138 |
| | 135-175 | Emerald Silt (Facies A) | 34.0 | GSC-3068 |
| 79-011-10 | 25-55 | Emerald Silt (Facies A) | 28.17 + 3.5 - 2.4 | GX-8542 |
| | 180-210 | Emerald Silt (Facies A) | 30.545 + 4.2 - 3.2 | GX-8543 |
| | 350-380 | Emerald Silt (Facies A) | 26.235 + 1.6 - 1.2 | GX-8544 |
| | 550-580 | Emerald Silt (Facies A) | 36.3 ± 0.98 | GSC-3152 |

| | | | | |
|---|---|---|---|---|
| 79-011-11 | 80-120 | Emerald Silt (Facies B) | 26.9 ± 0.65 | GSC-3164 |
| | 220-350 | Emerald Silt (Facies B) | 27.31 ± 0.65 | GX-8545 |
| | 320-350 | Emerald Silt (Facies B) | 26.32 + 1.8 - 1.35 | GX-8546 |
| | 560-590 | Emerald Silt (Facies B) | 30.9 ± 0.54 | GSC-3231 |
| 79-011-12 | 30-60 | Emerald Silt (Facies B) | 20.75 + 1.2 - 1.05 | GX-8547 |
| | 190-220 | Emerald Silt (Facies B) | 17.715 + 0.8 - 0.6 | GX-8548 |
| | 425-455 | Emerald Silt (Facies B) | 35.0 ± 1.6 | GSC-3251 |
| | 570-600 | Emerald Silt (Facies B) | 27.3 ± 0.6 | GSC-3244 |
| 79-011-2v | 40-70 | Emerald Silt (Facies A) | 30.0 | GX-8538 |
| | 160-190 | Emerald Silt (Facies A) | 27.31 + 1.4 - 1.1 | GX-8539 |
| | 233-280 | Emerald Silt (Facies A) | 36.5 ± 1.37 | GSC-2995 |
| 79-011-3v | 60-90 | Emerald Silt (Facies A) | 24.5 + 1.7 - 1.3 | GX-8540 |
| | 232-285 | Emerald Silt (Facies A) | 35.4 ± 0.81 | GSC-2987 |
| 79-011-4v | 38-78 | Emerald Silt (Facies A) | 39.5 ± 1.39 | GSC-2983 |
| 82-003-4s | 25-50 | LaHave Clay | 14.465 ± 0.55 | GX-8799 |
| | 163-185 | Emerald Silt (Facies B) | 23.6 + 1.2 - 1.0 | GX-8800 |
| | 235-255 | Emerald Silt (Facies B) | 19.0 + 1.05 - 0.9 | GX-8801 |
| 82-003-6s | 25-55 | Emerald Silt (Facies B) | 24.82 + 1.4 - 1.15 | GX-8796 |
| | 175-200 | Emerald Silt (Facies B) | 37.0 | GX-8797 |
| | 275-300 | Emerald Silt (Facies B) | 30.8 + 3.0 - 2.0 | GX-8798 |
| 82-003-7s | 60-95 | Emerald Silt (Facies B) | 33.1 + 5.5 - 2.7 | GX-8793 |
| | 185-210 | Emerald Silt (Facies A) | 37.0 | GX-8794 |
| | 280-320 | Emerald Silt (Facies A) | 37.0 | GX-8795 |

TABLE IV - SEDIMENT BANDING COMPILATION

MIDDLE BANK AREA

| Core No. | No. of Couplets /m in core | | No. of Seismic Reflections per 10m | Average Reflection distance m | Average No. of seismic reflections m | No. of Couplets in Core/seismic reflection |
|---|---|---|---|---|---|---|
| | High | Low | | | | |
| 79-7P | 64 | 34 | 28 | 0.35 | 2.8 | 22.8 |
| 79-4V | 48 | 37 | 32 | 0.31 | 3.2 | 14.9 |
| 79-3V | 44 | 29 | 30 | 0.33 | 3.0 | 14.6 |
| 79-9P | 51 | 34 | 30 | 0.33 | 3.0 | 17.0 |
| 79-2V | 43 | 31 | 30 | 0.33 | 3.0 | 14.3 |
| 79-2P | 45 | 23 | 23 | 0.44 | 2.3 | 20.1 |
| 79-6P | 70 | 37 | 20 | 0.50 | 2.0 | 35.3 |
| 79-10P | 40 | | 25 | 0.40 | 2.5 | 15.9 |
| 79-11P | 32 | | 30 | 0.33 | 3.0 | 10.6 |

GULF OF MAINE

| | | | | | | |
|---|---|---|---|---|---|---|
| 76-2 | 15 | | | | | |
| 76-5 | 27 | | 26 | 0.38 | 2.6 | 10.2 |
| 76-3 | 19 | | 14 | 0.70 | 1.4 | 13.7 |
| 76-4 | 30 | | | | | |

TABLE V - CHRONOLOGY OF EVENTS AND THICKNESS OF SEDIMENTS

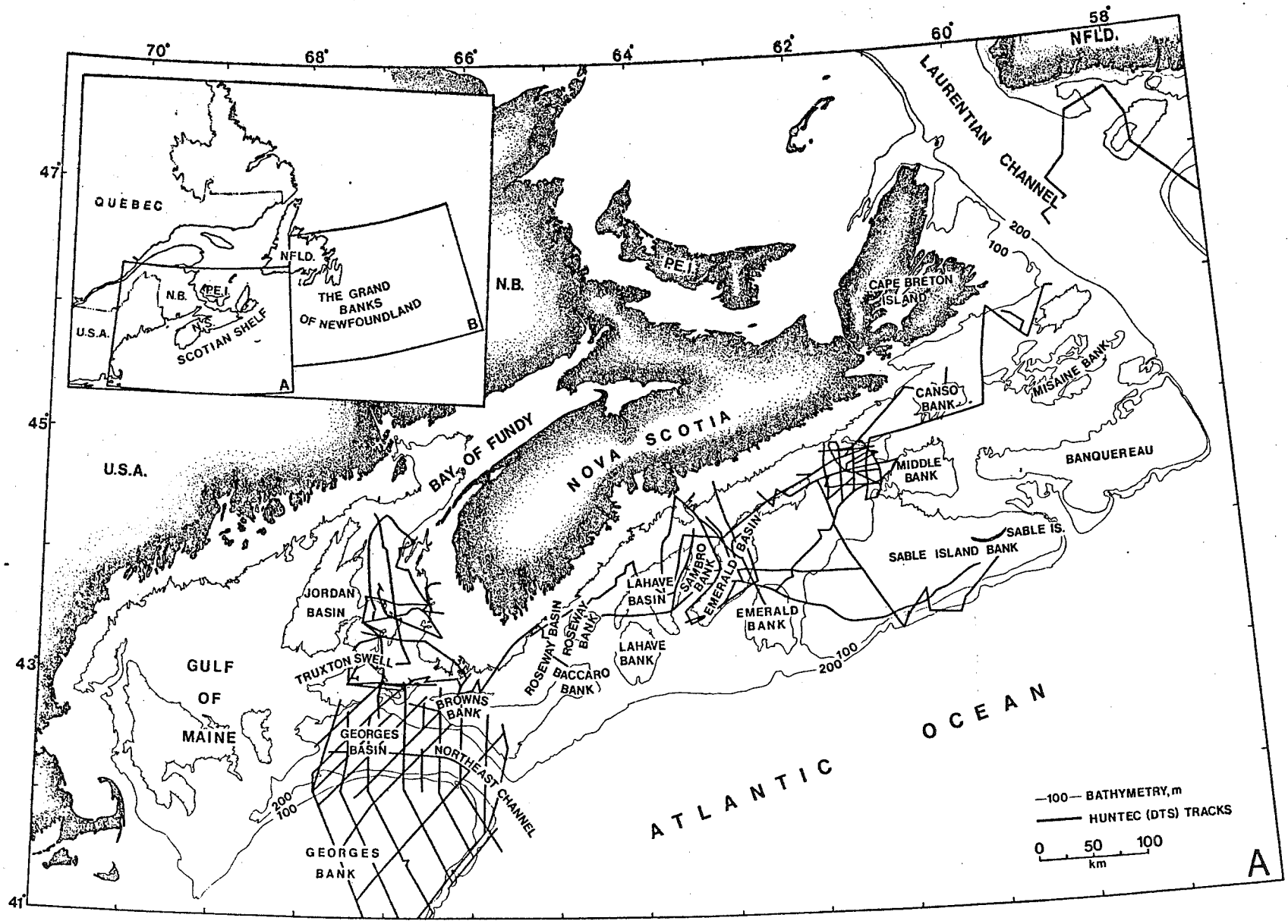
OFFSHORE SOUTHEASTERN CANADA

| Event | Gulf of Maine | Emerald Basin | Laurentian Channel | Placentia Bay |
|---|---|---|---|---|
| Ice sheet lift off, lift-off moraine development | 38,500 | 46,500 | 40,500 | 29,500 |
| Till tongue and moraine development | 31,000-26,500 | Tongue 2-9 44,000-34,500 Tongue 4, main moraine complex -42,000 | 35,000 | 28,500 |
| Emerald Silt Facies "A" | 38,500-26,500 | 46,500-32,500 | 40,500-31,500 | 29,500-28,000 |
| Emerald Silt Facies "B" | 26,500-13,500 | 32,500-16,500 | 31,500-23,000 | 28,000-15,500 |
| LaHave Clay | 13,500-present | 16,000-present | 23,000-present | 15,500-present |
| Thickness Emerald Silt Facies "A" Composite Section | 23m | 48m | 7.5m | 2m |
| Thickness Emerald Silt Facies "B" (Composite Section) | 13m | 22m | 6m | 13m |

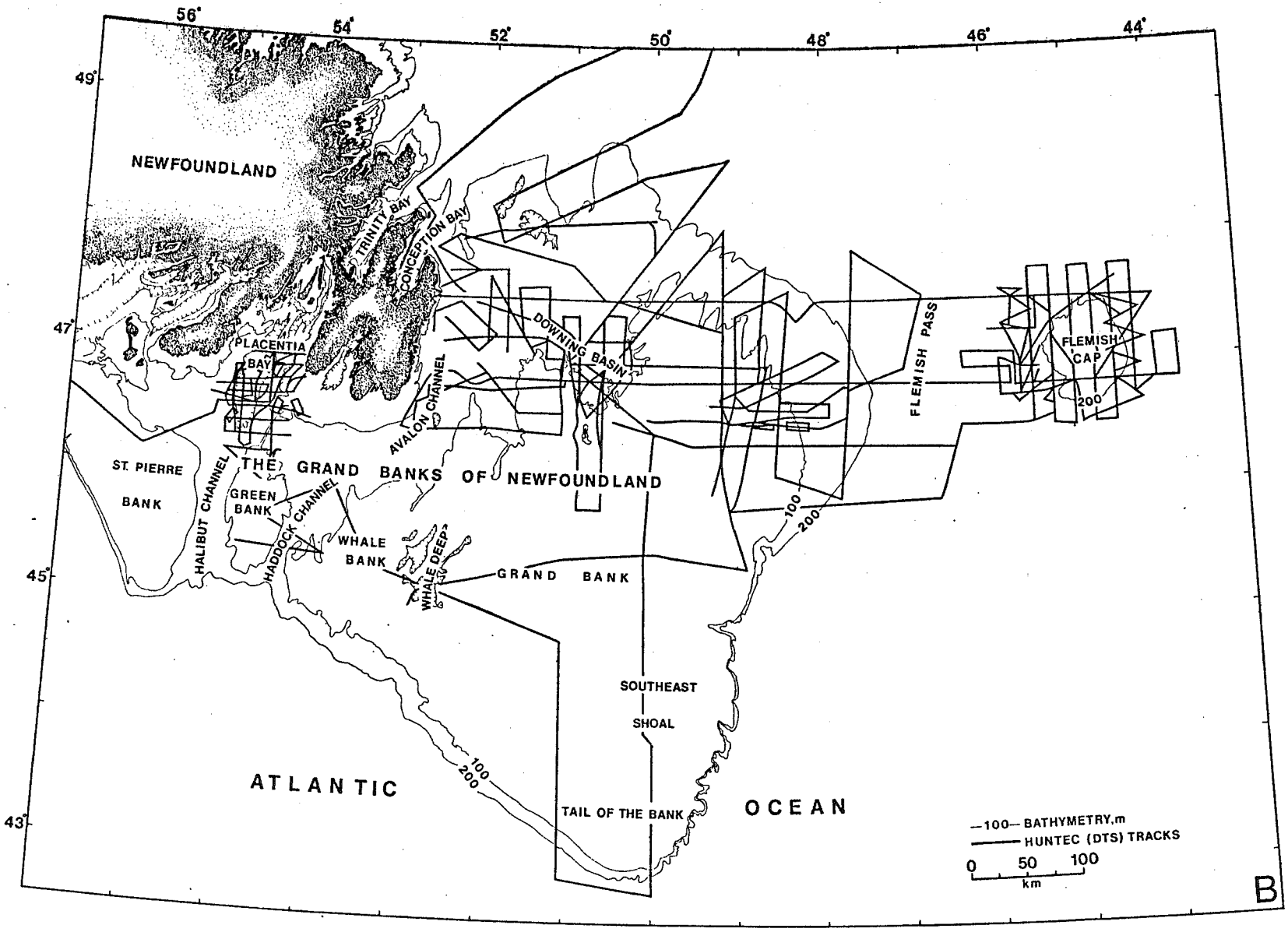
TABLE VI - GEOTECHNICAL PROPERTIES

Cruise: HUDSON 79-011

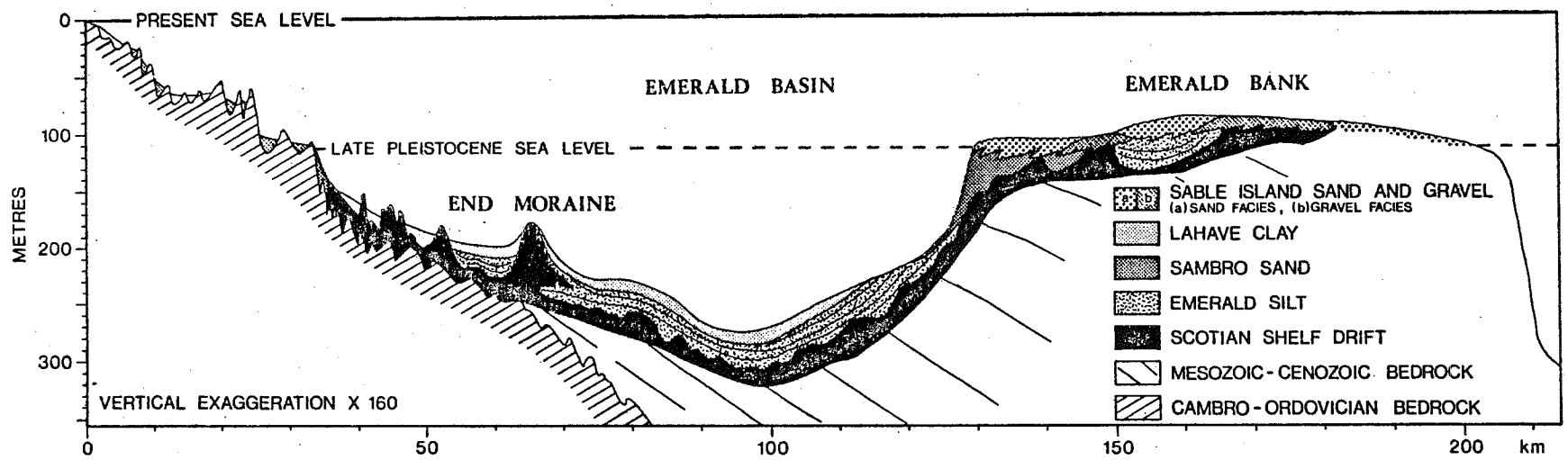
| Sample Designation: | 1 P Emerald Silt | 10 P Emerald Silt | 8 P Glacial Till | 7 P Emerald Silt |
|---|---|---|---|---|
| Water Depth: | 142m | 133m | 111m | 91m |
| Stratigraphic Order: Stratigraphic depth increase to the right | | | | |
| Physiographic control: | Basin Sample | Basin Sample | Bank Sample | Bank Sample |
| Water (W) | 39.6 | 48.3 | 32.2 | 36.0 |
| Plasticity Index ($I_p$) | 23.2 | 25.5 | 19.5 | 22.7 |
| Liquidity Index ($I_L$) | 0.71 | 0.78 | 0.38 | 0.43 |
| Unconfined Compression Strength ($S_u$) | 19 | 19 | 45 | 35 |
| Preconsolidated Stress ($P_c'$) | | | 150 | 100 |
| Median Diameter, mm ($M_d$) | 0.014 | 0.0002 | 0.004 | 0.002 |
| Core Length, m | 8.43 | 5.99 | 1.85 | 0.90 |

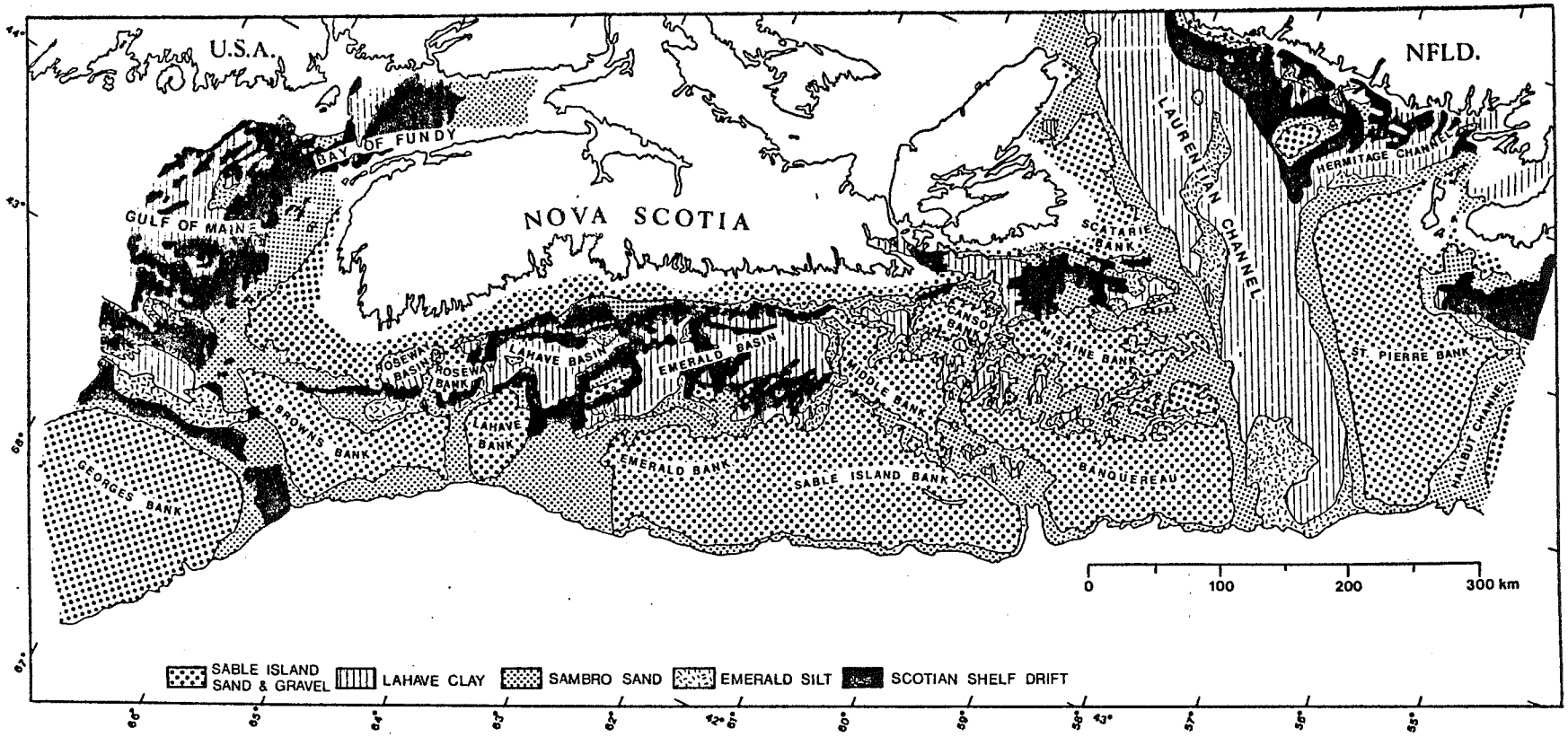

70°
68°
66°
64°
62°
60°
58°
47°
45°
43°
41°
NFLD.
LAURENTIAN CHANNEL
QUEBEC
NFLD.
N.B.
P.E.I.
N.S.
U.S.A.
SCOTIAN SHELF
THE GRAND BANKS OF NEWFOUNDLAND
A
B
N.B.
P.E.I.
CAPE BRETON ISLAND
200
100
MISAINE BANK
CANSO BANK
U.S.A.
BAY OF FUNDY
NOVA SCOTIA
MIDDLE BANK
BANQUEREAU
SABLE IS.
SABLE ISLAND BANK
SAMBRO BANK
EMERALD BASIN
LAHAVE BASIN
EMERALD BANK
JORDAN BASIN
ROSEWAY BASIN
ROSEWAY BANK
LAHAVE BANK
BACCARO BANK
TRUXTON SWELL
GULF OF MAINE
BROWNS BANK
GEORGES BASIN
NORTHEAST CHANNEL
GEORGES BANK
ATLANTIC OCEAN
—100— BATHYMETRY, m
HUNTEC (DTS) TRACKS
0 50 100
km
A

1 a

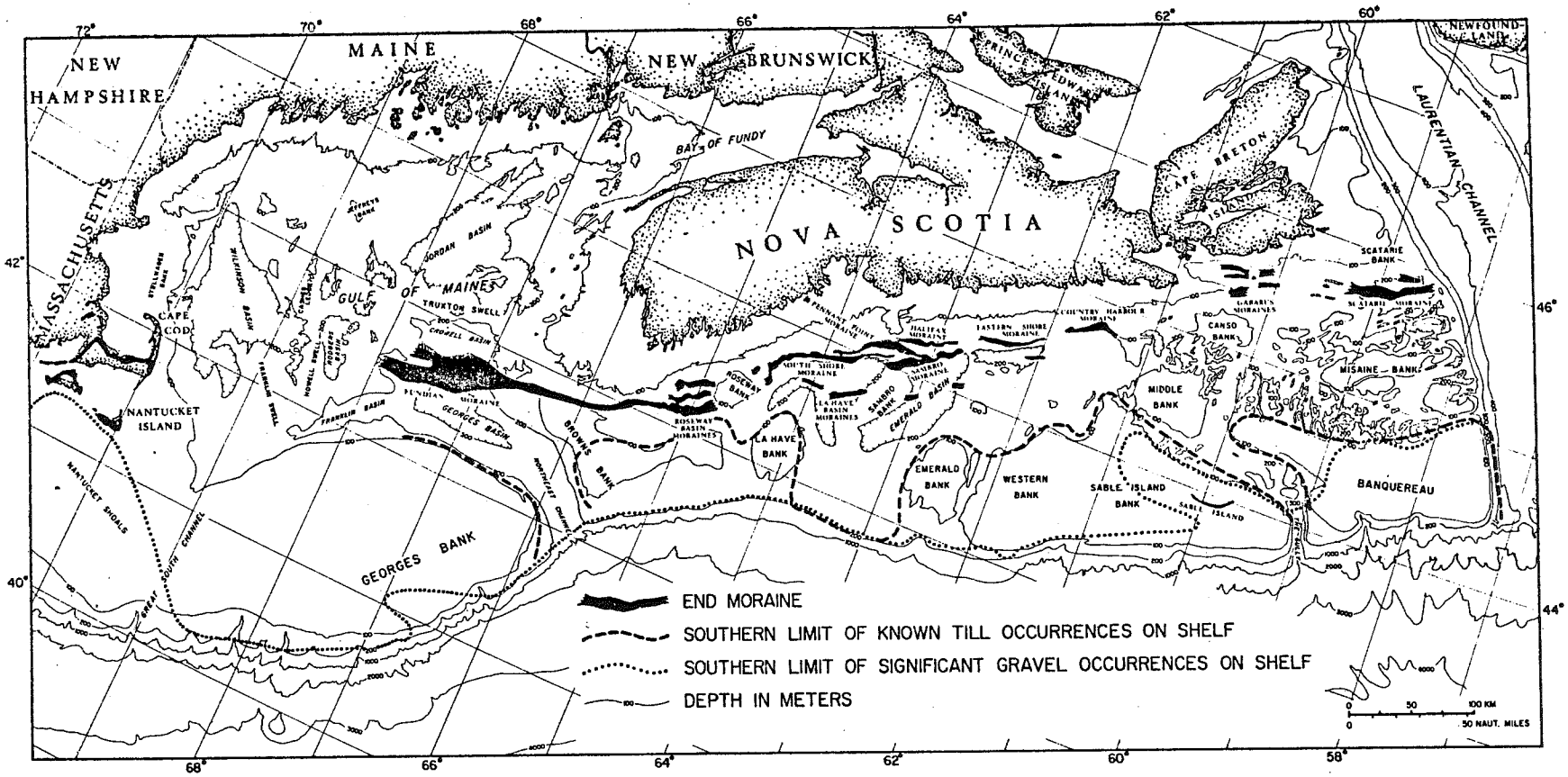

56°
54°
52°
50°
48°
46°
44°
49°
47°
45°
43°
NEWFOUNDLAND
TRINITY BAY
CONCEPTION BAY
PLACENTIA BAY
AVALON CHANNEL
DOWNING BASIN
FLEMISH PASS
FLEMISH CAP
200
ST. PIERRE BANK
HALIBUT CHANNEL
THE GRAND BANKS OF NEWFOUNDLAND
GREEN BANK
HADDOCK CHANNEL
WHALE BANK
WHALE DEEP
GRAND BANK
100
200
SOUTHEAST SHOAL
ATLANTIC
OCEAN
TAIL OF THE BANK
—100— BATHYMETRY,m
HUNTEC (DTS) TRACKS
0 50 100
km
B

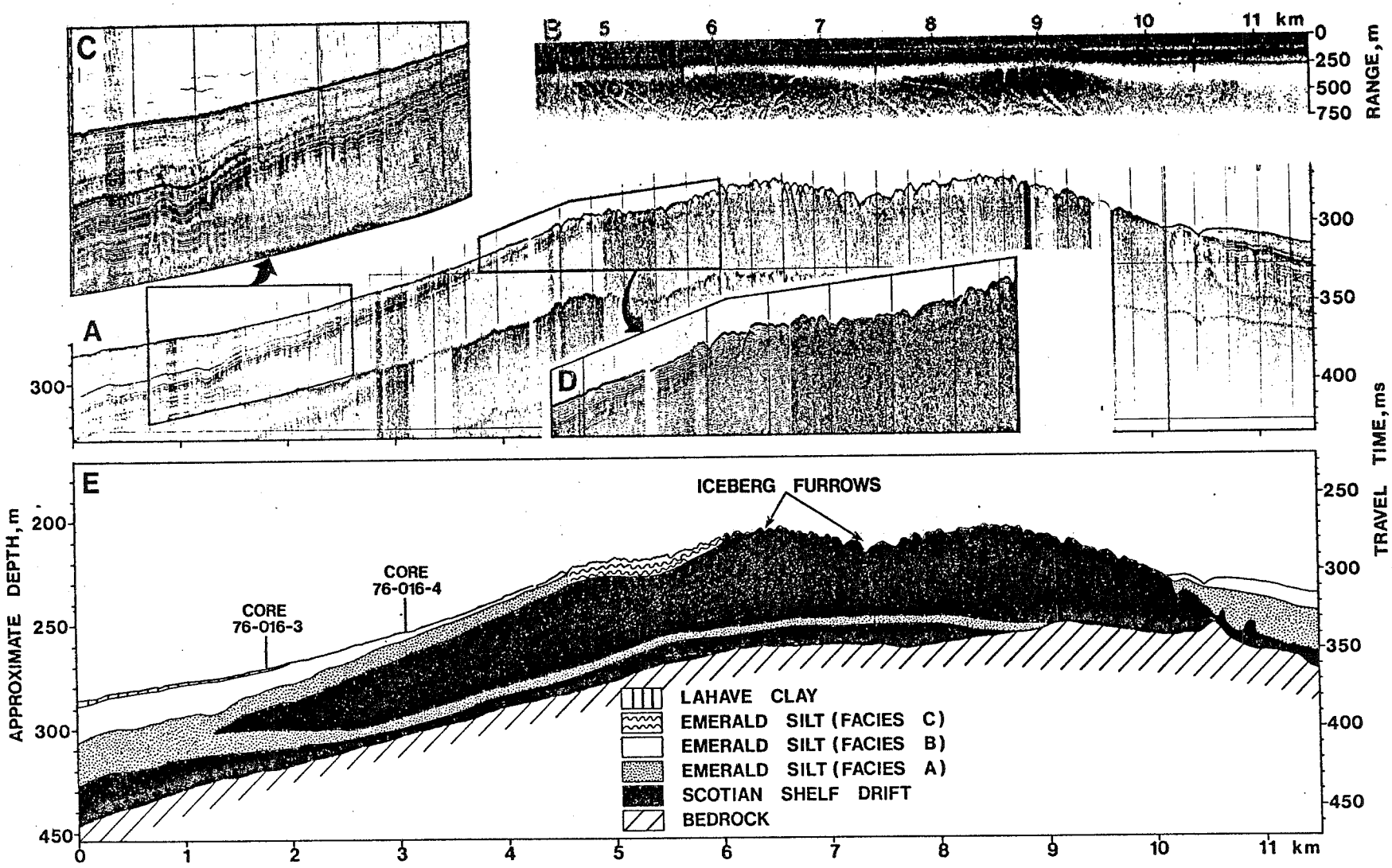
1b


PRESENT SEA LEVEL
LATE PLEISTOCENE SEA LEVEL
EMERALD BASIN
EMERALD BANK
END MORAINE
METRES
0
100
200
300
VERTICAL EXAGGERATION X 160
0
50
100
150
200
km
SABLE ISLAND SAND AND GRAVEL
(a) SAND FACIES, (b) GRAVEL FACIES
LAHAVE CLAY
SAMBRO SAND
EMERALD SILT
SCOTIAN SHELF DRIFT
MESOZOIC - CENOZOIC BEDROCK
CAMBRO - ORDOVICIAN BEDROCK




SURFICIAL GEOLOGY OF SCOTIAN SHELF AND ADJACENT AREAS


NEW HAMPSHIRE
MAINE
NEW BRUNSWICK
PRINCE EDWARD ISLAND
NEWFOUNDLAND
LAURENTIAN CHANNEL
BAY OF FUNDY
NOVA SCOTIA
CAPE BRETON ISLAND
MASSACHUSETTS
CAPE COD
NANTUCKET ISLAND
GULF OF MAINE
JORDAN BASIN
TRUXTON SWELL
CROWELL BASIN
FRANKLIN BASIN
PLUMBIAN MORAINE
GEORGES BASIN
GEORGES BANK
NANTUCKET SHOALS
GREAT SOUTH CHANNEL
NORTHEAST CHANNEL
BROWNS BANK
ROSEWAY BANK
ROSEWAY BASIN MORAINES
LA HAVE BANK
LA HAVE BASIN MORAINES
SOUTH SHORE MORAINE
SANDBO BANK
EMERALD BASIN
HALIFAX MORAINE
EASTERN SHORE MORAINE
COUNTRY HARBOR MORAINE
EMERALD BANK
WESTERN BANK
SABLE ISLAND BANK
SABLE ISLAND
MIDDLE BANK
CANSO BANK
SCATARIE BANK
SCATARIE MORAINE
GABARUS MORAINES
MISAINE BANK
BANQUEREAU
EASTERN CHANNEL
END MORAINE
SOUTHERN LIMIT OF KNOWN TILL OCCURRENCES ON SHELF
SOUTHERN LIMIT OF SIGNIFICANT GRAVEL OCCURRENCES ON SHELF
DEPTH IN METERS
0 50 100 KM
0 50 NAUT. MILES


C
B
5
6
7
8
9
10
11 km
0
250
500
750
RANGE, m
A
300
D
300
350
400
TRAVEL TIME, ms
E
ICEBERG FURROWS
CORE
76-016-4
CORE
76-016-3
APPROXIMATE DEPTH, m
200
250
300
450
250
300
350
400
450
LAHAVE CLAY
EMERALD SILT (FACIES C)
EMERALD SILT (FACIES B)
EMERALD SILT (FACIES A)
SCOTIAN SHELF DRIFT
BEDROCK
0
1
2
3
4
5
6
7
8
9
10
11 km

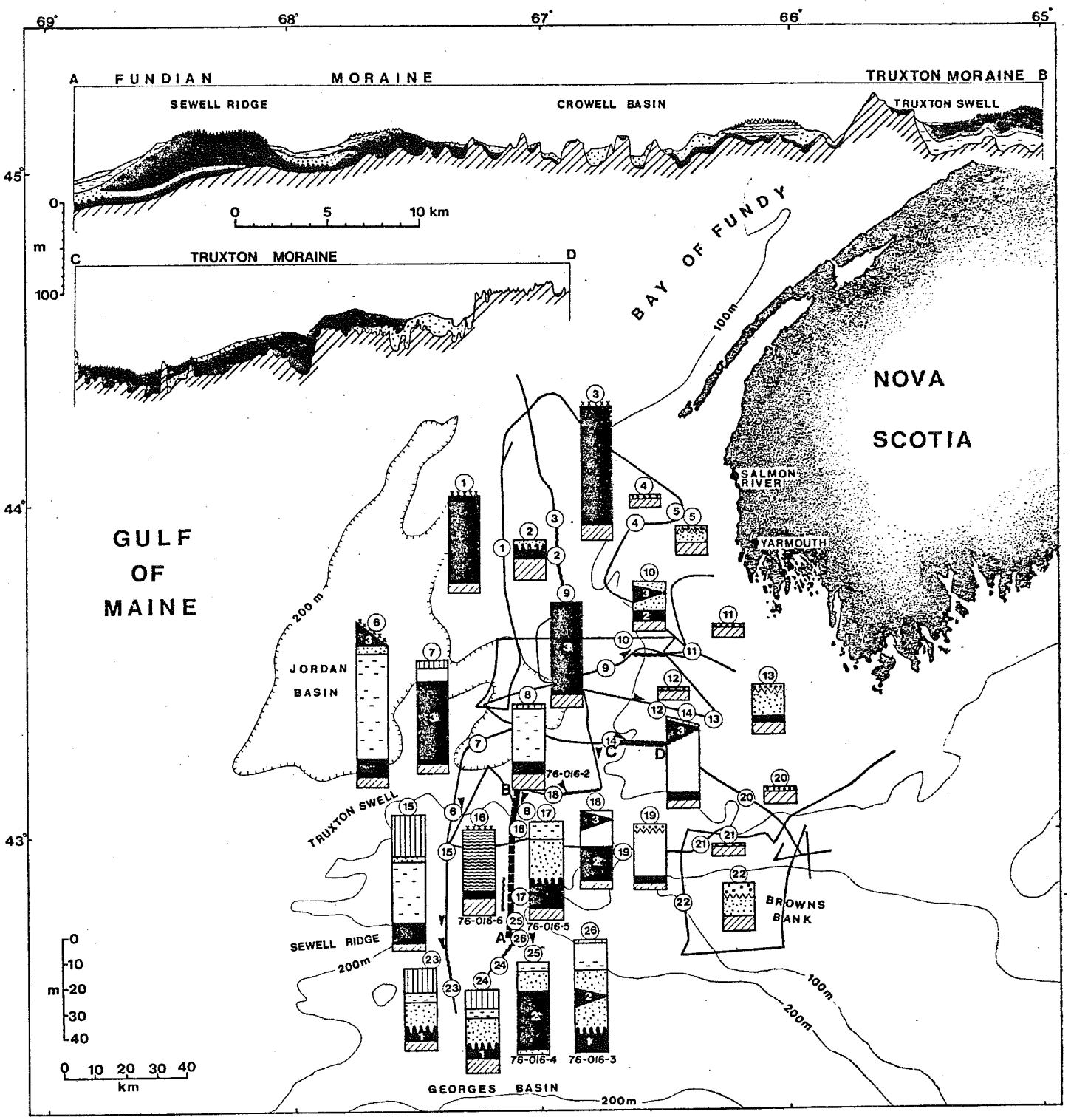
## LEGEND (FIG 6b – h)

