QUATERNARY PALEOCLIMATES AND SEDIMENTATION SOUTHWEST OF THE GRAND BANKS

by

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TABLE OF CONTENTS

Page

ABSTRACT

Piston cores, 8-12 m long, were collected from the tops of seamounts southwest of the Grand Banks. They were examined for climatic signature and change in the sedimentation pattern. The cores consist of alternating clays and foram-nanno ooze. They were dated by C14 paleomagnetism and foraminifera and coccolith biostratigraphy. Foraminifera and carbonate cycles were used for paleoclimatic determinations. They indicate that Eastern Canada was deglaciated three times during the last 600,000 years B.P. Delaciation occurred at 11,000 years B.P., 130,000 years B.P. and 600,000 years B.P. (although the 600,000 years B.P. date is open to question and could be younger). They probably correspond to the classical Holocene Epoch, Sangamon and Yarmouthian Interglacials. In addition three interstadials occurred in the Wisconsinan and two in the Illinoian stage.

The data from the cores allows reconstruction of the following events. With the advance of continental ice, the Gulf Stream moved towards the southeast and there was southward penetration of a cold water front. With the decrease of temperature, bottom current activity also increased. Biogenic productivity was greatly reduced and the faunal and floral assemblages changed. Advance of the ice sheets on the continental shelf and eustatic lowering of the sea level

resulted in increased clastic sediment supply on the outer-continental margin. The provenance of the clays deposited during the glacial stages can be determined from their mineralogy. The Early Illinoian sedimentation pattern was controlled by the advance of an ice sheet across Newfoundland. The Laurentian ice sheet eroded the Gulf of St. Lawrence and the Laurentian Channel in the Late Illinoian. Sediment contribution from the Wisconsin ice sheet was relatively small.

LIST OF FIGURES

Page

LIST OF PLATES

LIST OF TABLES

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CHAPTER I

INTRODUCTION

The Quaternary has been a period of extensive glaciation accompanied by eustatic oscillation of sea level. On the continental margin of Eastern Canada, as in many other places in the world, there has been rapid variation in the influx of sediments. During times of lowered sea level in the Pleistocene, vast parts of the continental shelves of Eastern Canada were exposed to glaciation; in places subjected to erosion while in others large amounts of sediment were deposited. With sea level lower and ice sheets covering many parts of the continental shelves (Curray, 1965), sedimentation was under quite different conditions than at present. Most of the canyons at the shelf break were active and possibly the site of frequent initiation of turbidity currents. Moreover, glacial meltwater from the shelves and the Laurentian Channel supplied large amounts of fine grained sediment to the continental margin (Heezen and Hollister, 1972). The result was a fairly uniform distribution of fine-grained sediments over a large part of the outer continental margin, complicated by coarse-grained sediments transported down the canyons by turbidity currents and, in some cases, being ice rafted. The retreat of the ice sheets and the rapid rise of sea level left most of the sediment on shelves in disequilibrium with the new environment. In places the shelf was covered by a transgressive sheet of sand, but much of the sediment was left on the storm-dominated shelves, where it could be reworked and substantially influence the input of fine-grained sediments on the outer continental margin. Furthermore, during glacial periods the Norwegian Sea Overflow Water was more important than at present and probably the Western Boundary Undercurrent and the Labrador Current played a more important role. The Gulf Stream was possibly deflected further southeast and played a relatively less important role in the sedimentation in the area south of Grand Banks (Pastovreh et al., 1975).

The pattern of biogenic sedimentation also drastically changed on the outer continental margin. The areas which now receive only foram-nanno ooze and some ice rafted sediments, during glacial times received mostly fine-grained terrigenous sediments and some ice rafted material.

The present study has been based on several cores collected from seamounts southwest of the Grand Banks. The project was designed to provide information about the Late Quaternary history and processes of sedimentation on the outer continental margin off the Southern Grand Banks. More specifically, some of the questions for which answers might be sought were:

1. How has the climate in the Atlantic Canada changed during the last 600,000 years?

2. How has the sedimentation on the continental margin changed from glacial to interglacial stages?

3. How intensive were the various Wisconsin and earlier glacial advances in Eastern Canada and how far seaward did the ice sheets extend?

4. Did the Gulf Stream and Labrador Current change their position with change in climate?

5. Is it possible to correlate the climatic changes recorded in this area with those studied in the Mid-Atlantic Ridge and possibly also with Caribbean cores?

6. Was the ocean's response to climatic change on the continent spontaneous?

7. Was there any significant change in carbonate dissolution with change in climate?

8. The colour of the clay in the cores changes with depth. Are these changes due to difference in source of the clay or due to diagenesis?

9. What is the source of the sediments in this area and do they tell anything about the erosion of the continent, especially from which direction the ice sheets advanced?

10. What are the major process of sedimentation, especially of the fine grained sediments?

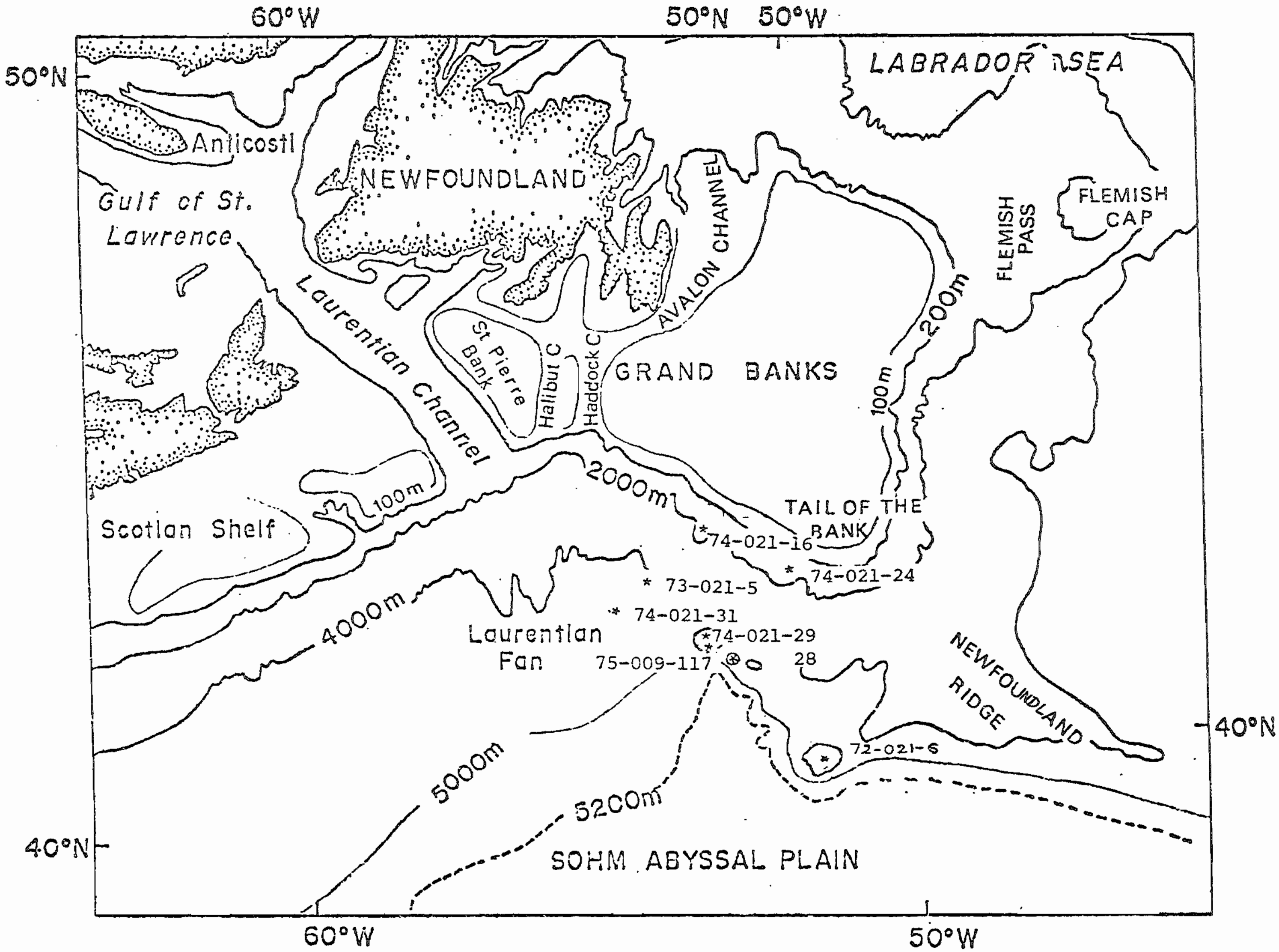

Figure 1. Location of cores

Material

Long piston cores were collected from the continental slope and continental rise off the Grand Banks, during 1974 and 1975 cruises on C.S.S. Hudson. This thesis reports the preliminary results of this study. Since this first phase of the work is more concerned with stratigraphy, only the cores collected from the tops of seamounts, where the rate of sedimentation is very slow, have been examined.

Table I gives the location, water-depth, length and core diameter of the cores. Cores 74-021-29 and 74-021-28 were collected from the same seamount and core 75-009-117 from another seamount 10 km southeast from the first one. The distance of these seamounts from the shelf break is about 65 km. Water depth on the top of these seamounts is slightly more than 3 km and they are about 1 km above the regional seafloor. The tops of these seamounts are covered with thick sediments.

A few other nearby long cores were briefly examined (Fig. 1, Table I). Cores 74-021-16 and 24 are from the Grand Banks margin (Benteau, 1975). Core 72-021-6 is from another seamount south of the Grand Banks (Piper, 1975), and cores 73-011-5 and 31 are from the nearby Laurentian Fan (Stow, 1975 and Edgar, personal communication). Since these cores are described in above mentioned references they are not described in this thesis.

TABLE 1

LIST OF LONG PISTON CORES

Core No.	Length	Core Diameter	Water Depth	Position		Note
				Lat °N	Long °W	
74-021-29	10 m	2 ½"	3140 m	42°18.0'	53°00.0'	Top of seamount
74-021-28	7 m	2 ½"	3328 m	42°21.0'	53°00.4'	Side of seamount
75-009-117	11 m	2¾"	3300 m	42°06.0'	52°44.8'W	Top of seamount
74-021-31	5 m	2 ½"	4646 m	42°27.7'	53°55.9'	Almost of Laurentian Fan, Stow 1975
73-011-5	11 m	2 ½"	3885 m	43°05.3'	53°41.2'	Continental Rise off Grand Banks, Edgar (unpub. notes)
72-021-6	10 m	2 ½"	3780 m	40°20.8'	51°30.0'	Top of seamount Piper, 1975
74-021-16	10 m	2 ½"	1609 m	43°30.7'	52°53.1'	Grand Banks Slope Benteau, 1975
74-021-24	12 m	2 ½"	1404 m	42°59.4'	51°39.6'	Grand Banks Slope Benteau, 1975

REGIONAL SETTING AND PREVIOUS WORK

MAINLAND EASTERN CANADA

Mainland Eastern Canada lies within the Appalachian orogen and comprises mostly Paleozoic rock with some metamorphosed Precambrian rocks exposed on Cape Breton Island, Southern New Brunswick (King _et al._, 1975) and in Newfoundland (Williams _et al._, 1974). The chain has been affected by a series of orogenic disturbances, but the time and intensity of orogeny varies at different places.

In Nova Scotia the pre-orogenic (flysch) assemblage is represented mostly by the Meguma Group, consisting of at least 10 km of alternating layers of quartz metawacke and slate and was presumably deposited in a Cambro-Ordovician deep sea fan complex built offshore from Morocco (Schenk, 1975). The Meguma belt shows evidence of shoaling upwards from deep-sea fans to a paralic lithosome. In Southern Nova Scotia, the pre-orogenic rocks are composed of Late Cryptozoic and Early Paleozoic shelf sedimentary rocks and volcanics. The post-orogenic Late Devonian to Late Triassic rocks (molasse) of the mainland, exposed mostly in Northern Nova Scotia (Avalon Platform), consist of redbeds and minor carbonates all deposited in intermontane basins of a rift-valley setting (Schenk, 1975).

Insular Newfoundland forms the northeastern termination of the exposed Appalachian system and is composed mainly of Late Precambrian and Paleozoic rocks. The central part of Newfoundland represents a former oceanic crust, at least in part, whereas the western and eastern parts of Newfoundland are characterized by continental basement complexes (Williams et al., 1974). Predominant rocks of the Newfoundland Appalachians are graywacke, argillite, carbonate and polydeformed metasediments. Underwater sampling on the Grand Banks suggests that the Precambrian rocks of the Avalon zone continue beneath the surface (Lilly, 1965), and possibly extend up to Flemish Cap (Pelletier, 1971). Post-orogenic sediments in Newfoundland are represented by a Carboniferous molasse sequence which occur within linear graben-like basins in Western Newfoundland. Triassic rocks are not known from Newfoundland.

The flysch sequence of Nova Scotia and the Paleozoic rocks of Newfoundland were deposited in the Proto-Atlantic ocean which apparently opened in the Late Precambrian and was finally closed in the Late Carboniferous with the Hercynian Disturbance. Opening of the present Atlantic is marked by Late Triassic basalt flows in Nova Scotia. A thick Jurassic to Tertiary succession on the Scotian Shelf and Grand Banks was deposited after the opening of present Atlantic (Bartlett and Smith, 1971, McIver, 1972, Jansa and Wade, 1975) and represents a second geosynclinal cycle (King, 1975).

Much of both Newfoundland and Nova Scotia are covered by Pleistocene glacial till.

CONTINENTAL MARGIN

The geology of the continental margin off the east coast of Canada has been reviewed recently by Stanley et al. (1972), Emery and Uchupi (1972), Keen (1974), Keen and Keen (1974), and Grant (1975). Syntheses of the Quaternary history and processes of sedimentation in this area have been attempted by Piper (1975), and Pastouret et al. (1975). Much of the following outline is derived from these sources.

Geographically the continental margin of Eastern Atlantic Canada can be divided into three parts, the Scotian Margin, the Laurentian Channel and Fan, and the Grand Banks.

Scotian Margin

The Scotian Shelf extending from Northeast Channel to the Laurentian Channel, is about 700 km long and its width increases from 100 km off the southern coast of Nova Scotia to 250 km off the coast of Cape Breton. Stanley and Cok (1968) divided the Scotian shelf into three zones; the inner, central and outer shelf. The inner shelf is characterized by a rough topography which can be closely related with mainland physiography and geology. The inner shelf is covered by relatively thin Quaternary deposits, mostly gravel-sand-mud admixtures. The central shelf is characterized by a number of irregular basins, suggesting glacial erosion, possibly guided by pre-glacial fluvial topography (Emery and Uchupi, 1972; King, 1969). These basins and troughs are commonly floored with mud and mud-sand admixtures. The outer shelf is generally broad and flat. It is shallow towards the east and forms two large banks, Banquereau Bank and Sable Island Bank. Most of the outer shelf is covered with sand and sand-gravel admixtures (Stanley and Cok, 1968). The central and outer shelf is underlain by Mesozoic to Tertiary sediments (King, 1975).

The shelf break generally occurs at about 120 m water depth, and the slope-rise junction between 1200 m to 1700 m. The upper slope is generally covered by spill-over sand from the outer shelf and the lower slope with hemipelagic sediments, whereas the continental rise is generally floored by slump and slides, turbidites, contourites hemipelagic and pelagic sediments. The Quaternary history of this area suggests a seaward extension of the shelf edge, slope, and rise by sediment accumulation. Holocene sediments on the slope and rise are mostly gray and vary in thickness from 1 m to 5 m, but most of the Pleistocene sediments are brownish-red in colour (Stanley *et al.*, 1972).

The Laurentian Channel

The Laurentian Channel, separating the Scotian Shelf from the Grand Banks, connects the Gulf of St. Lawrence with the Atlantic ocean. The channel has an average width of about 100 km and a length of 400 km from the edge of the shelf to the Cabot Strait, with an average water depth of 400 m. The origin of the channel is possibly related to the regional structure but its topography has been later modified by glacial erosion (Shepard, 1931, 1973; Nota and Loring 1964; Loring and Nota, 1973; King, 1970). Holocene sediments of the Laurentian Channel are generally grayish silty clays, which are underlain by reddish brown Pleistocene glacial-marine sediments (Conolly *et al.*, 1967). Thicknesses of the Pleistocene and Holocene sediments are quite variable but in general the Quaternary succession is more than 6 m thick and the thickness decreases towards the open ocean. The Laurentian Channel is perhaps the most important source of sediments to the outer continental margin off the east coast of Canada. At the end of this channel it forms a large deep-sea fan, about 500 km long and 300 km wide extending from the continental rise to the Sohm Abyssal Plain (Piper, 1975b). The Laurentian Fan is cut by numerous channels. Most of the active channels are floored with gravel or sand, whereas the interchannel areas and abandoned channels are covered with sand and mud. Pleistocene sediments of the Laurentian fan are generally red in colour, alternating with relatively thin gray sediments. The red sediments are believed to be deposited during glacial stages and the grayish foram rich sediments during interstadials (Stow, 1975). Holocene sediments are in general grayish in colour and rich in forams. The average rate of sedimentation on the Laurentian fan is about 15 cm/1000 years during the Wisconsin glaciation and 10cm/1000 years during the Interstadials and Interglacials (Stow, 1975).

Newfoundland Margin

The continental shelf off the southern Newfoundland coast is known as the Grand Banks. Like the Scotian Shelf, it can be divided into three zones; the inner, central and outer shelf. The inner shelf has a very irregular topography dominated by elongate troughs, trending southwest parallel to the regional structure of Newfoundland. These troughs are much deeper than the average depth of the Grand Banks--generally about 300 m, but occasionally reaching 440 m (Uchupi, 1970). The points of greatest depth are near the landward ends of the troughs where many small fiordlike tributaries occur. The troughs are believed to be the result of glacial erosion along structural zones of weakness (Emery and Uchupi 1972). This zone extends from 72 km to 150 km offshore and large areas are covered with gravels. Most of these gravels were deposited during Pleistocene lower sea level by glacial ice or meltwater streams and were subsequently reworked (Sen Gupta and McMullen, 1969). This zone is dominated by a very prominent channel, the Avalon Channel, which separates the main Grand Bank from the Avalon Peninsula. On the southern side of Burin Peninsula, the St. Pierre Channel separates the St. Pierre Bank from the Island. Near the southern end of Avalon Peninsula, the Avalon Channel is divided into two channels, the Haddock Channel and the Halibut Channel, and cut through the outer shelf right into the continental slope.

The central part of Grand Banks also has quite irregular topography, dominated by shoals and basins. In this part of the shelf, deposition was perhaps more significant than erosion. Surface sediments are mostly sand and sand-gravel admixture with some local concentrations of biogenic sand (Sen Gupta and McMullen, 1969). The basins are generally floored with sand-gravel-mud admixtures. The central shelf is about 100 km wide and runs parallel to the inner shelf, but the topography becomes very rough towards the north. Water depth in this part of the shelf ranges from 80 m up to 300 m, but the depth is less than 15 m at the Virgin Rocks and Eastern Shoals. Outcrops of Precambrian rocks at Virgin Rocks possibly indicate that the glacial erosion was not insignificant on this part of the shelf (Emery and Uchupi, 1972).

The outer shelf is broad and flat with water depths ranging from 50 m to 100 m, and the surface sediments are mostly clean sand with scattered pebbles, probably dropped by icebergs (Emery and Uchupi, 1972). The area is more or less similar to the outer shelf off Nova Scotia. The sands on the outer shelf were probably derived by in situ reworking of the Pleistocene substrate during the Holocene transgression and the presence of gravel as far as the shelf break suggests that the Pleistocene ice sheet once extended across the shelf (Slatt, 1974). Most of the sediments of the outer shelf are enriched with biogenic calcium carbonate, primarily barnacle valves with lesser amounts of mollusc, echinoid and foraminifera fragments. The barnacle valves were possibly deposited during a Pleistocene lower stand of sea level and were later reworked in a beach-like environment, as suggested by well-rounded and polished fragments (Müller and Milliman, 1973). Echinoids, foraminifera and some mollusc shells are fresh in appearance and may have been deposited subsequent to the barnacles. Müller and Milliman (1973) maintain that since this regressive sediments has remained relatively rich in carbonate, it indicates a general lack of glacial and fluvial sediment influx during the Late Wisconsin glaciation or in the Holocene.

The continental slope off the Grand Banks is cut by numerous submarine canyons but almost none of them can be traced to water depths shallower than 200 m. The break between the shelf and slope occurs at the same depth as on the Scotian margin, except on the northern Grand Banks which has a broad and gently sloping continental slope and rise, and does not have any well developed submarine canyons. Very little is known about the sediments from the slope and rise. Two cores collected from the southern Grand Banks (Benteau, 1975) suggest that during the glacial stage sedimentation was coarser, mostly sand, presumably emplaced by sand spillover from the shelf or by turbidity currents, while during the interstadials mostly silt and clay were deposited.

Evidence from geophysical investigations suggests that the sediments on the continental rise south of Grand Banks are quite disturbed, possibly due to plate motion and large slumps from the slope (Emery and Uchupi, 1972). Detailed sedimentological data from this area are lacking, but recent studies (Piper and Slatt, 1976) suggest that while substantial amounts of fine grained sediment have been transported by bottom currents, the role of turbidity currents was never insignificant.

The Newfoundland Ridge extends southeastward from the Tail of the Banks, separating the Sohm Abyssal Plain from the Newfoundland basin, which lies to the north. The width of this ridge varies from 240 km in the northwest to 100 km near its southeastern limit at 39°N, 44°W. Water depth over the crest of the ridge gradually increases to the southeast. Spur Ridge trends southwest from the Tail of the Bank almost at right angles to the Newfoundland Ridge. Water depth over the crest of this ridge also ranges from 3000 m where it shoulders against the Newfoundland Ridge to 4000 m at its most southwesternly extension at 40° 10 N, 52° 10 W (Renwick, 1973).

PHYSICAL OCEANOGRAPHY

Surface circulation patterns in the area studied are dominated by the Gulf Stream and the Labrador Current, but the coastal water and the slope water also play important roles. The Gulf Stream is a major circulatory system of the North Atlantic ocean. The current is relatively narrow and well defined between the Florida and Cape Hatteras and is known as the Florida Current. Between Cape Hatteras and Grand Banks it becomes broad, markedly irregular and meandering. In the proper sense the name Gulf Stream applies to this part of the current, but loosely the Gulf Stream refers to both the sections (Emery and Uchupi, 1972). The Gulf Stream appears to reach the bottom of the ocean (Fuglister, 1963) but the velocity of the current falls very rapidly below 500 m from the surface water. The surface velocity is also quite variable; velocities up to 200 cm/sec. along meanders have been reported. Near Cape Hatteras the Gulf Stream departs from the continental slope and flows along the continental rise, above the area having bottom contours of 4000 m to 5000 m. In this part of the Atlantic the Gulf Stream tends to meander with amplitudes of nearly 40 miles and a wavelength of 200 miles. These large scale meanders move and change shape relatively slowly having a period of the order of a month (Fuglister, 1963).

The Gulf Stream meets the Labrador Current near the Tail of the Grand Banks (Figure 2). The Labrador Current has its beginning at the Davis Strait where the southflowing Baffin Land Current, which has followed the western boundary of Baffin Bay from the Arctic Ocean crosses the Strait to join the West Greenland Current. The new current is then reinforced by the water moving outward from Hudson Bay and Foxe Basin through Hudson Strait. It splits into two parts when it encounters the northern edge of the Grand Banks and one part flows over the Grand Banks, especially through the Avalon Channel, while the other part continues towards the southeast along the continental slope and pushes the Gulf Stream further away from the Grand Banks. An expansion or intensification of the Labrador Current in this area is often accompanied by a tongue-like extension of cold water which may extend southeasterly as far as 41°N (Hachey, 1961).

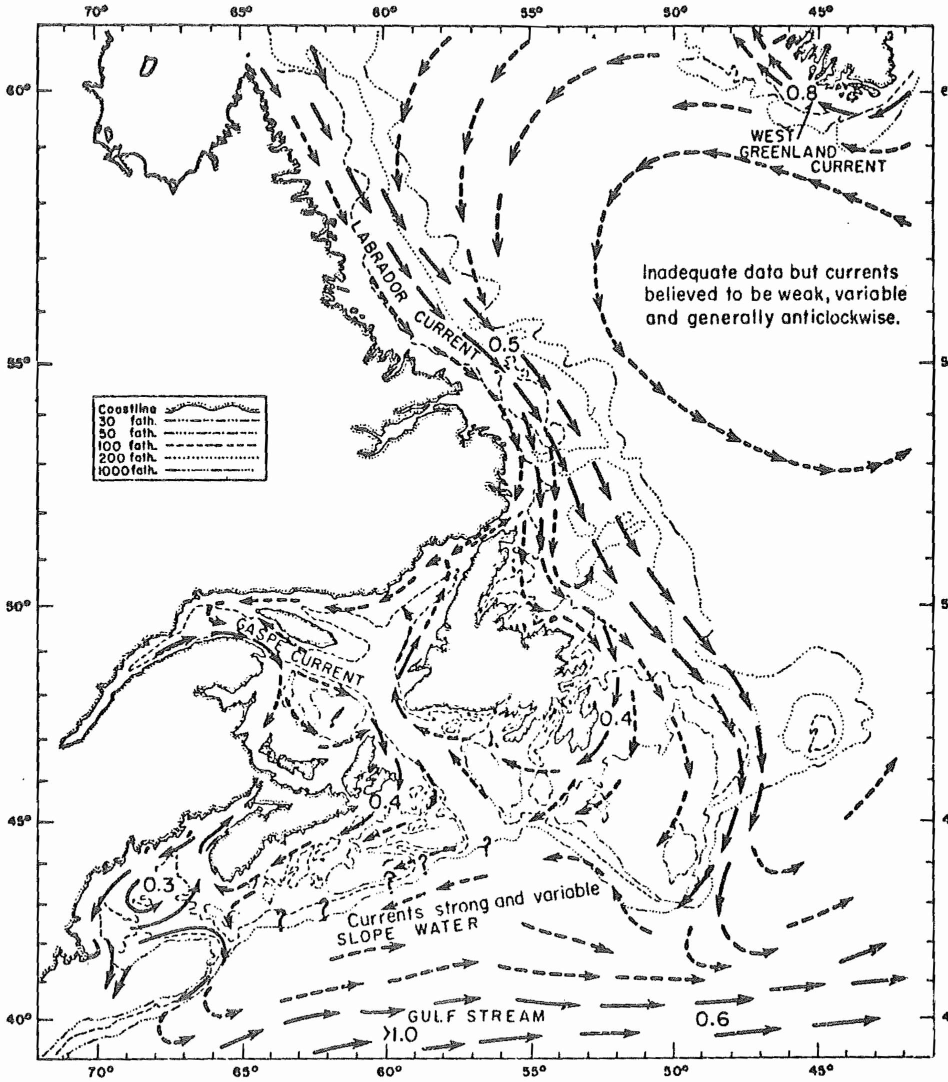

Figure 2. Typical circulation of surface waters around the Grand Banks during spring and summer. Solid arrows: current direction relatively persistent; broken arrows: current direction less persistent. Numbers: approximate speed of current in knots.

(After, Templeman, 1966)

An apparent weakening of the Labrador Current may cause a retreat of cold water northward and an intrusion of Gulf Stream water onto the Grand Banks. During the period from March to May, the cold arctic water is at maximum flood and spreads southward enveloping the entire Grand Banks for a brief period in the surface layer and for more extended periods in the greater depths. As summer advances, the influence of the arctic waters becomes less obvious, and it is generally restricted to the deeper layers.

The term "slope water" has been applied to the band of water between the edge of the continental shelf and the Gulf Stream. It is a continuous band of water from Cape Hatteras to the Grand Banks. It is characterized as mixed water involving a coastal, Labrador current and Gulf Stream contributions.

The term "Norwegian Sea Overflow Water" has been proposed for the cold southward flow countercurrent to the Gulf Stream (Emery and Uchipi, 1972). The Norwegian Sea Overflow Water includes the Labrador water plus some of the Iceland-Scotland Overflow Water and Denmark Strait Overflow Water. Generally, the deep bottom component of the Norwegian Sea Overflow is referred as the Western Boundary Undercurrent. The physical oceanographic evidence would end the Western Boundard Undercurrent at Cape Hatteras, beyond which the Norwegian Sea Overflow Water cannot be identified. But photographic and geological evidence indicates that the Western Boundary Undercurrent continues at least as far as the Bahamas. The current is believed to be strong enough to transport at least fine-grained sediments and to produce ripple marks and scour marks on the bottom of the ocean (Heezen and Hollister, 1972).

In addition to the circulation of surface and subsurface water, physical oceanographic evidence suggest the presence of stratification in the water of the North Atlantic. At least six major density interfaces can be identified within the upper 3 km of water south of the Grand Banks.

Ice Movement

Ice Movement in the area south of the Grand Banks is mostly influenced by the Labrador Current. Most of these icebergs originate from Baffin Bay and East Greenland. According to Hachey (1971), the average travel time of a berg from the time of calving until arrival at the Tail of the Grand Banks is about three years. Icebergs reach as far south as 40°N.

Ice formed in the Gulf of St. Lawrence during the winter months discharges through Cabot Strait in the spring. The discharge of ice normally commences in February and is accelerated in March and April. The ice that forms in the Gulf of St. Lawrence is mostly in the form of ice pack. Ice pack coming through the Laurentian Channel is strongly influenced by the Labrador Current and is carried towards the south. Under present climatic conditions the area south of the Tail of the Bank is not influenced by the Gulf of St. Lawrence ice, but in the Pleistocene conditions could have been different (Piper, 1975a).

QUATERNARY STRATIGRAPHY

Although there have been extensive studies of the Quaternary of North America (Wright and Frey, 1965), the chronology of glaciation is still far from clear, mainly because of the problem of dating different glacial events. In recent years, with the study of Marine Sediments, there has been substantial improvement in this field, but as direct lithologic correlation between the continent and the sea is not possible, one has to depend on time correlation which require precise dating of strata both on land and sea.