### COLUMN SECTIONS

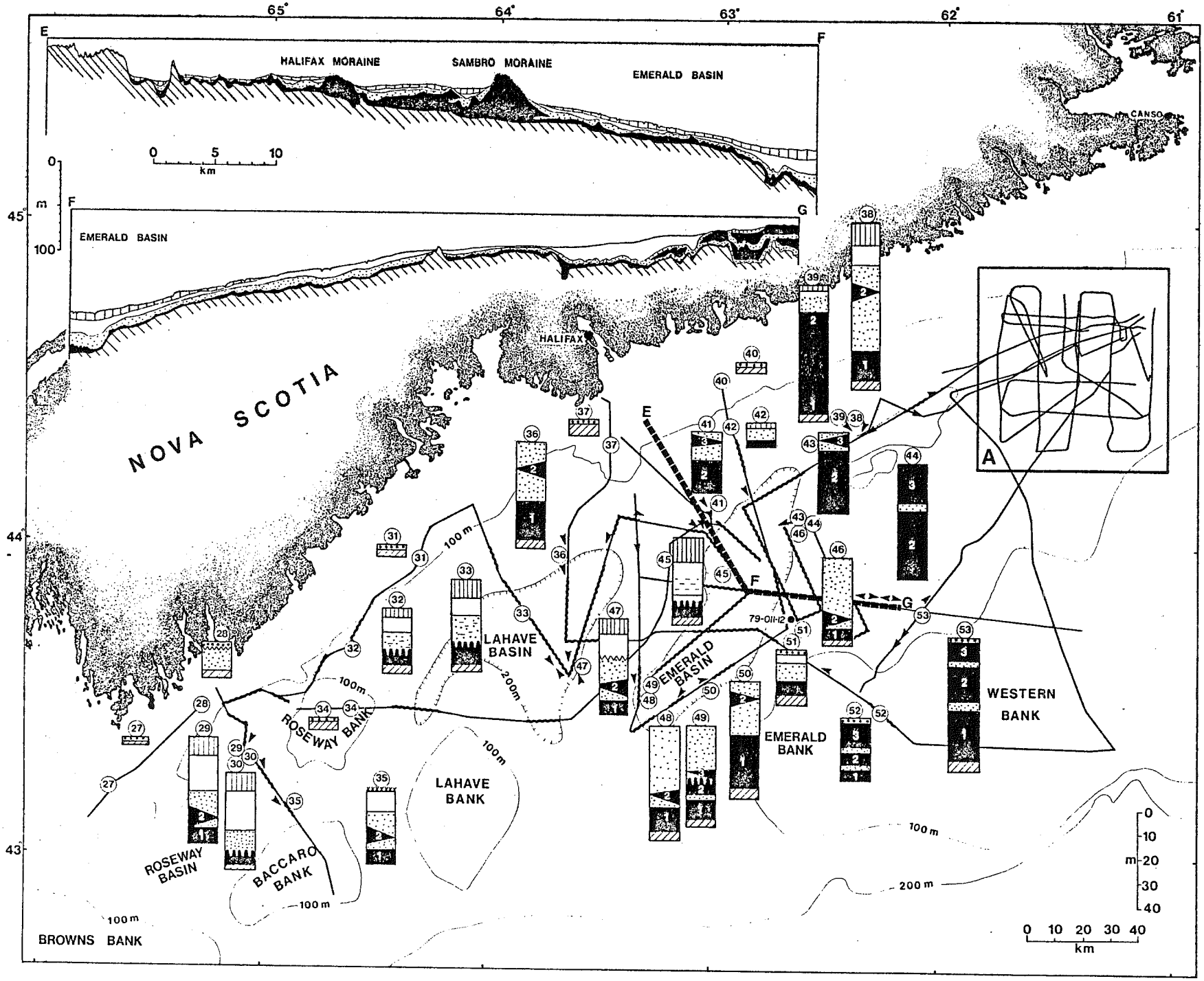
- SABLE ISLAND SAND AND GRAVEL AND SAMBRO SAND
- LAHAVE CLAY
- EMERALD SILT FACIES C
- EMERALD SILT FACIES B
- EMERALD SILT FACIES A
- SCOTIAN SHELF-NEWFOUNDLAND SHELF DRIFT (TILL)
- BEDROCK UNDIFFERENTIATED
- TILL TONGUE
- UNCONFORMITY
- LIFT-OFF MORAINES
- ICEBERG FURROWS
- DROPSTONES

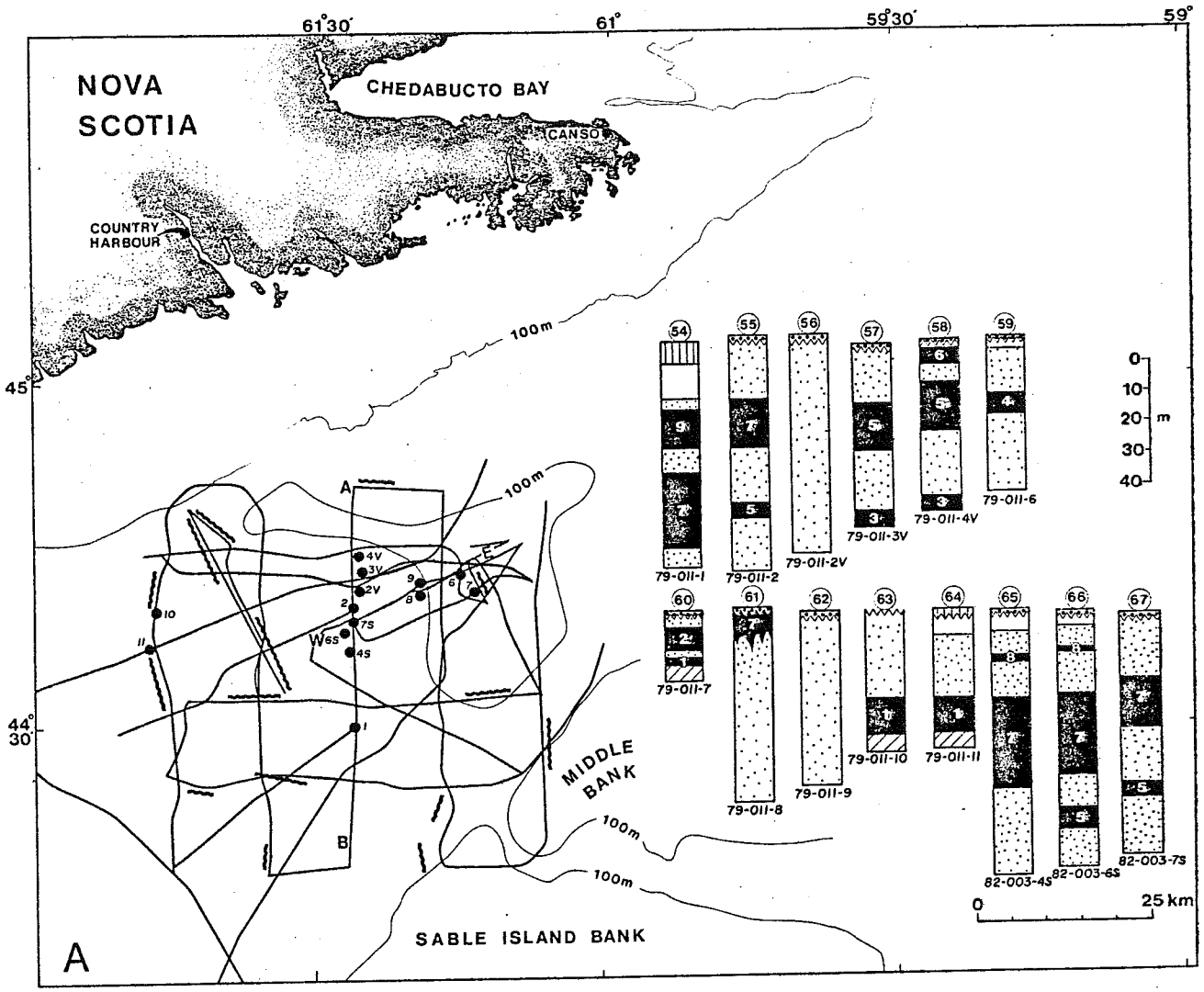
### MAP

- (1) COLUMN SECTION POSITION AND NUMBER
- TILL TONGUE POSITION AND DIRECTION
- LIFT-OFF MORAINE
- -100m- BATHYMETRY
- HUNTEC DTS TRACK
- CROSSECTION
- CORE POSITION

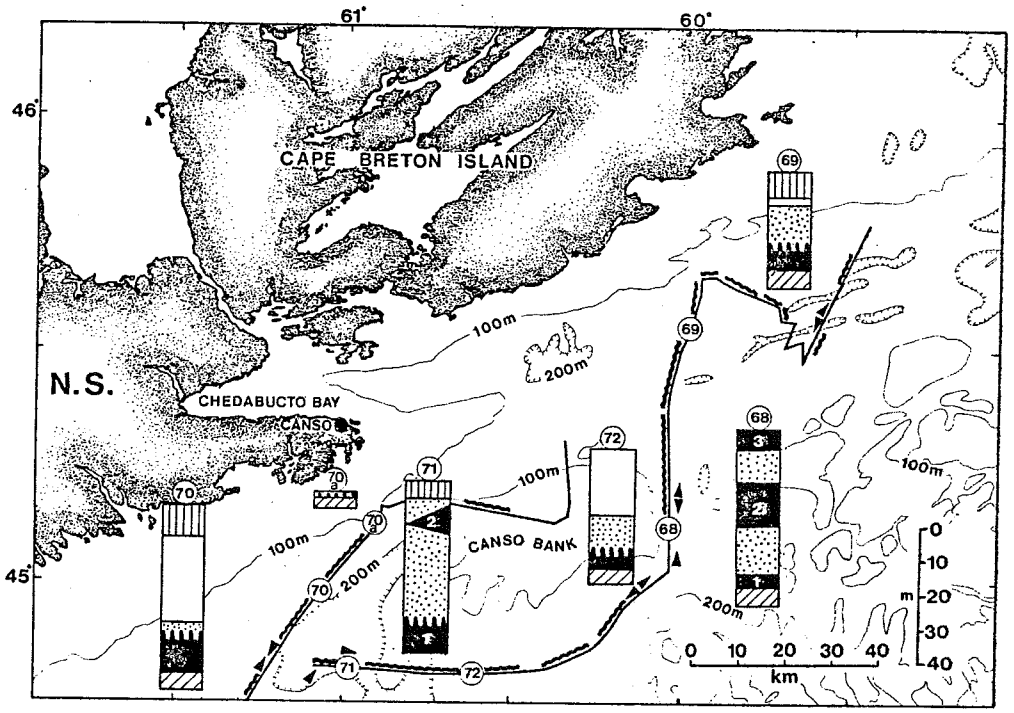

69°
68°
67°
66°
65°
45°
44°
43°
A FUNDIAN MORAINE
TRUXTON MORAINE B
SEWELL RIDGE
CROWELL BASIN
TRUXTON SWELL
0 5 10 km
0
m
100
C TRUXTON MORAINE D
BAY OF FUNDY
100m
NOVA SCOTIA
SALMON RIVER
YARMOUTH
GULF OF MAINE
200 m
JORDAN BASIN
TRUXTON SWELL
76-016-2
76-016-6
76-016-5
76-016-4
76-016-3
SEWELL RIDGE
200m
BROWNS BANK
100m
200m
0
10
20
30
40
m
0 10 20 30 40
km
GEORGES BASIN
200m

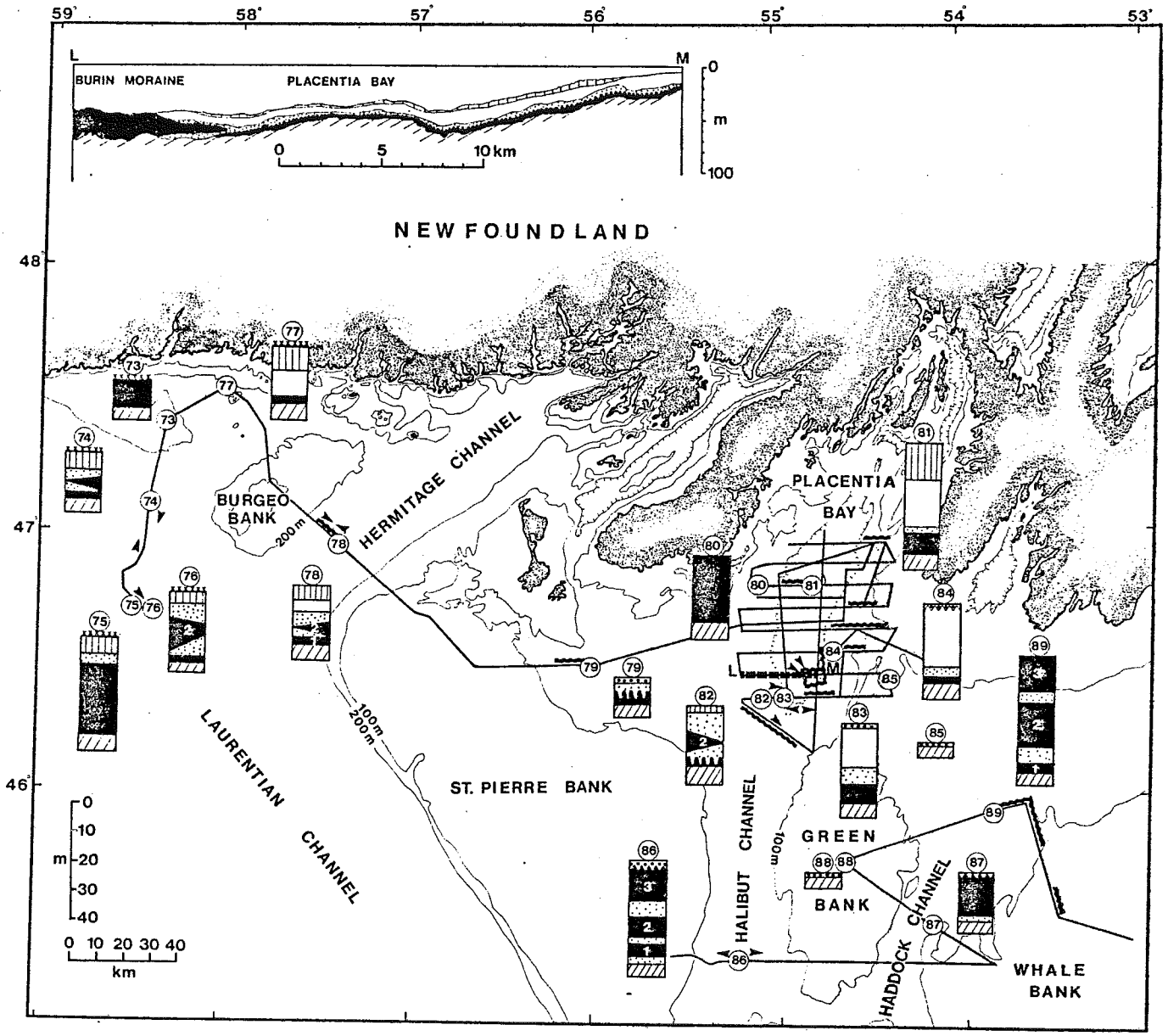
6 b


65°
64°
63°
62°
61°
45°
44°
43°
E
F
G
HALIFAX MORAINE
SAMBRO MORAINE
EMERALD BASIN
0 5 10
km
0
m
100
EMERALD BASIN
HALIFAX
CANSO
NOVA SCOTIA
A
LAHAVE BASIN
EMERALD BASIN
ROSEWAY BANK
LAHAVE BANK
EMERALD BANK
WESTERN BANK
ROSEWAY BASIN
BACCARO BANK
BROWNS BANK
79-011-12
100 m
200 m
0 10 20 30 40
km
0
10
m 20
30
40

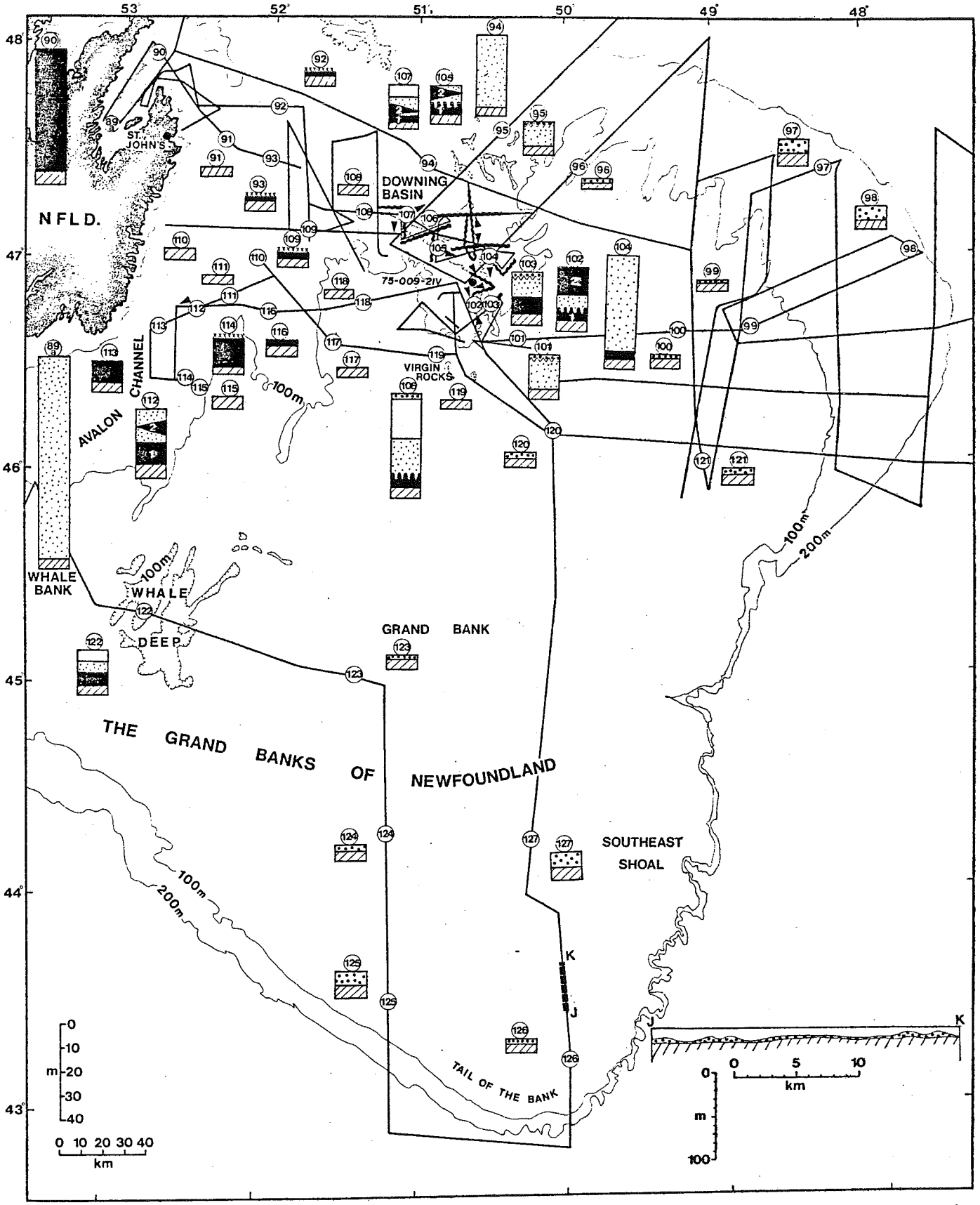

61°30'
61°
59°30'
59°
NOVA SCOTIA
CHEDABUCTO BAY
CANSO
COUNTRY HARBOUR
100m
45°
44° 30'
A
B
C
MIDDLE BANK
SABLE ISLAND BANK
79-011-1
79-011-2
79-011-2V
79-011-3V
79-011-4V
79-011-6
79-011-7
79-011-8
79-011-9
79-011-10
79-011-11
82-003-4S
82-003-6S
82-003-7S
0
10
20
30
40
m
0
25 km

6 d

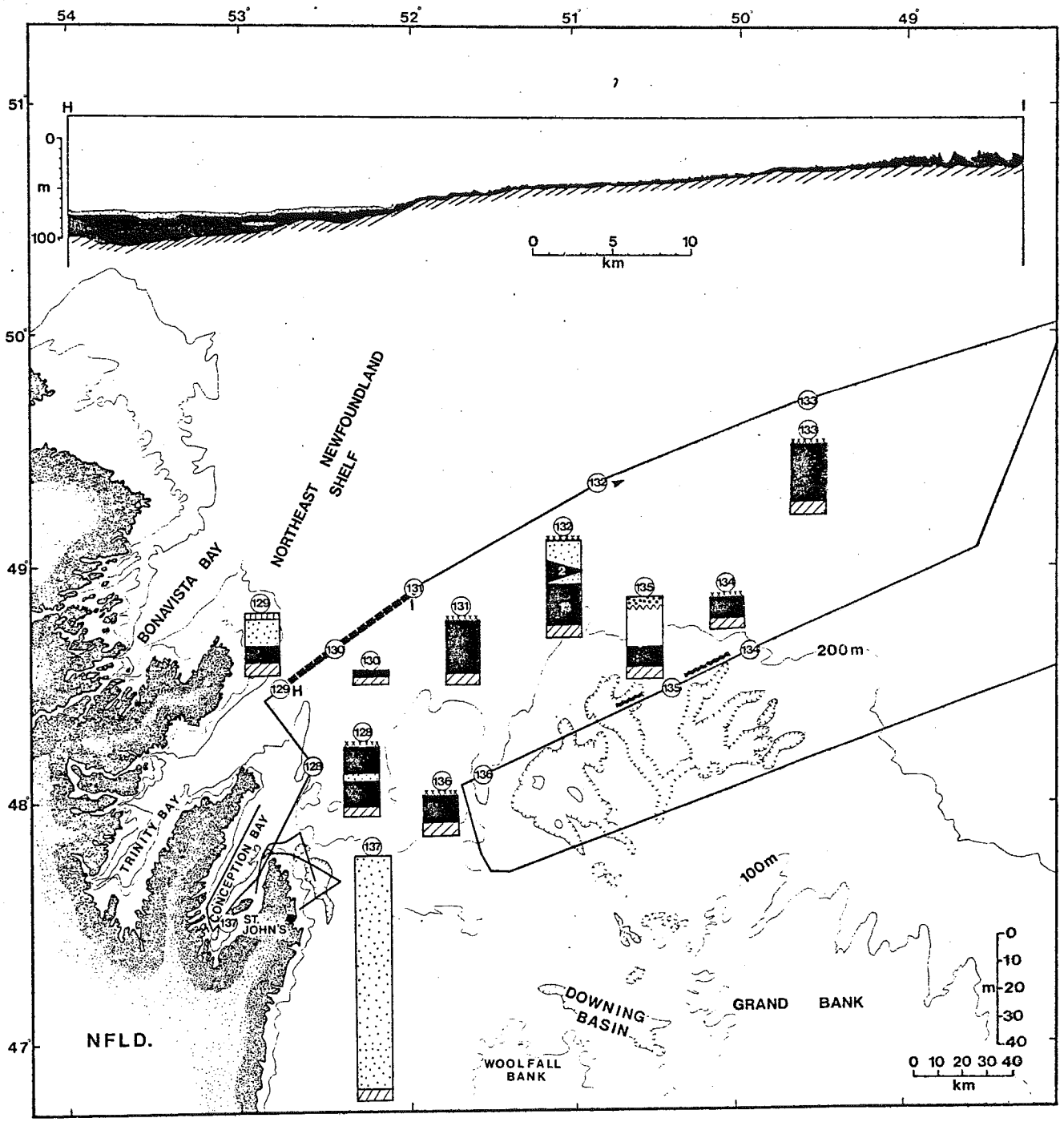

61°
60°
46°
45°
CAPE BRETON ISLAND
N.S.
CHEDABUCTO BAY
CANSO
CANSO BANK
100m
200m
69
68
72
71
70
0 10 20 30 40
km
0
10
20
30
40
m

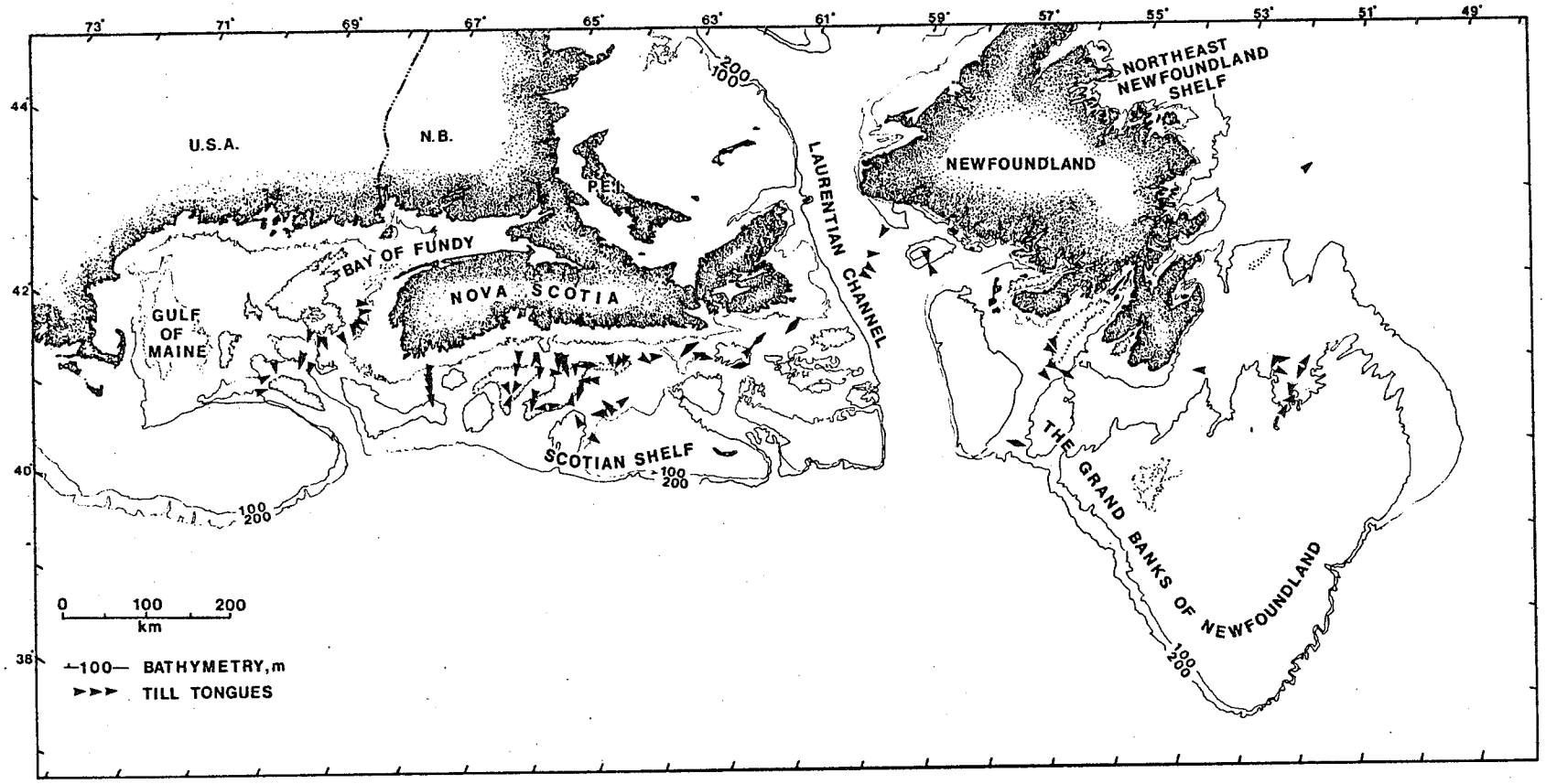

59°
58°
57°
56°
55°
54°
53°
48°
47°
46°
L
M
BURIN MORAINE
PLACENTIA BAY
0
m
100
0
5
10 km
NEWFOUNDLAND
BURGEO BANK
HERMITAGE CHANNEL
PLACENTIA BAY
LAURENTIAN CHANNEL
ST. PIERRE BANK
HALIBUT CHANNEL
GREEN BANK
HADDOCK CHANNEL
WHALE BANK
200 m
100 m
200 m
100 m
0
10
m 20
30
40
0 10 20 30 40
km

6 f

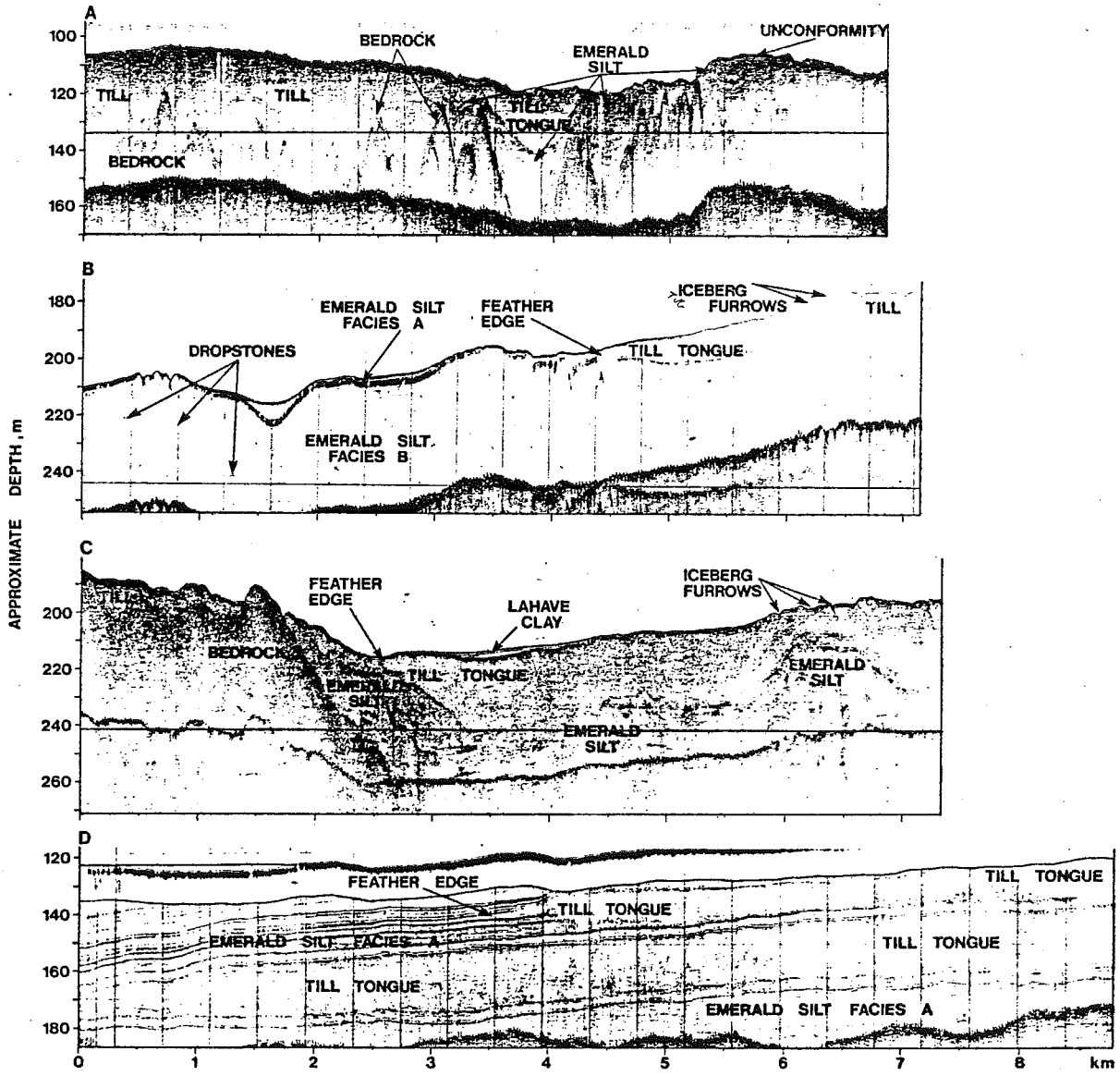

53°
52°
51°
50°
49°
48°
48°
47°
46°
45°
44°
43°
NFLD.
ST. JOHN'S
DOWNING BASIN
75-009-21V
VIRGIN ROCKS
AVALON CHANNEL
100m
WHALE BANK
WHALE DEEP
GRAND BANK
THE GRAND BANKS OF NEWFOUNDLAND
SOUTHEAST SHOAL
TAIL OF THE BANK
200m
K
J
0 10 20 30 40
km
m
0 10 20 30 40
0 5 10
km
m
0
100

6g

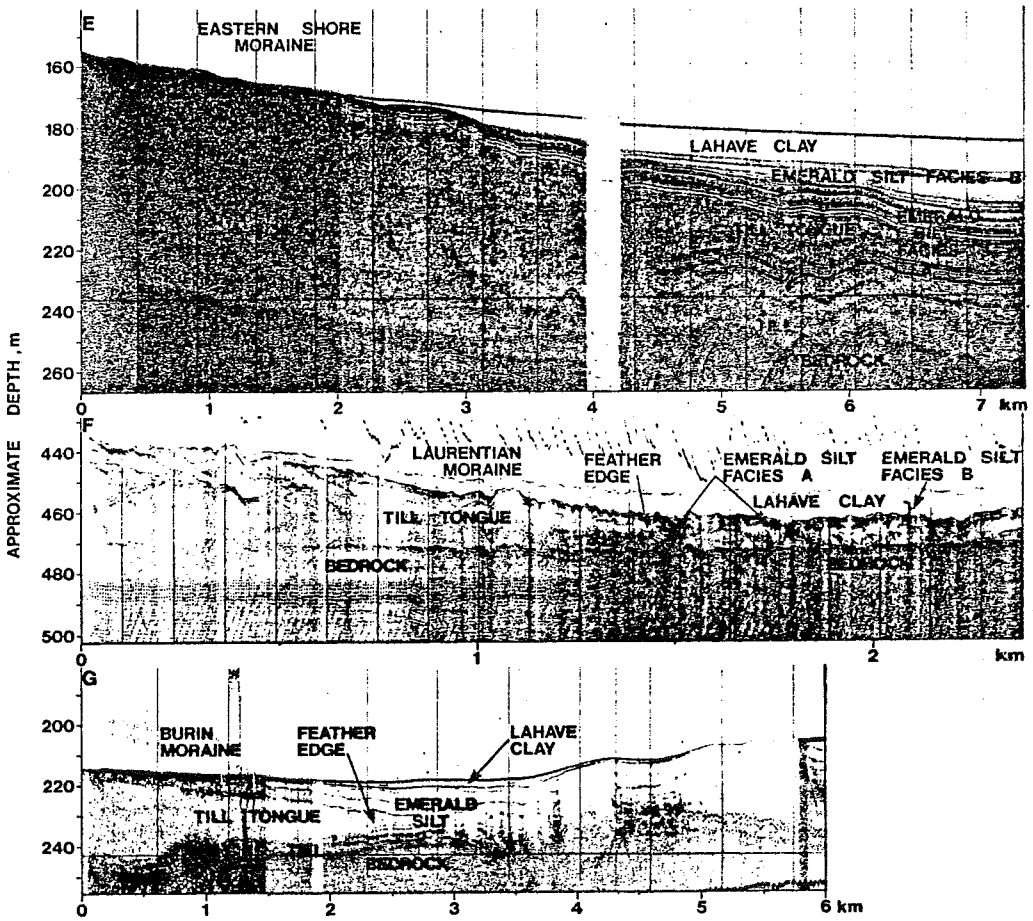

54
53°
52°
51°
50°
49°
51°
50°
49°
48°
47°
H
I
0
m
100
0 5 10
km
NORTHEAST NEWFOUNDLAND SHELF
BONAVISTA BAY
TRINITY BAY
CONCEPTION BAY
ST. JOHN'S
NFLD.
DOWNING BASIN
WOOLFALL BANK
GRAND BANK
200m
100m
129
130
131
132
133
134
135
136
137
128
0
10
m 20
30
40
0 10 20 30 40
km