There is no general agreement on the exact correlation of the various glacial episodes of North America and Europe

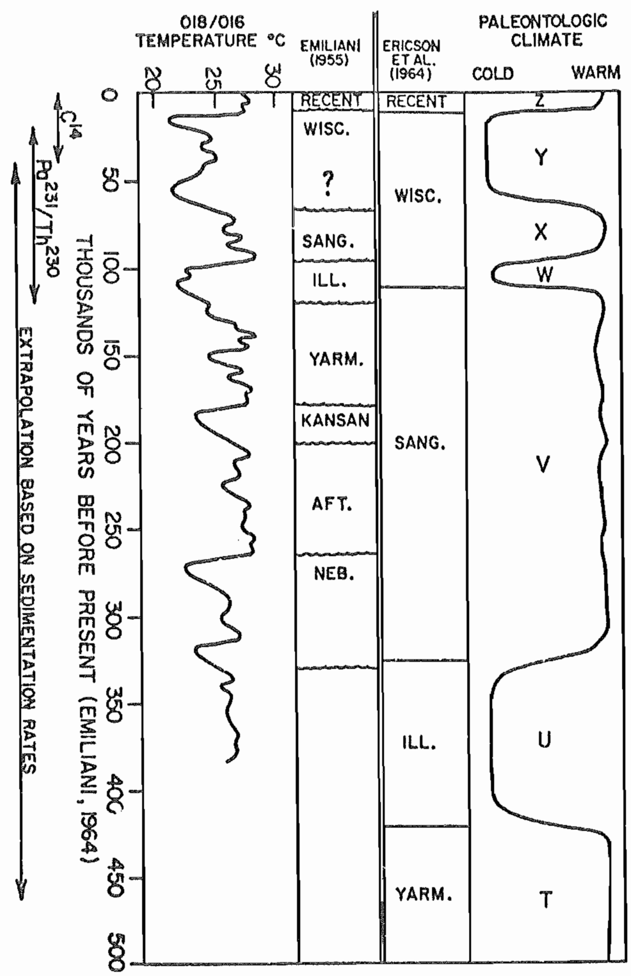

Figure 3. Comparison of climatic curves of Emiliani (1955, 1964) and Ericson et al. (1964). Although the two generally agree in their assignment of ages to various levels in the core, they seriously disagree as to (1) the climate characterizing the period from 120,000 to 300,000 and (2) the correlation with the continental glacial sequences. For example, Ericson et al. consider the "x" zone to be an interstadial with the Wisconsin, and Emiliani correlates it with the Sangamon interglacial. Emiliani assigns the Nebraskan to a level in the core with an age of about 300,000 years, and Ericson et al. to one with an age of 1,500,000 years. (Broecker, 1965).

with the paleoclimatic cycles recorded in the deep-sea sediments. With the exception of the loess sequences (Kukla, 1972) the record on land is essentially a discontinuous one. The various glacial stages preserved on the land as terraces and till deposits have been subsequently cut and denuded during the interveining interglacial stages as well as eroded by subsequent glacial advances. Thus the record is extremely patchy and the preserved part of a given glacial stage in many instances represents a small fraction of the total time involved in the expansion and reduction of the ice sheet.

The result obtained from the study of deep-sea sediments by different workers are also quite variable. The geochemical work of Arrhenius (1952) on the deep-sea cores from the equatorial Pacific and the isotopic work by Emiliani (1955, 1966) suggest the occurrence of numerous, quasiperiodic variations in isotopic parameters with almost constant amplitudes. These variations were shown by absolute dating to be related to the glacial/interglacial cycles of the northern latitudes and led to the conclusion that approximately eight major glaciations had occurred during the past 400,000 years and, more generally, that there were twenty glaciations per million years (Emiliani, 1971). Emiliani (1955, 1965) places the Plio-Pleistocene boundary at the beginning of cold fluctuations of sea temperature determined from oxygen isotope ratios. This time scale places the beginning of the glacial Pleistocene at 420,000 years and the base of the Pleistocene Epoch at 800,000 years.

Another time scale based on deep-sea sediments has been proposed by Ericson and his associates. Although the main difference in approach from that of Emiliani is simply in the methods used for identifying the onset of sea-temperature fluctuations (from changes in foraminiferal assemblages), the two resulting time scales are completely different (Fig. 3). Ericson et al. (1964) estimated the base of the Pleistocene, as well as the initiation of glaciation, at 1.5 million years. Following the establishment of a paleomagnetic time scale they extended the Pleistocene glacial chronology to 1.87 million years B.P. (Ericson and Wollin, 1968). Berggren (1972) identified the initiation of major cooling in the northern hemisphere at three million B.P., in the Mid-Pliocene.

These two time scales and the problem of dating them radiometrically and correlating them with continental glaciation are critically discussed by Broecker (1965) and Cooke (1973). Both of these authors favour a "long chronology" as proposed by Ericson and his group rather than the "short chronology" of Emiliani.

The Plio/Pleistocene boundary is based on a type section in Italy and it has been dated by magnetic stratigraphy at about 2 m.y. Most workers use an approximate correlation with the base of the Olduvai Paleomagnetic event as the most convenient working definition of the Plio/Pleistocene boundary in the ocean (e.g. Berggern, 1972).

Quaternary of North Atlantic

Perhaps the most significant contribution to the study of the Quaternary of the Atlantic is the work done by Ericson and his associates (Ericson *et al.*, 1964, 1968). They claim to find the evidence of four major glacials of land in the Equatorial Atlantic cores and correlate them with the Wisconsin, Illinoian, Kansan and Nebraskan glaciations. Their correlations needs confirmation, but most of their zones can be recognized in almost all Atlantic cores, even in those from higher latitudes (Kennett, 1970). Recent work on the North Atlantic by McIntyre and Ruddiman, 1972 and Ruddiman and McIntyre (in press) suggest about 7 major climatic changes in the North Atlantic during the last 600,000 years.

Berggren (1972) reviewed the Quaternary studies of the North Atlantic and concluded from the previous work and from the work on DSDP leg 12, that in North Atlantic the first appearance of Globorotalia truncatulinoides and the disappearance of Discoaster correspond approximately to the Plio/Pleistocene boundary.

Quaternary of Eastern Canada

The Quaternary stratigraphy of Eastern Canada is fairly incomplete and patchy. Data from land suggest two prominent widespread interstadials within the Wisconsin stage: the St. Pierre Interstadial, at about 70,000 years B.P. and the Port Talbot Interstadial approximately at 45,000 years B.P.. Another less prominent interstadial, the Plum Point Interstadial occurred at about 22,000 years B.P. (Dreimanis, 1971). Only during the St. Pierre Interstadial was there deglaciation of most of the St. Lawrence lowland (Gadd, 1971).

Prest and Grant (1969) suggested that the Wisconsin ice sheets in Atlantic Canada were mostly of local origin. The major ice caps developed in Quebec, in New Brunswick and possibly also in Newfoundland (Brookes, 1970) and Nova Scotia (Grant, 1971). McDonald (1971) suggested that the earliest Wisconsin glacial advance was more severe in Eastern Canada but post St. Pierre glacial phases were both more extensive than the earliest Wisconsin glaciation and affected areas farther west. Prest and Grant (1969) concluded that Laurentide ice was not active over the Maritime provinces in Wisconsin, as had generally been believed previously, and that the growth of Appalachian glaciers during the build-up of the last continental ice sheet may have effectively barred Laurentide ice from most parts of the region. The Laurentian Channel served as an outlet that diverted Laurentide ice through Cabot Strait to the Atlantic Ocean.

CHAPTER II

GENERAL METHODS

The cores on which this study is based were collected from the tops of seamounts southwest of the Grand Banks, on two Bedford Institute of Oceanography and Dalhousie University cruises on C.S.S. Hudson in 1974 and 1975. Core locations and their water depths are shown in Figure 4. Navigation was by Loran-C, controlled by satellite navigation. The Loran-C provided continuous position data, while the satellite navigation was used to check whether the Loran-C was locked into the correct signal peak. The accuracy of the ship's position was within ± 200 m in an east-west direction and ± 800 m in a north-south direction.

Cores 74-021-28 and 74-021-29 were collected from the same seamount, but core 29 came from the flat surface on the top of the seamount, whereas core 28 was from the flank. The seamount is about 1 km above the sea bed. On this cruise an Alpine 2½" I.D. piston corer fitted with ½" cable and an 1800 lb head was used. An Alpine 1 " I.D. gravity corer was used as a tripping device for the corer. Gravity cores suggest that considerable amounts of sediment are missing from the top of the piston cores, presumably because of an incorrect length of trip wire. Moreover, core 28, being from the side of the seamount, is very likely to represent slump and spill over sediments.

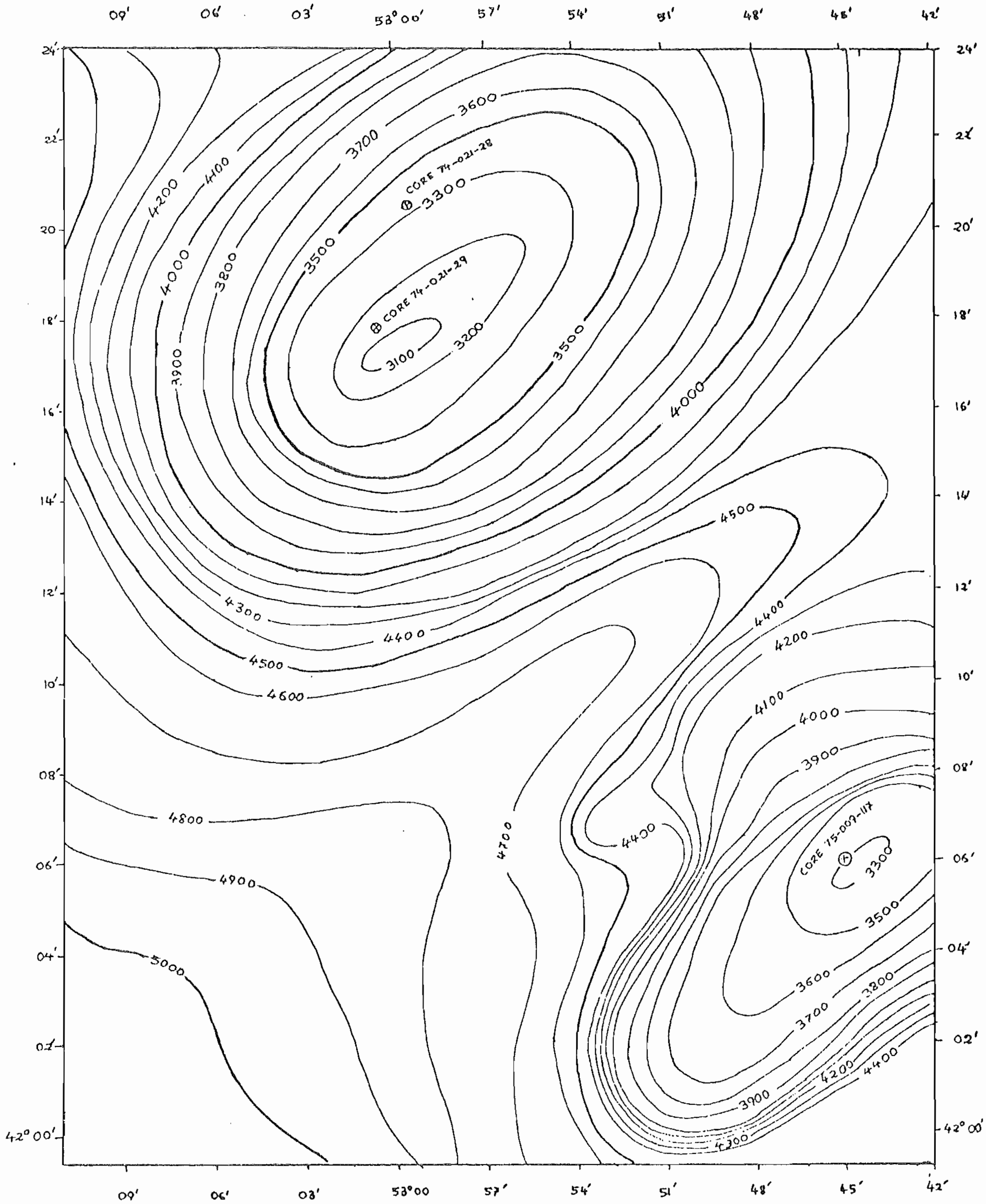

Figure 4: Bathymetric Chart showing core location (depth in meters)

Core 75-009-117 was collected from another seamount 10 km southwest of and 150 m deeper than the first one, using a Benthos $2^3/_4$" I.D. piston and $2^3/_4$" gravity corer. Both core penetration and preservation are very good and the correlation between the gravity and piston cores suggest that almost nothing is missing from the top of the piston core.

Cores were cut into 1.5 m lengths and stored in a cold room at 4°C for periods of up to four months before they were split.

Cores were first X-radiographed and then split in their plastic liners. It was found essential to X-radiograph the cores before splitting, since the mud is very sticky and the cores may be considerably damaged. The split faces were cleaned and described centimeter by centimeter (Appendix). Smear slides were prepared at 1 to 10 cm intervals to examine the lithology closely. X-radiographs were logged systematically for sedimentary structures. Immediately after splitting, one half--the archive half--was sealed in a plastic sleeve with a small amount of water and stored in the cold room. The other half--the working half--was also similarly sealed and stored in the cold room.

Samples were collected systematically for carbonate analysis, grain size, paleomagnetism, study of foraminifera and coccoliths, clay mineralogy and silt and sand mineralogy. Selected samples were also taken for C 14 dating, impregnated thin-sections and for the study of carbonate dissolution.

Bulk carbonate content was determined at 10 cm intervals by volumetric methods as described by Black et al. (1965). Foraminifera samples were taken at 5 cm intervals or less from parts of the core with high foraminiferal concentrations. From homogenous muds with very rare foraminifera, samples were taken at 50 cm intervals and less where there was any marked change in the physical properties of the mud. Samples were dried and soaked in water with about 10 cc calgon to disperse the clay and then were washed in a 63μ sieve with distilled water until the washings became clean. In some cases it was essential to use H_2O_2 but in that case sodium borate was used as a buffer and very light ultrasonic treatment was given. In most cases, however, no H_2O_2 or ultrasonic treatment were used because of their effect on the ultrastructure of the foram tests.

The relative abundance of different species in the samples was determined and their paleoclimatic significance was evaluated. Coiling directions of Globigerina pachyderma and Globorotalia truncatulinoides were used to have a better control over the relative changes in climate. Generally 100 individuals were counted.

The porosity of _G. pachyderma_ and _G. bulloides_, also believed to a paleoclimatic indicator, was measured with a Cambridge 600 SEM.

The effects of carbonate dissolution were measured on the ultrastructure of _G. pachyderma_ and _G. bulloides_.

Standard sieve and pipette techniques have been used for grain size analysis. Paleomagnetic properties were measured on a DSM-1 Spinner Magnetometer.

Untreated and glycolated samples of the less than 2μ fraction of clay were run on a Phillips PW 4620 X-ray diffractometer. Grain mounts and thin-sections were prepared from the sand fraction. Impregnated thin-sections were used to study the structure of the silt laminae. Smear slides were used to determine the relative abundance of coccoliths and other biogenic components of the sediment.

DESCRIPTION OF CORES

The cores consists of three basic lithofacies, (1) foram-nanno ooze, (2) gray mud, (3) red mud. However, another intermediate type of lithofacies, a mixture of mud and foram-nanno ooze is also present, mostly in the upper part of the cores. The nomenclature used here is after DSDP (Weser, 1974). The term mud is used for all sediments consisting predominantly of clay and silt. Colours are described from the GSA Rock-Color chart.

The lower part of the stratigraphic succession cored consists mostly of gray mud. This is overlain by red mud and the uppermost part of the sequence again consists mostly of gray mud. Beds of foram-nanno ooze 5 - 45 cm thick are found throughout the succession, apparently unrelated to variation in colour of the mud.