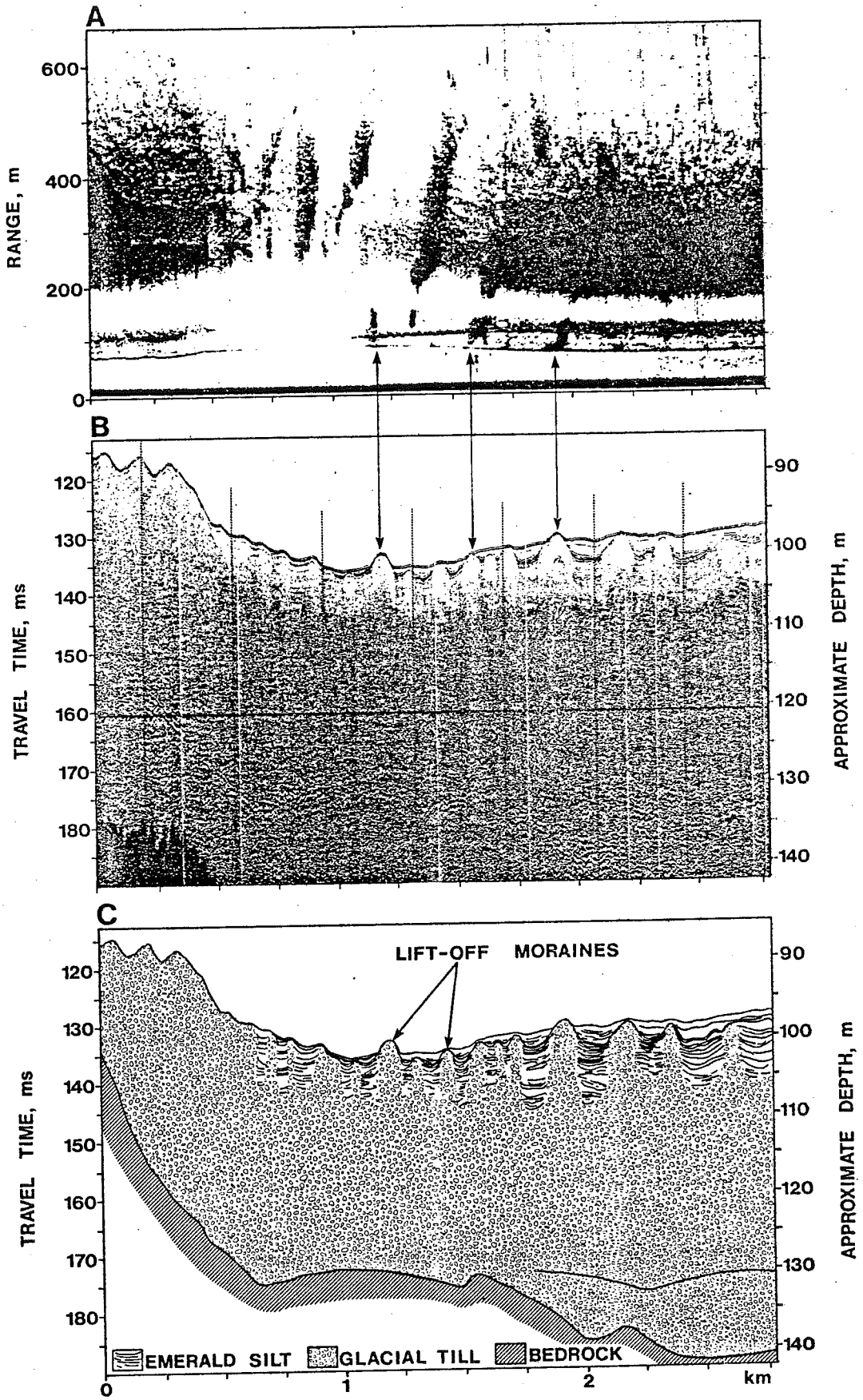

73°
71°
69°
67°
65°
63°
61°
59°
57°
55°
53°
51°
49°
44°
42°
40°
38°
U.S.A.
N.B.
P.E.I.
BAY OF FUNDY
NOVA SCOTIA
GULF OF MAINE
SCOTIAN SHELF
LAURENTIAN CHANNEL
NEWFOUNDLAND
NORTHEAST NEWFOUNDLAND SHELF
THE GRAND BANKS OF NEWFOUNDLAND
100
200
0 100 200
km
100 BATHYMETRY, m
TILL TONGUES

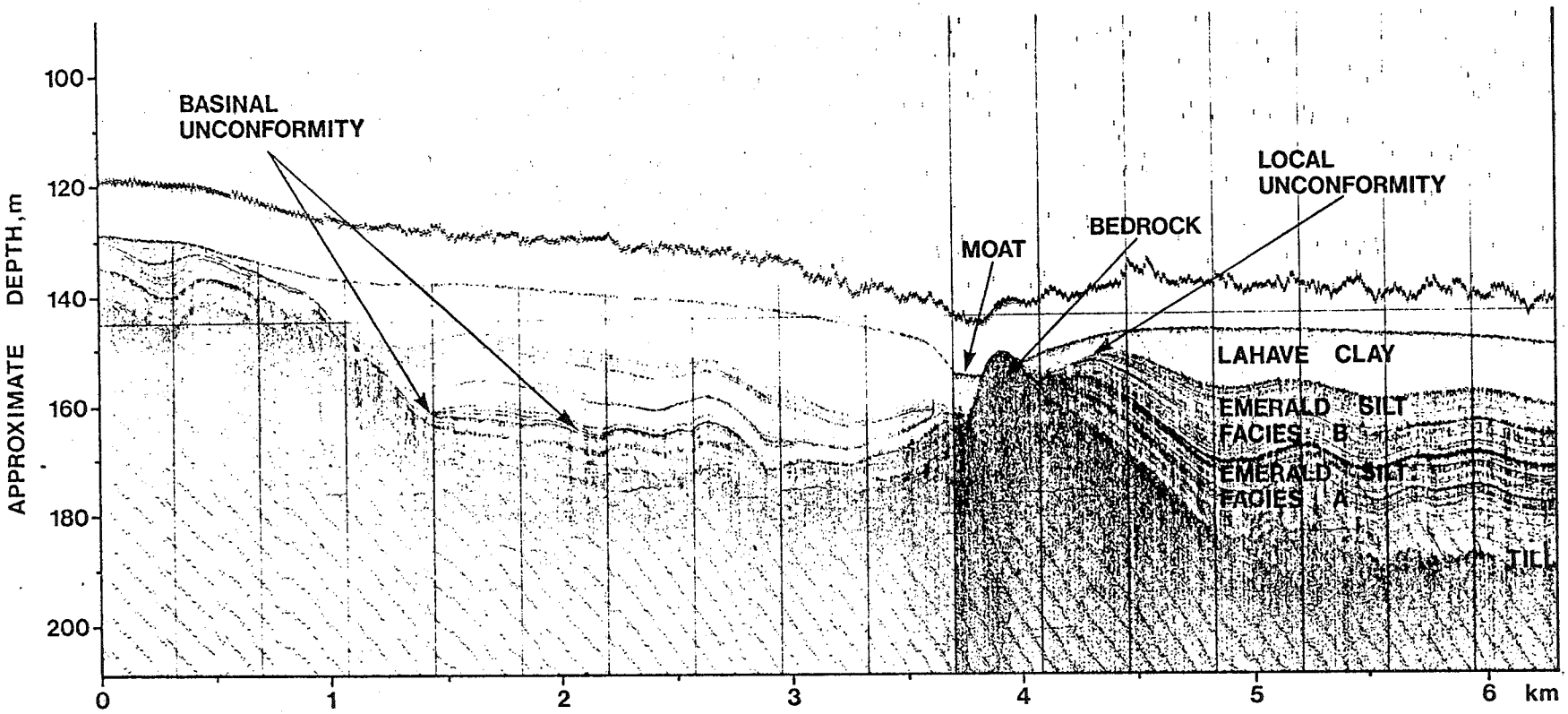
7a

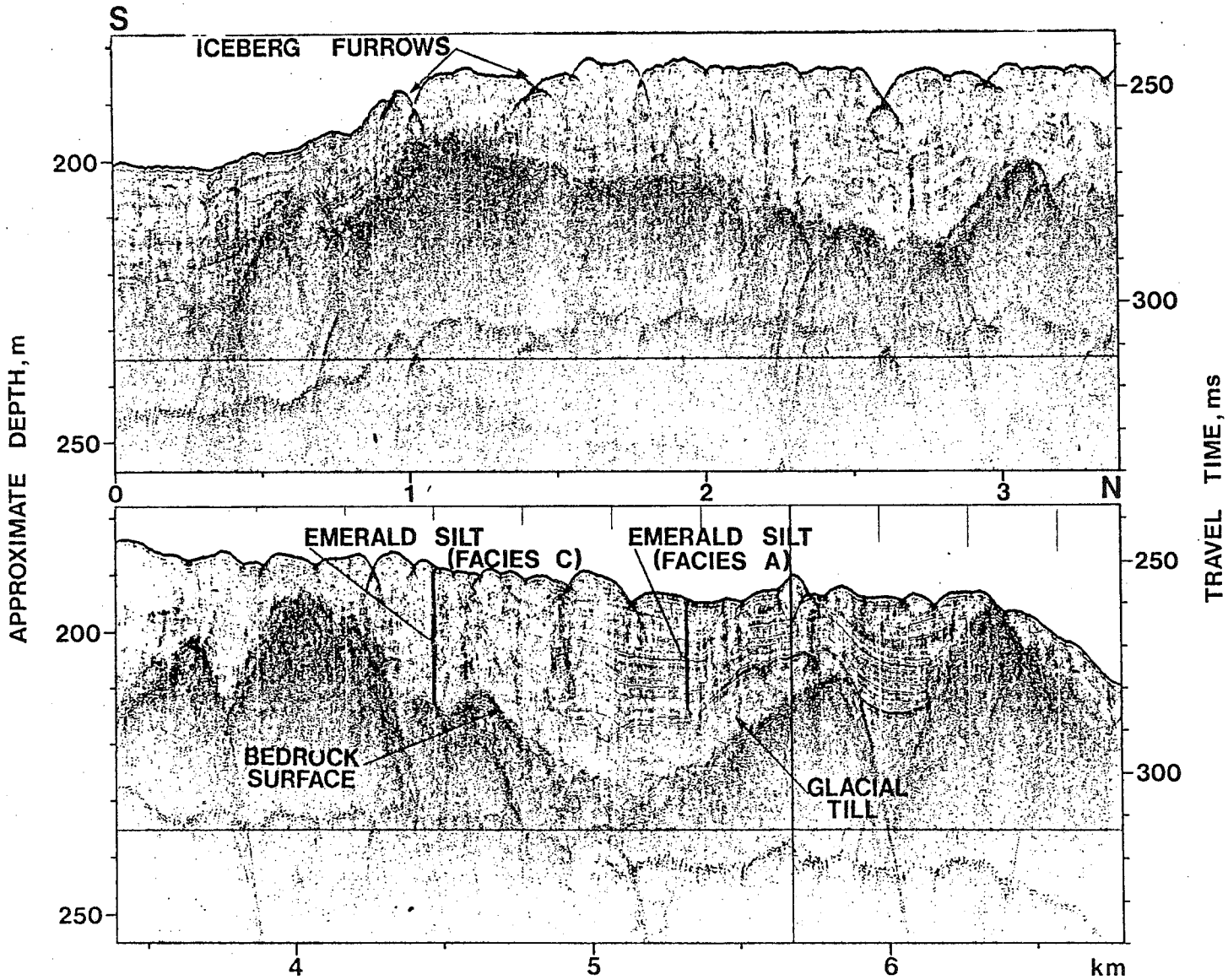

A
100
120
140
160
BEDROCK
UNCONFORMITY
EMERALD SILT
TILL
TILL
TILL TONGUE
BEDROCK
B
180
200
220
240
EMERALD SILT FACIES A
FEATHER EDGE
ICEBERG FURROWS
TILL
DROPSTONES
TILL TONGUE
EMERALD SILT FACIES B
APPROXIMATE DEPTH, m
C
200
220
240
260
FEATHER EDGE
ICEBERG FURROWS
LAHAVE CLAY
BEDROCK
TILL TONGUE
EMERALD SILT
EMERALD SILT
EMERALD SILT
D
120
140
160
180
FEATHER EDGE
TILL TONGUE
TILL TONGUE
EMERALD SILT FACIES A
TILL TONGUE
TILL TONGUE
EMERALD SILT FACIES A
0
1
2
3
4
5
6
7
8
km

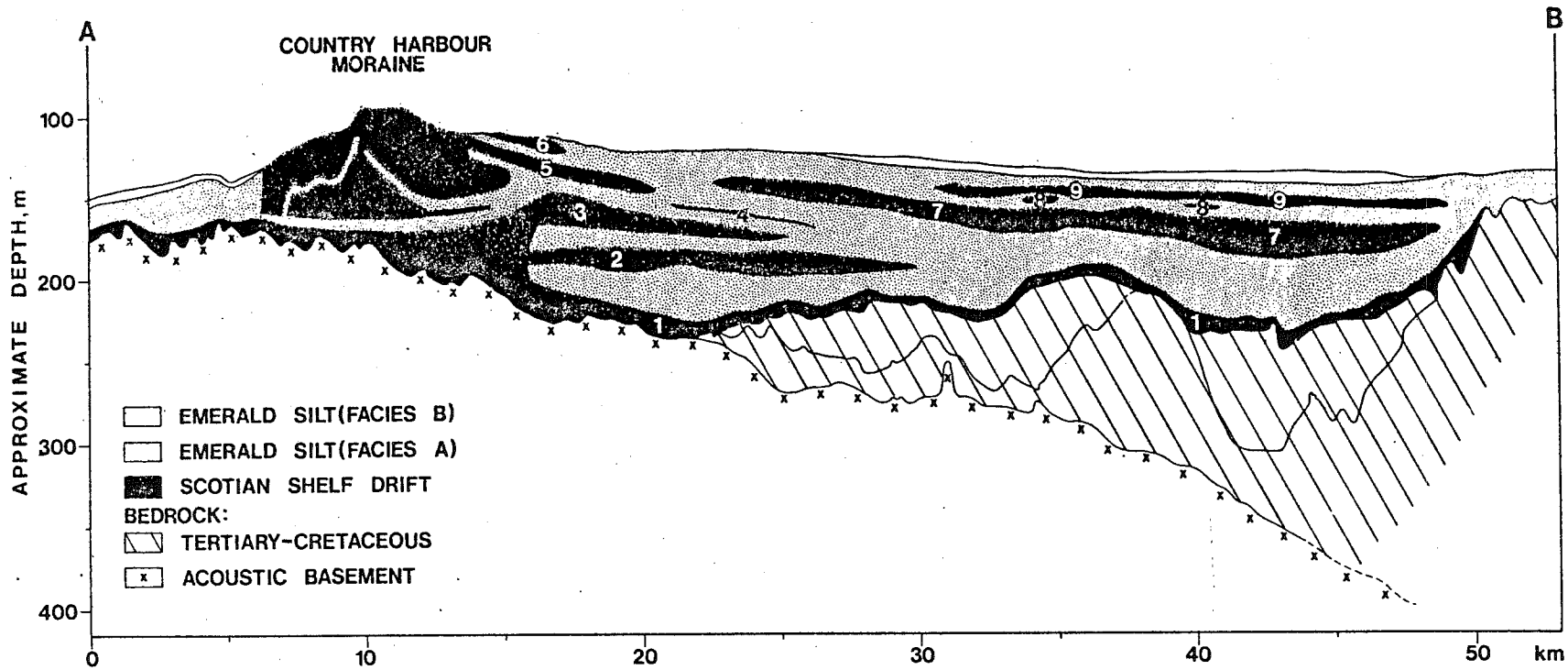
7 b

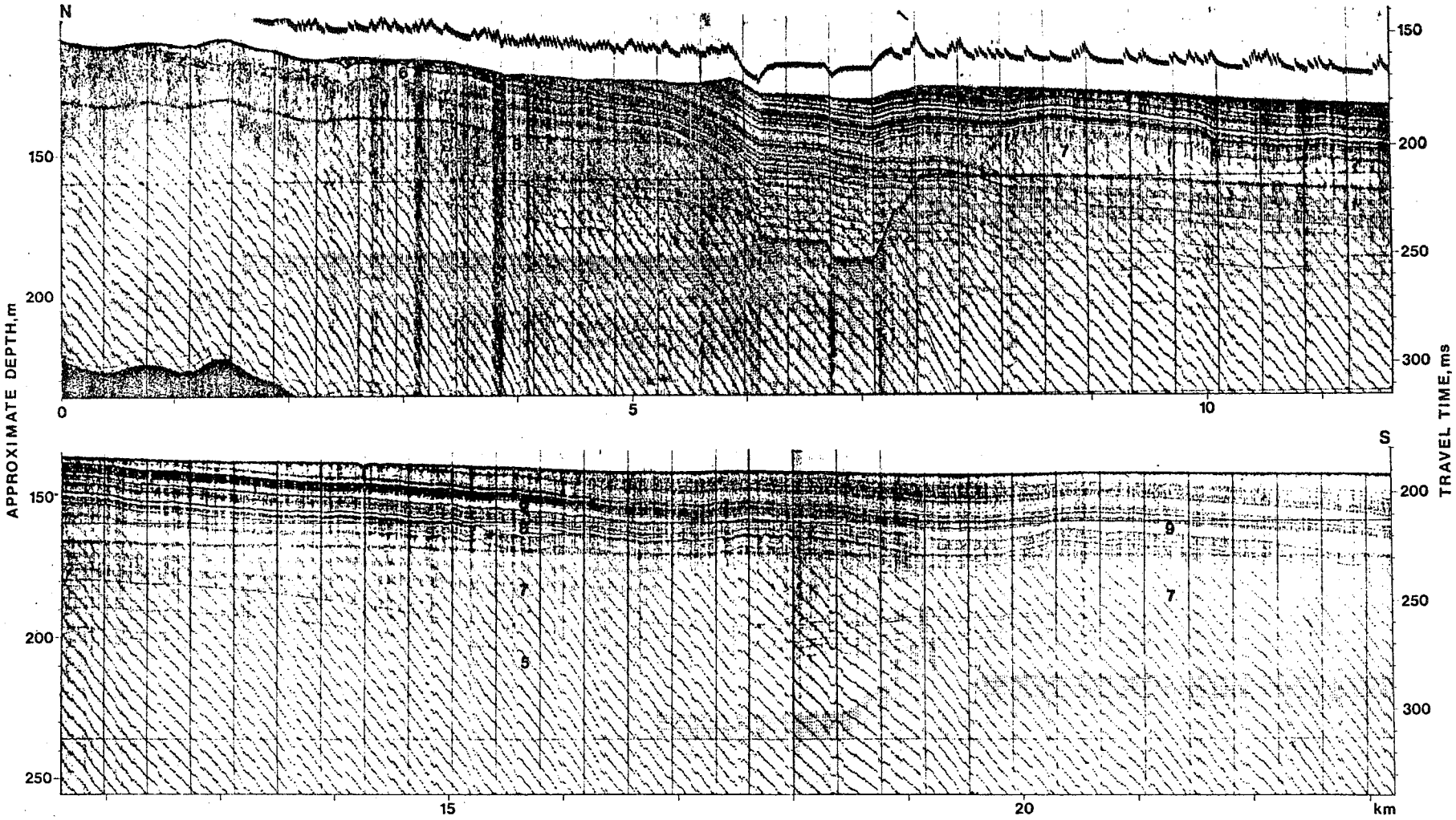

E
EASTERN SHORE MORAINE
LAHAVE CLAY
EMERALD SILT FACIES B
BEDROCK
APPROXIMATE DEPTH, m
160
180
200
220
240
260
0
1
2
3
4
5
6
7 km
F
LAURENTIAN MORAINE
FEATHER EDGE
EMERALD SILT FACIES A
EMERALD SILT FACIES B
LAHAVE CLAY
TILL TONGUE
BEDROCK
440
460
480
500
0
1
2
km
G
BURIN MORAINE
FEATHER EDGE
LAHAVE CLAY
EMERALD SILT
TILL TONGUE
BEDROCK
200
220
240
0
1
2
3
4
5
6 km

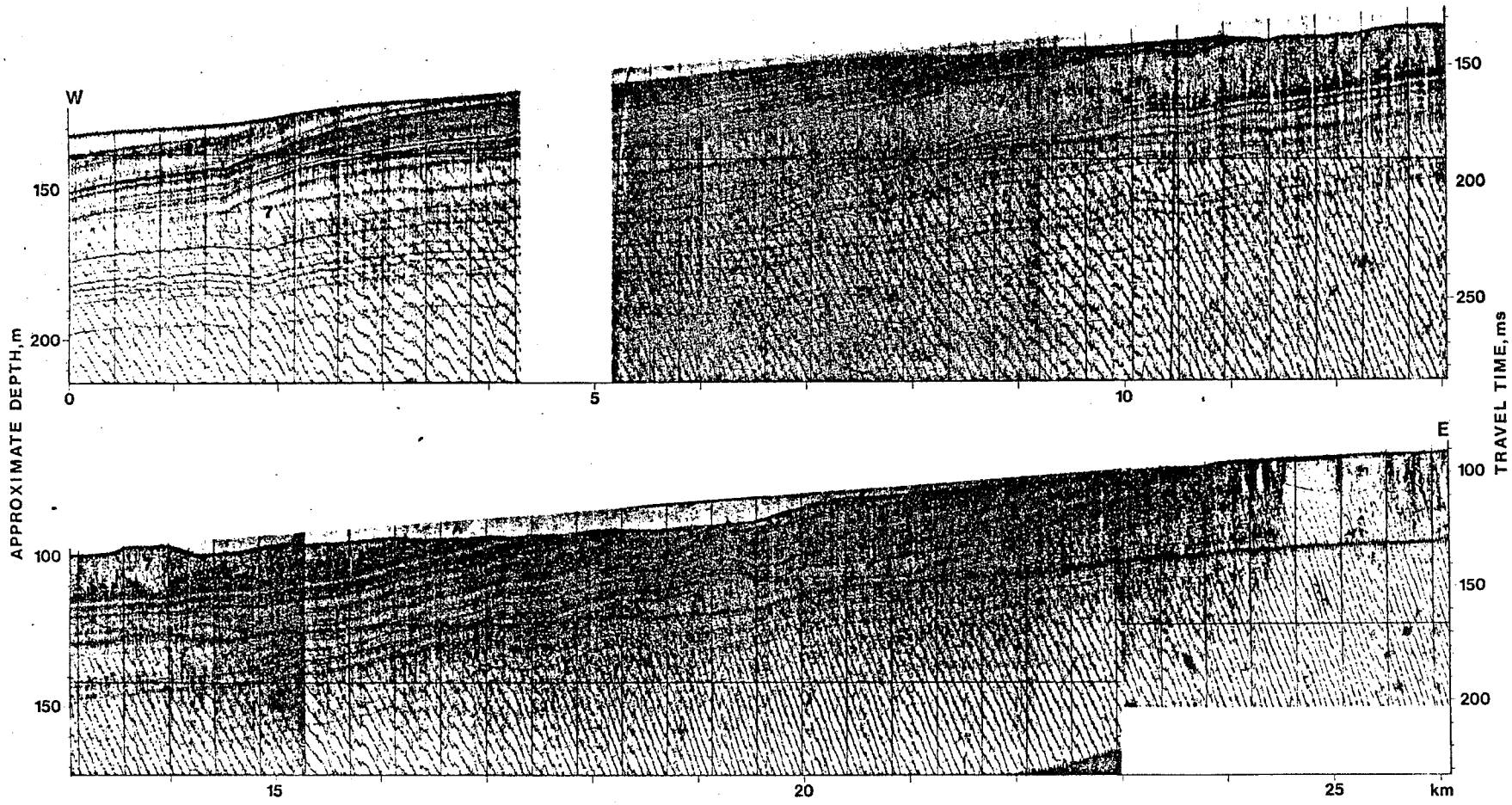

A
RANGE, m
600
400
200
0
B
TRAVEL TIME, ms
120
130
140
150
160
170
180
APPROXIMATE DEPTH, m
90
100
110
120
130
140
C
LIFT-OFF MORAINES
EMERALD SILT
GLACIAL TILL
BEDROCK
0
1
2
km

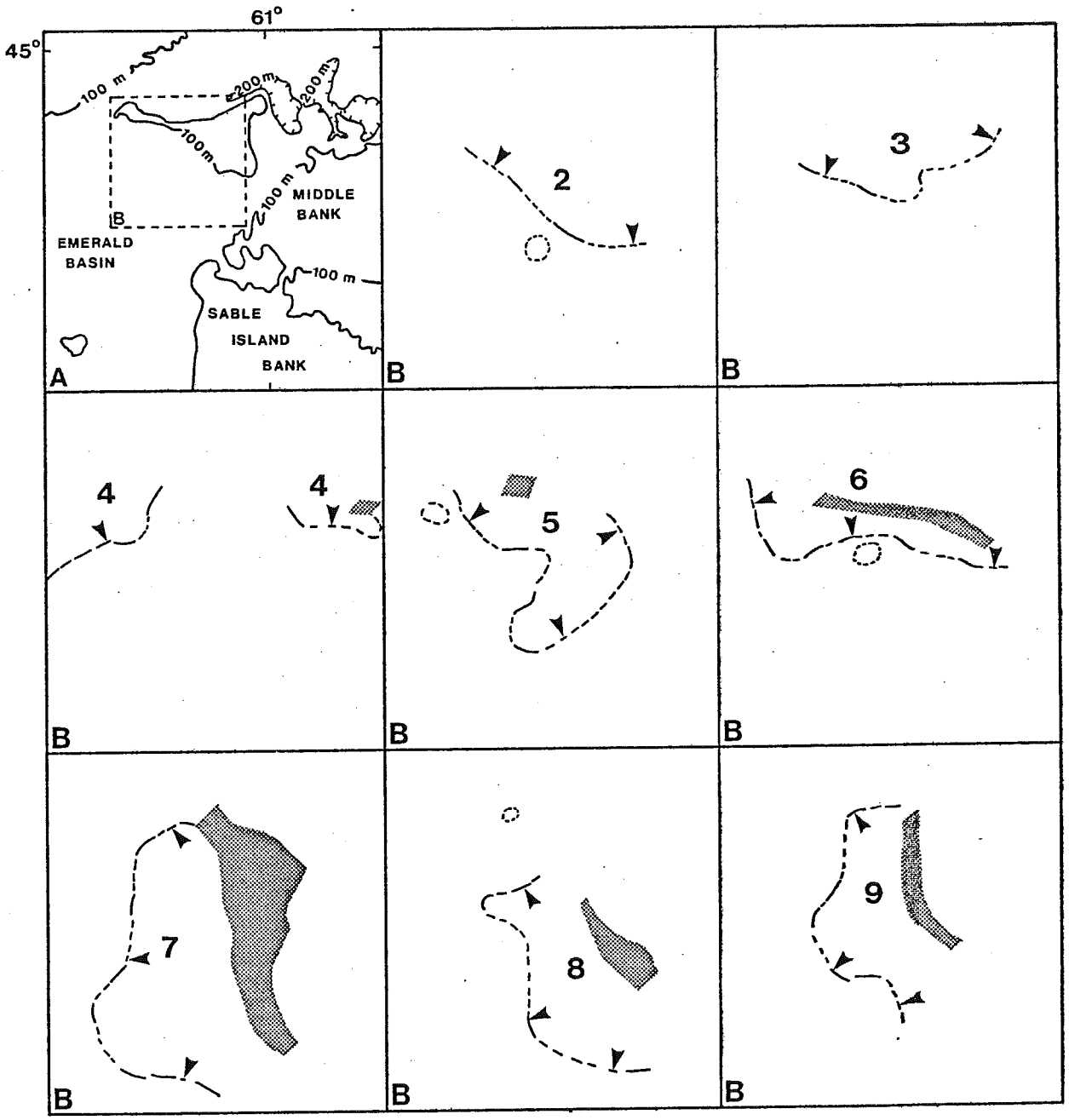

BASINAL
UNCONFORMITY
LOCAL
UNCONFORMITY
BEDROCK
MOAT
LAHAVE CLAY
EMERALD SILT
FACIES B
EMERALD SILT
FACIES A
TILL
APPROXIMATE DEPTH, m
100
120
140
160
180
200
0
1
2
3
4
5
6
km


S
ICEBERG FURROWS
200
250
300
0
1
2
3
N
EMERALD SILT (FACIES C)
EMERALD SILT (FACIES A)
BEDROCK SURFACE
GLACIAL TILL
4
5
6
km
APPROXIMATE DEPTH, m
TRAVEL TIME, ms


A
B
COUNTRY HARBOUR
MORAINE
APPROXIMATE DEPTH, m
100
200
300
400
0
10
20
30
40
50
km
1
2
3
4
5
6
7
8
9
EMERALD SILT(FACIES B)
EMERALD SILT(FACIES A)
SCOTIAN SHELF DRIFT
BEDROCK:
TERTIARY-CRETACEOUS
ACOUSTIC BASEMENT


N
S
APPROXIMATE DEPTH, m
TRAVEL TIME, ms
150
200
250
300
0
5
10
15
20
km
6
5
7
8
9


W
E
APPROXIMATE DEPTH, m
TRAVEL TIME, ms
150
200
100
150
200
250
100
150
200
0
5
10
15
20
25
km

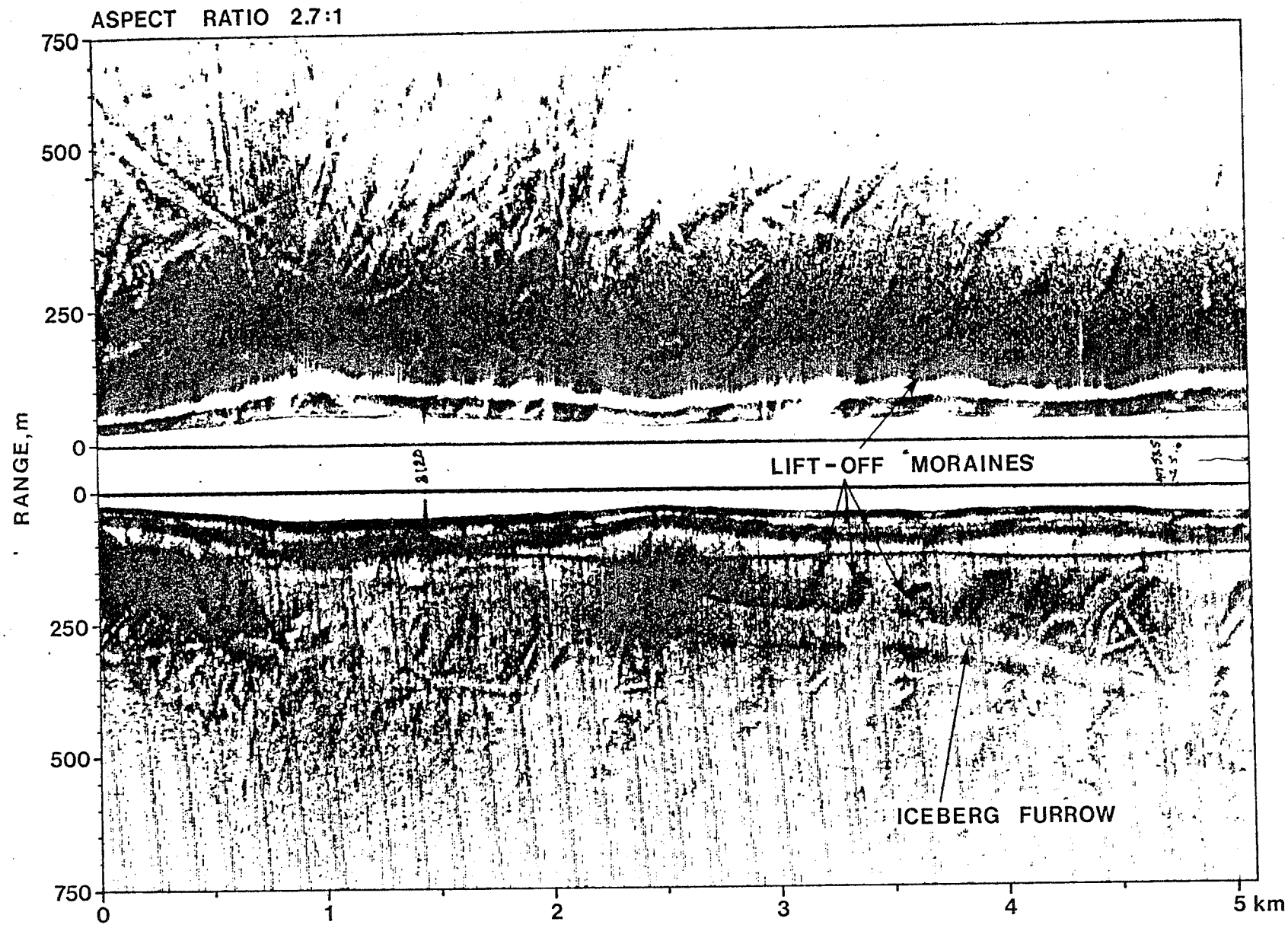


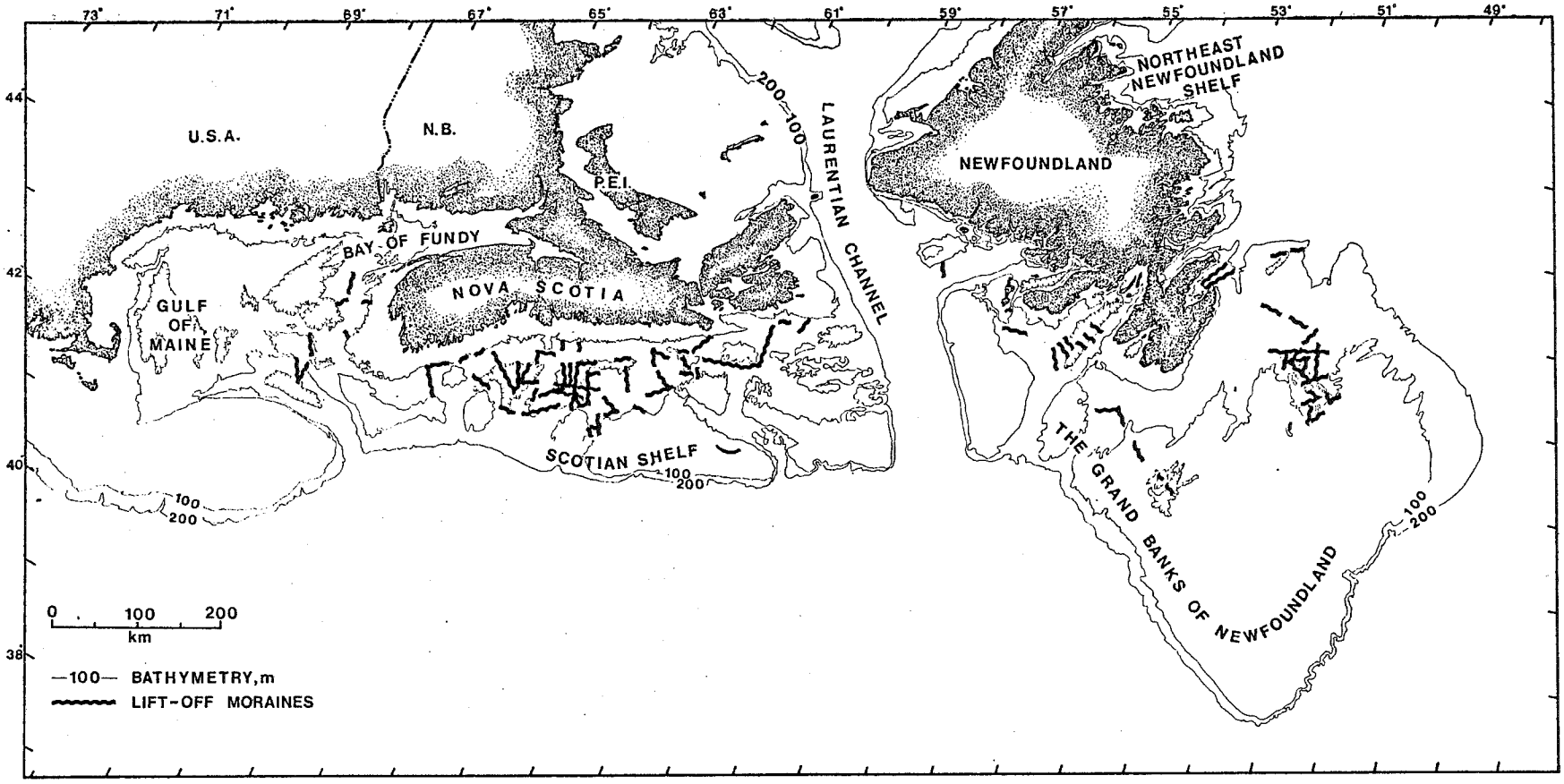

FEATHER EDGE OF TILL TONGUES: DEFINED ———
APPROXIMATE - - - - - -

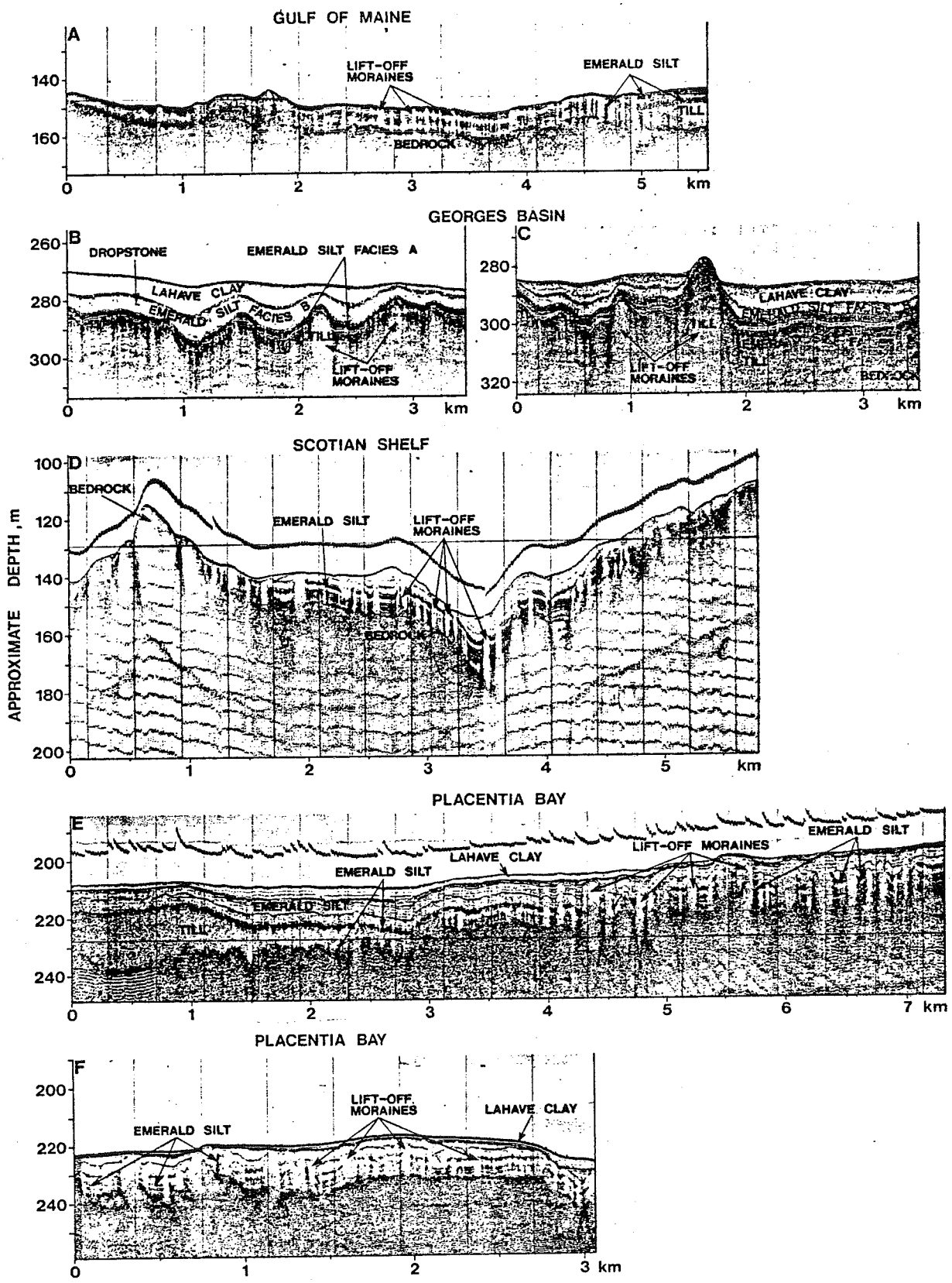
ERODED TILL TONGUE AT SEABED

DIRECTION OF TILL TONGUE THINNING

TILL TONGUE NUMBER 2


ASPECT RATIO 2.7:1
750
500
250
0
0
250
500
750
RANGE, m
LIFT-OFF MORAINES
ICEBERG FURROW
0
1
2
3
4
5 km


73°
71°
69°
67°
65°
63°
61°
59°
57°
55°
53°
51°
49°
44
42
40
38
U.S.A.
N.B.
P.E.I.
BAY OF FUNDY
NOVA SCOTIA
GULF OF MAINE
SCOTIAN SHELF
LAURENTIAN CHANNEL
NEWFOUNDLAND
NORTHEAST NEWFOUNDLAND SHELF
THE GRAND BANKS OF NEWFOUNDLAND
100
200
0 100 200
km
—100— BATHYMETRY, m
LIFT-OFF MORAINES