DESCRIPTION OF LITHOFACIES

Foram-Nanno Ooze

This occurs as relative thin beds (5 to 45 cm) alternating with gray and red mud. The sediment is light grayish in colour and consists mostly of planktonic foraminifera and coccoliths with a small amount of ice rafted pebbles and other terrigenous material. The carbonate content is quite variable, but is typically around 50% with a maximum of 75%. In general, beds of foram-nanno ooze have gradational tops and bases. The size and amount of pebbles are quite variable. Most of these pebbles have sharp angular edges and appear to be freshly broken. The pebble concentration is fairly high near the base of the beds and then gradually decreases towards the top. Petrologically these pebbles consist mostly of metasedimentary rocks, limestones, basalt and granite. Both the foraminifera concentration and the carbonate content is generally highest at the middle part of these beds. The major detrital components of the sand fraction are quartz, with little clastic carbonate, feldspar, glauconite and amphibole. The glauconite and the detrital carbonate content are rather higher than in the other lithofacies. The clay mineralogy of this facies consists of illite, chlorite and considerable amounts of montmorillonite and kaolinite (Figure 8).

Red Mud

This is grayish red (10R 4/2) to pale brown (5YR 5/2) in colour and generally contains less than 1% sand. The red mud occurs only at one stratigraphic horizon in the cores, forming a unit about 3 m thick. The sand size particles are mostly forams. The detrital component consists of quartz, small amount of clastic carbonate, feldspar, ilmenite and trace amounts of amphibole, epidote and other heavy minerals. This facies in general is relatively rich in ilmenite.

Most of this facies is almost structureless but locally there are fine silt laminae and bands. The bands are generally less than 1 cm thick and show very slight changes in grain size and in mineralogy from the surrounding sediments. They can be distinguished only in x-radiographs and thin sections. In general, the silt laminae are less than 1 mm thick with only a few having thicknesses up to 2 mm. Most of these laminae are fairly continuous; however, about 10% are lenticular and discontinuous. The laminae appear in groups and solitary silt laminae are relatively rare. Illite, chlorite and kaolinite are the most common minerals in this facies.

Gray Mud

This is brownish gray to olive gray in colour and occurs in the upper part as well as in the lower part of the cores. Dark yellowish brown (10YR 4/2) muds are included in this facies. The amount of sand present is very low, normally not more than 1%. Sometimes, especially in the upper part of the stratigraphic succession this gray clay is mixed with foram-nanno ooze, resulting in an increase of sand size particles. Major sedimentary structures visible on the clean core face are fine, thin, closely grouped or solitary silt laminae. The laminae are more or less similar to those present in the red mud. Other structures includes some light gray and some black mottles. In the x-radiograph this facies shows a few thin pebble horizons. Mineralogy of the sand fraction is similar to other facies except that the amphibole content is higher. Illite and chlorite are the dominant clay minerals.

Muddy Foram-Nanno Ooze

At some horizons a facies transitional between the foram-nanno ooze and either red or gray mud occurs, comprising a homogenous mixture of subequal amounts of each lithology. This facies occurs mostly stratigraphically in the upper gray horizon (Figure 5). Carbonate content of this facies is fairly high but lower than the foram-nanno ooze. X-radiographs show the presence of pebbles and granules throughout the bed but with higher pebble concentrations being near the base.

SEQUENCE OF LITHOFACIES

The tops of the piston cores may be missing, but since piston core 75-009-117 correlates perfectly with the gravity trip core, the top of this core is presumed complete. Taking core 75-009-117 as the "key" core, a sequence of lithofacies can be established in it, which is divided into lithostratigraphic units as follows:

22 cm	Foram-nanno ooze	Unit A
18 cm	Muddy foram-nanno ooze	Subunit B1
16 cm	Foram-nanno ooze	Subunit B2
26 cm	Gray mud	Subunit B3
17 cm	Muddy foram-nanno ooze	Subunit B4
50 cm	Gray mud	Subunit B5
35 cm	Foram-nanno ooze	Subunit B6
50 cm	Gray mud	Subunit B7
45 cm	Foram-nanno ooze	Unit C
350 cm	Red mud	Unit D
15 cm	Foram-nanno ooze	Unit E
120 cm	Gray mud	Subunit F1
10 cm	Foram-nanno ooze	Subunit F2
160+ cm	Gray mud	Subunit F3

In the case of cores 74-021-28 and 29 the piston and gravity cores do not correlate with each other, suggesting that the tops of these cores are missing. Nevertheless, the

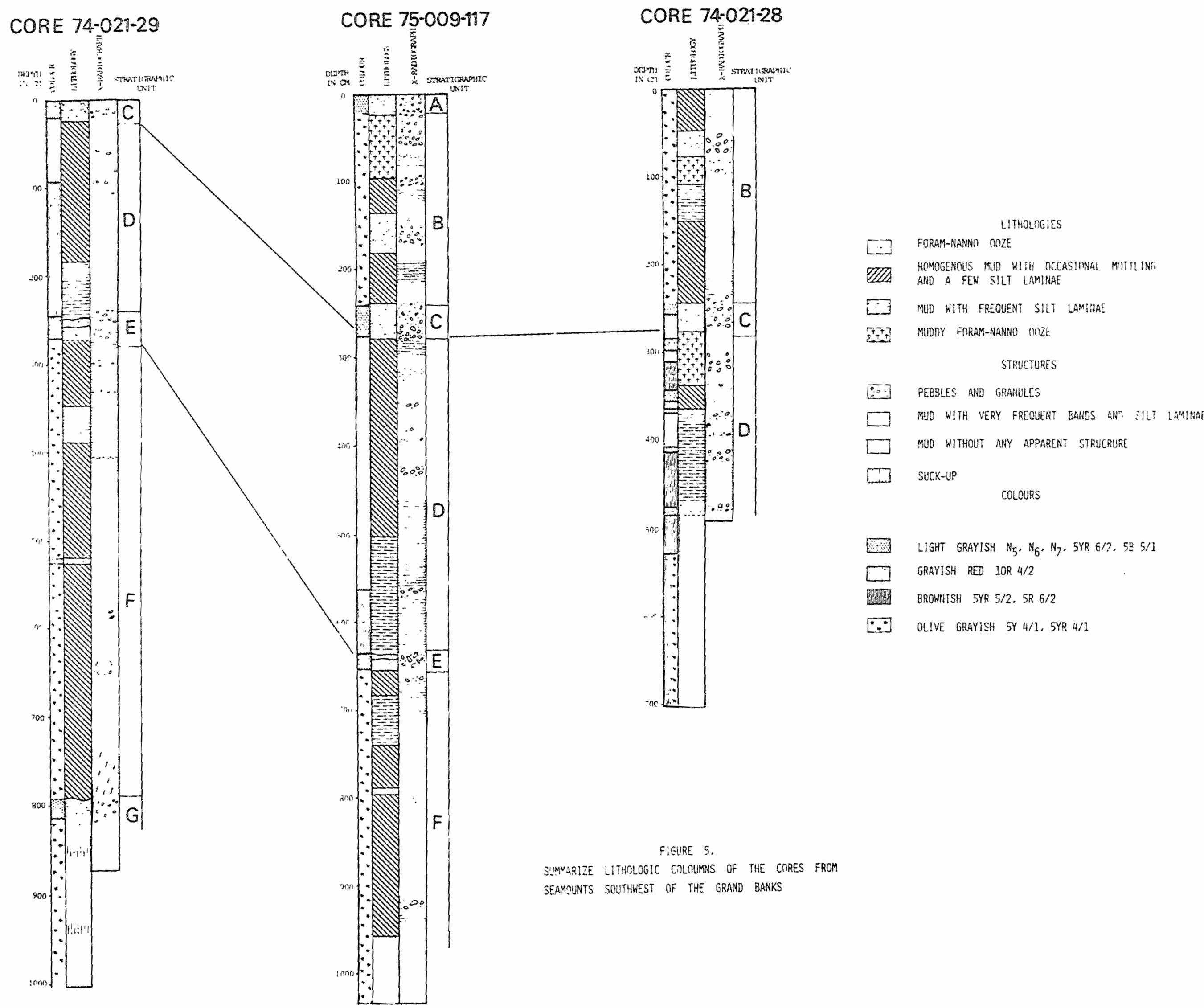

FIGURE 5.
SUMMARIZE LITHOLOGIC COLOUMNS OF THE CORES FROM SEAMOUNTS SOUTHWEST OF THE GRAND BANKS

probable correlation of these two cores with core 75-009-117 is shown in Figure 5. It is mainly a lithologic correlation, since the foram-nanno ooze horizons at the top of the red mud facies (unit C) and at the base (unit E) serve as excellent datum planes. This correlation is further confirmed by the paleontological analysis described in Chapter V. As the core 74-021-29 has greater penetration, we can use this core to extend the sequence of lithofacies further downward:

10	cm	Foram-nanno ooze	Subunit F2
250	cm	Gray mud	Subunit F3
30	cm	Foram-nanno ooze	Unit G

In subsequent discussion, core 75-009-117 is used as the type section of units A through F1 and 74-021-29 for F2 through G. Depths of particular units in the cores refer to these type sections.

DESCRIPTION OF STRATIGRAPHIC UNITS

Unit A

Unit A occurs at the top of all gravity cores and in the upper part of core 117, but is missing from cores 28 and 29. The thickness of this unit generally varies from 20 cm to 25 cm and it consists mostly of yellowish gray (5Y 8/1) to light olive gray (5Y 6/1) foram-nanno ooze with varying amounts of pebbles and granules. This unit is particularly sandy and the percentage of sand and pebble sized material ranges from 40 to 50%; silt varies within 20 to 30% and clay is about 25%.

X-radiographs show pebbles throughout the unit, but a higher pebble concentration near the base. The proportion of clay is also higher near the base of this unit and gradually decreases upwards. The concentration of both foraminifera and coccoliths increases towards the top of this unit, with the average percentage of coccoliths about 10-15% of the total volume of the sediments. Carbonate content near the base is about 30%, increasing to about 70% at the top of the unit.

Unit B

Unit B consists of light brown to light olive gray mud alternating with foram-nanno ooze. The complete section of Unit B is present in core 117, and the upper part in gravity core 117G. This unit is completely missing from core 29. This unit has been divided into seven subunits.

Subunit B1

Subunit B1 is present in all gravity cores and in core 117, and directly underlies unit A. The thickness of this Subunit varies from 15 cm to 25 cm, with an average of 18 cm. It consists of muddy foram-nanno ooze of light brown (5YR 6/4) to pale red (5R 6/2) colour with some light grayish mottles. This subunit is less sandy than unit A, with the percentage of sand and granules in the range of 25 to 35%, of which about 75% is biogenic carbonate.

This subunit is almost structureless even in x-radiographs, except for a few scattered granules. However, granule concentrations increase towards the upper and lower contacts. Average carbonate content of this subunit is about 20%: this also increases towards the upper and lower contacts. Coccolith concentrations is much lower than in unit A.

Subunit B2

Subunit B2 is light olive gray (5Y 6/1) muddy foram-nanno ooze with greenish gray (5GY 6/2) mottles. The colour however, changes to pale yellowish brown (10YR 6/2) near the base. This unit is present in core 117 and in gravity core 117G, and possibly partly in core 28G. It is about 16 cm thick. It starts at the base of subunit B1 and continues to a depth of 58 cm in core 117. The upper contact of this subunit is fairly sharp and can be readily picked out. The lower contact is rather gradual but can be easily recognized in the x-radiograph by the presence of a pebble horizon. The pebble concentration gradually decreases upwards. A relatively pebble free horizon appears at a depth of 50 cm (core 117) in the area where the core shows frequent greenish mottles. These mottles are fairly common in the lower part of this subunit. Just near the base, the sediment is mostly muddy and pale yellowish in colour. Foram and coccolith concentrations gradually increase upwards and reach the highest concentration at about 46 cm depth in core 117. Above this horizon the concentration again decreases. A few rare thin beds of clay alternate with the foram-nanno ooze. The thicknesses of such beds is generally less than 1 cm. Grain size is quite variable in this unit. Generally the percentage of sand-sized material and pebbles varies from 20 to 25%. The percentage of clay is highest at the base, lowest near the middle and then increases again towards the top of the Subunit. Carbonate content and coccolith concentrations also show the same pattern.

Subunit B3

Subunit B3 is a pale yellowish brown (10YR 6/2) mud with light brown (5 YR 6/4) mottles. It is present in core 117G and 117, is about 22 cm thick, and extends from 58 cm to 80 cm depth. The upper part of this unit is almost structureless except for a few mottles. X-radiographs show the presence of a 3 cm thick granule and coarse sandy horizon at a depth of 70 to 73 cm and below this level the clay is interlaminated with silt laminae. These silt laminae are quite disturbed and visible only in x-radiographs. The thickness of these laminae is generally less than 1 mm and most of them usually have a sharp top and base. A few are lenticular. This subunit is mostly muddy, the percentage of sand being in the range of 5 to 10%. Both the carbonate content and coccolith concentration are rather low.

Subunit B4

Subunit B4 occurs below subunit B3 in cores 117G and 117, and extends to a depth of 110 cm, with an average thickness of about 30 cm. It consists of light olive gray (5Y 5/2) muddy foram-nanno ooze, occasionally mottled with brownish mud. Both the upper and lower contacts of this unit are rather gradual. The lower contact is more easily recognized in x-radiographs, rather than on the clean core face, since it corresponds to the disappearance of pebbles. The upper part of this subunit is mostly muddy with frequent silt laminae, visible only in x-radiographs. The middle and lower parts show a scattered distribution of pebbles with highest pebble concentration between 90 to 100 cm, and a gradual decrease towards both the base and the top of the subunit. Foram-nanno concentrations also show the same pattern. This unit is fairly sandy, with the percentage of pebbles and sand varying from 10 to 20%. Most of the sand size fraction is however foraminifera. The highest carbonate content of 30% is reached at 95 cm, with a gradual decrease to about 17% towards the top and base of the subunit. Coccolith concentration also shows a similar pattern, with an average concentration of about 7 to 9%.

Subunit B5

Subunit B5 is a light olive gray (5Y 5/2) mud with about 5% sand. This unit is present in cores 117G and 117, between the depths of 110 cm and 140 cm. In x-radiographs the upper part of this subunit shows frequent bands and sharp silt laminae but the lower part is almost structureless homogenous mud. No pebbles were observed either on the clean core face or in x-radiographs. Foraminifera and cocco-ltih concentration is very low in this subunit and so is the carbonate content (about 17%). This subunit is believed to be present also in the upper part of core 74-021-28, but it is very disturbed.

Subunit B6

Subunit B6 occurs in core 117 at a depth of 140 cm to 175 cm and in core 28 between 50 cm and 90 cm depth, the average thickness being about 35 cm. It consists of light olive gray (5Y 5/2) to light gray (N7) foram-nanno ooze and pebbles. X-radiographs show a higher pebble concentration near the base of this subunit with a gradual decrease towards the top. Foram-nanno concentration is highest between 155 cm and 165 cm and it gradually decreases towards the upper and lower contacts. This subunit is quite sandy, with the sand

and pebble size fraction varying between 30 to 45%. Most of the sand size fraction consists of foraminifera, with small amounts of quartz. This subunit is fairly rich in calcium carbonate: the highest carbonate content of 55% is reached in the middle part of this subunit and gradually decreases towards the contacts.

Subunit B7

Subunit B7 is light olive gray (5Y 5/2) homogenous mud and is about 60 cm thick in core 117. In core 28, this sub-unit shows a slight change in colour towards brownish gray (5YR 4/1). It is thicker in core 28. This subunit contains comparatively small amounts of sand, generally not more than 5%. X-radiographs show silt laminae and bands in core 117 at 190 to 210 cm, but no pebbles were noticed. Most of these silt laminae are very sharp and their thickness is generally less than 1 mm. Core 28 shows frequent bands and silt laminae throughout the entire length and a few granules near the top, of thie subunit. This subunit is very low both in carbonate content and coccolith concentration.

Unit C

Unit C consists of light gray foram-nanno ooze with pebbles and granules. The average thickness of this unit is about 45 cm and it occurs in all three piston cores, but is especially well preserved in core 117. This unit separates the upper gray mud of unit B from the red mud of unit D. Both the foraminifera and coccolith concentrations are highest in the middle part of this unit and gradually decrease upwards and downwards. Pebble concentration shows a pattern similar to that in other foram-nanno ooze horizons, with the highest concentration near the base. A few thin bands of clay alternate with the foram-nanno ooze, especially near the base and top of this unit. The thickness of such bands is generally less than 1 cm, but in core 28 these bands are slightly thicker. This unit has a very high carbonate content, reaching 77% in the middle part.

Unit D

Unit D is grayish red (10R 4/2) to pale brown (5YR 5/2) mud with very small amounts of sand, normally less than 1%. It occurs in all the piston cores. The thickness of this unit is variable, the average being 350 cm. The upper part of this unit is almost structureless, except for a few light grayish mottles. The lower part is laminated with fine silt laminae, most of which are less than 1 mm thick and have sharp upper and lower contacts. A few however are up to 2 mm thick and rather lenticular. X-radiographs show a few pebble and granule horizons in the upper part of this unit. Carbonate content is very low, generally about 15% but it is slightly higher near the grayish mottles. Coccolith concentration is also very low, and in some cases they are almost absent.

Unit E

Unit E is medium gray (N5) to pale brown (5YR 5/2) foram-nanno ooze. It is about 30 cm thick and occurs in cores 29 and 117. In contrast to the other foram-nanno horizons, this unit has a very sharp top and gradual base. One very prominent muddy horizon divides this unit into two parts, with the lower part much thicker than the upper one. In both parts the foraminifera and coccolith concentration gradually increases upwards and then there is a sudden change to mud. X-radiographs shows only a few pebbles and granules, slightly concentrated near base and top of this unit. This unit is quite sandy, with the percentage of sand and pebbles varying from 10 to 30%. Maximum carbonate content is about 35%.

Unit F

Unit F is brownish gray (5YR 4/1) to olive gray (5Y 4/1) mud with some mottles and fine silt laminae. One foram-nanno ooze horizon divides it into two parts and hence for simplicity this unit has been divided into three subunits.

Subunit F1

Subunit F1 is a brownish gray (5YR 4/1) clay, about 2 m thick and occurs in cores 29 and 117. With the exception of the middle part of this subunit, which is thinly laminated with fine silt laminae, the rest is massive. Occasionally this subunit shows small black mottles or thin black bands, some of which are along silt laminae. In x-radiographs this subunit shows frequent bands and sharp silt laminae, but no pebbles or granules. This subunit contains very small amounts of sand, usually not more than 1%, and the carbonate content is also very low.

Subunit F2

Subunit F2 is a thin band of medium gray (N5) foram-nanno ooze. It is about 8 cm thick and is present in cores 29 and 117. It has sharp upper contact and gradational base. It is relatively sandy, containing about 10-20% sand and granules. X-radiographs show the presence of very small amounts of granules throughout the subunit. Average carbonate content of the subunit is about 25%.

Subunit F3

Subunit F3 is olive gray (5Y 4/1) mud with very small amounts of sand. It is about 250 cm thick and present near the base of cores 29 and 117. The base is seen only in core 29. In x-radiographs this subunit shows frequent bands and sharp silt laminae, and a few very thin pebble horizons.

Both the foraminifera and coccolith concentrations are very low and in some cases coccoliths are almost completely absent. Black mottles and bands are present. Carbonate content is very low.

Unit G

Unit G is light gray (N7) to light brownish gray (5YR 6/1) foram-nanno ooze and occurs in core 29. It is about 30 cm thick with a very sharp upper contact, the lower contact seems to be gradual but was not seen. A few sharp silt laminae appear near the base ofthis unit, just above the level where the core is sucked in. Most of these laminae have a sharp upper and lower contact and are fairly continuous. A few are however, discontinuous and lenticular. The average thickness of the laminae is about 0.5 mm. X-radiographs show scattered pebble distribution throughout the unit. Both foraminifera and coccolith concentrations gradually increase towards the top of this unit. The highest carbonate content is about 70% and is reached near the top of this unit.

CHAPTER III

PETROLOGY

The predominant components of these sediments are firstly clay and secondly biogenic carbonate. In most cases detrital sand and pebbles constitute only a minor fraction.

CLAY MINERALOGY

Analytical procedure

The clay mineral fraction was examined by x-ray diffraction. Samples were prepared by techniques similar to that described by Griffin *et al.*, 1968. Sufficient amount of sample was taken and about 10 ml 10% hydrogen peroxide and 200 ml distilled water were added. The samples were boiled to remove labile organic carbon. On cooling they were wet sieved through a 63μ sieve and the fine fraction was collected in a litre cylinder. Sufficient 1% calgon solution was added to disperse the clay and distilled water was added to bring the volume to 1 litre. After stirring, the suspension was allowed to settle for 16 hours. Then the top 20 cm of the suspension was separated. This consisted of the less than 2μ fraction. About 50 cc of 10% acetic acid was added to the <2μ fraction to remove carbonate. The samples were then washed twice by centrifuging with concentrated calcium chloride solution added as necessary to flocculate.

Iron oxide was removed by the techniques described by Mehra and Jackson (1960). The samples were treated with sodium citrate and sodium bicarbonate, heated to approximately 85°C, then sodium dithionite was added and the samples were stirred for about 15 minutes and were again washed twice by centrifuging.

Samples were run on a Phillips PW 4620 x-ray diffractometer using the smear-on-glass-slide technique, in part at Memorial University of Newfoundland, and at Dalhousie University. The untreated slide was x-rayed at 1° 2θ min. over the range 2° to 15°. The samples were then treated with glycol in a vaccuum desiccator overnight at 80°C and were x-rayed from 2° to 3-° at 1° 2θ/min. A slow scan of the glycolated samples was run at 0.25° 2θ/min. through the range from 24° to 26° to resolve the ∿ 3.5 Å kaolinite-chlorite peaks.

The major groups of clay minerals found in the cores were illite, kaolinite, chlorite and montmorillonite. They were identified by the processes described by Carroll (1970) and Biscay (1965). Mixed-layered clays and different polytypes were not treated separately. In addition to the clay minerals which constitute the major part of the <2μ sample, minor amounts of other minerals were also present but only the quartz, feldspar and amphibole were present in sufficent amounts to permit any quantitative study. Biscaye's (1965) methods of quantifying the abundances of major clay minerals was used.

Results and Discussion

Clay minerals were studied in detail from the 74-021-29 and 75-009-117. The clay in the cores can be divided into four major groups, the red mud (unit D), upper gray mud (unit B), lower gray mud (unit F) and the light grayish clay present with the foram-nanno ooze. (units A, C and G). The most common clay mineral is illite, which is followed by chlorite and relatively small amounts of kaolinite and montmorillonite. Table 2 shows the relative distribution of different clay minerals in the cores.

The red mud contains an average of about 60% illite, 15% chlorite, 15% kaolinite and 10% montmorillonite. In contrast, the olive gray muds show lower kaolinite content. The mineralogy of olive gray mud in both units B and F is similar with an average clay mineral content of about 65% illite, 20% chlorite, 5% kaolinite and 10% montmorillonite.

There are two possible explanations for these mineralogical changes. First they are due to diagenesis, or second they are related to different source areas of the clays, and possibly to different sedimentation process. If they are due to diagenesis, then the mineralogical changes

TABLE 2

CLAY MINERAL PERCENTAGES (BASED ON CLAYS AS 100% OF <2 μ FRACTION) ALL GLYCOLATED

Core No.	Depth in cm	Description	Montmorillonite	Illite	Chlorite	Kaolinite
75-009-117	30	Muddy red foram-nanno ooze	16	55	14	15
75-009-117	100	Gray mud	7	63	30	--
75-009-117	205	Gray mud	6	73	16	6
75-009-117	239	Gray mud	12	66	16	6
74-021-28	100	Gray mud	8	64	19	9
74-021-28	290	Gray mud	-	79	21	--
74-021-29	3	Light gray foram-nanno ooze	16	42	20	22
74-021-29	6	Light gray foram-nanno ooze	13	47	24	16
74-021-29	10	Light gray foram-nanno ooze	20	63	21	12
74-021-29	37	Red mud	9	60	14	17
74-021-29	44	Red mud	11	57	20	12
74-021-29	50	Red mud	11	58	22	10
74-021-29	80	Red mud	18	54	15	14
74-021-29	129	Red mud	9	57	26	9
74-021-29	189	Red mud	5	64	20	12
74-021-29	236	Red mud	10	54	23	14
74-021-29	248	Red mud	12	52	24	12

TABLE 2 (continued)

Core No.	Depth in cm	Description	Montmor-illonite	Illite	Chlo-rite	Kaoli-nite
74-021-29	262	Light gray foram-nanno ooze	20	41	22	17
74-021-29	489	Gray mud	-	71	21	8
74-021-29	500	Gray mud	17	56	20	7
74-021-29	589	Gray mud	5	65	24	4
74-021-29	617	Gray mud	4	74	13	9
74-021-29	669	Gray mud	3	71	26	-
74-021-29	678	Gray mud	9	71	13	7
74-021-29	791	Light gray foram-nanno ooze	29	42	14	16

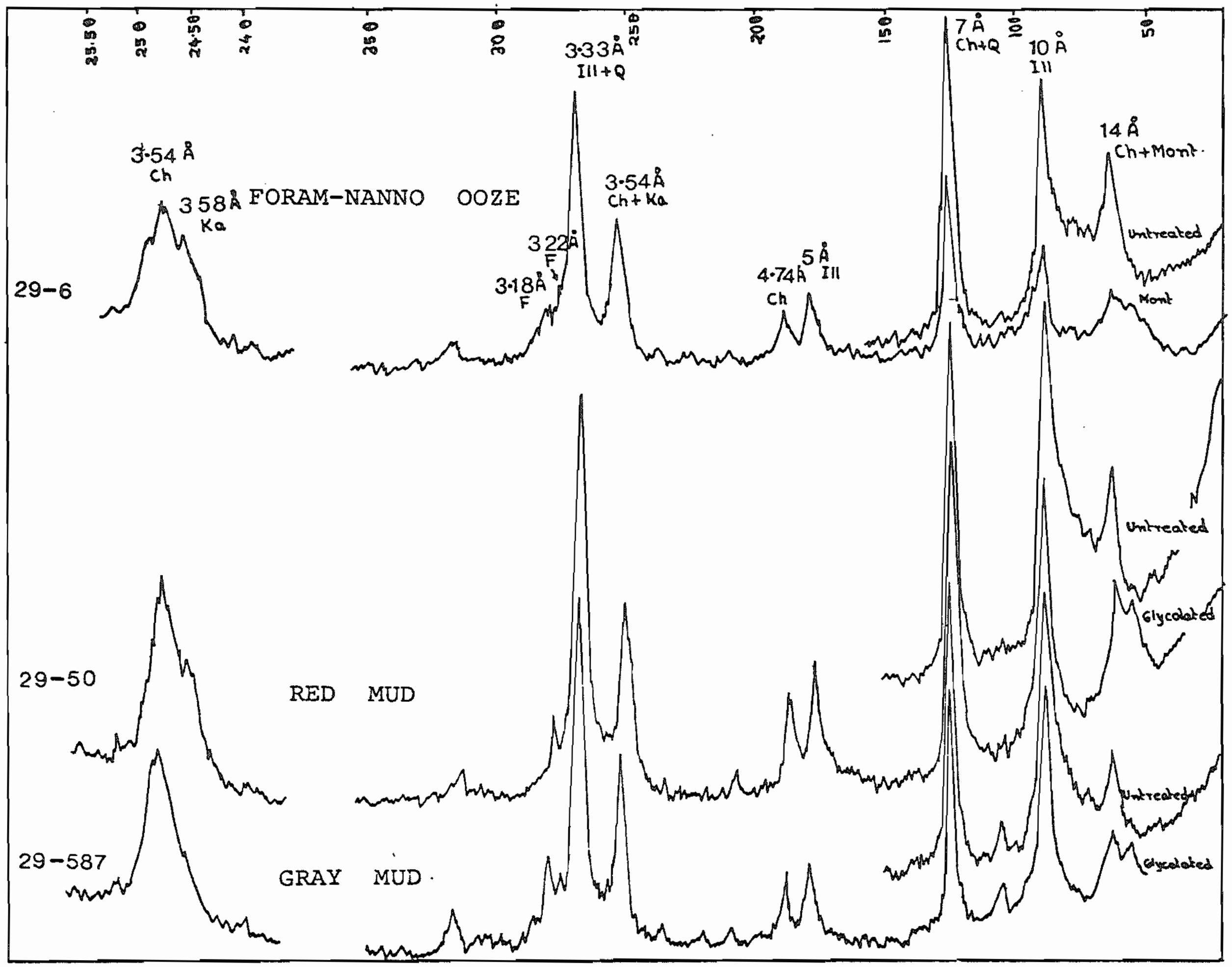

Figure 6: X-ray diffraction of oriented mounts of typical clay from foram-nanno ooze, gray mud and red mud. K. Kaolinite, Mnt. montmorillonite, Ill. illite, Ch. chlorite, Q. quartz, F. feldspar.

can be expected to be more or less related with depth in the core and with time. Since the cores are still uncompacted and the general lithology is fairly similar, it is unlikely that different types of pore fluid are mainly responsible for differential diagenesis. If diagenesis is related to time and depth, then the lower part of the cores should be more affected, although at such shallow depths the effect of burial diagenesis would be very low. Montmorillonite and mixed-layer clays are most affected during early diagenesis (Dunoyer de Segonzac, 1970). Our x-ray diffraction patterns do not show any change in mixed-layer clays that can be related to depth. Furthermore, there is no systematic change in the crystallinity of kaolinite and illite (Figure 6).

The second possibility is that the change is due to difference in the source area. It is almost impossible that early diagenesis could change the proportion of chlorite, illite and kaolinite in the cores. Variation of chlorite, illite and kaolinite in the red and gray mud, therefore, can be considered to be directly related to different source area. The presence of higher amounts of amphibole as well as feldspar and quartz in the olive gray mud also supports the idea of a different source.

It has been suggested that sources of kaolinite are lower Carboniferous and Triassic redbeds of the Maritime

Provinces and some offshore Mesozoic and Tertiary Strata (Allen and Johns, 1960; Barnett and Abbott, 1966; Stanley _et al._, 1972; and Piper and Slatt, 1976). Source of montmorillonite are Triassic redbeds of the Maritime Provinces (Stanley _et al._, 1972) and offshore Cretaceous and Cenozoic strata and weathering products of some basic and ultrabasic rocks (Piper and Slatt, 1976). Both montmorillonite and kaolinite are absent or rare in sediments from Labrador and Newfoundland and the adjacent inner shelf. In Newfoundland, only one kaolinite rich glacial drift sample has been found; it occurs on the Port-au-Port Peninsula, which is underlain by Paleozoic carbonates (Piper and Slatt, 1976). Montmorillonite and kaolinite in sediments from the outer continental shelf from Labrador to Nova Scotia are believed to be derived from underlying coastal plain strata (Piper and Slatt, 1976).