GULF OF MAINE
A
LIFT-OFF MORAINES
EMERALD SILT
TILL
BEDROCK
140
160
0
1
2
3
4
5
km
GEORGES BASIN
B
DROPSTONE
EMERALD SILT FACIES A
LAHAVE CLAY
EMERALD SILT FACIES B
TILL
LIFT-OFF MORAINES
260
280
300
0
1
2
3
km
C
LAHAVE CLAY
TILL
LIFT-OFF MORAINES
BEDROCK
280
300
320
0
1
2
3
km
SCOTIAN SHELF
D
BEDROCK
EMERALD SILT
LIFT-OFF MORAINES
BEDROCK
APPROXIMATE DEPTH, m
100
120
140
160
180
200
0
1
2
3
4
5
km
PLACENTIA BAY
E
EMERALD SILT
LIFT-OFF MORAINES
LAHAVE CLAY
EMERALD SILT
EMERALD SILT
TILL
200
220
240
0
1
2
3
4
5
6
7 km
PLACENTIA BAY
F
LIFT-OFF MORAINES
LAHAVE CLAY
EMERALD SILT
200
220
240
0
1
2
3 km

LITHOLOGY:

- MUD
- SANDY MUD
- GRAVELLY-SANDY MUD
- SAND
- SHELLS
- PYRITIZED ORGANIC FRAGMENTS

X-RADIOGRAPHIC DESCRIPTION:

- RHYTHMIC BANDS, ALTERNATING SILT AND CLAY
- GRAVEL

SEDIMENTARY STRUCTURES:

- STRONG / WEAK — RHYTHMICALLY BANDED
- STRONG / WEAK — RHYTHMICALLY BANDED, WISPY
- STRONG / WEAK — LAMINATED
- BIOTURBATED

SEISMIC EVENTS:

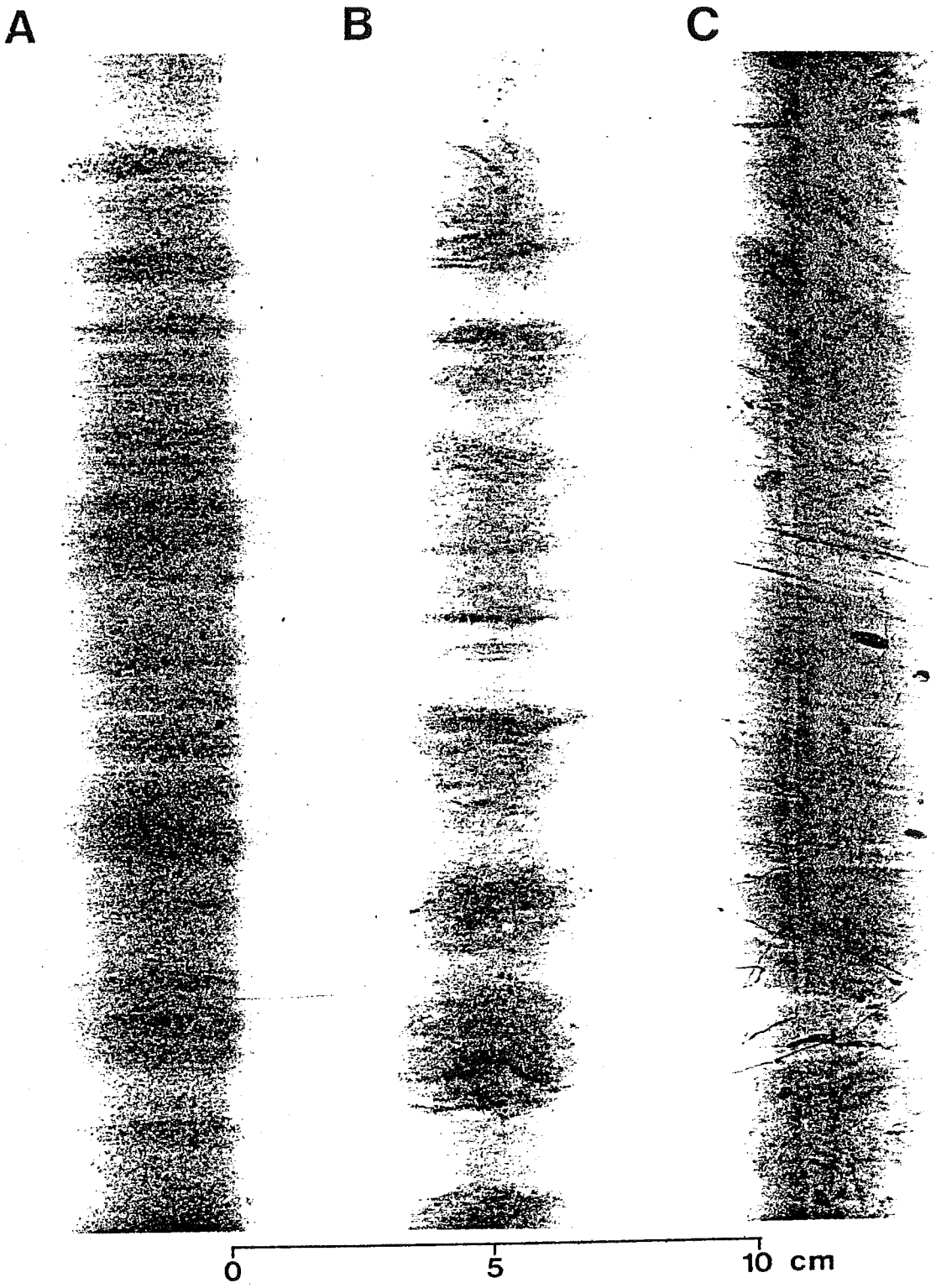
- STRONG
- MODERATE
- WEAK
- UNCONFORMITY

GRAIN SIZE:

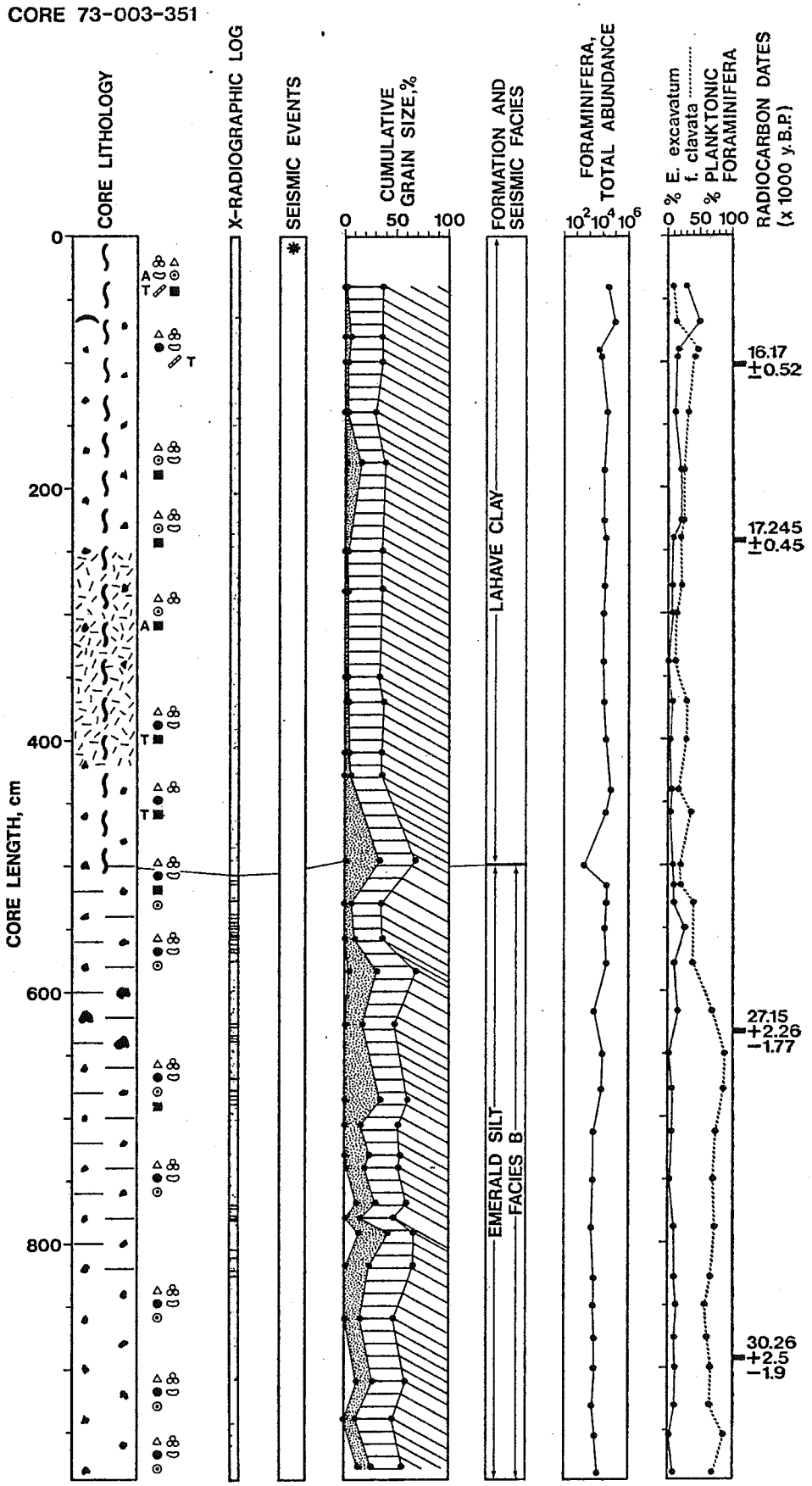
- CLAY
- SILT
- SAND
- GRAVEL

SAND FRACTION MICROSCOPIC ANALYSIS:

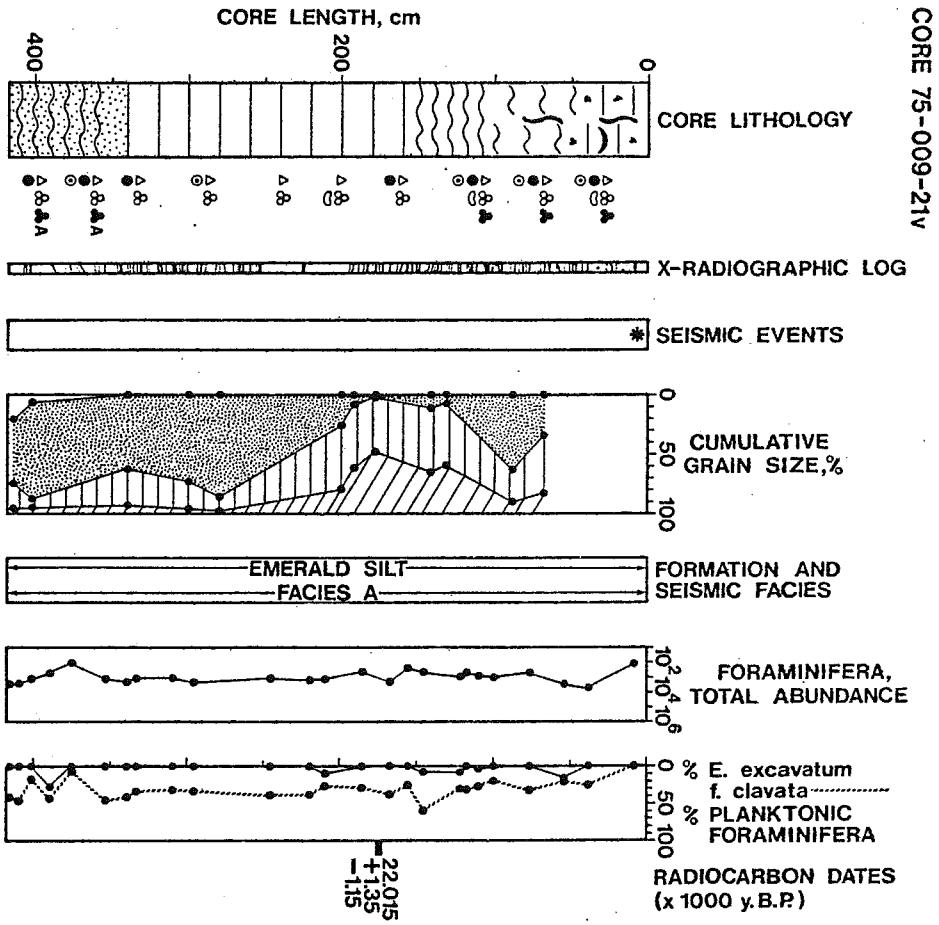
- FORAMINIFERA
- ERODED FORAMINIFERA
- OSTRACODS
- CARBONACEOUS FRAGMENTS
- SILICLASTICS (QUARTZ, FELDSPAR, MICA ETC.)
- QUARTZ WITH OVERGROWTH
- LITHIC FRAGMENTS
- PYRITE FRAGMENTS AND CONCRETIONS
- T TRACE
- A ABUNDANT

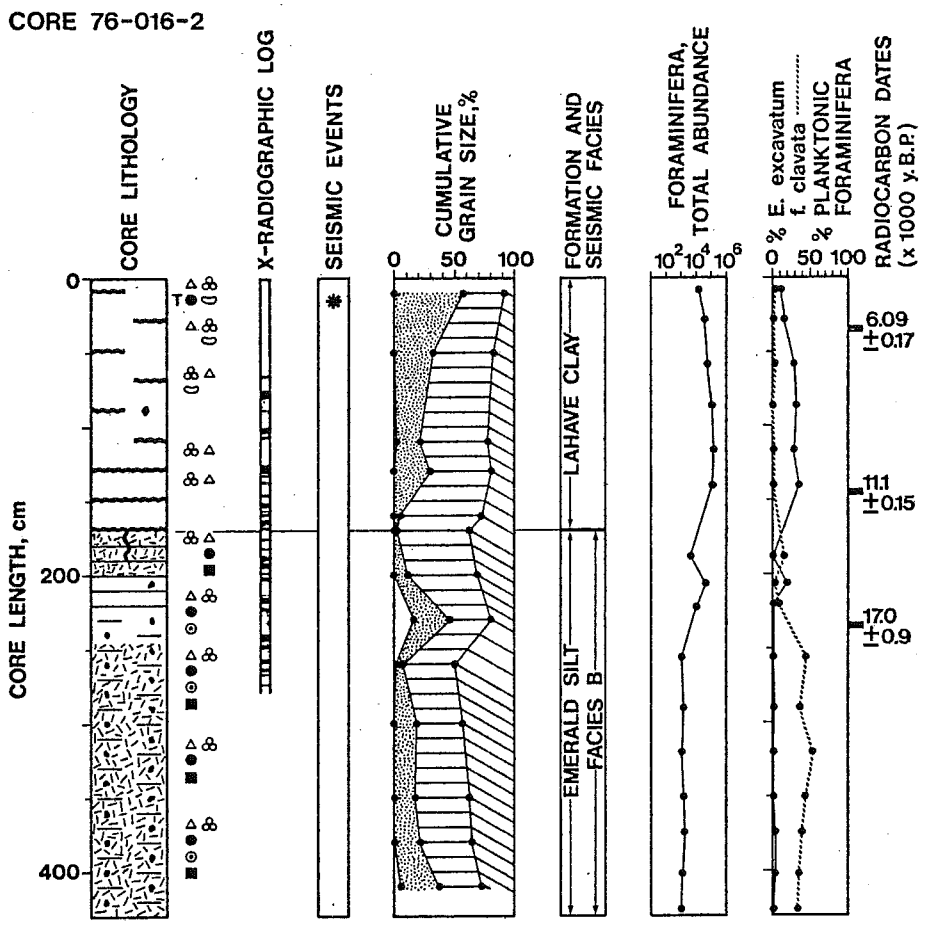


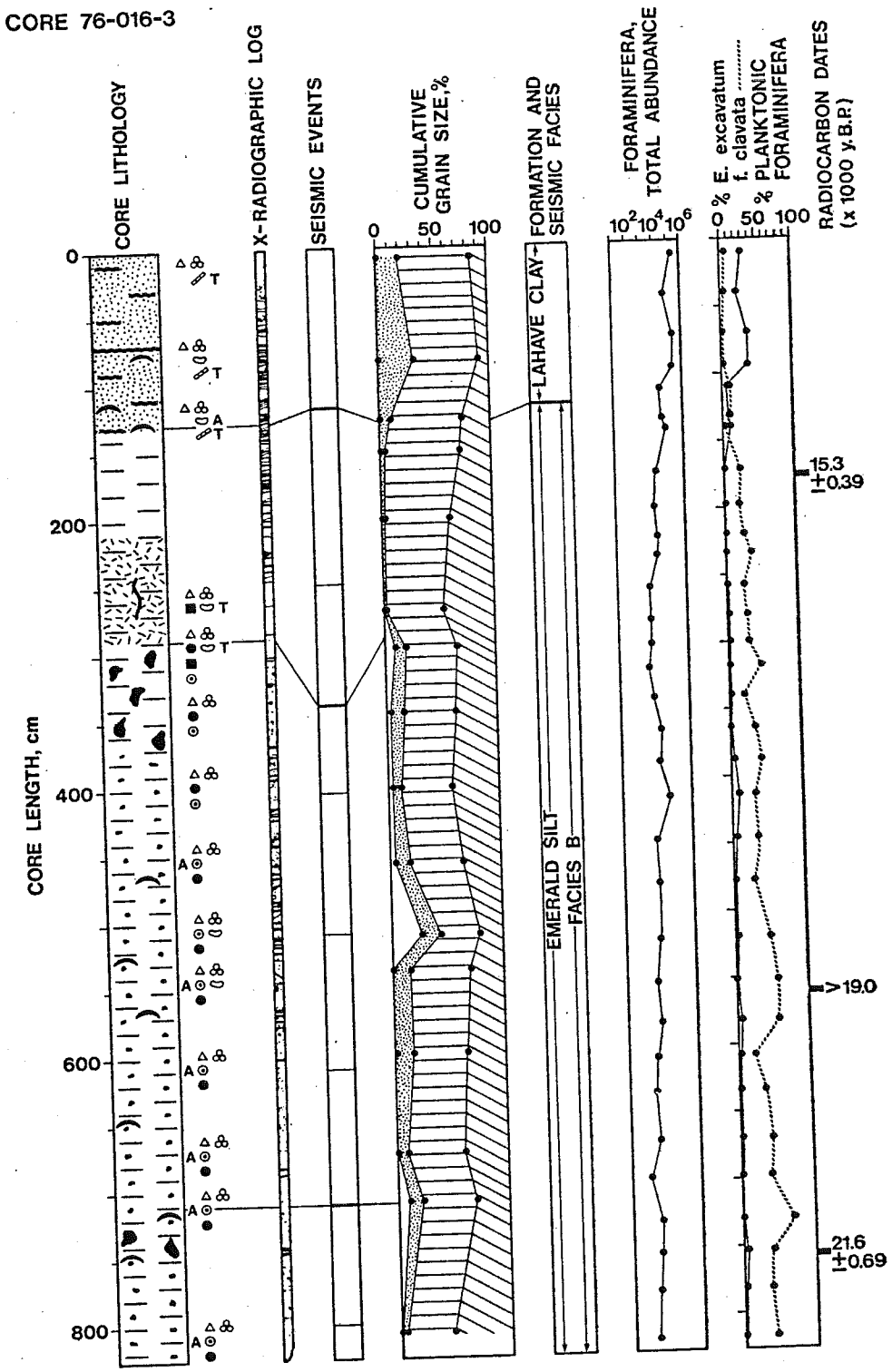

17aa

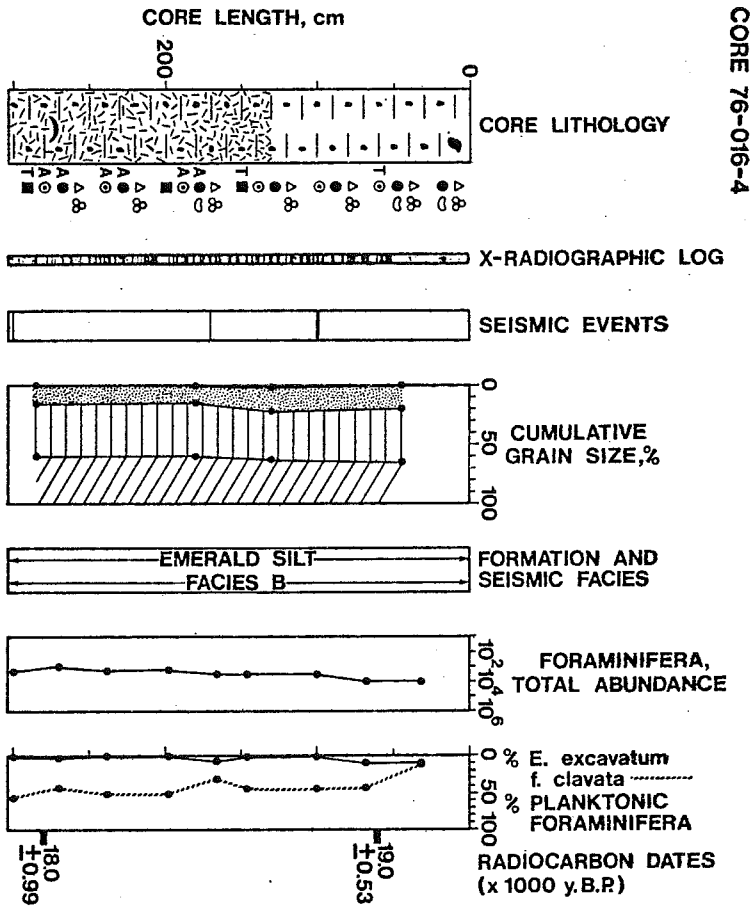

CORE 73-003-351
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0 50 100
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$ $10^4$ $10^6$
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
0 50 100
RADIOCARBON DATES (x 1000 y.B.P.)
CORE LENGTH, cm
0
200
400
600
800
A
T
LAHAVE CLAY
EMERALD SILT
FACIES B
16.17 ±0.52
17.245 ±0.45
27.15 +2.26 −1.77
30.26 +2.5 −1.9

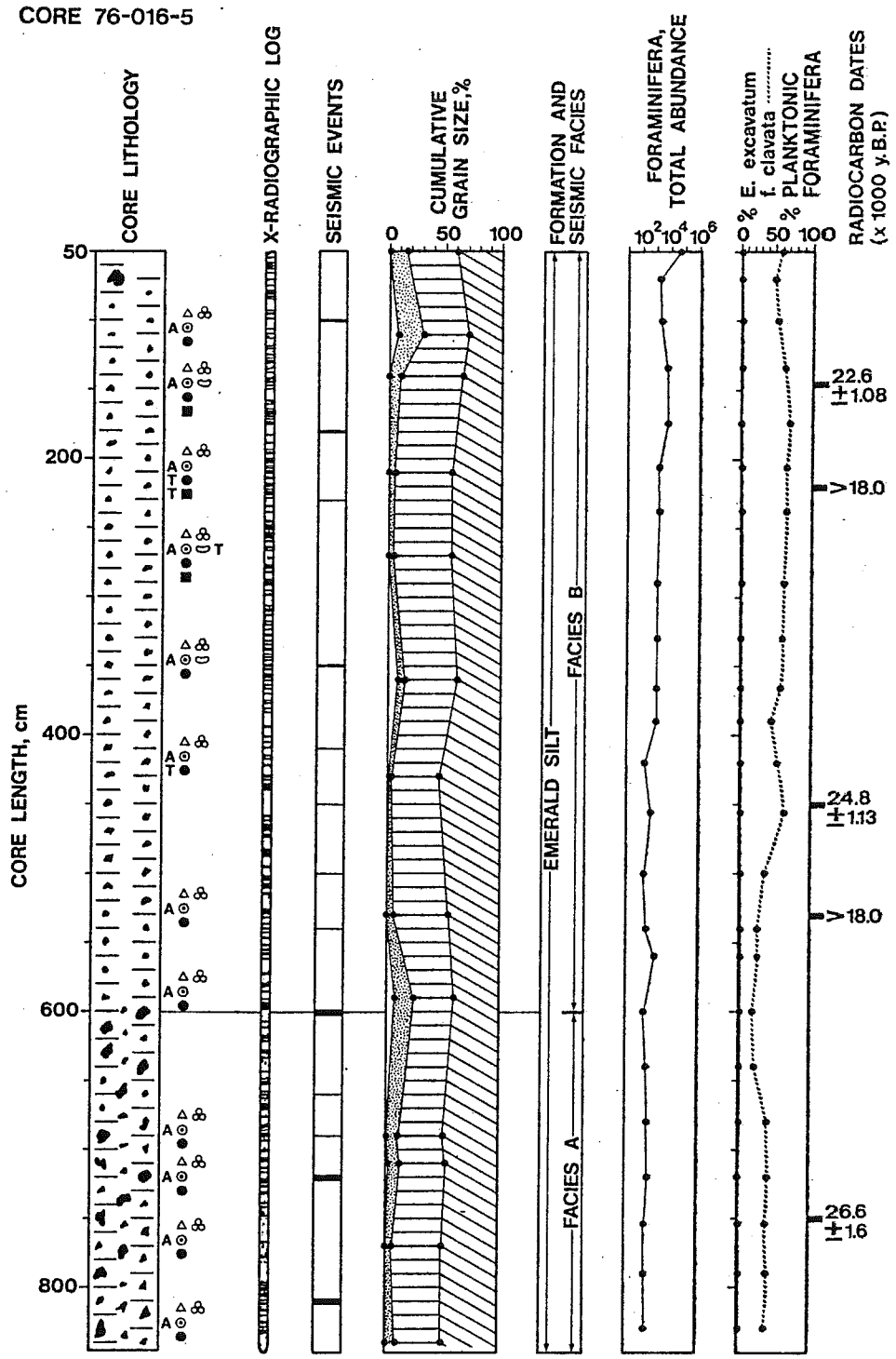
17 b

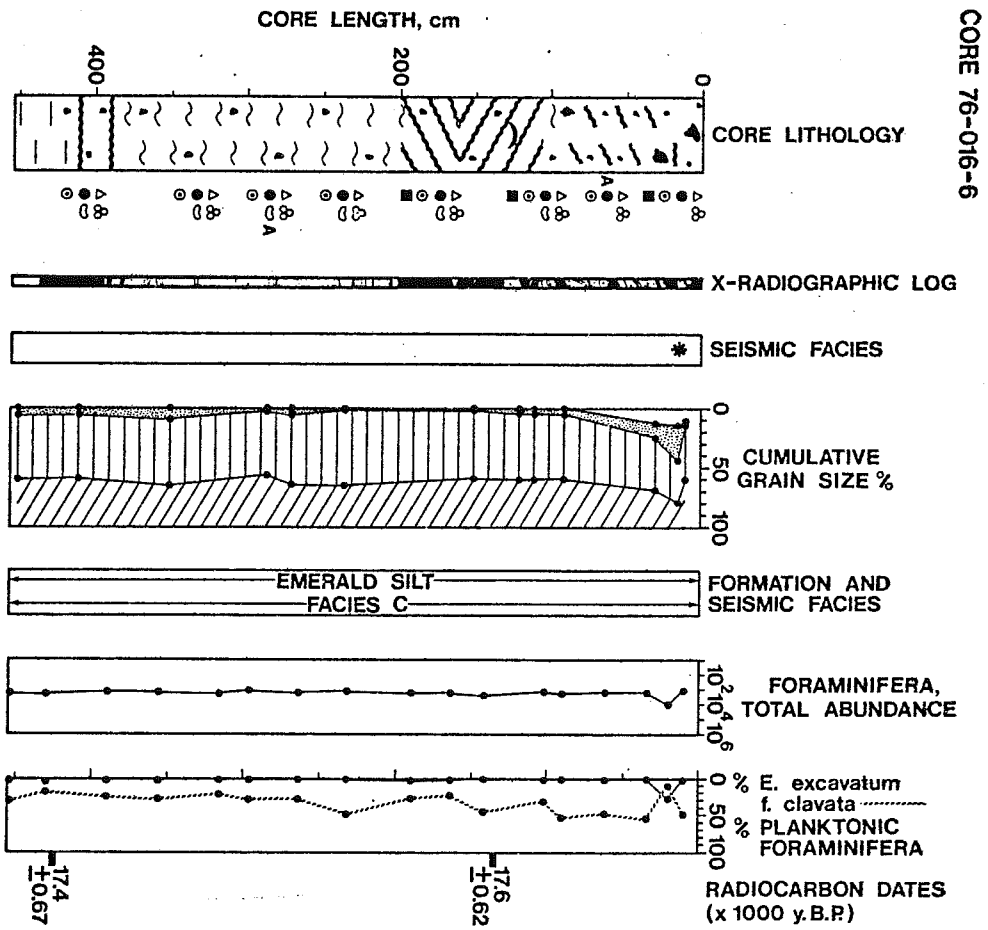

CORE 75-009-21v
CORE LENGTH, cm
400
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0
50
100
EMERALD SILT
FACIES A
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$
$10^4$
$10^6$
% E. excavatum f. clavata
% PLANKTONIC FORAMINIFERA
22,015 +1.35 −1.15
RADIOCARBON DATES (x 1000 y.B.P.)