Similarity between the mineralogy of illite and kaolinite rich red mud and red till on mainland Nova Scotia indicates a possible genetic relationship. It is further supported by the distribution of the red till in the Maritime Provinces (Figure 9), and the red mud in the North Atlantic. Red till is found in Nova Scotia, New Brunswick and in the Gulf of St. Lawrence (Conolly _et al._, 1967) and the red mud in the North Atlantic is not found north of the Tail of the Bank. Cores from the Laurentian Fan and Laurentian Channel

(Conolly *et al*., 1967; Piper, 1975b) show that both the intensity of red colour and amount of red mud increases towards the Laurentian Channel, suggesting it is the possible source. Heezen and Hollister (1972) suggested that the major input of red clay in the Northwestern Atlantic is controlled by the Laurentian Channel, and red mud is carried further south by the Western Boundary Undercurrent.

Presumably most of the red mud came through the Laurentian Channel with some from Nova Scotia, either carried by glacial met water or later reworked from the till left on the shelf. The gradual change in the red colour towards the sea (Heezen and Hollister, 1972) and increase of montmorillonite (Piper and Slatt, 1976) possibly indicate that these clays have been mixed with the clay eroded from the Mesozoic and Cenozoic strata of the Scotian Shelf and Grand Banks and in deep water presumably also with clay brought by the Western Boundary Undercurrent. Contributions from erosion of Mesozoic and Cenozoic strata is believed to be low, because the Quaternary history of the Scotian margin (Stanley *et al*., 1972) and Grand Bank margin suggests a gradual progradation of the slope and rise rather than erosion.

The so-called gray "Newfoundland type till" rich in illite and locally with substantial chlorite is found throughout much of Newfoundland and Cape Breton Island and reflects a predominance of Precambrian and Lower Paleozoic igneous

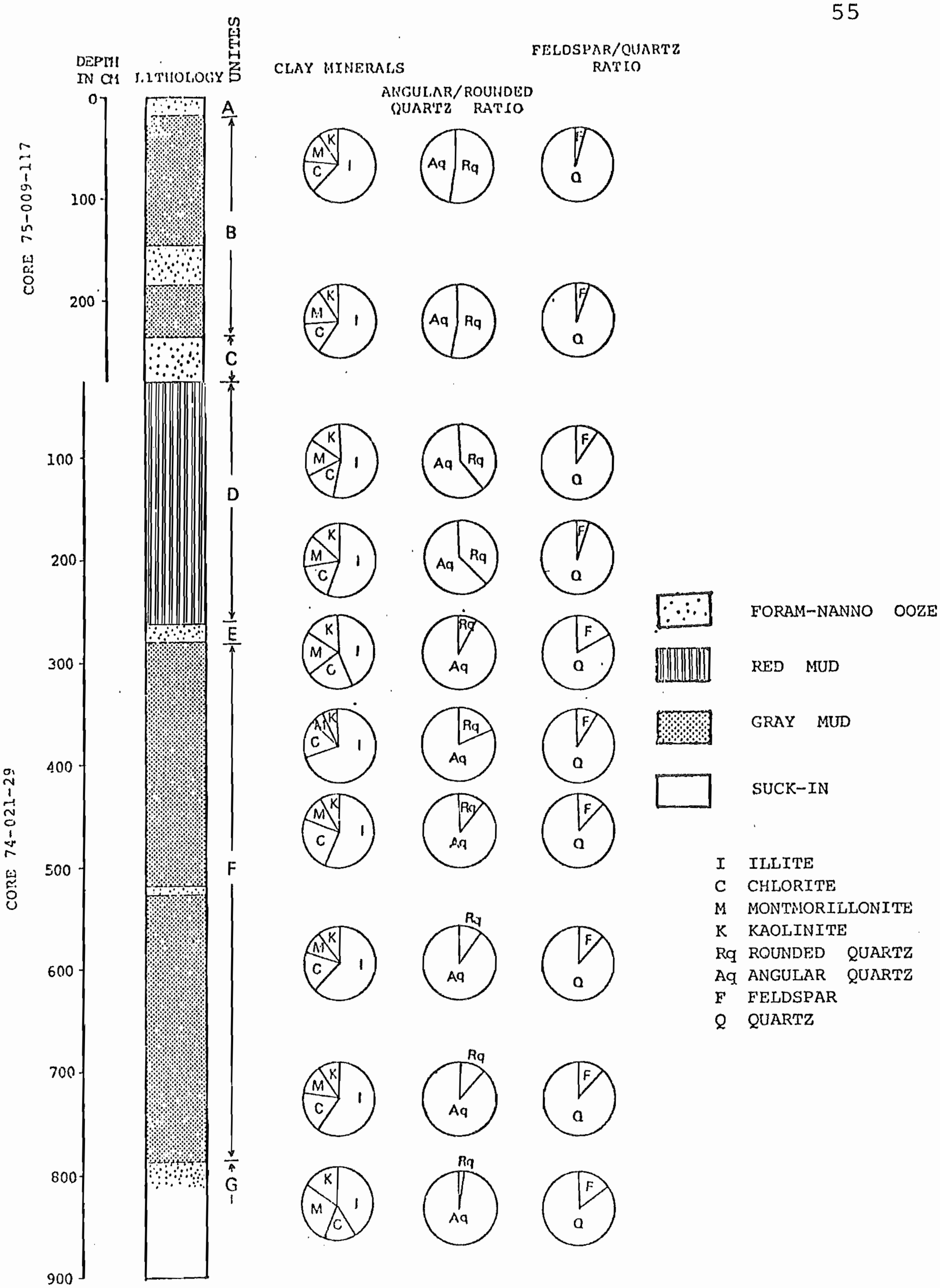

Figure 7: Distribution of clay minerals and round Quartz grains and Feldspar in the stratigraphic sequence.

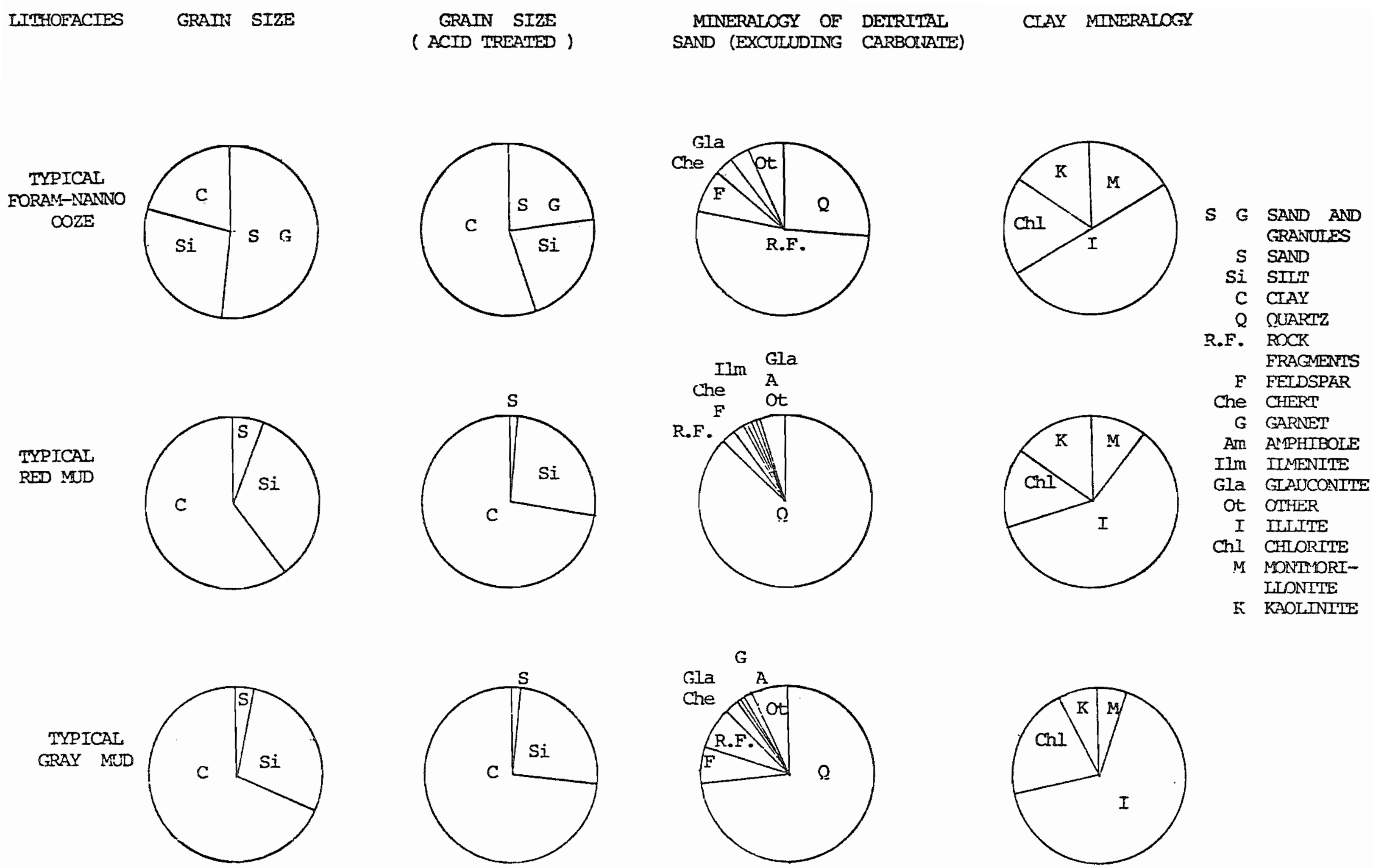

Figure 8: Grain size distribution, sand mineralogy and clay mineralogy of three major lithofacies, i.e. foram-nanno ooze, red mud and gray mud.

and sedimentary source rocks. The "Newfoundland type till" is also found on the coast of Labrador and Baffin Island, but shows a slight increase in the illite content (Piper and Slatt, 1976). The distribution of illite and chlorite rich gray till in northeastern Canada is presumably controlled by the advance and retreat of the ice sheets in Cape Breton, Newfoundland, Labrador and Baffin Island. The dissimilar chlorite content between Late Wisconsin sediments from Newfoundland and Labrador reflects a lack of Laurentide ice on the island at that time (Piper and Slatt, 1976).

Similarity in the clay mineralogy of gray mud on the South of Grand Banks and "Newfoundland type till" suggests the same origin. It is further confirmed by the silt mineralogy. However, the presence of montmorillonite and small amounts of kaolinite indicate possible mixing with the clay eroded from the Grand Banks Cenozoic strata, and possibly also with the clay brought by the Western Boundary Undercurrent. Distribution of this gray mud in the North Atlantic may have been influenced also by the surface circulation of the ocean. It is possible that the high kaolinite and montmorillonite content of the light gray clay associated with foram-nanno ooze is at least partly controlled by the Gulf Stream as suggested by Pastouret _et al_. (1975).

MINERALOGY OF SILT

Silt mineralogy was largely determined from smear slides. In some samples the silt size fraction was examined under the SEM. In others, microsieves were used to separate the 30 μ to 10 μ size fraction, which was then examined in grain mounts under a petrographic microscope. The predominant components of the silt size fraction throughout the cores are carbonate and quartz, followed by feldspar, amphibole, glauconite, hematite, ilmenite, rutile, mica, sphene, and opaques. Most of the silt size carbonate fraction is biogenic, mostly broken foram tests and a few coccoliths, and the percentage varies according to the total carbonate content of the sediments. The 10 to 30 μ size fraction separated by wet sieving, a process requiring ultrasonic treatment, was very low in carbonates, presumably because most of the carbonates were broken down by the ultrasonic treatment. Silt size limestone fragments are found, but are much less abundant than the biogenic silt carbonate. Quartz and feldspar are higher in the gray mud than in the red mud. In the foram-nanno horizons, this fraction is relatively low. Glauconite is present in both red and gray mud, but is commonest in the foram-nanno ooze. The amount of amphibole is higher in the gray mud than in the red mud and foram-nanno ooze. The percentage of hematite, ilmenite and rutile is highest in the red mud.

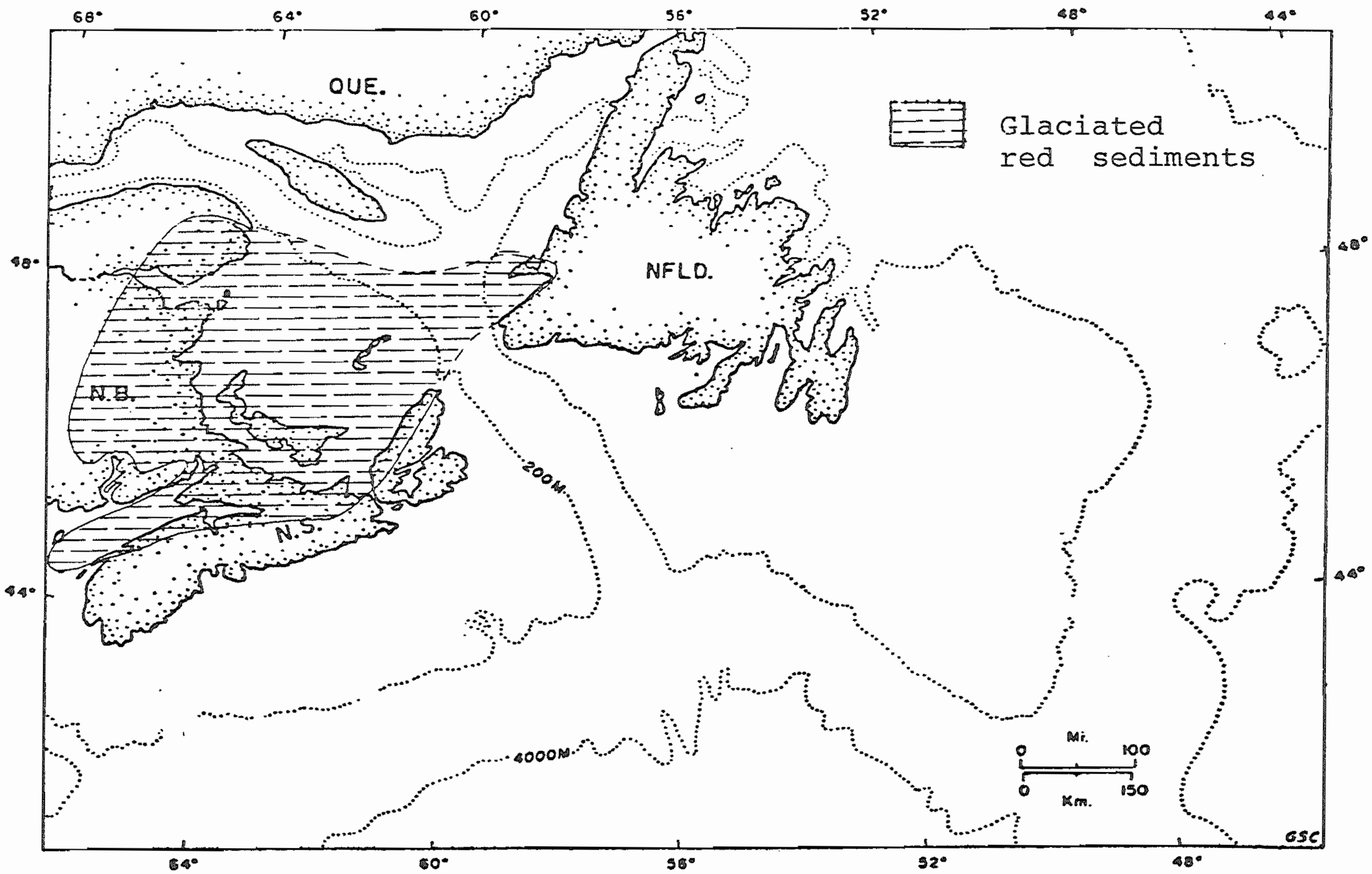

Figure 9: Distribution of glaciated red sedimentary rocks in Maritime Provinces (after Heezen and Hollister, 1972).