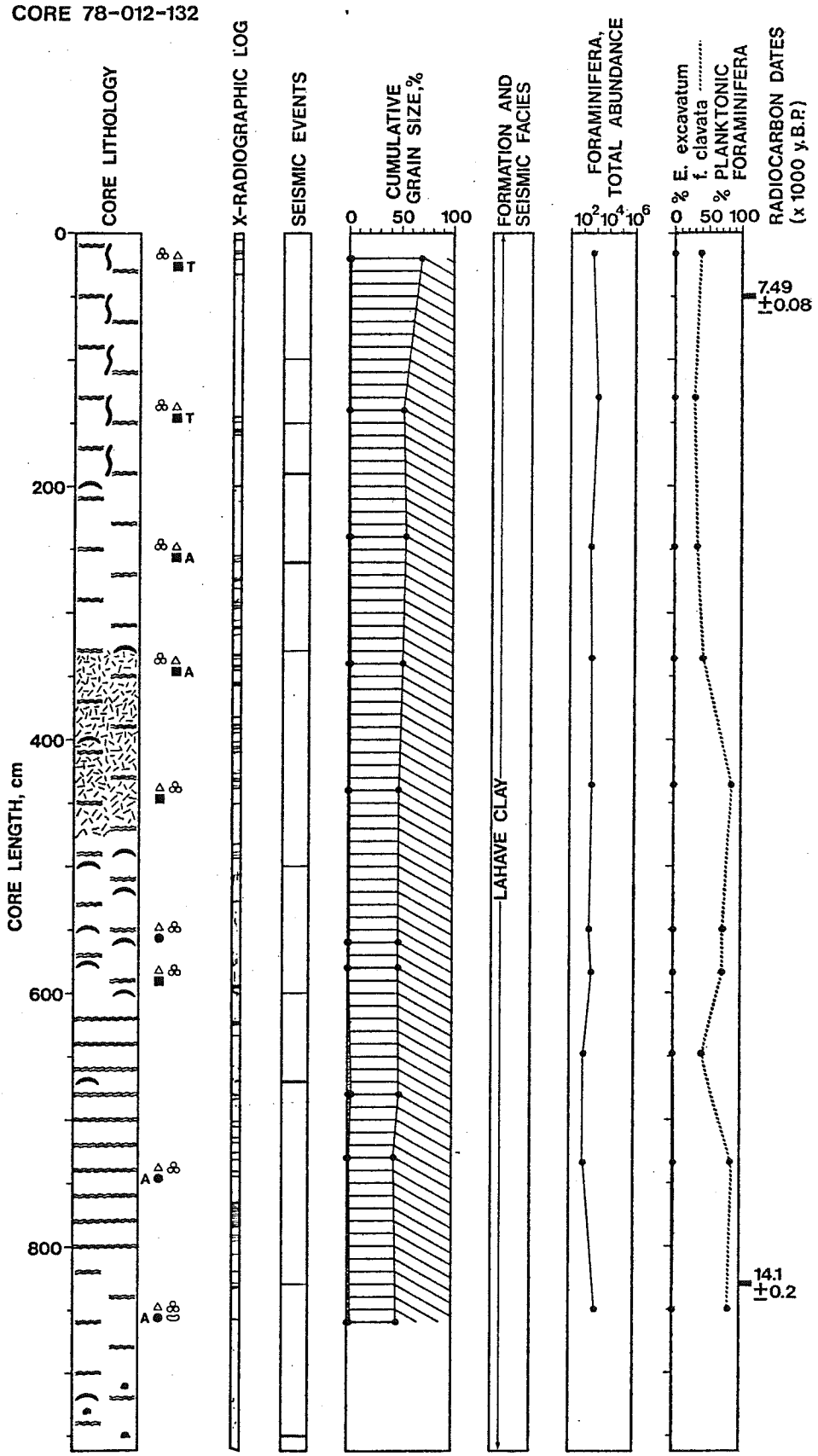

CORE 76-016-2
CORE LENGTH, cm
0
200
400
CORE LITHOLOGY
T
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0 50 100
FORMATION AND SEISMIC FACIES
LAHAVE CLAY
EMERALD SILT
FACIES B
FORAMINIFERA, TOTAL ABUNDANCE
10^2 10^4 10^6
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
0 50 100
RADIOCARBON DATES (x 1000 y.B.P.)
6.09 ±0.17
11.1 ±0.15
17.0 ±0.9

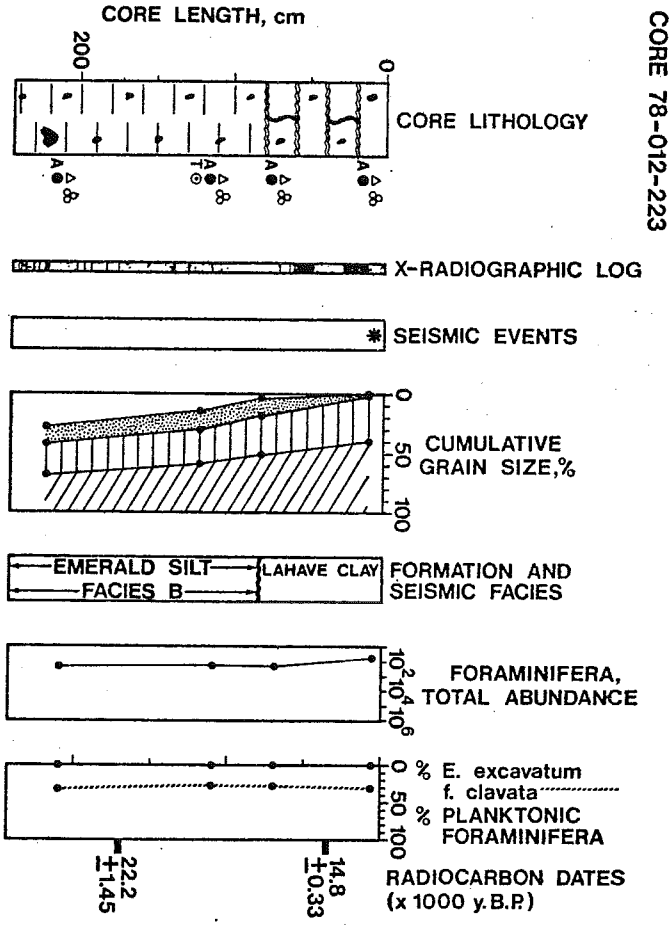

CORE 76-016-3
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0 50 100
FORMATION AND SEISMIC FACIES
LAHAVE CLAY
EMERALD SILT
FACIES B
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$ $10^4$ $10^6$
% E. excavatum f. clavata
% PLANKTONIC FORAMINIFERA
0 50 100
RADIOCARBON DATES (x 1000 y.B.P.)
15.3 ±0.39
>19.0
21.6 ±0.69
CORE LENGTH, cm
0
200
400
600
800
T
A

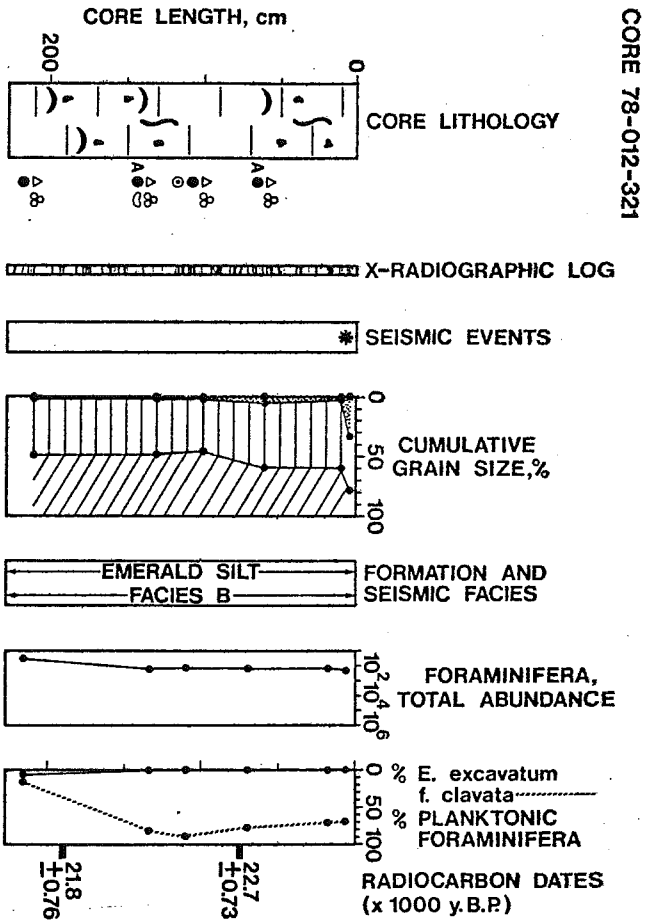

CORE 76-016-4
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
0
50
100
CUMULATIVE GRAIN SIZE,%
EMERALD SILT
FACIES B
FORMATION AND SEISMIC FACIES
$10^2$
$10^4$
$10^6$
FORAMINIFERA, TOTAL ABUNDANCE
0
50
100
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
18.0 ±0.99
19.0 ±0.53
RADIOCARBON DATES (x 1000 y.B.P.)

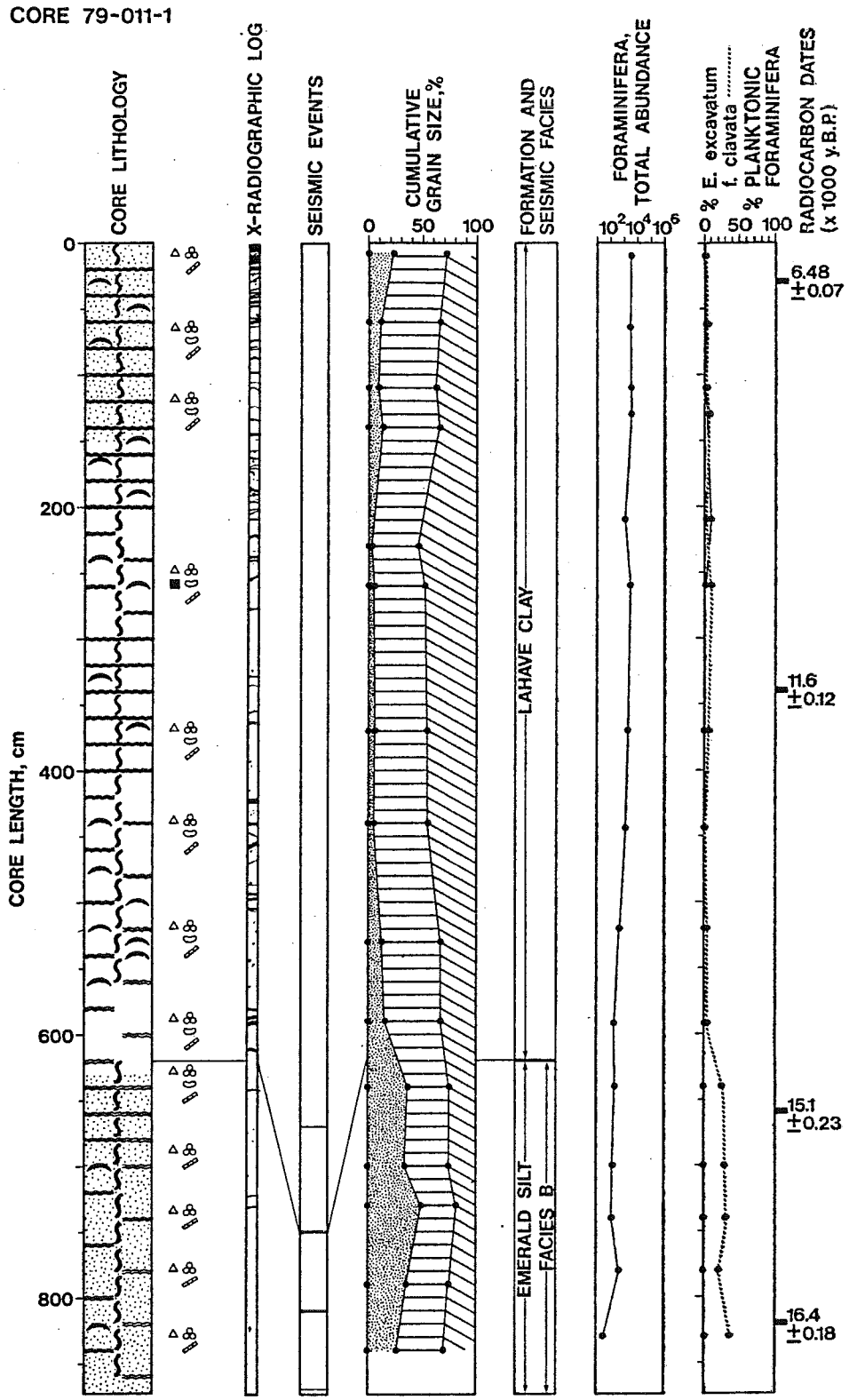

CORE 76-016-5
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0 50 100
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$ $10^4$ $10^6$
% E. excavatum
% f. clavata
% PLANKTONIC FORAMINIFERA
0 50 100
RADIOCARBON DATES (x 1000 y.B.P.)
CORE LENGTH, cm
50
200
400
600
800
EMERALD SILT
FACIES B
FACIES A
22.6 ±1.08
>18.0
24.8 ±1.13
>18.0
26.6 ±1.6

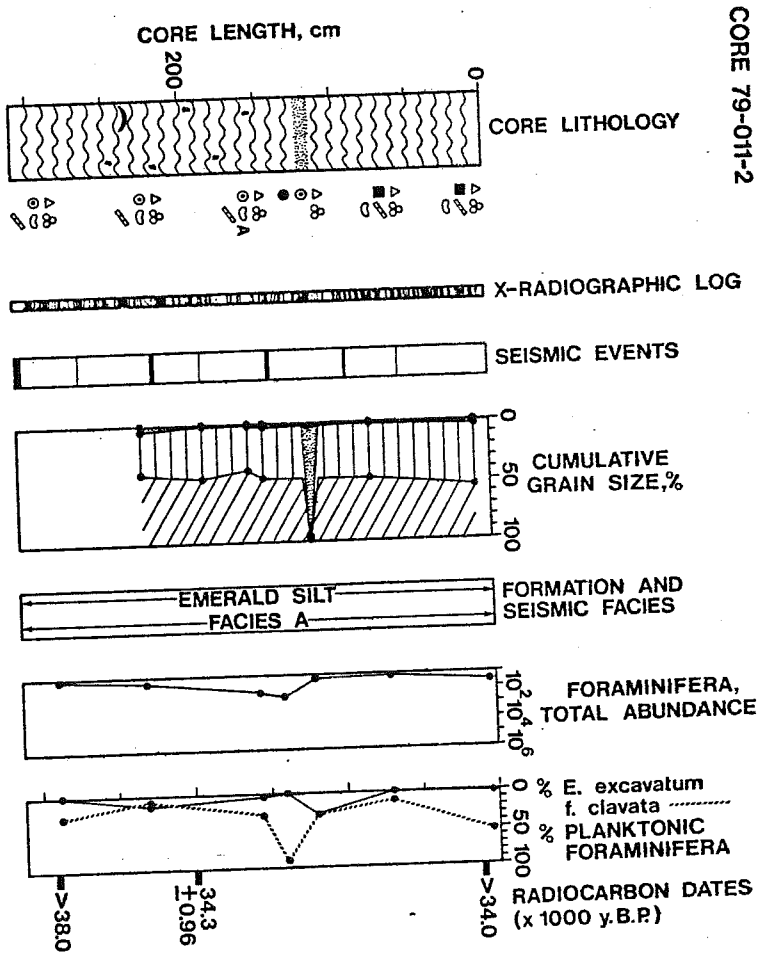

CORE 76-016-6
CORE LENGTH, cm
0
200
400
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC FACIES
CUMULATIVE GRAIN SIZE %
0
50
100
EMERALD SILT
FACIES C
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$
$10^4$
$10^6$
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
0
50
100
17.4 ± 0.67
17.6 ± 0.62
RADIOCARBON DATES (x 1000 y.B.P.)

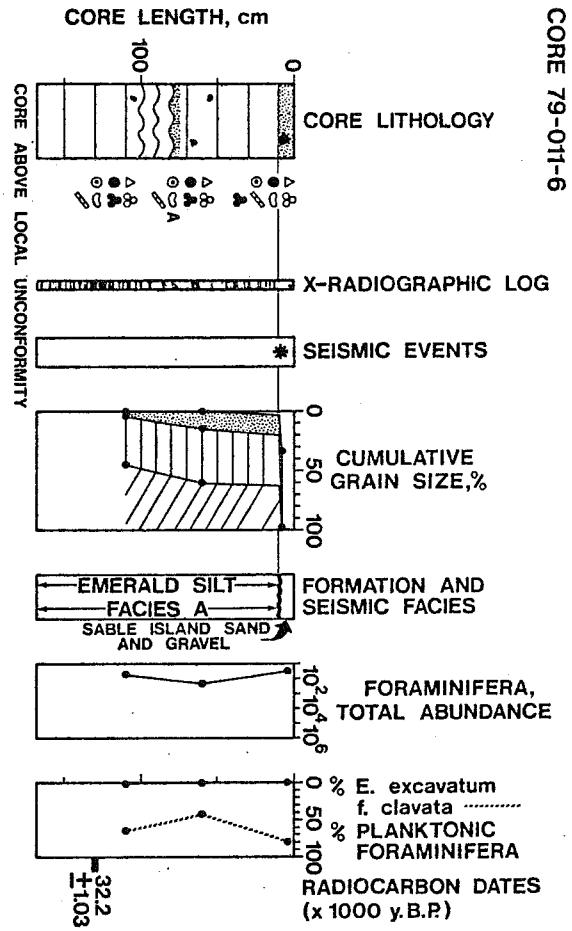

CORE 78-012-132
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
RADIOCARBON DATES (x 1000 y.B.P.)
CORE LENGTH, cm
0
200
400
600
800
0
50
100
$10^2$ $10^4$ $10^6$
LAHAVE CLAY
7.49 ±0.08
14.1 ±0.2
T
A

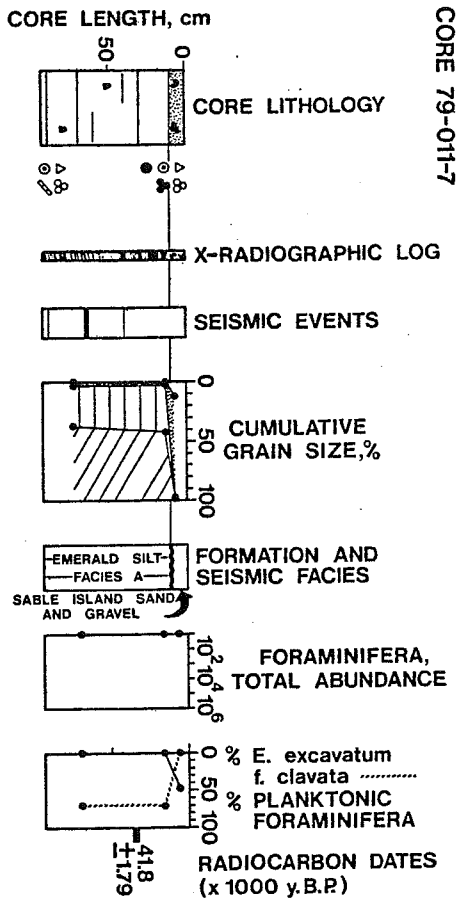

CORE 78-012-223
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0
50
100
EMERALD SILT
FACIES B
LAHAVE CLAY
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$
$10^4$
$10^6$
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
22.2 ±1.45
14.8 ±0.33
RADIOCARBON DATES (x 1000 y.B.P.)

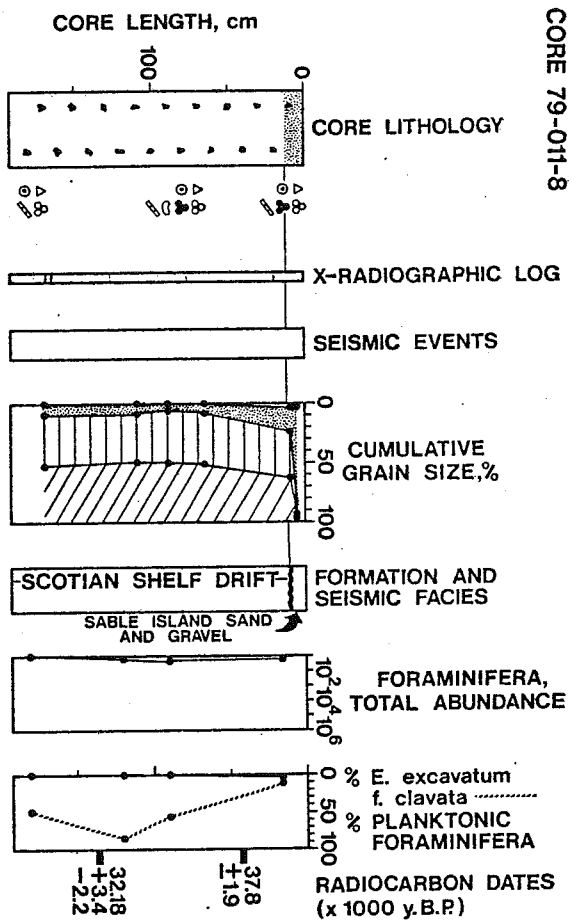

CORE 78-012-321
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
0
50
100
CUMULATIVE GRAIN SIZE,%
EMERALD SILT
FACIES B
FORMATION AND SEISMIC FACIES
$10^2$
$10^4$
$10^6$
FORAMINIFERA, TOTAL ABUNDANCE
0 % E. excavatum
f. clavata
50 100 % PLANKTONIC FORAMINIFERA
21.8 ±0.76
22.7 ±0.73
RADIOCARBON DATES (x 1000 y.B.P.)

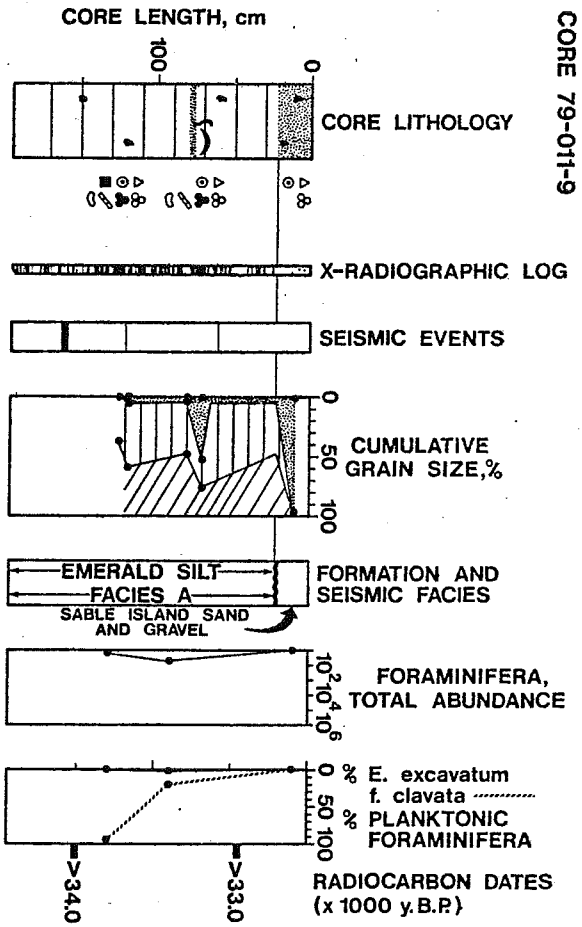

CORE 79-011-1
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0 50 100
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$ $10^4$ $10^6$
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
0 50 100
RADIOCARBON DATES (x 1000 y.B.P.)
CORE LENGTH, cm
0
200
400
600
800
LAHAVE CLAY
EMERALD SILT
FACIES B
6.48 ±0.07
11.6 ±0.12
15.1 ±0.23
16.4 ±0.18

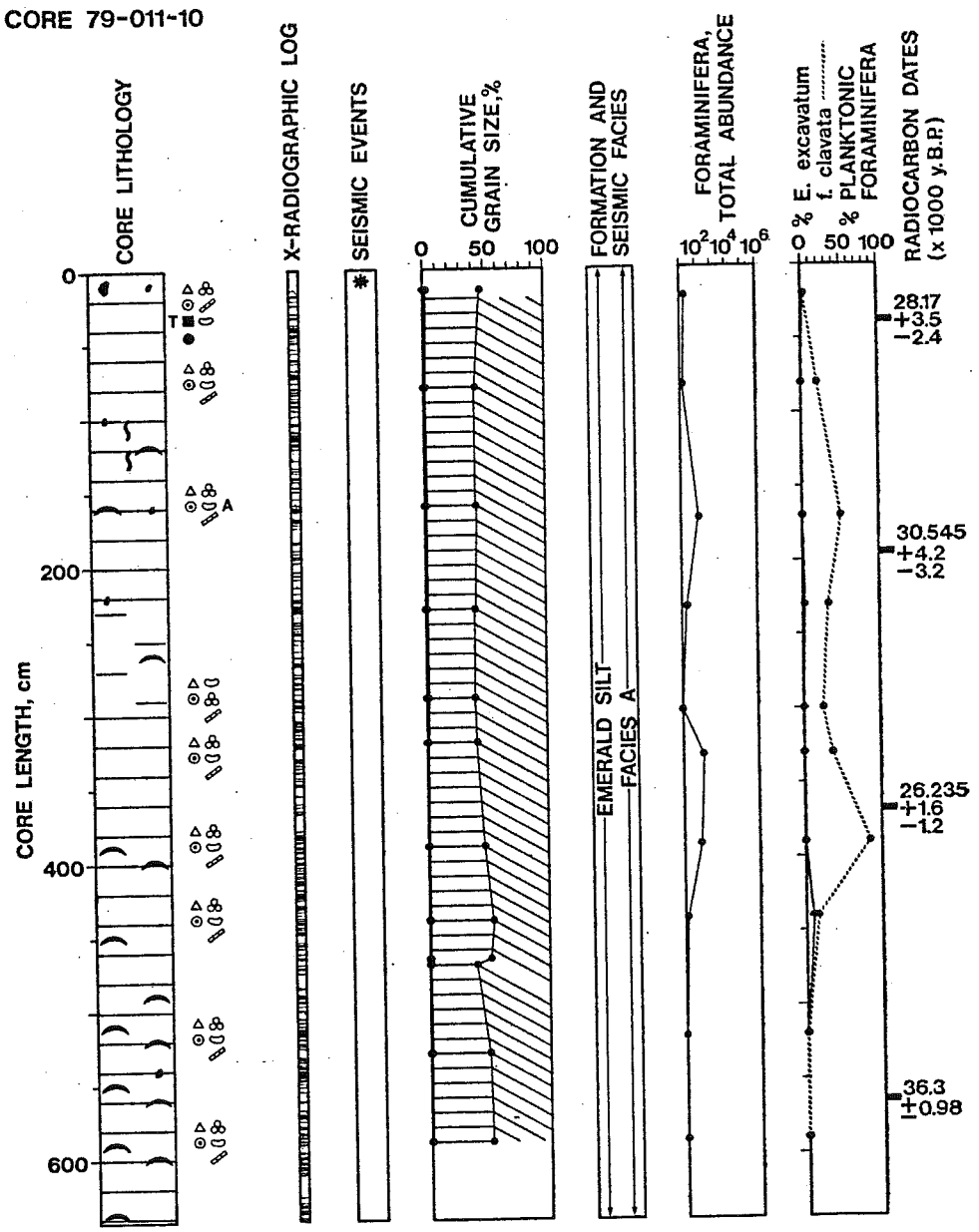

CORE 79-011-2
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0
50
100
FORMATION AND SEISMIC FACIES
EMERALD SILT
FACIES A
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$
$10^4$
$10^6$
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
RADIOCARBON DATES (x 1000 y.B.P.)
>38.0
34.3 ±0.96
>34.0

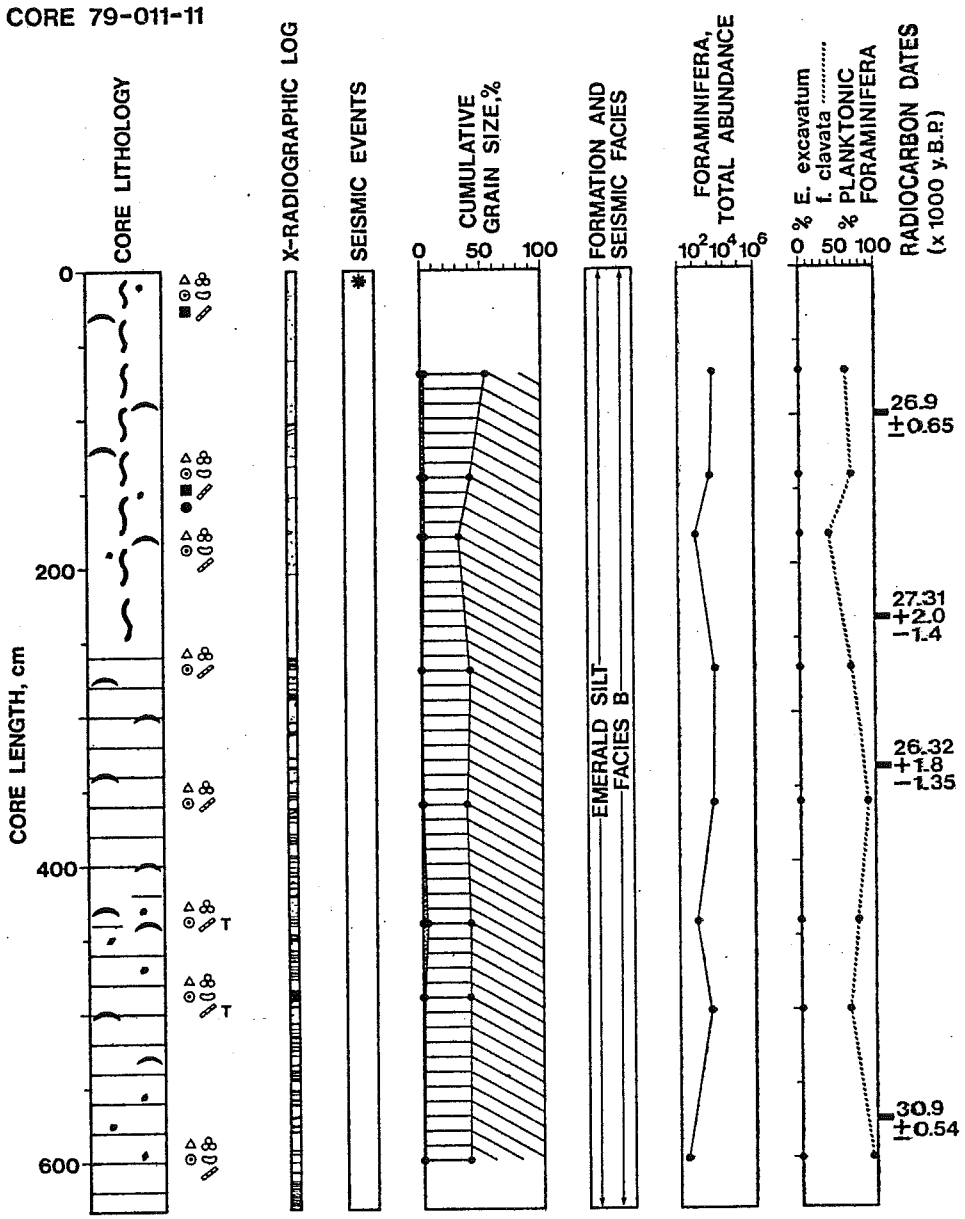

CORE 79-011-6
CORE LENGTH, cm
0
100
CORE LITHOLOGY
A
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CORE ABOVE LOCAL UNCONFORMITY
0
50
100
CUMULATIVE GRAIN SIZE,%
EMERALD SILT
FACIES A
SABLE ISLAND SAND AND GRAVEL
FORMATION AND SEISMIC FACIES
$10^2$ $10^4$ $10^6$
FORAMINIFERA, TOTAL ABUNDANCE
0 % E. excavatum
f. clavata
50 % PLANKTONIC FORAMINIFERA
100
32.2 ± 1.03
RADIOCARBON DATES (x 1000 y.B.P.)

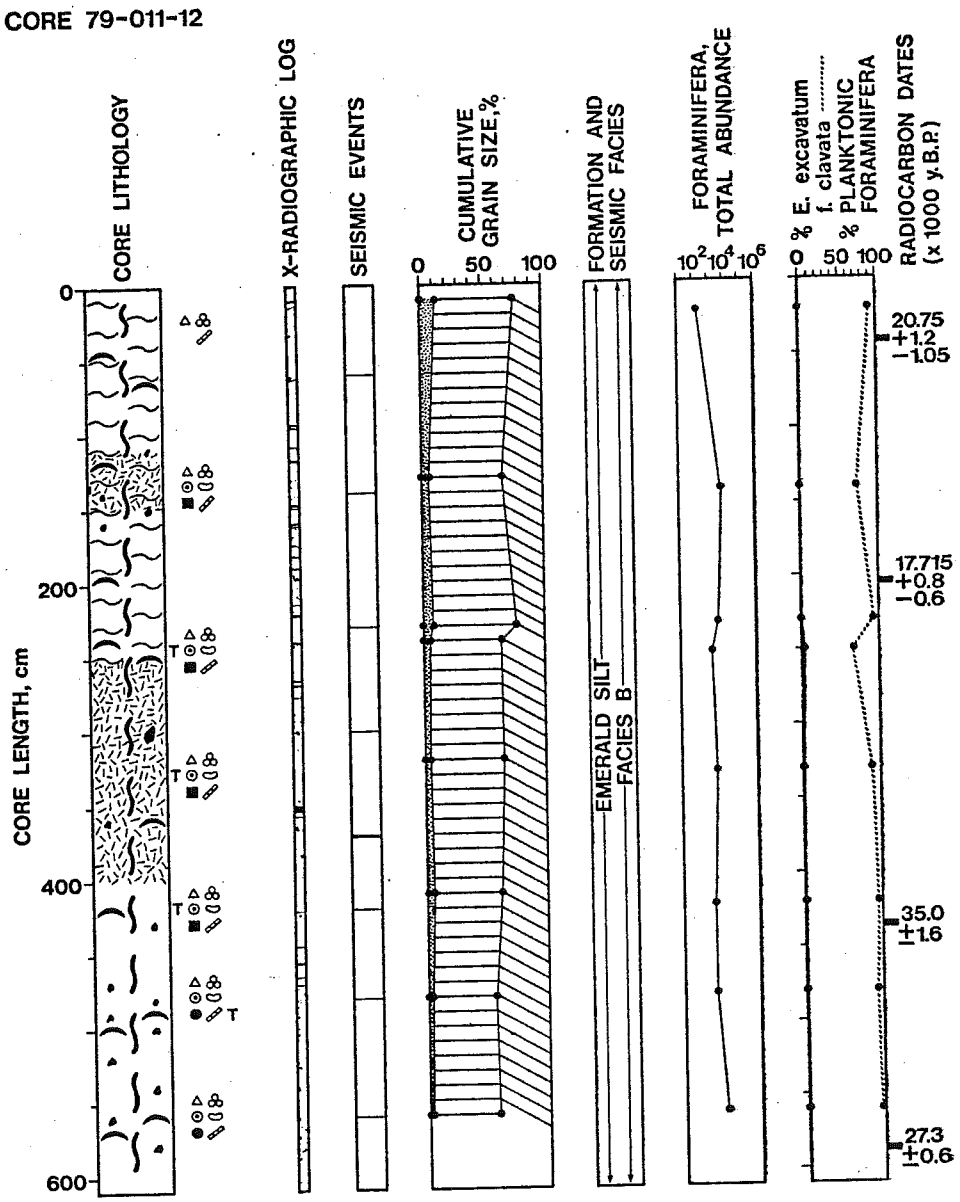

CORE 79-011-7
CORE LENGTH, cm
50
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
0
50
100
CUMULATIVE GRAIN SIZE,%
EMERALD SILT
FACIES A
SABLE ISLAND SAND AND GRAVEL
FORMATION AND SEISMIC FACIES
$10^2$
$10^4$
$10^6$
FORAMINIFERA, TOTAL ABUNDANCE
0 % E. excavatum f. clavata
50 % PLANKTONIC FORAMINIFERA
100
41.8 ± 1.79
RADIOCARBON DATES (x 1000 y.B.P.)

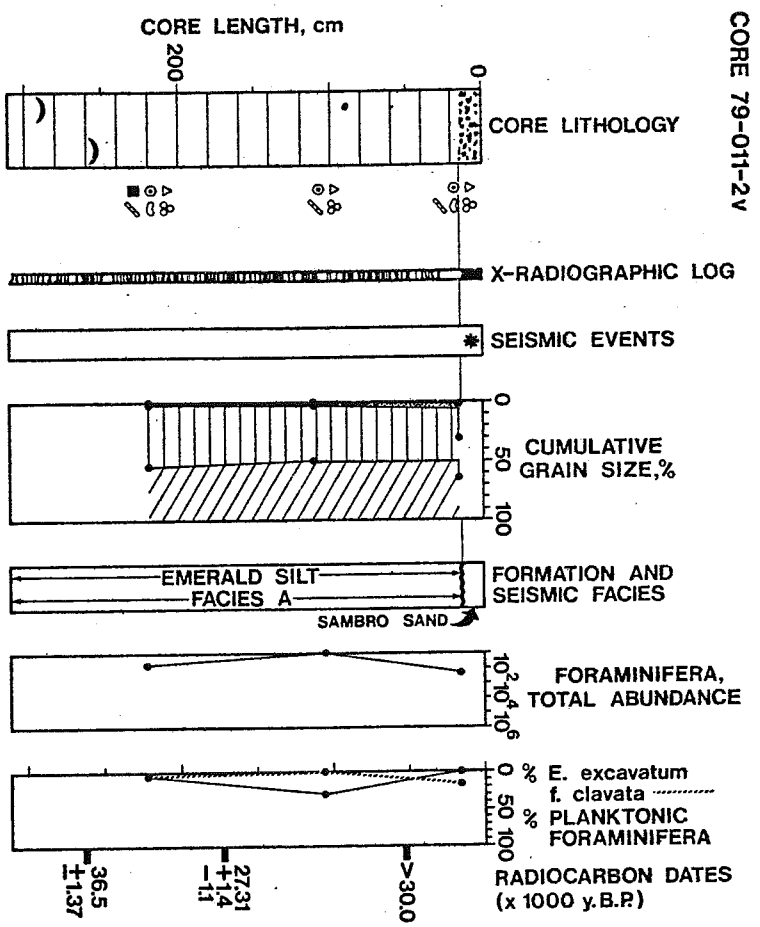

CORE 79-011-8
CORE LENGTH, cm
0
100
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0
50
100
SCOTIAN SHELF DRIFT
FORMATION AND SEISMIC FACIES
SABLE ISLAND SAND AND GRAVEL
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$
$10^4$
$10^6$
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
0
50
100
32.18 +3.4 −2.2
37.8 ±1.9
RADIOCARBON DATES (x 1000 y.B.P.)