MINERALOGY OF SAND

The muds contain very small amounts of sand, in general not more than 1 per cent, including the biogenic component of foraminifera. As a result very large samples were needed to make a mineralogical or textural study of the sand fraction. The sand fraction was first treated with hydrochloric acid, to remove the biogenic component, which unfortunately also removes the detrital carbonate. Grain mounts and thin sections were prepared, and the relative frequency of different minerals was determined by counting 300 to 350 grains. Sometimes because of the very small quantity of sand, point counting was not possible and instead the number of grains per unit area was determined. Separation of heavy minerals was not possible because of very small quantity of sample available.

Two remarkable changes in the mineralogy of the sand fraction were observed. First, the ratio of rounded to angular quartz grains in the mud increases towards the tops of the cores. Second, feldspar becomes less common and more weathered towards the tops of the cores (Figure 7). Both suggest that the sand in the muds in the upper part of the cores are more mature than in the lower part. This may be due to either a greater distance of transport from the source area, or a longer period of reworking on the continental shelf before final deposition on the seamounts.

Surface textures of the quartz grains are variable. The majority of them are angular and appear to be freshly broken, with conchoidal fracture and parallel striations and groove marking (Plate 1) which are very characteristic of glacial environments (Krinsley and Doornkamp, 1973). These type of grains are quite common in the gray mud of unit F and in the foram-nanno ooze. The second type of grain is subangular with pitting and solution marks, and sometimes with precipitation, possibly suggesting slight reworking before final deposition. This type of quartz grain is fairly common throughout the cores. The third varity is almost perfectly rounded, sometimes with conchoidal fractures, indicating impact in a high energy zone (Plate 2), possibly on a beach. This type of grain is present in the gray clay of unit B, and sometimes also in the red mud of unit D.

Among the feldspars, plagioclase is dominant, but microcline and orthoclase are also common and some of the feldspar shows perthitic intergrowths. There is no direct correlation between mineralogy of sand size feldspar and the type of clay, but in general feldspars in the gray clay of unit F and in the foram-nanno horizons are comparatively fresh.

PLATE 1

Scanning electron microscope photographs of quartz grains exhibiting features that were produced in the source area.

A-B. Quartz grain with conchoidal breakage and parallel ridges, indicating glacial origin. Note the lack of weathering and diagenesis. This type of grain is very common in the lower gray mud (unit F) and in foram-nanno ooze.

C-D. Slightly reworked glacial quartz grains with conchoidal breakage and solution and precipitation features. Adhering particles in photo C are possibly calcite.

<u>NOTE:</u> The black bar at the bottom of each photograph is the scale. In A, B, C this is 20 μ, in D it is 40 μ.

PLATE 2

A-B. Well rounded grain with conchoidal breakage and V-shaped depression possibly indicating a beach environment. This type of grain is fairly common in the upper gray mud (unit B).

C-D. Clastic biogenic carbonate, shell fragments with sponge burrowing.

<u>Scale</u>: A = 200 μ; B and C = 40 μ; D = 4 μ

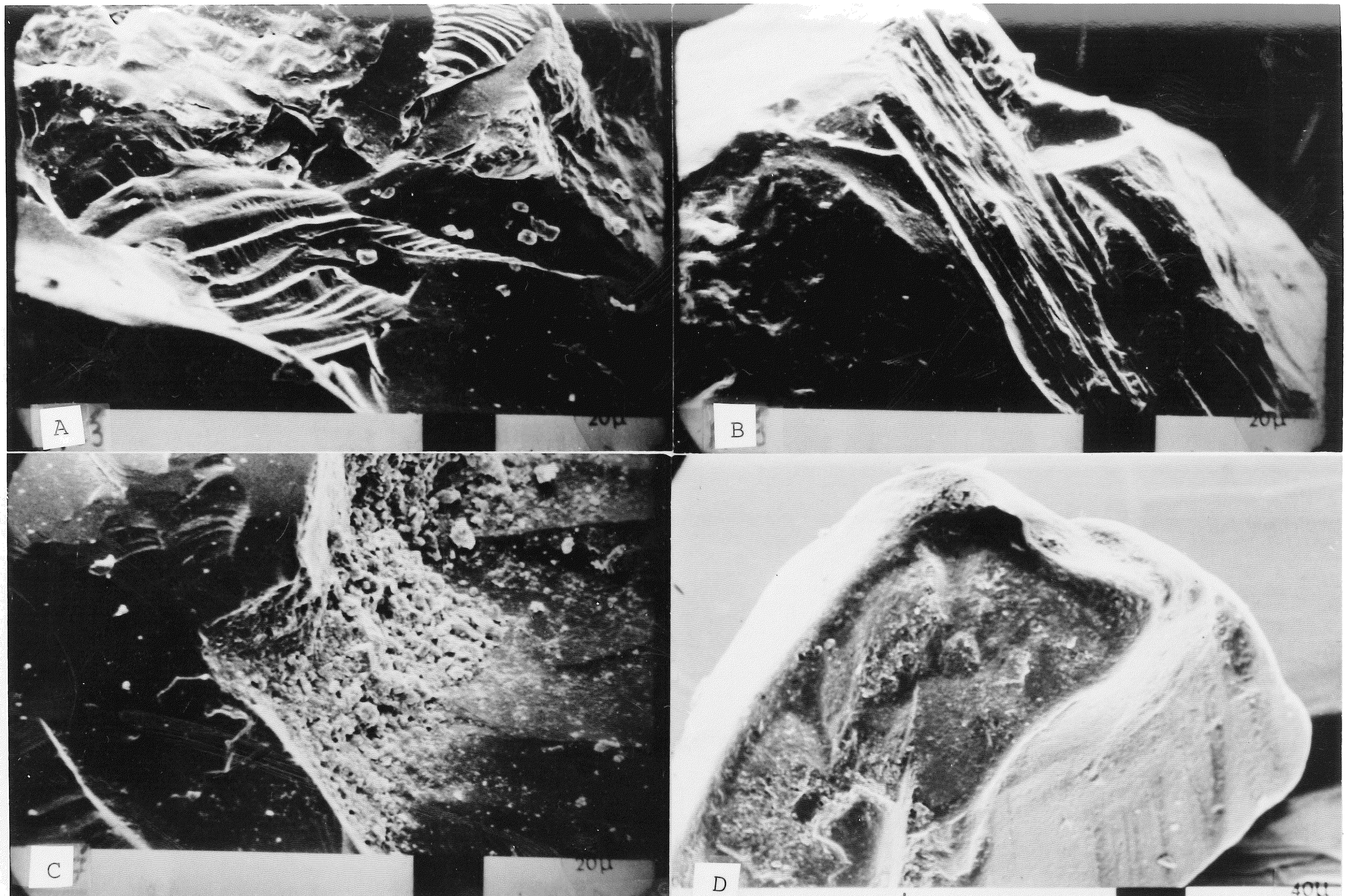

PLATE 1

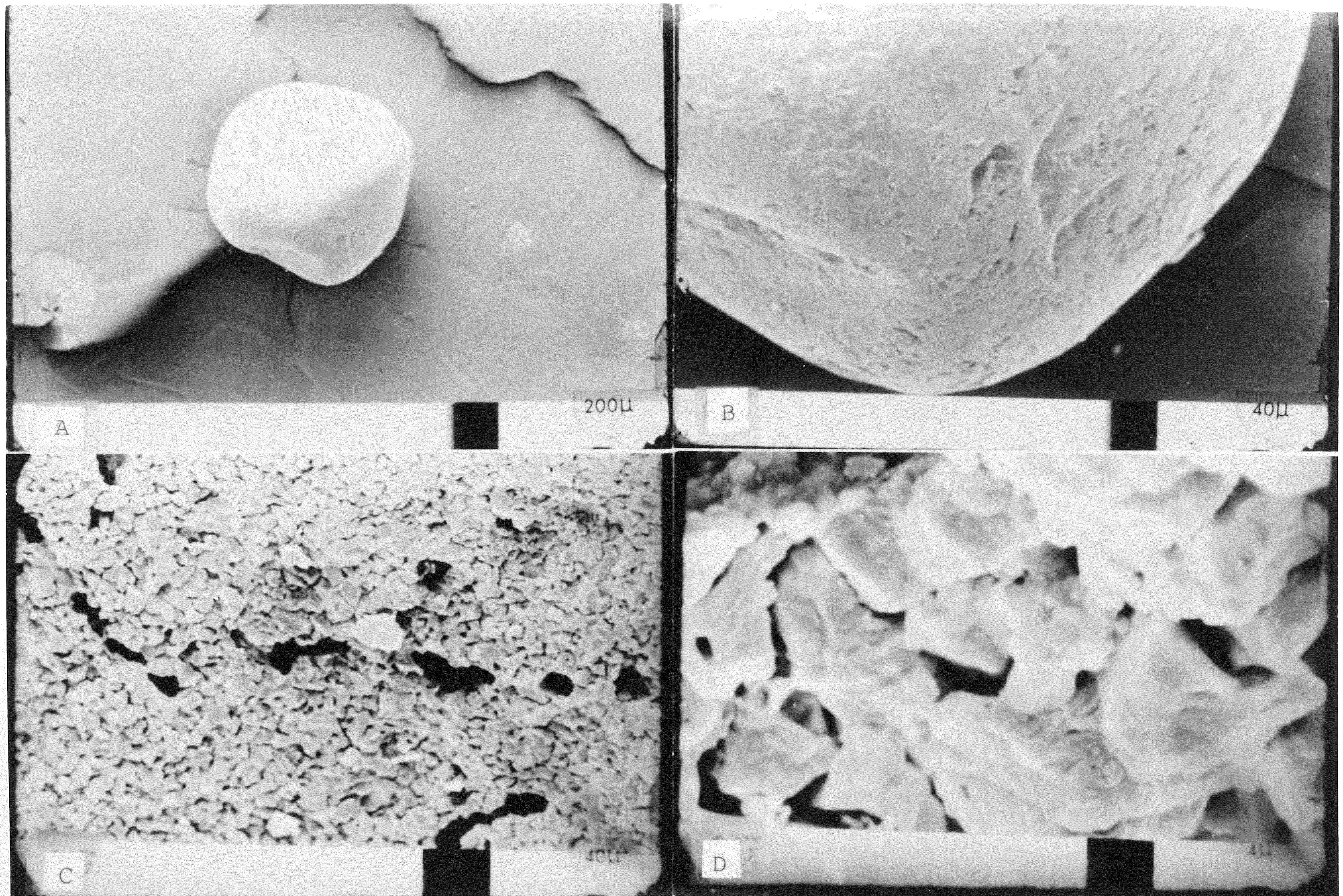

PLATE 2

Small amounts of sand size detrital carbonate are present throughout the cores, the percentage is slightly higher in the foram-nanno ooze than in the muds. The carbonate clasts can be divided into two different types. The first type consists of bio-clasts, possibly broken bivalve shell fragments with some sponge burrowing (Plate 2) and is present both in the mud and in foram-nanno ooze. The second type of detrital carbonate is limestone fragments which are generally common in the foram-nanno ooze (Plate 3).

Amphiboles are mostly actinolitic hornblende, commonest in the gray mud of unit F. Small amounts of glauconite are present throughout the core. Trace amounts of biotite, ilmenite, muscovite, sphene, rutile, garnet and chert are also present in the sand fraction (Figure 8).

The sand on the top of these seamounts might have been transported either by icebergs or in suspension following storms on the Grand Banks. The fine size and uniform distribution of the sand in the mud and the absence of pebbles suggests suspension transport. The sands on the Grand Banks consists mostly of quartz, feldspar, shell fragments, glauconite, and hornblende (Slatt, 1973; Slatt *et al.*, 1972 and personal communication, 1976 (see appendix)). The presence of glauconite and shell fragments in the sand fraction of the muds suggests that the Grand Banks is the major source

PLATE 3

A-B. Detrital carbonate, micritic limestone.

C-D. Detrital carbonate, fragment of chalk.

Scale: A = 100 μ ; B = 4 μ ; C = 100 μ ; 4 μ

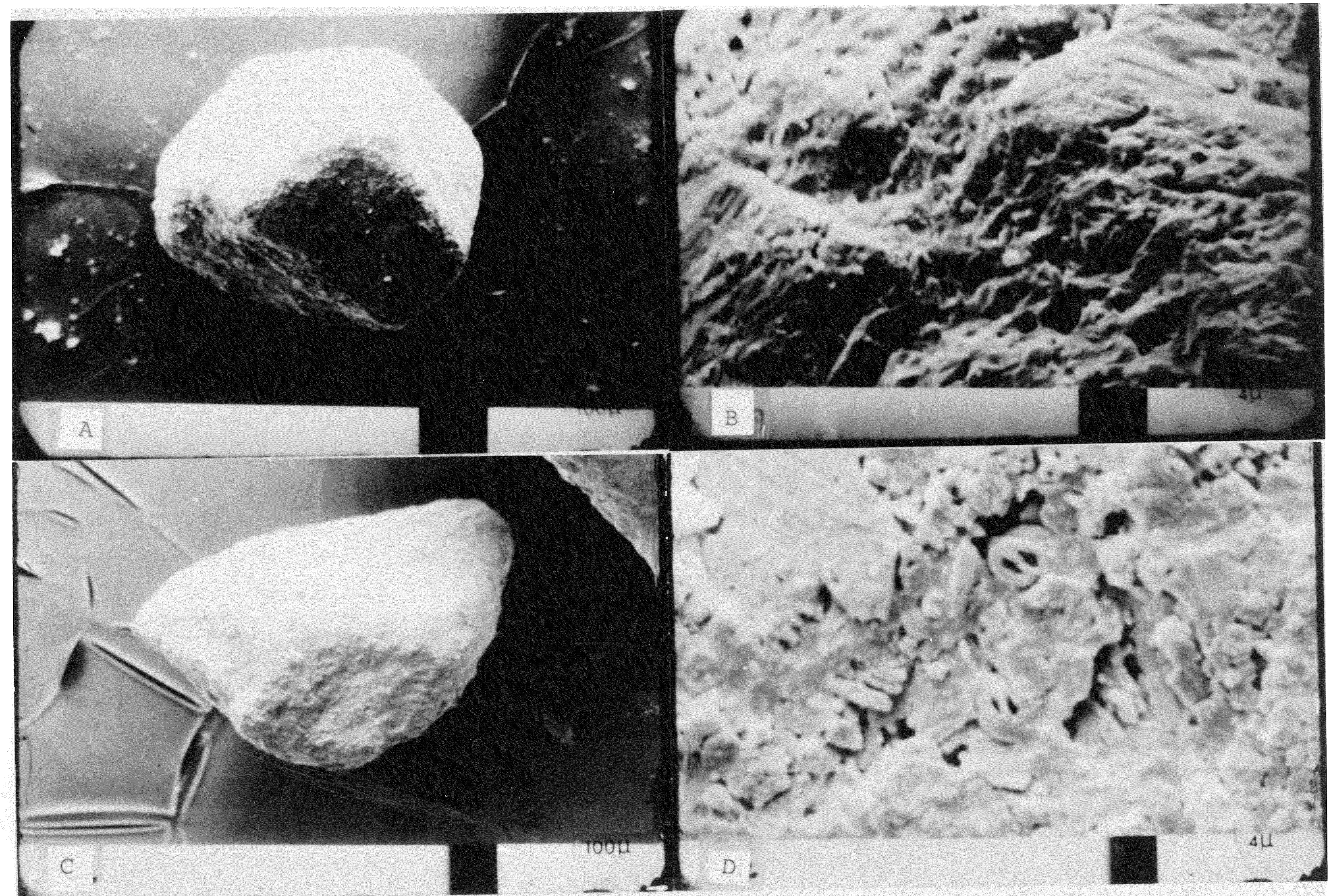

PLATE 3

of sand. Small amount of sand from other sources were possibly deposited from icebergs. Since the amount of such sand in the mud is very low and it has been diluted by the Grand Banks sand, it is difficult to suggest a possible source. Red siltstone fragments and ilmenite in some horizons in the red mud possibly indicate a Laurentian Channel ice source.