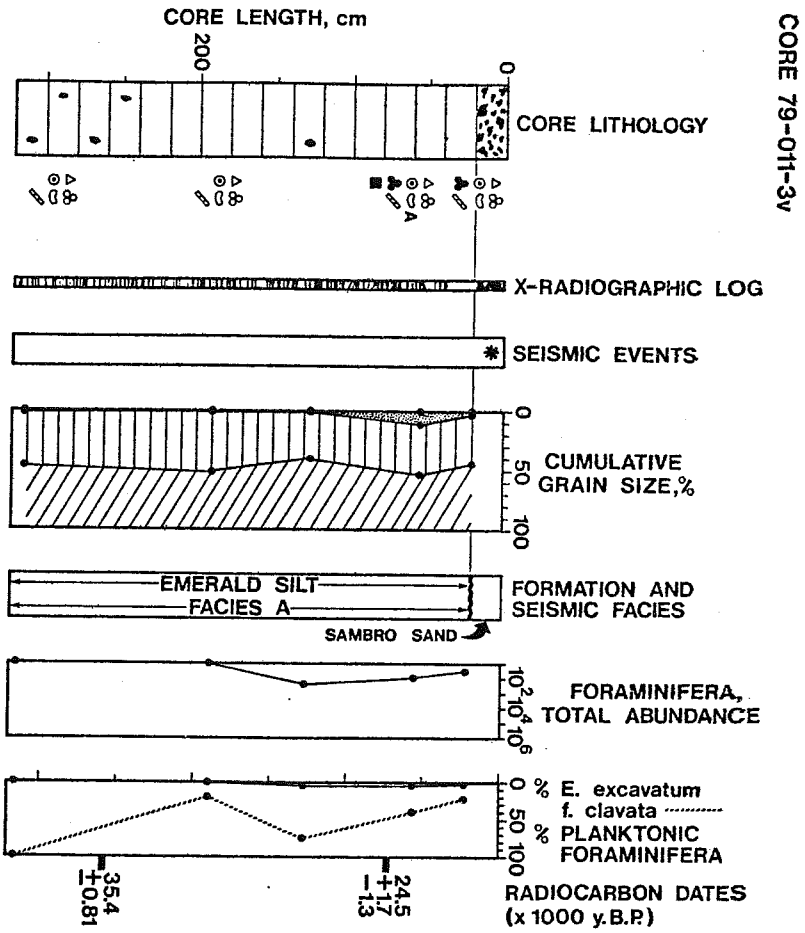

CORE 79-011-9
CORE LENGTH, cm
100
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
0
50
100
CUMULATIVE GRAIN SIZE,%
EMERALD SILT
FACIES A
SABLE ISLAND SAND AND GRAVEL
FORMATION AND SEISMIC FACIES
$10^2$
$10^4$
$10^6$
FORAMINIFERA, TOTAL ABUNDANCE
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
V34.0
V33.0
RADIOCARBON DATES (x 1000 y.B.P.)

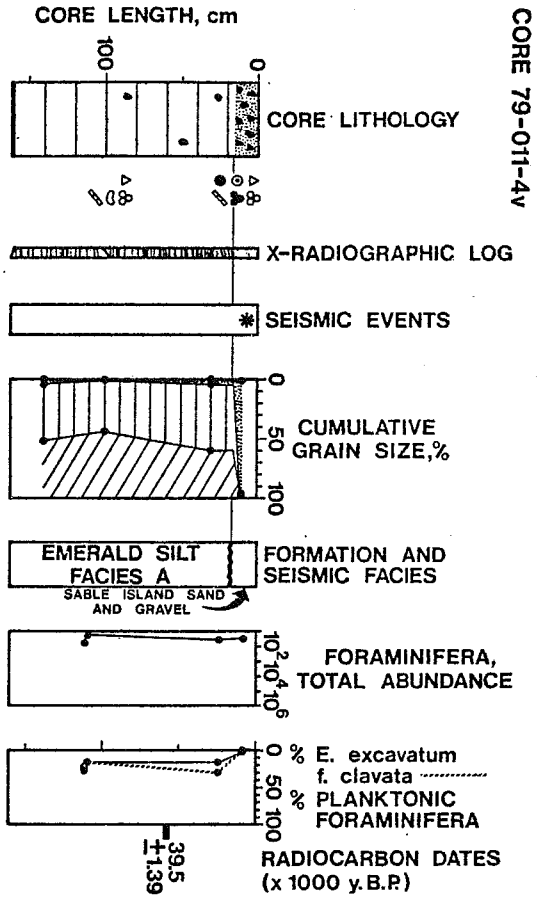

CORE 79-011-10
CORE LENGTH, cm
0
200
400
600
CORE LITHOLOGY
T
A
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0 50 100
FORMATION AND SEISMIC FACIES
EMERALD SILT
FACIES A
FORAMINIFERA, TOTAL ABUNDANCE
10^2 10^4 10^6
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
0 50 100
RADIOCARBON DATES (x 1000 y.B.P.)
28.17 +3.5 −2.4
30.545 +4.2 −3.2
26.235 +1.6 −1.2
36.3 ±0.98

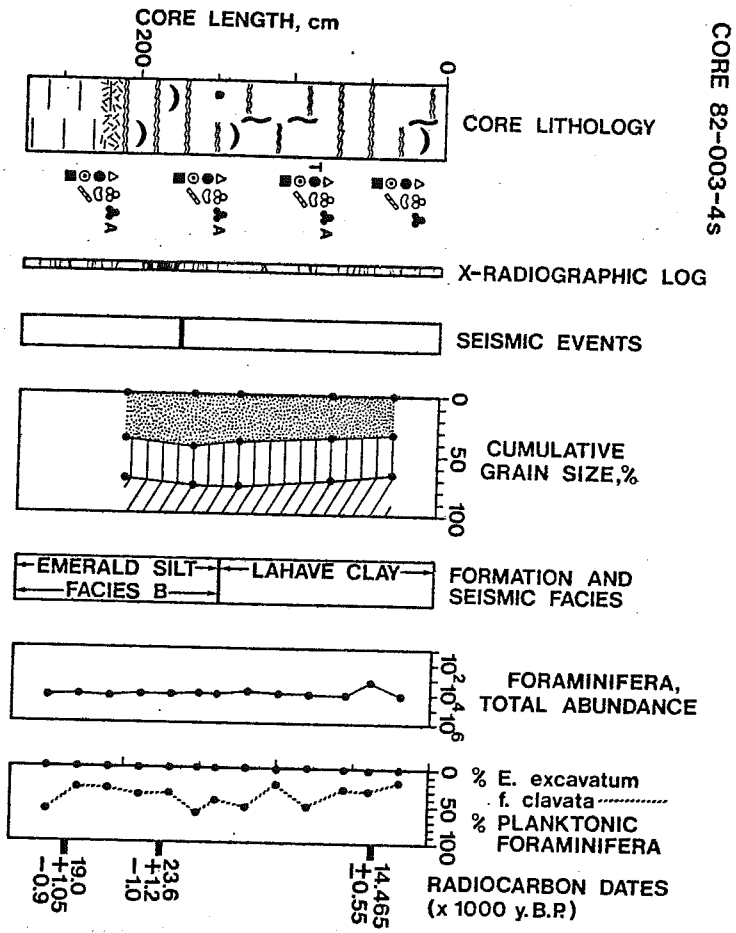

CORE 79-011-11
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
RADIOCARBON DATES (x 1000 y.B.P.)
CORE LENGTH, cm
0
200
400
600
T
T
0
50
100
EMERALD SILT
FACIES B
$10^2$ $10^4$ $10^6$
0
50
100
26.9
±0.65
27.31
+2.0
−1.4
26.32
+1.8
−1.35
30.9
±0.54

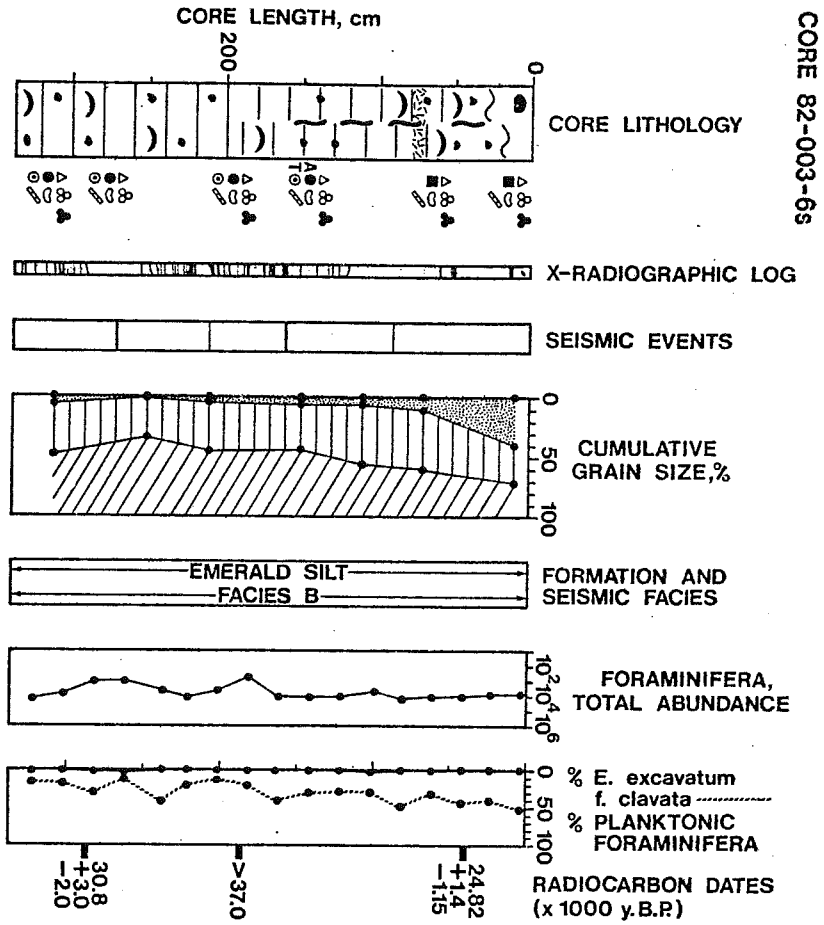

CORE 79-011-12
CORE LENGTH, cm
0
200
400
600
CORE LITHOLOGY
T
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0 50 100
FORMATION AND SEISMIC FACIES
EMERALD SILT FACIES B
FORAMINIFERA, TOTAL ABUNDANCE
10² 10⁴ 10⁶
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
0 50 100
RADIOCARBON DATES (x 1000 y.B.P.)
20.75 +1.2 −1.05
17.715 +0.8 −0.6
35.0 ±1.6
27.3 ±0.6

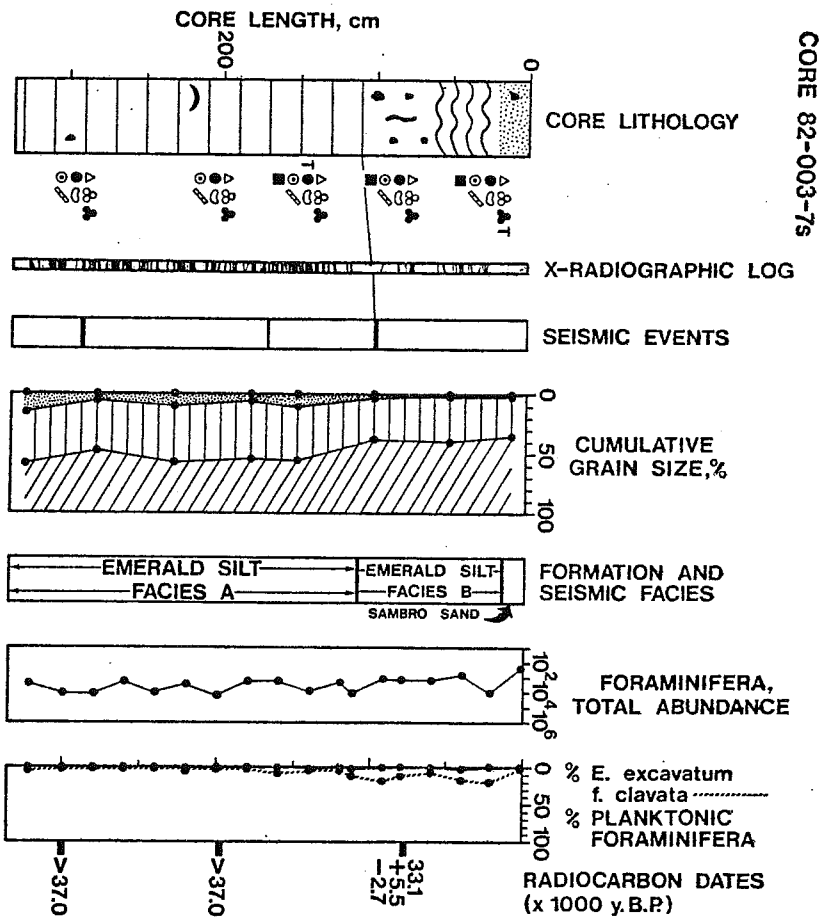

CORE 79-011-2V
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
0
50
100
CUMULATIVE GRAIN SIZE,%
EMERALD SILT
FACIES A
SAMBRO SAND
FORMATION AND SEISMIC FACIES
$10^2$ $10^4$ $10^6$
FORAMINIFERA, TOTAL ABUNDANCE
0 % E. excavatum
f. clavata
50 100 % PLANKTONIC FORAMINIFERA
36.5 ± 1.37
27.31 +1.4 -1.1
>30.0
RADIOCARBON DATES (x 1000 y.B.P.)

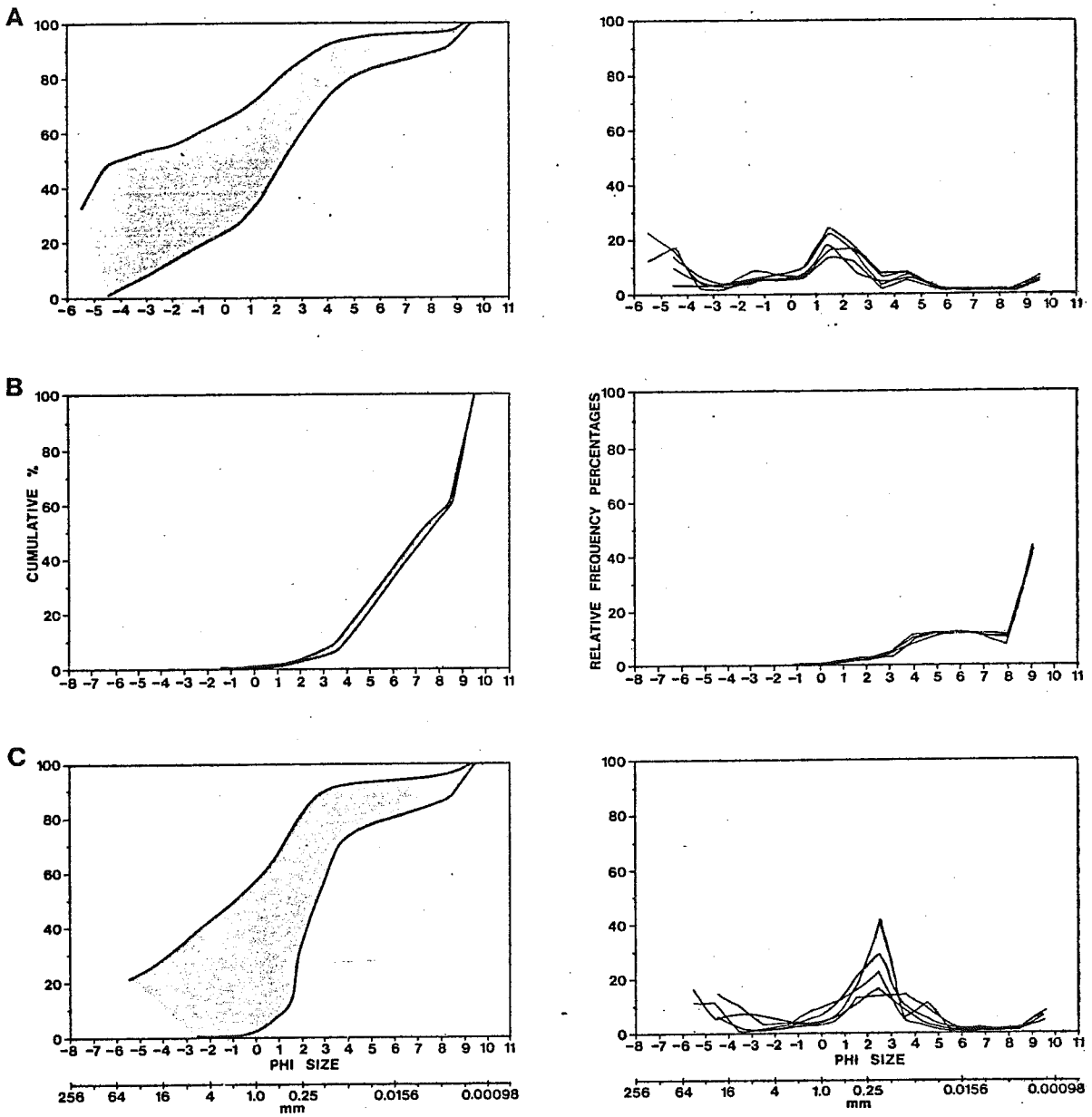

CORE 79-011-3v
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0
50
100
EMERALD SILT
FACIES A
SAMBRO SAND
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$ $10^4$ $10^6$
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
0 50 100
35.4 ±0.81
24.5 +1.7 −1.3
RADIOCARBON DATES (x 1000 y.B.P.)

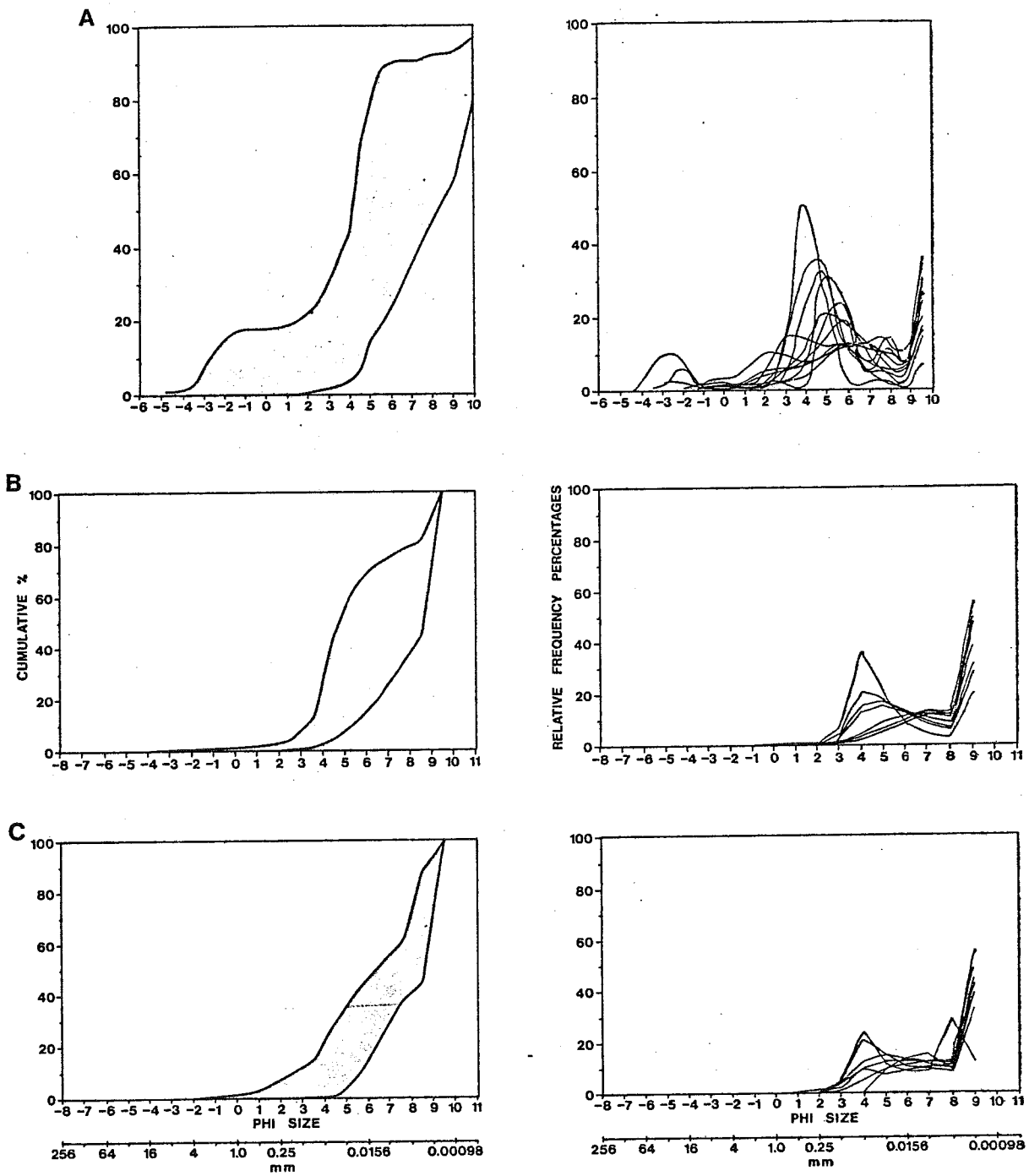

CORE 79-011-4v
CORE LENGTH, cm
100
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
0
50
100
CUMULATIVE GRAIN SIZE,%
EMERALD SILT FACIES A
SABLE ISLAND SAND AND GRAVEL
FORMATION AND SEISMIC FACIES
$10^2$ $10^4$ $10^6$
FORAMINIFERA, TOTAL ABUNDANCE
0 % E. excavatum
f. clavata
0 50 100 % PLANKTONIC FORAMINIFERA
39.5 ±1.39
RADIOCARBON DATES (x 1000 y.B.P.)

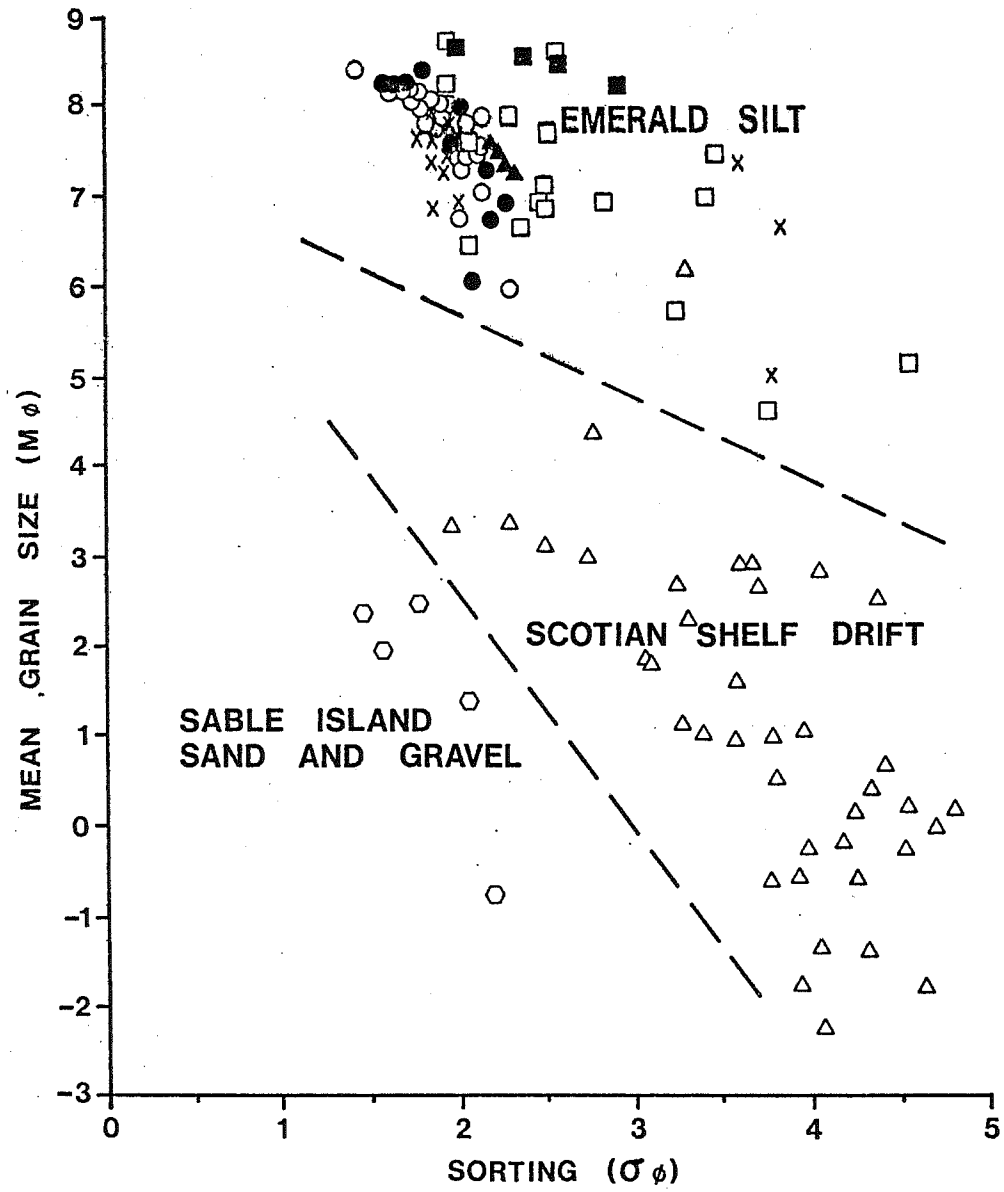

CORE 82-003-4s
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0
50
100
EMERALD SILT
FACIES B
LAHAVE CLAY
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
19.0 +1.05 −0.9
23.6 +1.2 −1.0
14.465 ±0.55
RADIOCARBON DATES (x 1000 y.B.P.)


CORE 82-003-6s
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0
50
100
EMERALD SILT
FACIES B
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
$10^2$
$10^4$
$10^6$
% E. excavatum
f. clavata
% PLANKTONIC FORAMINIFERA
30.8 +3.0 −2.0
>37.0
24.82 +1.4 −1.15
RADIOCARBON DATES (x 1000 y.B.P.)

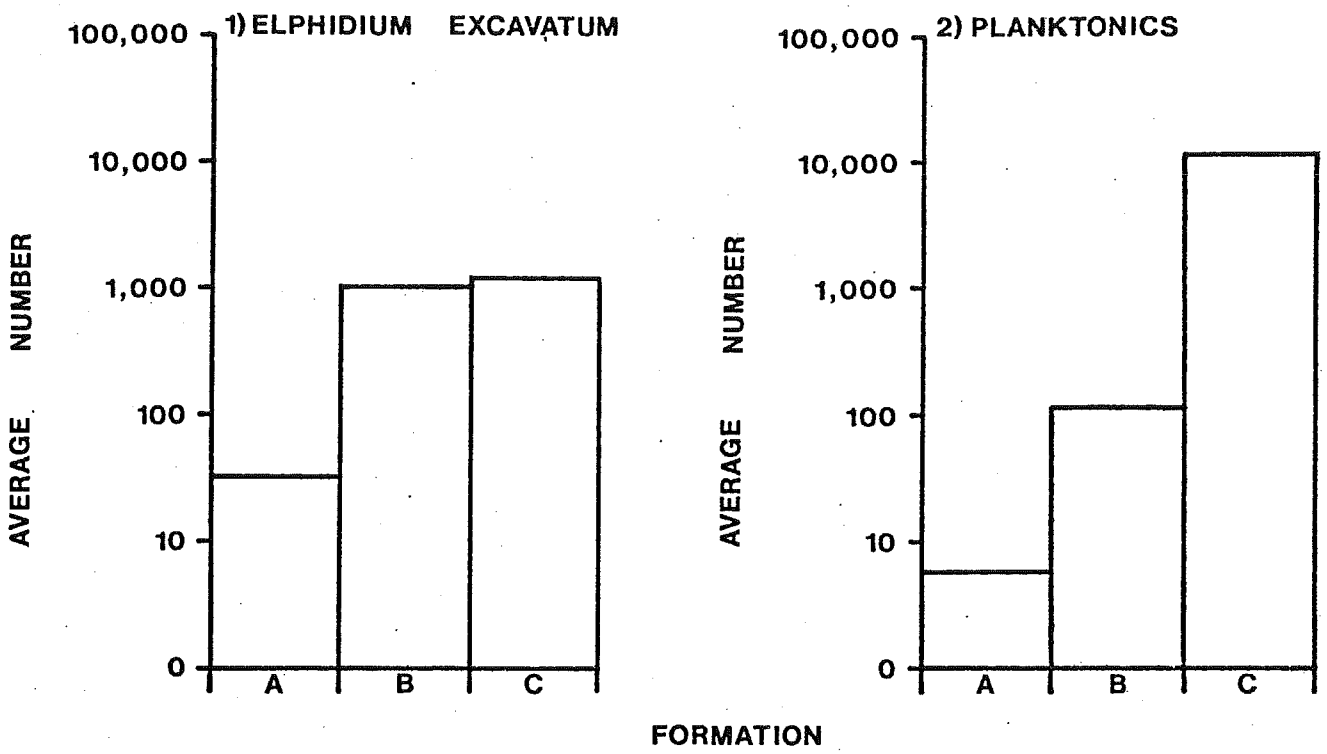

CORE 82-003-7s
CORE LENGTH, cm
200
0
CORE LITHOLOGY
X-RADIOGRAPHIC LOG
SEISMIC EVENTS
CUMULATIVE GRAIN SIZE,%
0
50
100
EMERALD SILT FACIES A
EMERALD SILT FACIES B
SAMBRO SAND
FORMATION AND SEISMIC FACIES
FORAMINIFERA, TOTAL ABUNDANCE
% E. excavatum f. clavata
% PLANKTONIC FORAMINIFERA
>37.0
>37.0
33.1 +5.5 −2.7
RADIOCARBON DATES (x 1000 y.B.P.)


A
B
C
CUMULATIVE %
RELATIVE FREQUENCY PERCENTAGES
PHI SIZE
mm
256 64 16 4 1.0 0.25 0.0156 0.00098


A
B
C
CUMULATIVE %
RELATIVE FREQUENCY PERCENTAGES
PHI SIZE
mm
256 64 16 4 1.0 0.25 0.0156 0.00098




| AREA | SCOTIAN SHELF DRIFT | EMERALD SILT FACIES A | EMERALD SILT FACIES B | EMERALD SILT FACIES C | SABLE IS. SAND AND GRAVEL |
|---|---|---|---|---|---|
| GULF OF MAINE | △ | ■ | □ | X | |
| EMERALD BASIN | ▲ TILL TONGUE | ○ | ● | | ⬡ |
| GRAND BANKS OF NEWFOUNDLAND | △ | | | | |

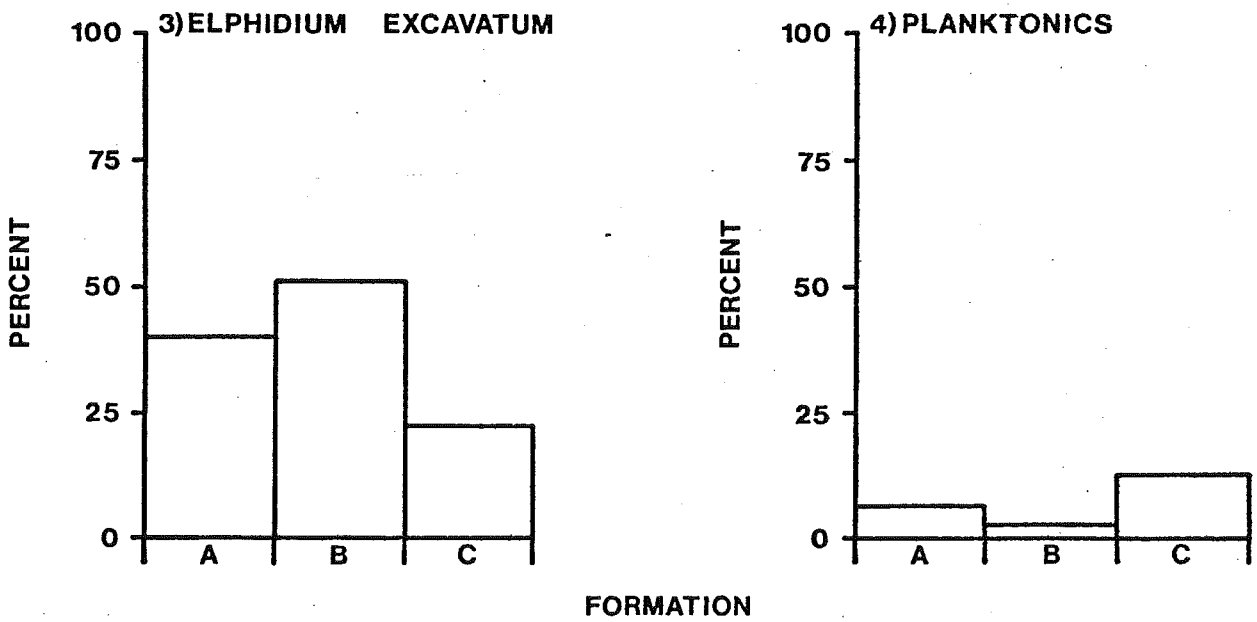


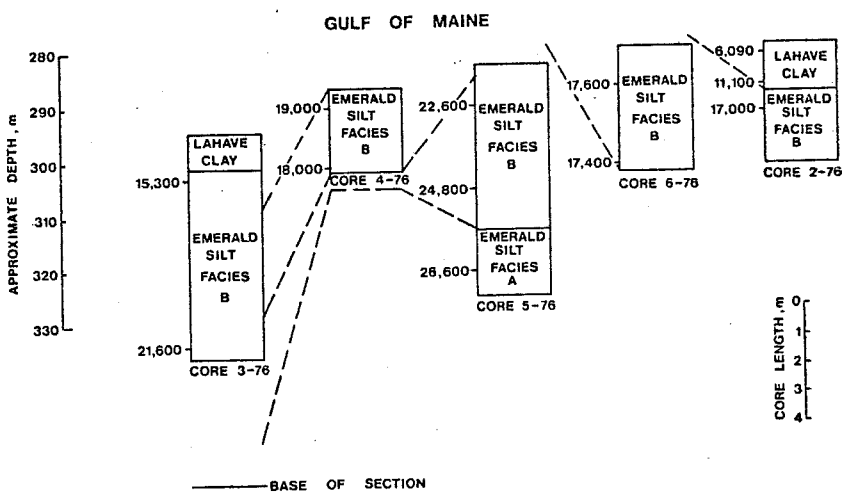

FORMATIONS:

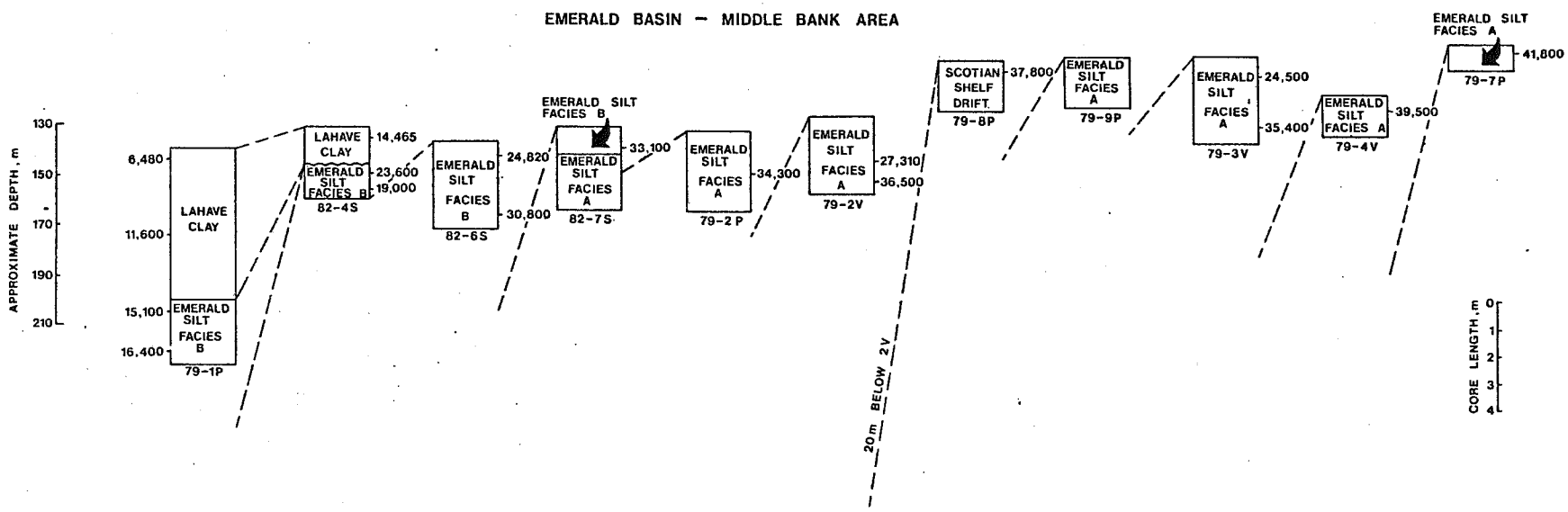
A – EMERALD SILT FACIES A

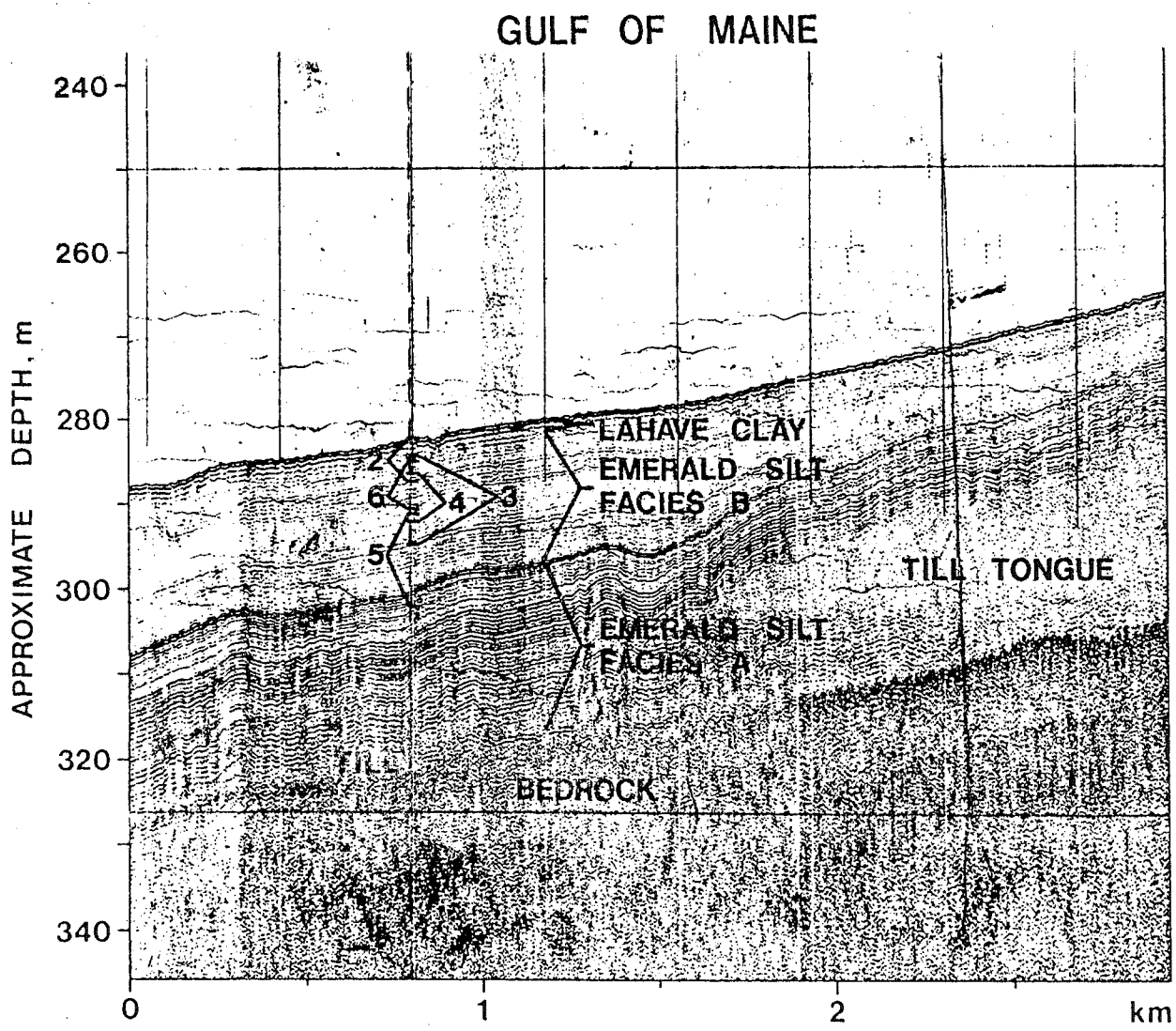
B – EMERALD SILT FACIES B

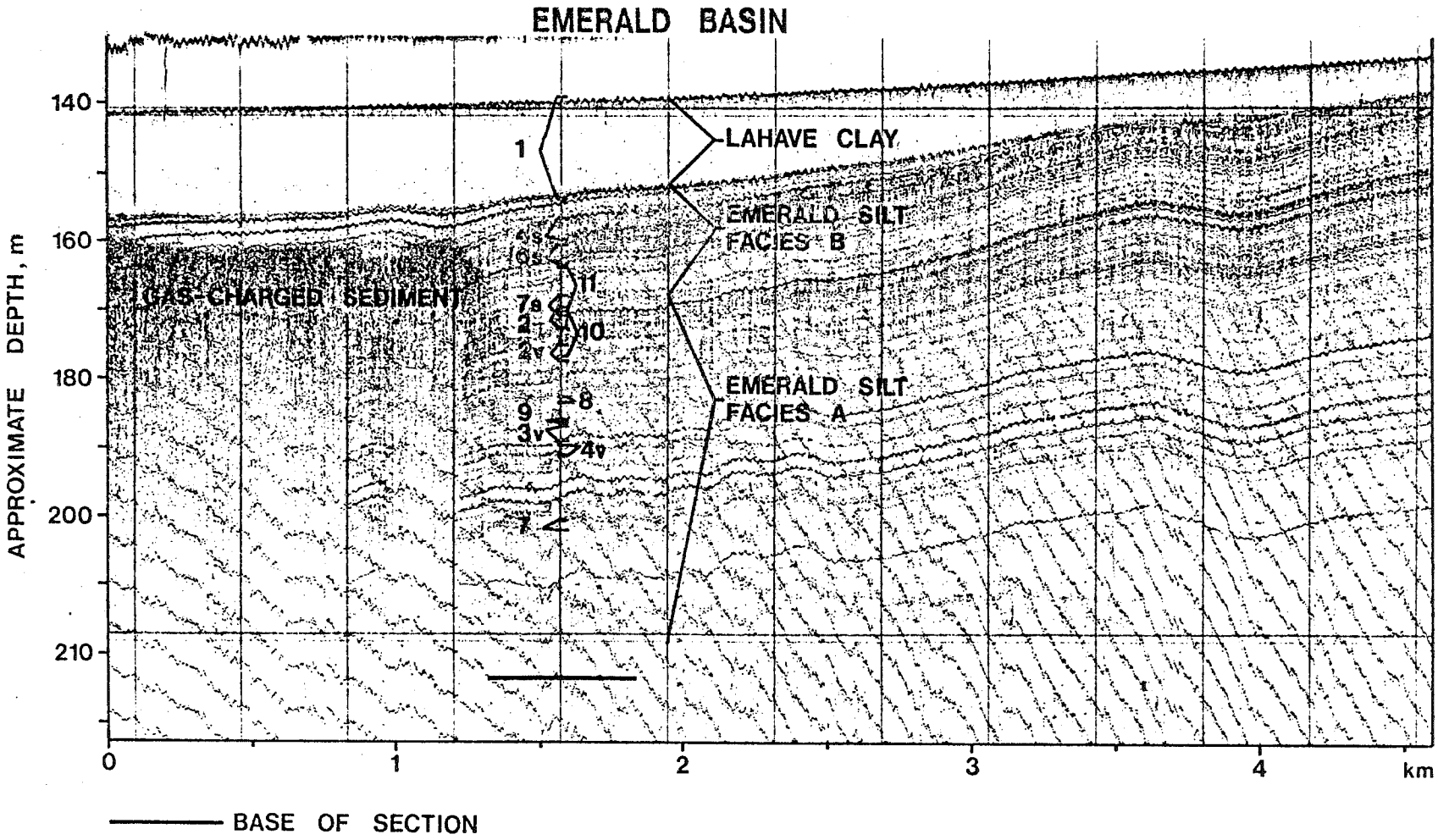
C – LAHAVE CLAY

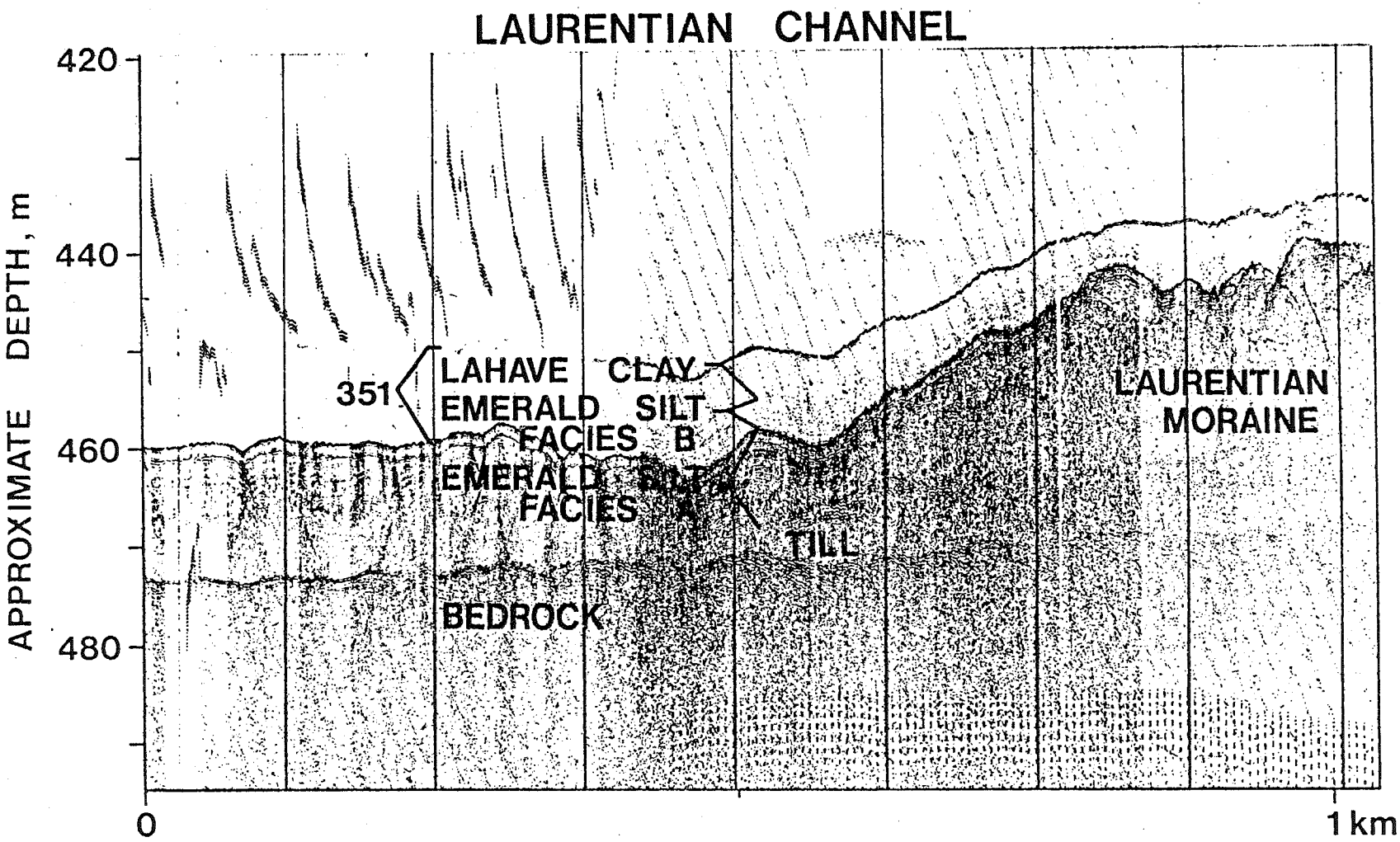


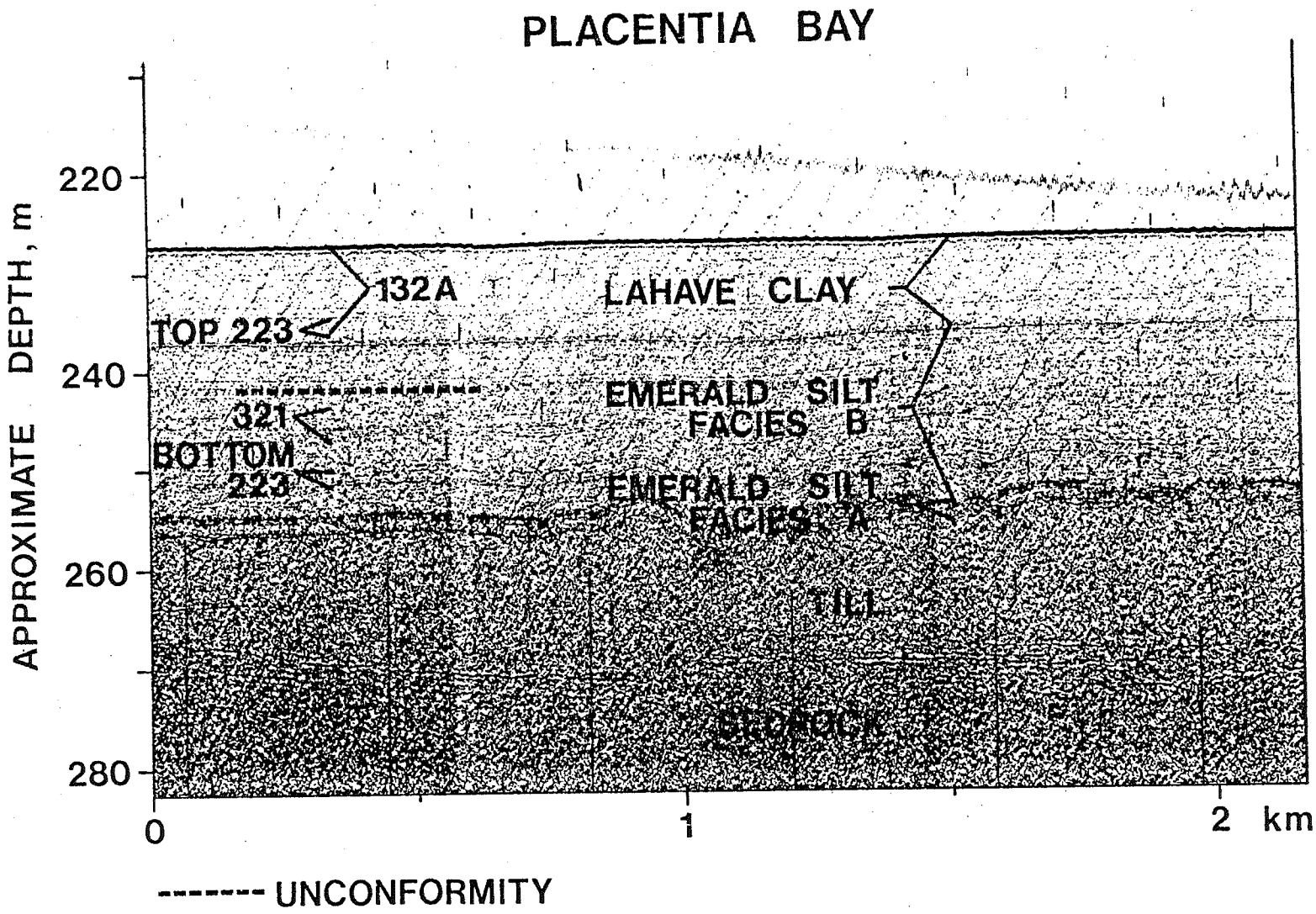

GULF OF MAINE
APPROXIMATE DEPTH, m
280
290
300
310
320
330
LAHAVE CLAY
EMERALD SILT FACIES B
15,300
21,600
CORE 3-76
19,000
18,000
EMERALD SILT FACIES B
CORE 4-76
22,600
24,800
26,600
EMERALD SILT FACIES B
EMERALD SILT FACIES A
CORE 5-76
17,600
17,400
EMERALD SILT FACIES B
CORE 6-76
6,090
11,100
17,000
LAHAVE CLAY
EMERALD SILT FACIES B
CORE 2-76
CORE LENGTH, m
0
1
2
3
4
BASE OF SECTION

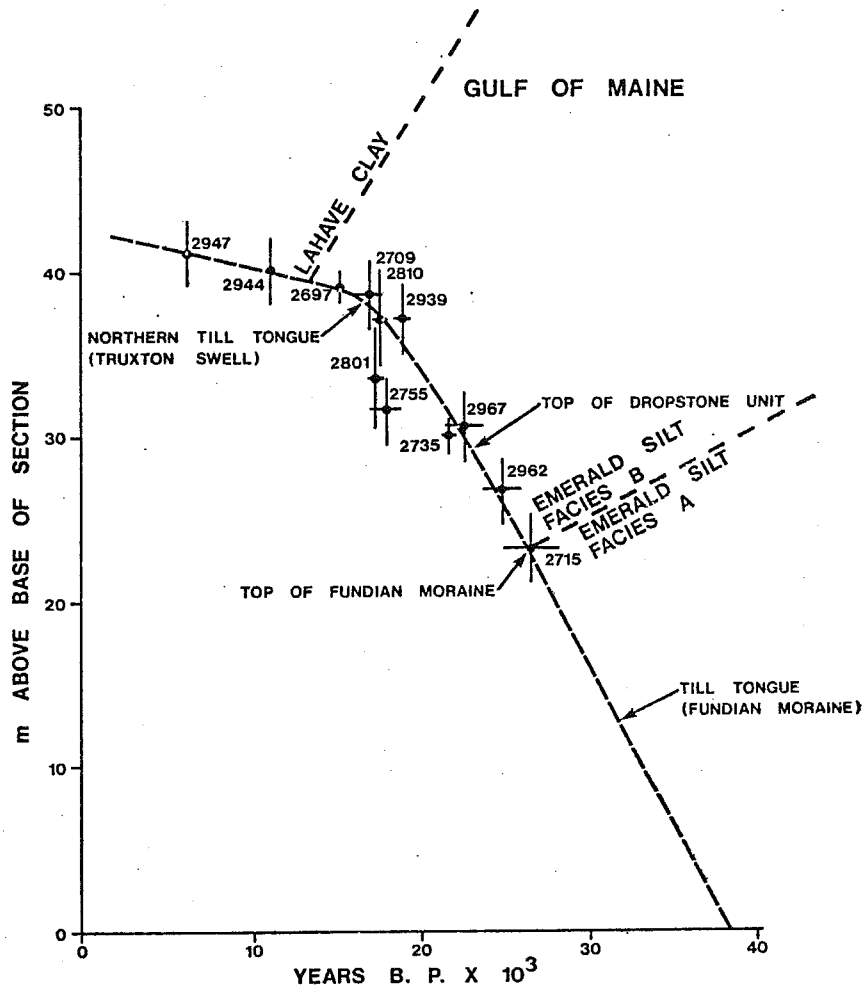

EMERALD BASIN – MIDDLE BANK AREA
APPROXIMATE DEPTH, m
130
150
170
190
210
LAHAVE CLAY
EMERALD SILT FACIES B
6,480
11,600
15,100
16,400
79-1P
LAHAVE CLAY
EMERALD SILT FACIES B
14,465
23,600
19,000
82-4S
EMERALD SILT FACIES B
24,820
30,800
82-6S
EMERALD SILT FACIES B
EMERALD SILT FACIES A
33,100
82-7S
EMERALD SILT FACIES A
34,300
79-2P
EMERALD SILT FACIES A
27,310
36,500
79-2V
20m BELOW 2V
SCOTIAN SHELF DRIFT
37,800
79-8P
EMERALD SILT FACIES A
79-9P
EMERALD SILT FACIES A
24,500
35,400
79-3V
EMERALD SILT FACIES A
39,500
79-4V
EMERALD SILT FACIES A
41,800
79-7P
CORE LENGTH, m
0
1
2
3
4

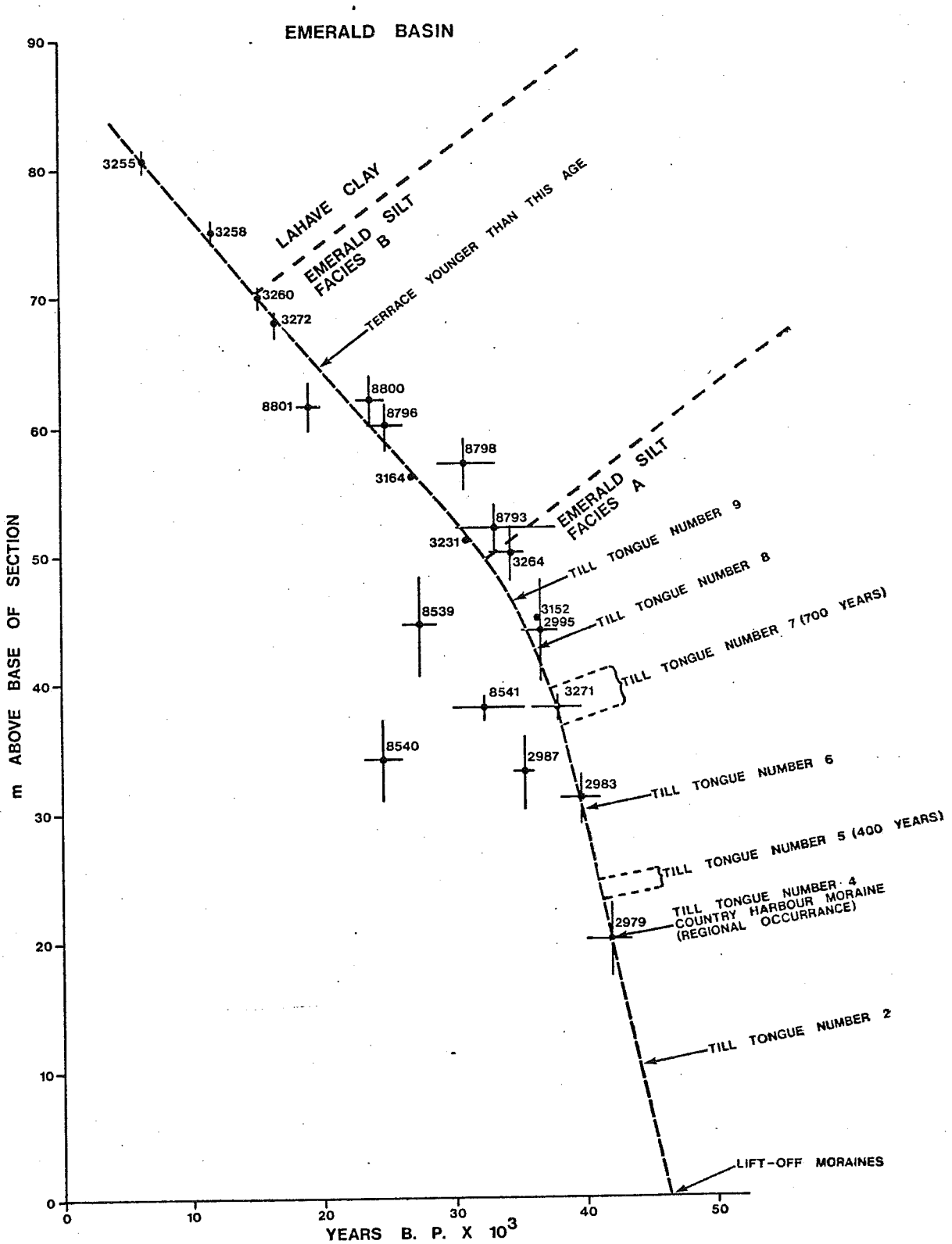

GULF OF MAINE
APPROXIMATE DEPTH, m
240
260
280
300
320
340
0
1
2
km
LAHAVE CLAY
EMERALD SILT
FACIES B
TILL TONGUE
EMERALD SILT
FACIES A
TILL
BEDROCK
2
3
4
5
6

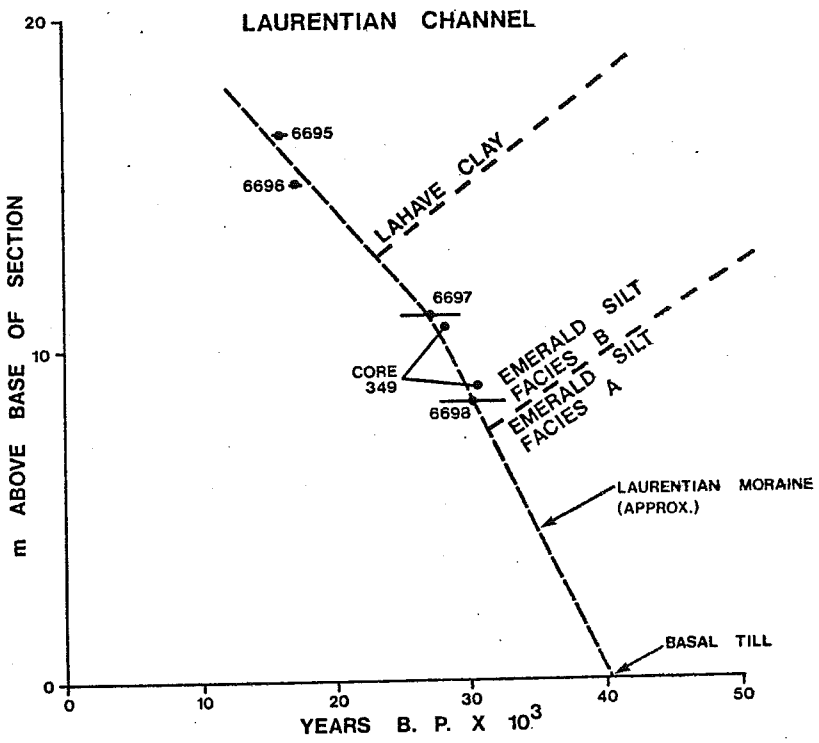

EMERALD BASIN
LAHAVE CLAY
EMERALD SILT FACIES B
EMERALD SILT FACIES A
GAS-CHARGED SEDIMENT
1
6s
10s
11
7a
2
10
4v
9
8
3v
4v
7
APPROXIMATE DEPTH, m
140
160
180
200
210
0
1
2
3
4
km
BASE OF SECTION

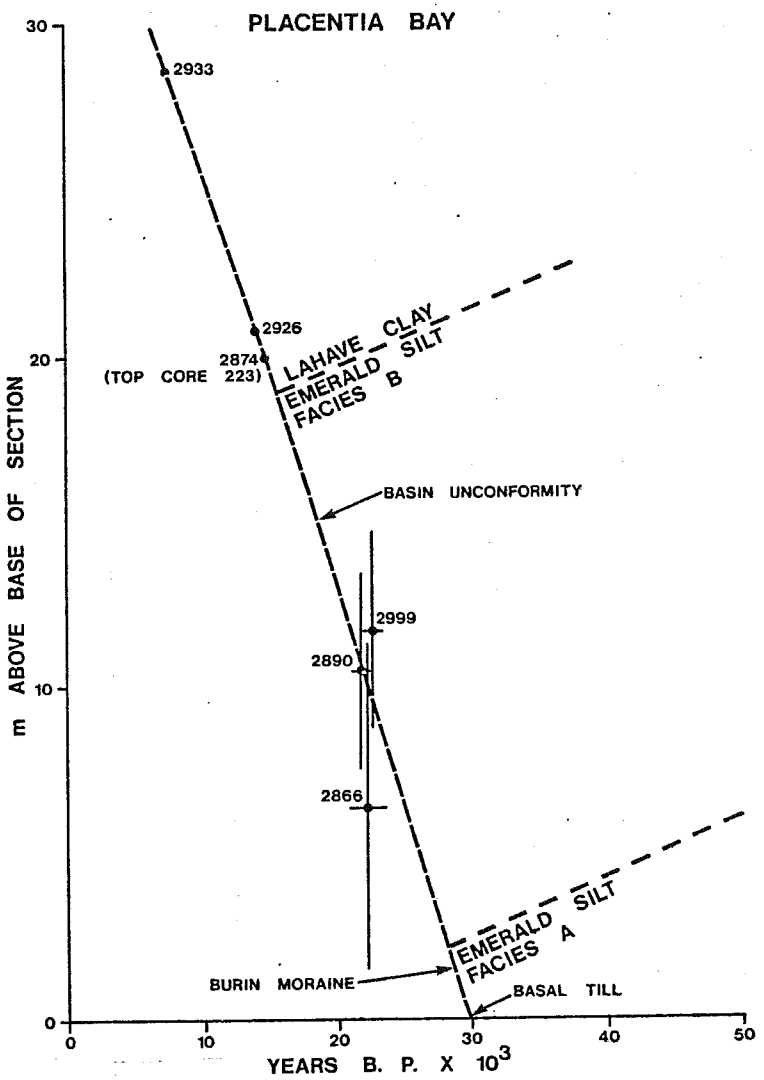

LAURENTIAN CHANNEL
APPROXIMATE DEPTH, m
420
440
460
480
351
LAHAVE CLAY
EMERALD SILT
FACIES B
EMERALD SILT
FACIES A
TILL
BEDROCK
LAURENTIAN
MORAINE
0
1 km

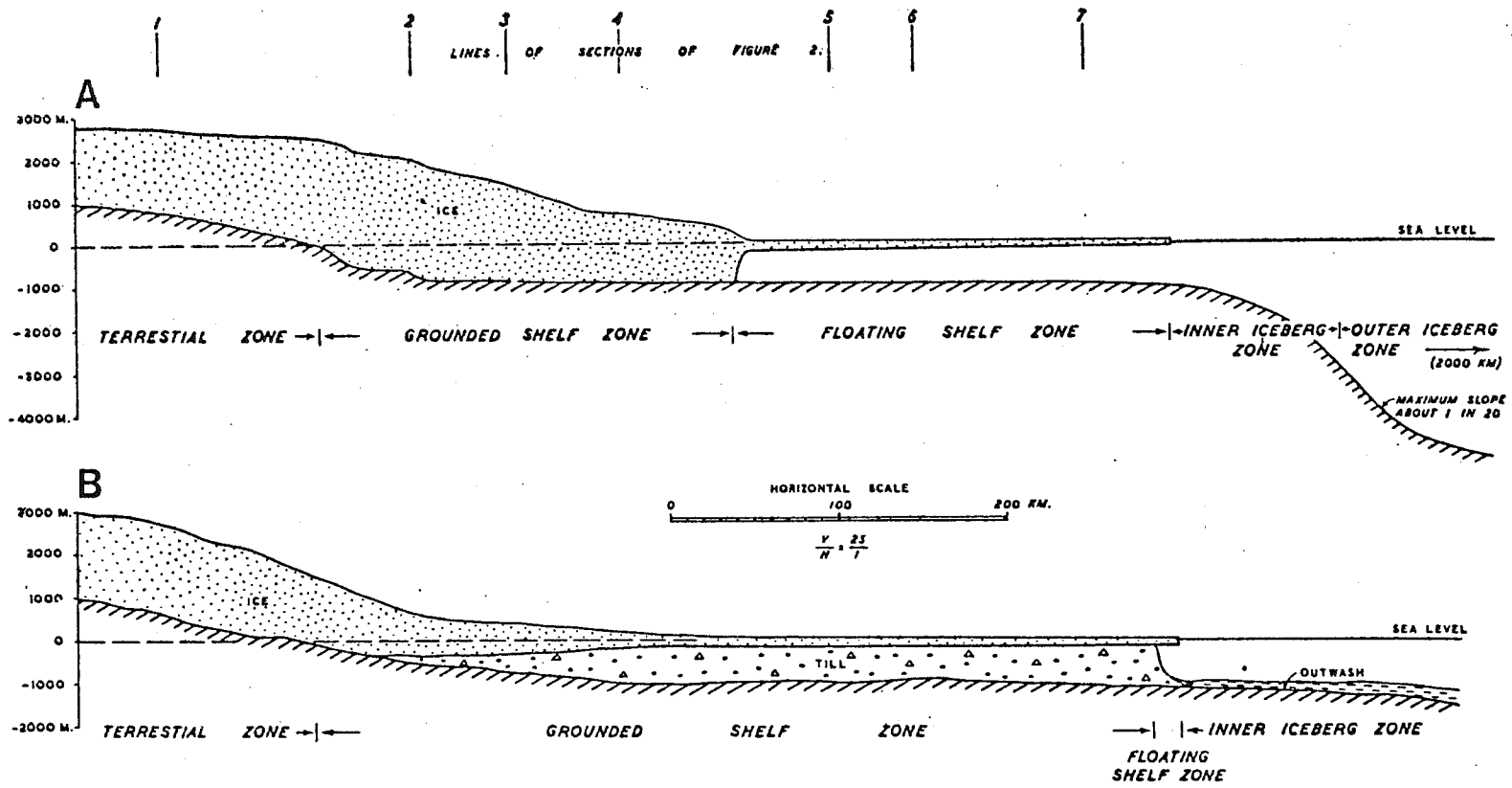

PLACENTIA BAY
APPROXIMATE DEPTH, m
220
240
260
280
132A
TOP 223
321
BOTTOM 223
LAHAVE CLAY
EMERALD SILT FACIES B
EMERALD SILT FACIES A
TILL
BEDROCK
0
1
2 km
------- UNCONFORMITY

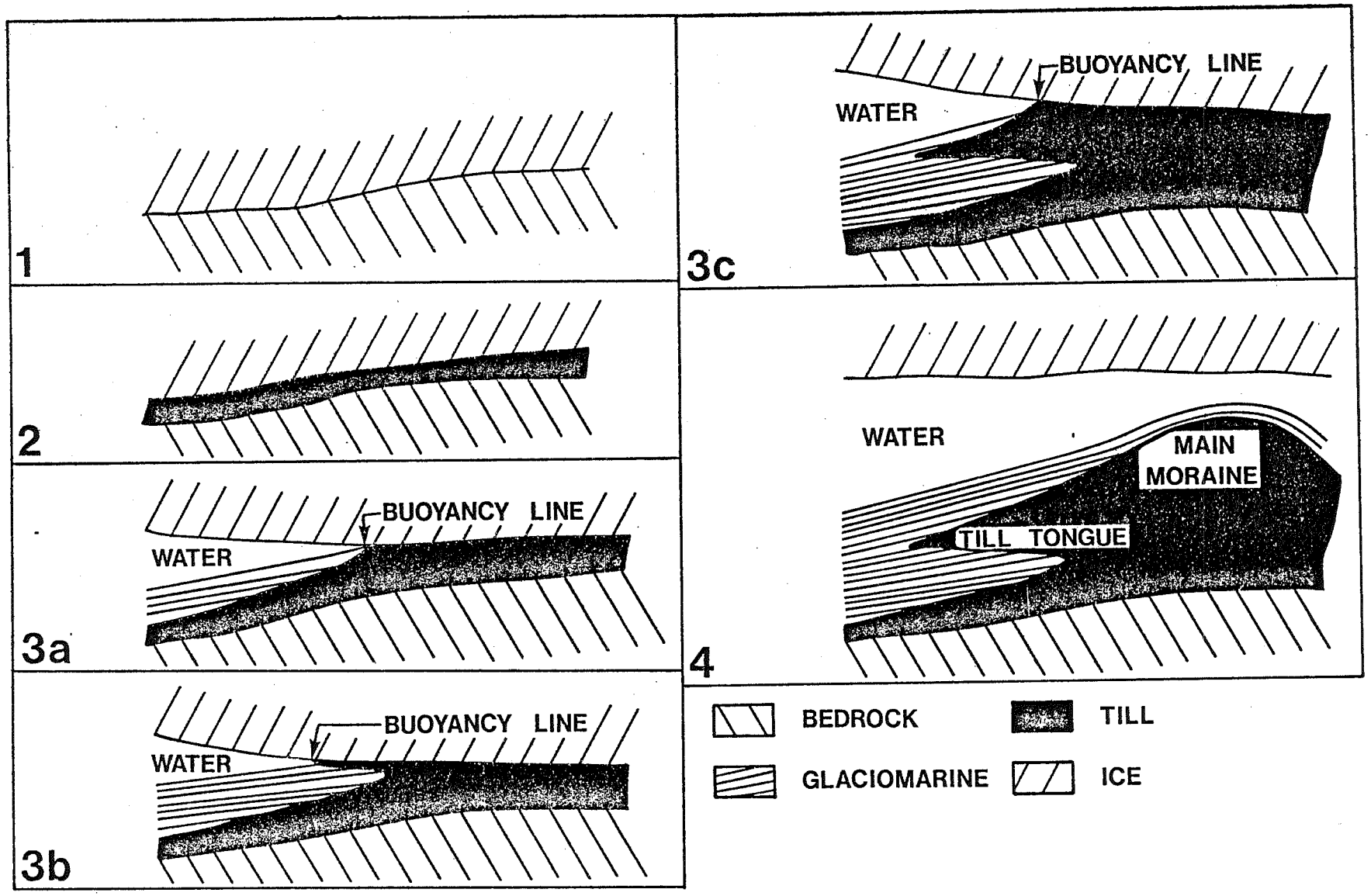

GULF OF MAINE
LAHAVE CLAY
2947
2944
2697
2709
2810
2939
NORTHERN TILL TONGUE
(TRUXTON SWELL)
2801
2755
2735
2967
TOP OF DROPSTONE UNIT
2962
EMERALD SILT
FACIES B
EMERALD SILT
FACIES A
2715
TOP OF FUNDIAN MORAINE
TILL TONGUE
(FUNDIAN MORAINE)
m ABOVE BASE OF SECTION
YEARS B. P. X 10^3
0
10
20
30
40
50

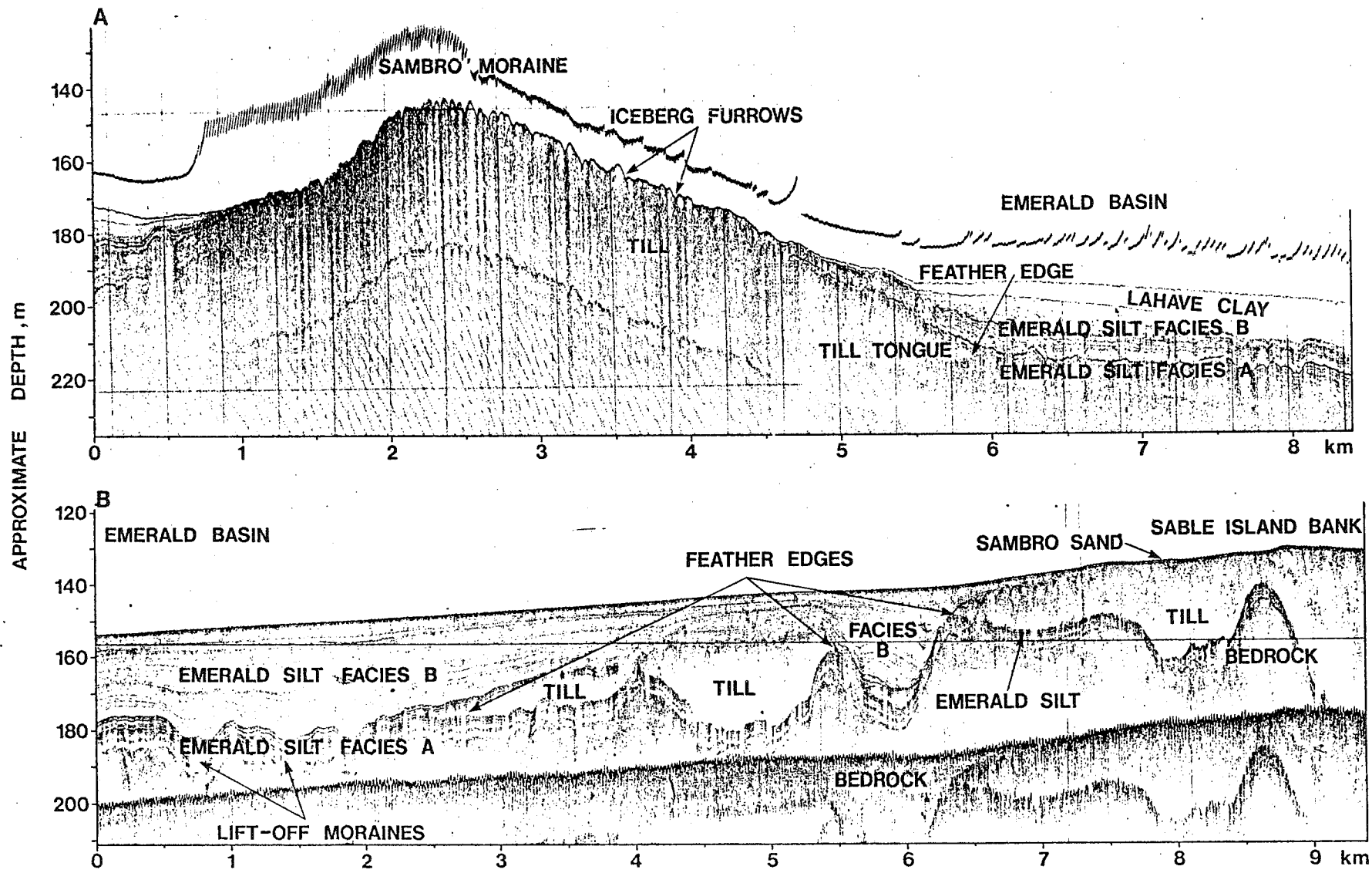

EMERALD BASIN
m ABOVE BASE OF SECTION
YEARS B. P. X 10^3
LAHAVE CLAY
EMERALD SILT FACIES B
TERRACE YOUNGER THAN THIS AGE
EMERALD SILT FACIES A
TILL TONGUE NUMBER 9
TILL TONGUE NUMBER 8
TILL TONGUE NUMBER 7 (700 YEARS)
TILL TONGUE NUMBER 6
TILL TONGUE NUMBER 5 (400 YEARS)
TILL TONGUE NUMBER 4
COUNTRY HARBOUR MORAINE
(REGIONAL OCCURRANCE)
TILL TONGUE NUMBER 2
LIFT-OFF MORAINES
3255
3258
3260
3272
8800
8796
8801
8798
3164
8793
3231
3264
3152
2995
8539
8541
3271
8540
2987
2983
2979

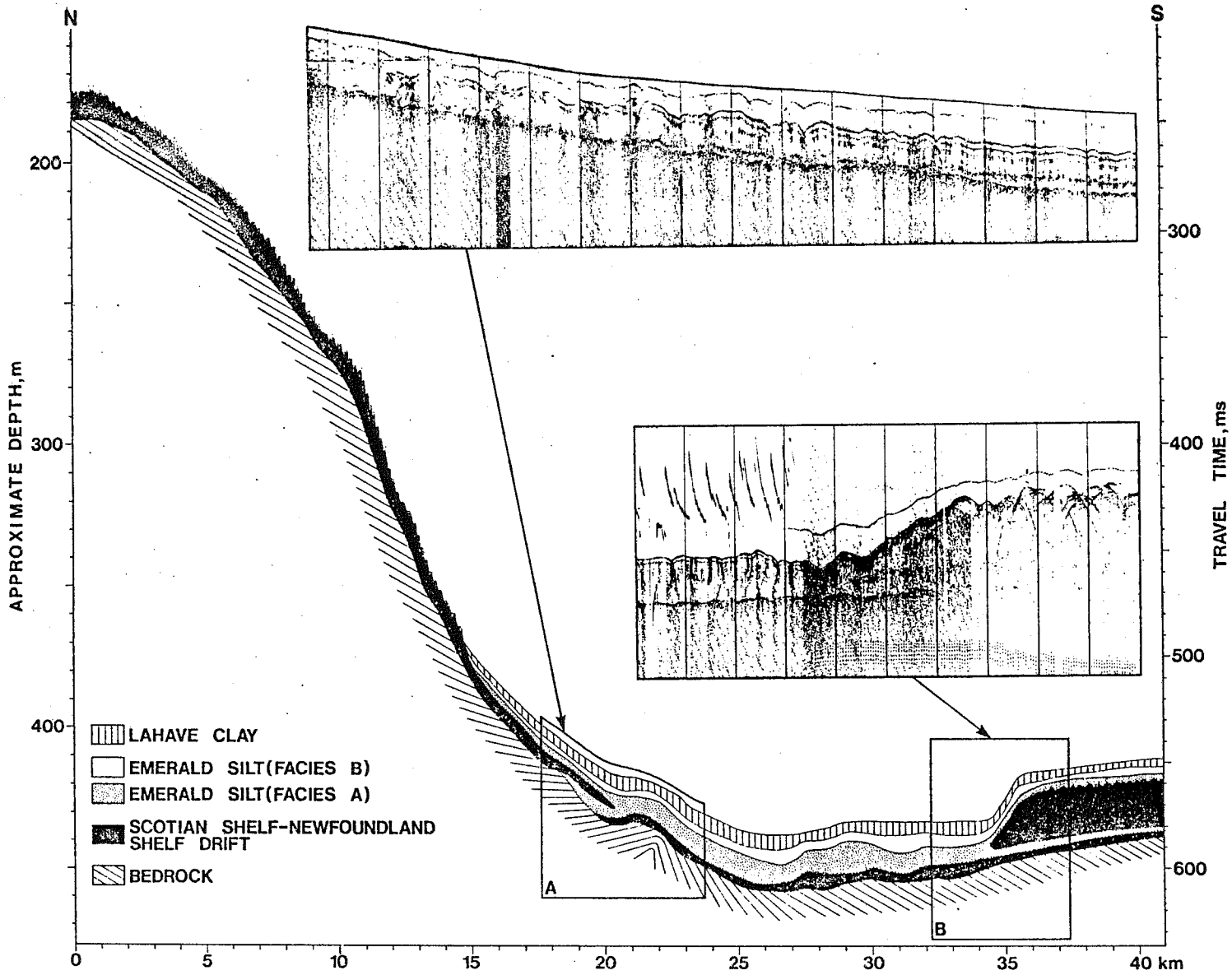

LAURENTIAN CHANNEL
m ABOVE BASE OF SECTION
20
10
0
6695
6696
LAHAVE CLAY
6697
CORE 349
6698
EMERALD SILT FACIES B
EMERALD SILT FACIES A
LAURENTIAN MORAINE (APPROX.)
BASAL TILL
0
10
20
30
40
50
YEARS B. P. X 10^3

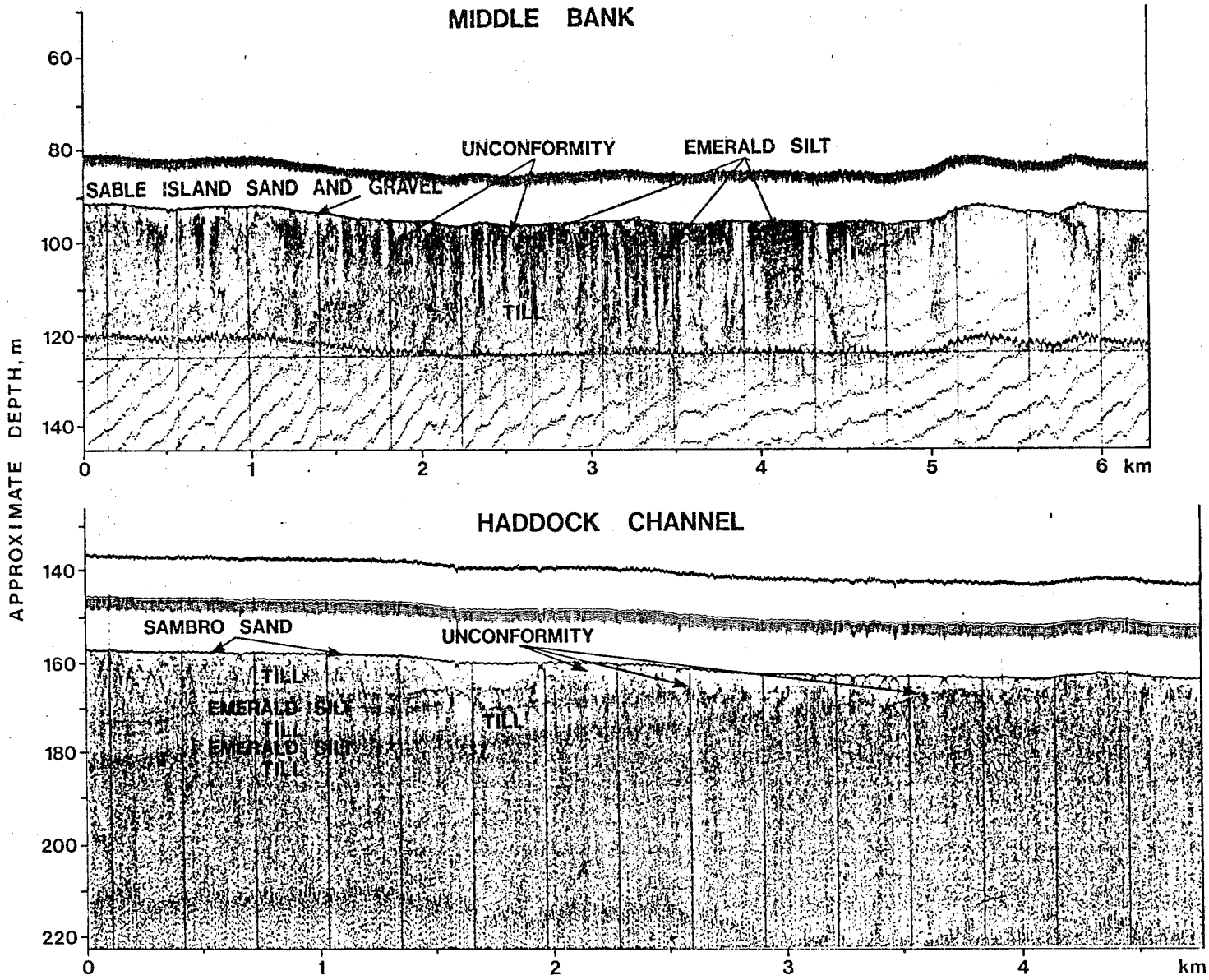

PLACENTIA BAY
2933
2926
2874
(TOP CORE 223)
LAHAVE CLAY
EMERALD SILT
FACIES B
BASIN UNCONFORMITY
2999
2890
2866
EMERALD SILT
FACIES A
BURIN MORAINE
BASAL TILL
m ABOVE BASE OF SECTION
YEARS B. P. X 10^3
0
10
20
30
40
50

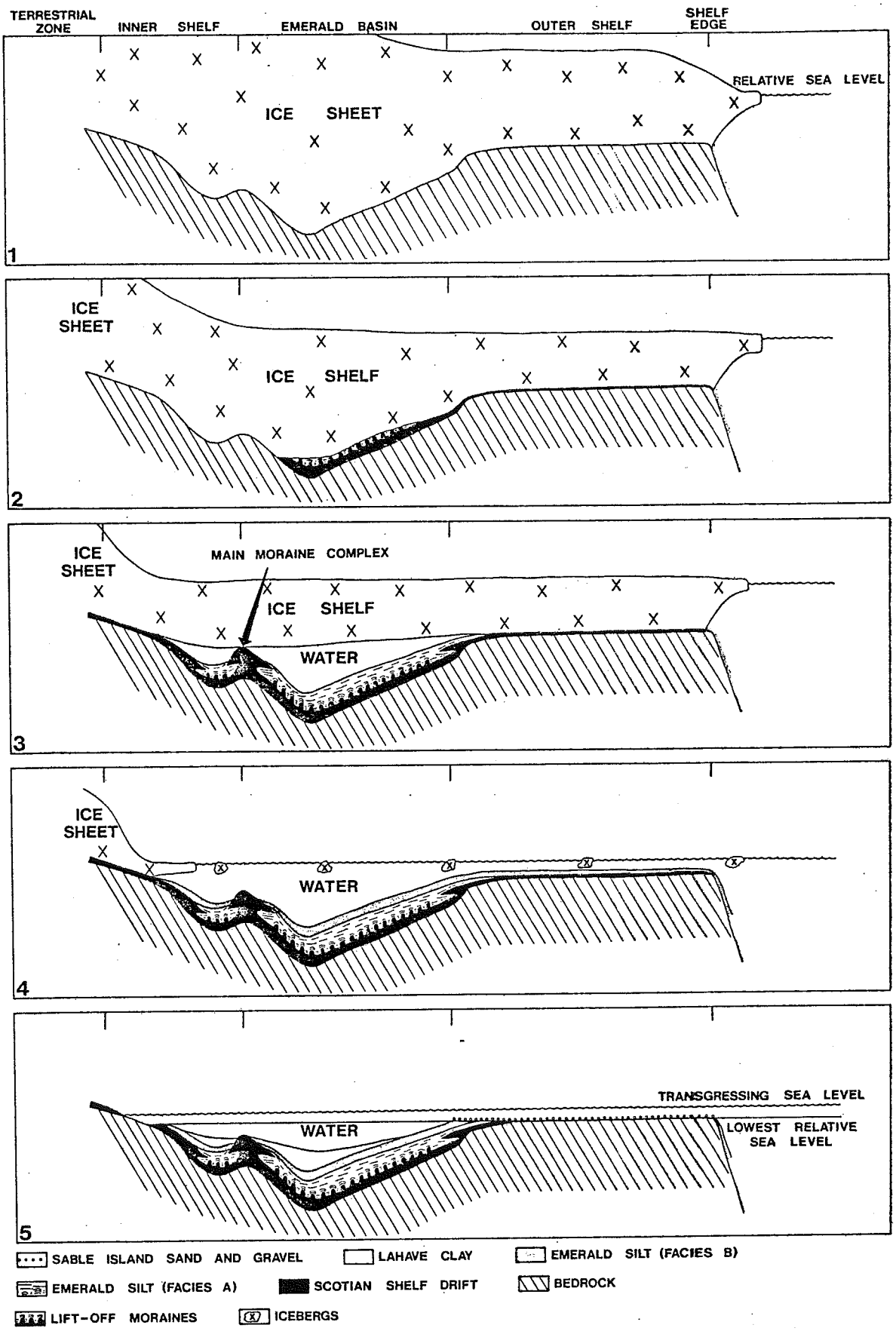

1
2
3
4
5
6
7
LINES OF SECTIONS OF FIGURE 2.
A
3000 M.
2000
1000
0
-1000
-2000
-3000
-4000 M.
ICE
SEA LEVEL
TERRESTIAL ZONE
GROUNDED SHELF ZONE
FLOATING SHELF ZONE
INNER ICEBERG ZONE
OUTER ICEBERG ZONE
(2000 KM)
MAXIMUM SLOPE ABOUT 1 IN 20
B
HORIZONTAL SCALE
0
100
200 KM.
V/H = 25/1
3000 M.
2000
1000
0
-1000
-2000 M.
ICE
TILL
SEA LEVEL
OUTWASH
TERRESTIAL ZONE
GROUNDED SHELF ZONE
INNER ICEBERG ZONE
FLOATING SHELF ZONE

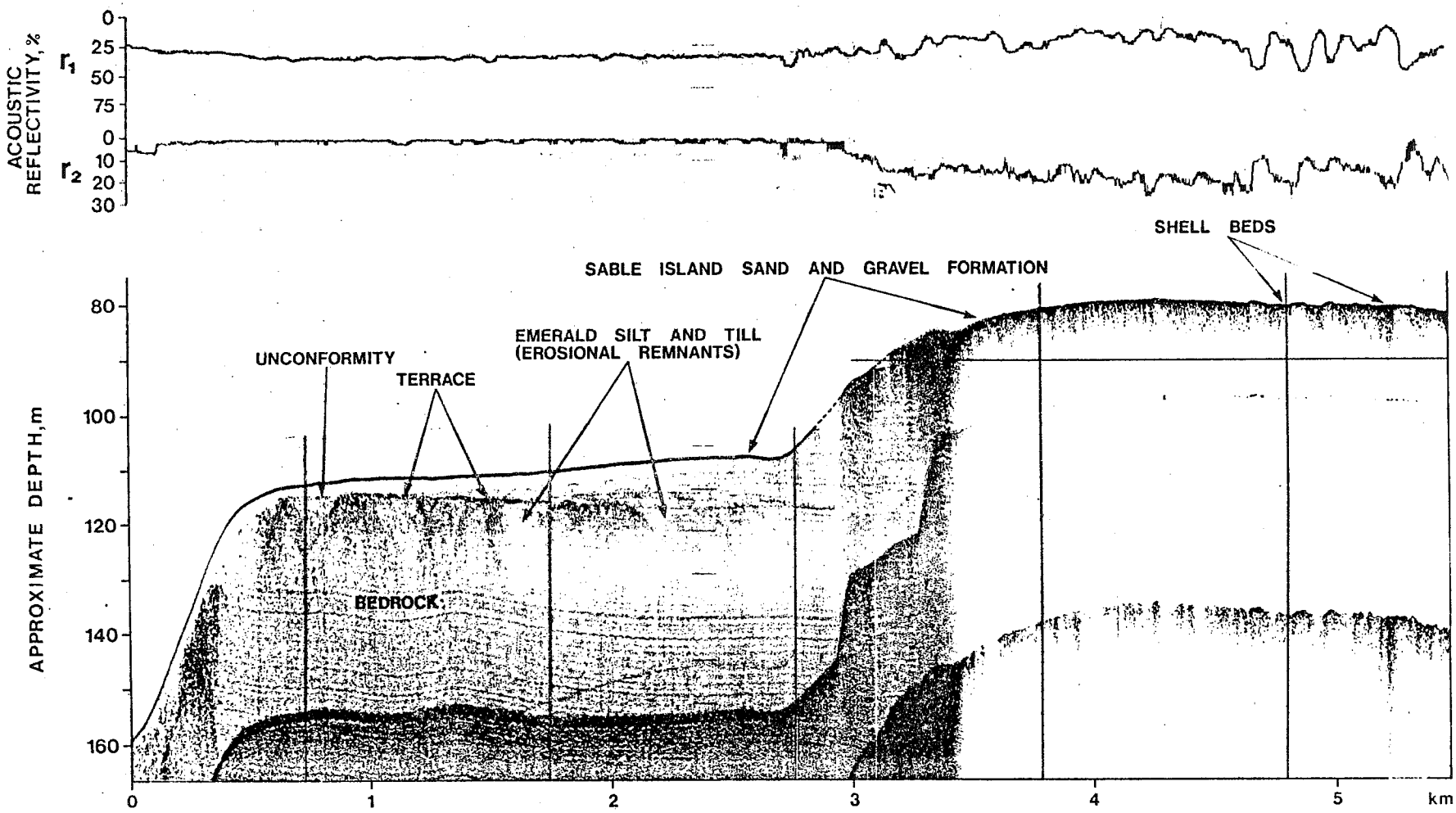

1
2
3a
BUOYANCY LINE
WATER
3b
BUOYANCY LINE
WATER
3c
BUOYANCY LINE
WATER
4
WATER
MAIN MORAINE
TILL TONGUE
BEDROCK
TILL
GLACIOMARINE
ICE

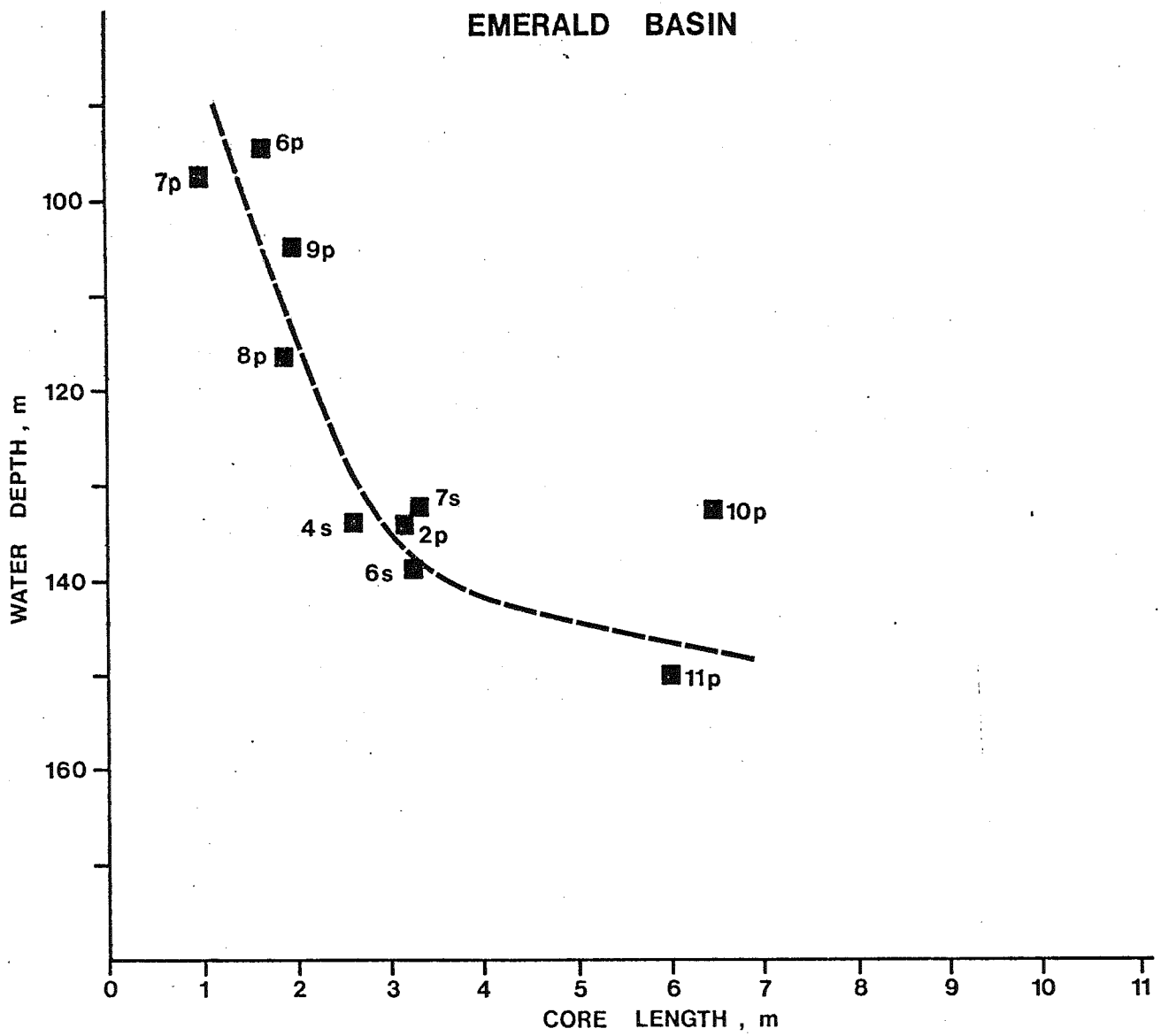

A
SAMBRO MORAINE
ICEBERG FURROWS
EMERALD BASIN
TILL
FEATHER EDGE
LAHAVE CLAY
EMERALD SILT FACIES B
TILL TONGUE
EMERALD SILT FACIES A
APPROXIMATE DEPTH, m
140
160
180
200
220
0
1
2
3
4
5
6
7
8
km
B
EMERALD BASIN
SAMBRO SAND
SABLE ISLAND BANK
FEATHER EDGES
FACIES B
TILL
BEDROCK
EMERALD SILT FACIES B
TILL
TILL
EMERALD SILT
EMERALD SILT FACIES A
BEDROCK
LIFT-OFF MORAINES
120
140
160
180
200
0
1
2
3
4
5
6
7
8
9
km

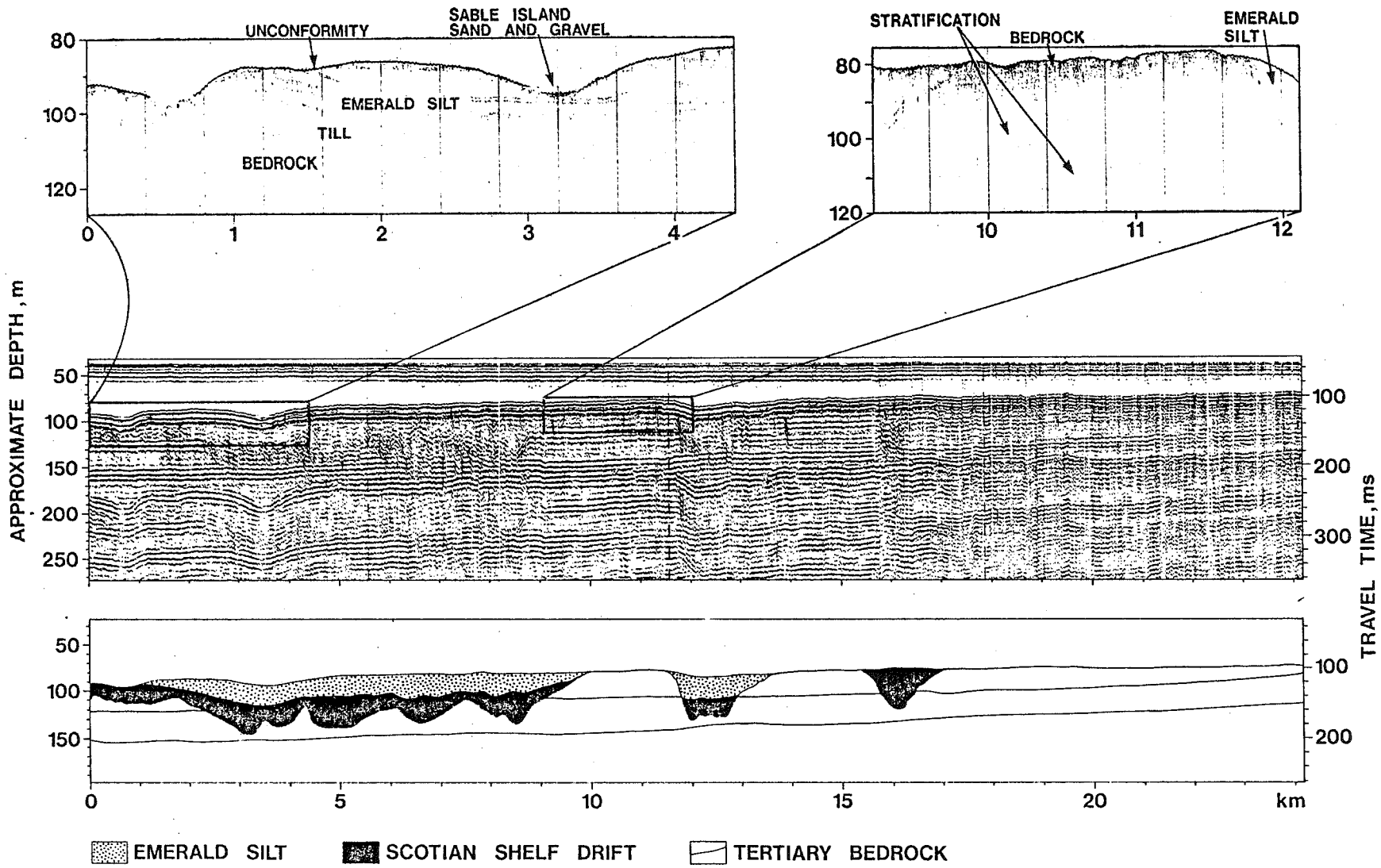

N
S
APPROXIMATE DEPTH, m
200
300
400
TRAVEL TIME, ms
300
400
500
600
LAHAVE CLAY
EMERALD SILT(FACIES B)
EMERALD SILT(FACIES A)
SCOTIAN SHELF-NEWFOUNDLAND SHELF DRIFT
BEDROCK
A
B
0
5
10
15
20
25
30
35
40 km

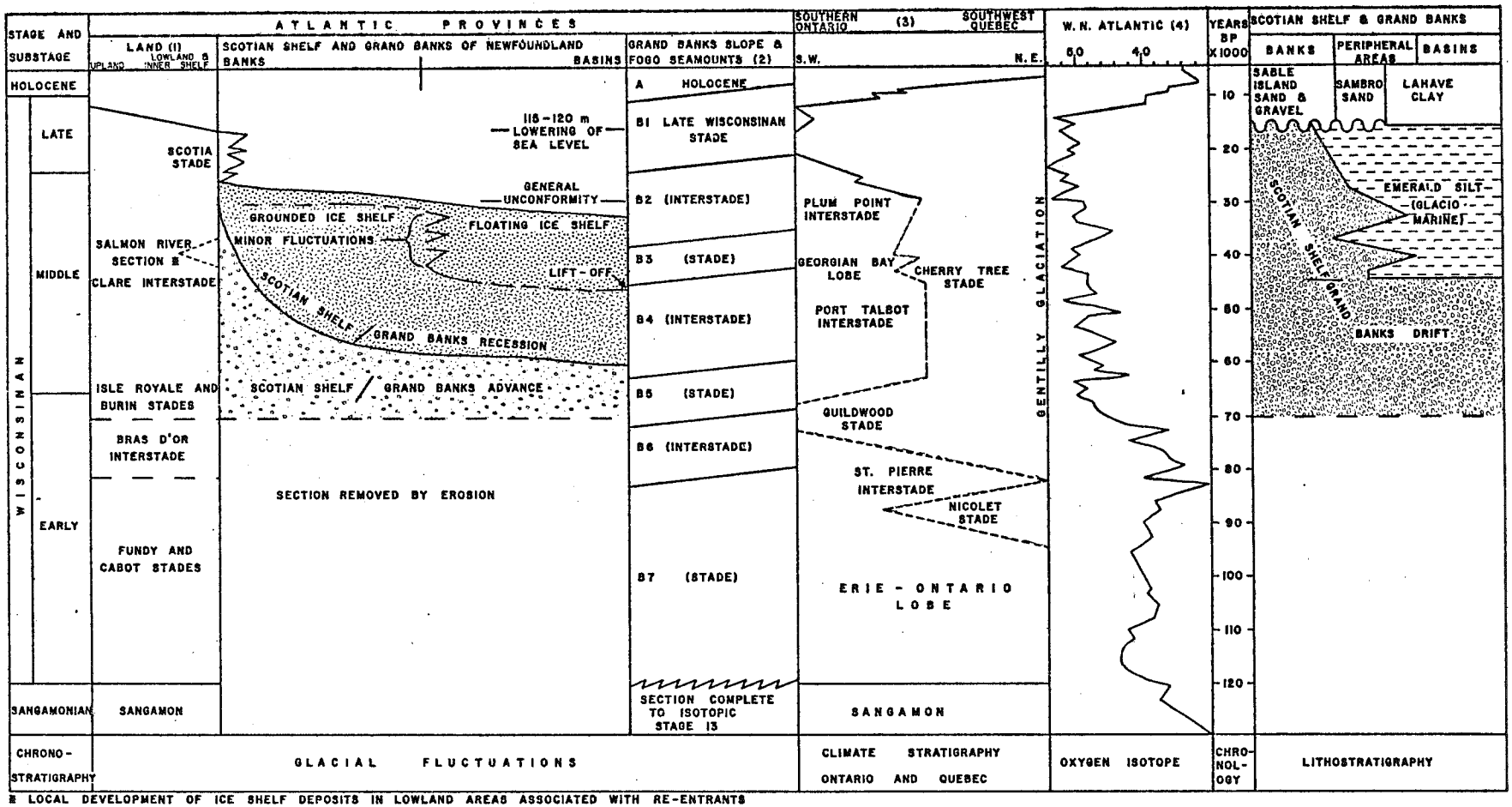

MIDDLE BANK
UNCONFORMITY
EMERALD SILT
SABLE ISLAND SAND AND GRAVEL
TILL
APPROXIMATE DEPTH, m
60
80
100
120
140
0
1
2
3
4
5
6 km
HADDOCK CHANNEL
SAMBRO SAND
UNCONFORMITY
TILL
EMERALD SILT
TILL
EMERALD SILT
TILL
TILL
140
160
180
200
220
0
1
2
3
4
km


TERRESTRIAL ZONE
INNER SHELF
EMERALD BASIN
OUTER SHELF
SHELF EDGE
RELATIVE SEA LEVEL
ICE SHEET
1
ICE SHEET
ICE SHELF
2
ICE SHEET
MAIN MORAINE COMPLEX
ICE SHELF
WATER
3
ICE SHEET
WATER
4
TRANSGRESSING SEA LEVEL
WATER
LOWEST RELATIVE SEA LEVEL
5
SABLE ISLAND SAND AND GRAVEL
LAHAVE CLAY
EMERALD SILT (FACIES B)
EMERALD SILT (FACIES A)
SCOTIAN SHELF DRIFT
BEDROCK
LIFT-OFF MORAINES
ICEBERGS


ACOUSTIC REFLECTIVITY, %
r1
0
25
50
75
r2
0
10
20
30
SHELL BEDS
SABLE ISLAND SAND AND GRAVEL FORMATION
EMERALD SILT AND TILL (EROSIONAL REMNANTS)
UNCONFORMITY
TERRACE
BEDROCK
APPROXIMATE DEPTH, m
80
100
120
140
160
0
1
2
3
4
5
km


EMERALD BASIN
WATER DEPTH, m
100
120
140
160
CORE LENGTH, m
0
1
2
3
4
5
6
7
8
9
10
11
6p
7p
9p
8p
7s
4s
2p
6s
10p
11p


UNCONFORMITY
SABLE ISLAND SAND AND GRAVEL
EMERALD SILT
TILL
BEDROCK
STRATIFICATION
BEDROCK
EMERALD SILT
APPROXIMATE DEPTH, m
TRAVEL TIME, ms
km
EMERALD SILT
SCOTIAN SHELF DRIFT
TERTIARY BEDROCK




* LOCAL DEVELOPMENT OF ICE SHELF DEPOSITS IN LOWLAND AREAS ASSOCIATED WITH RE-ENTRANTS