Loring and Nota (1973) and Piper (1975a) suggested subrounded rather etched garnet and zircon as an indicator of Laurentian Channel ice, but they are possibly not good indicators because the Newfoundland basin contains considerable amounts of garnet. The sand on the Grand Banks contains both garnet and zircon (McMullen, 1966).

Coarse and freshly broken quartz and basic rock fragments in the foram-nanno ooze are obviously ice rafted. Their occurrence with ice rafted pebbles also supports their ice rafted origin. Pebble petrology suggests Labrador Sea icebergs. Large shell fragment indicate that sea ice from the Grand Banks possibly contributed small amounts of sand during the Pleistocene lower sea levels.

PETROLOGY OF PEBBLES

Pebbles are mostly concentrated in the foram-nanno ooze horizons. They are very rare in the gray muds, but slightly more common in the red mud. Most of these pebbles have sharp angular edges, with almost no evidence of reworking. Some of them show parallel striations. A few pebbles are almost rounded, suggesting a reworked origin.

The petrology of these pebbles is quite variable. Generally the pebbles in the foram-nanno ooze are of metasedimentary and basic igneous rocks but some are granitic and limestones. Pebbles in the gray muds have a similar lithology.

The lithology of the pebbles in the red mud is slightly different. They consist mostly of granite, red siltstones and sandstones and limestones, as well as small amounts of metasedimentary and basic igneous rocks. It is obvious that these pebbles on the tops of seamounts are ice rafted but probably they have a variety of source areas.

Auffret (1968) suggested that the petrology of the pebbles and gravels on the Northern Grand Banks, Northeast Newfoundland shelf and Labrador shelf shows certain variations. Towards the north in the Labrador shelf and the Northeast Newfoundland shelf they are mostly of metamorphic and basic igneous rocks, but towards the south they are mostly of sandstone and quartzites. He could not separate ice rafted pebbles from relict or reworked pebbles derived from till or outwash. The pebble types dominant on the

Labrador shelf indicate that the Labrador sea icebergs appear to contain mostly metamorphic and igneous rocks with small amounts of Lower Paleozoic Limestone. A suit of pebbles dredged from the top of a seamount in the Newfoundland Basin at latitude 43°41.6'N and longitude 45°25.6'N was examined (see appendix). These pebbles, being from a seamount, are obviously ice rated. They are dominated by basic igneous and metamorphic rocks with some granitic rocks and limestone and small amounts of quartzite and sandstones.

Pebbles of the Laurentian Channel and Fan are dominated by relatively unmetamorphosed sandstones and crystalline rocks (mostly granitic) from the Canadian Shield (Conolly et al., 1967; Loring and Nota, 1975; Piper, 1975a). This indicates that the Laurentian Channel icebergs should be relatively rich in sedimentary rocks and possibly also in granitic rocks. Piper (1975a) suggested that the clasts of red siltstone and sandstone were transported by Laurentian Channel ice. He also suggested the possibility of simultaneous sediment supply from more than one source; a mixing of Laurentian Channel ice rafted pebbles with the Labrador sea ice rafted pebbles is very likely.

Pebbles of sedimentary rocks, especially red sandstones and siltstones, and granitic rocks associated with the red mud appear to come from Laurentian Channel ice. The pebbles of metamorphic and basic igneous rock may have come either from Northeastern Canada or from Greenland. Limestone was common both in the Labrador Sea and Laurentian Channel icebergs.

CARBONATE CONTENT

Carbonate content of the sediments was determined at 10 cm intervals throughout all the cores. Sometimes, when necessary samples were taken at closer intervals, bult dried samples were used to determine the carbonate content. Most of the analyses were done by volumetric method described by Black et al. (1965), but the analyses were repeated on a Leco Automatic Carbon Dioxide Titrator to check the accuracy. Amount of detrital carbonate was determined from smear slides.

Carbonate content of the sediments is shown plotted against the depth in the cores (Figures 16, 19, 21). The percentage of $CaCO_3$ in the sediment ranges from 11% to 77%, being lowest in mud and highest in foram-nanno ooze. The generalized curve shows eight carbonate highs. Three prominent carbonate maxima occur in units A, C and G. Among the remaining five carbonate maxima, two intermediate peaks (Figure 27) appear in the unit E and subunit B6. The amplitude of the remaining peaks is much less, the lowest being the one appearing in subunit F2.

CARBONATE DISSOLUTION

In order to make any meaningful interpretation of either the carbonate cycles or foraminiferal assemblages and their climatic significance it is essential to consider the effect of dissolution on the sediments. Selective solution can change the apparent latitudinal range of an assemblage (Berger 1968) as well as the total input of carbonate (Arrhenius, 1952). Most of the studies dealing with dissolution use the weight ratio of broken to unbroken foram tests to measure the intensity as well as the change in the degree of dissolution (Berger, 1970, 1972). Generally one gram of foram tests is used as a standard sample. The technique was found to be inconvenient to apply to the cores studied. Firstly, because the continental margin has a high detrital input and it is difficult to separate the broken tests from the other clastic particles. Secondly, the foram number in the mud is so low that it is alomst impossible to get one gram of foram tests.

Dissolution effects were measured on the tests of *Globigerina pachyderma** and *Globigerina bulloides*, because they are the most common species both in the mud and in foram-nanno ooze. The method was similar to that described by Bé *et al*. (1975). Neither hydrogen peroxide nor ultra-

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* The species has recently been described as Neogloboquadrina pachyderma, by Srinivasan, M. S. and Kennett, J. P. 1974, Secondary classification of the planktonic foraminifera Neogloboquadrina pachyderma as a climatic index, Science, vol. 186, p. 630-632.

sonic treatment were given to the samples used in the dissolution study. The extent of dissolution depends on how long the particle was in suspension before being finally deposited and how long it was exposed to the sediment-water interface, before being finally covered by other sediments. To balance the first effect approximately the same size particles were used for measurements. For G. pachyderma the 100μ to 150μ and for G. Bulloides the 300μ to 400μ size fraction were used. To check the second factor, differential solution on two opposite sides of the test was looked for, becasue the side facing the sediment-water interface will show more dissolution than the side facing the sediments.

At least 10 tests of each of the species were taken from different levels in the cores. In general, G. pachyderma has three well developed layers in the test wall, i.e. a microgranular layer, a subrhombic layer and a euhedral layer, but G. bulloides shows only two layers, a microgranular layer and a subrhombic layer (Plate 4). However, the outer layer of G. pachyderma may be either granulate or crystalline. In order to quantify the extent of dissolution, the amount of material removed from the different layers was determined. The series of photographs taken suggest that dissolution is very low, generally not more than 15% to 20% and no significant change in the extent of dissolution was observed between the muds and the foram-nanno ooze (Plates 5, 6, 7, 8).

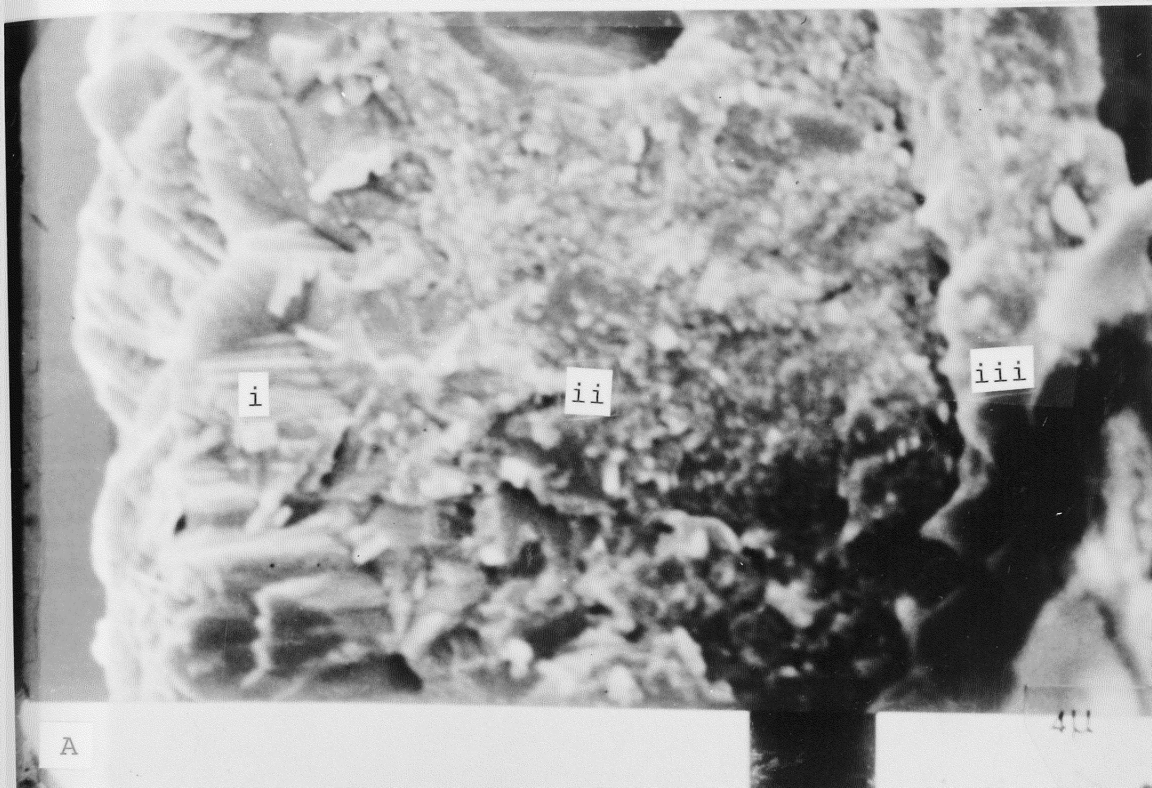
PLATE 4

A. Cross-sectional view of fully developed wall of Globi-gerina pachyderma. i) Euhedral layer, ii) Subrohombic layer, iii) Microgranular layer. Note fairly distinct boundary between each layer.

B. Wall structure of Globigerina bulloides. i) Subrhombic layer, ii) Microgranular layer. The boundary between these two layers is gradational and not distinct.

Scale: A and B = 4 μ

PLATE 4



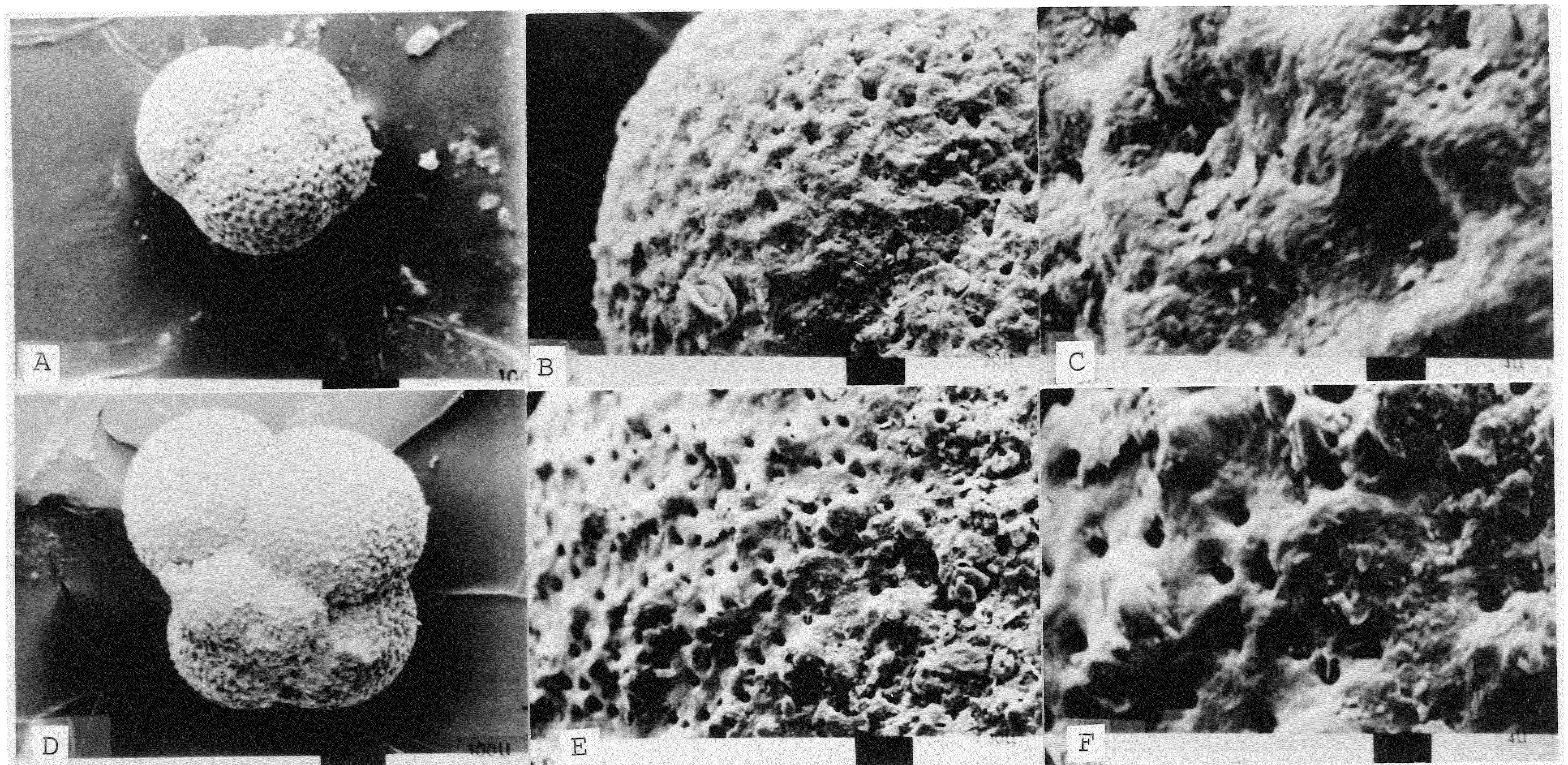


PLATE 5: Dissolution of foraminiferal tests, sample from the top of the core 74-021-29 depth 0 cm (unit C). A-C. effect of Globigerina pachyderma tests, note the outer layer partly removed by dissolution. D-F. Globigerina bulloides test, etched subrhombi layer. Total dissolution approximately about 20%.

Scale: A and D = 100 μ, B = 20 μ, C and F = 4 μ, E = 10 μ

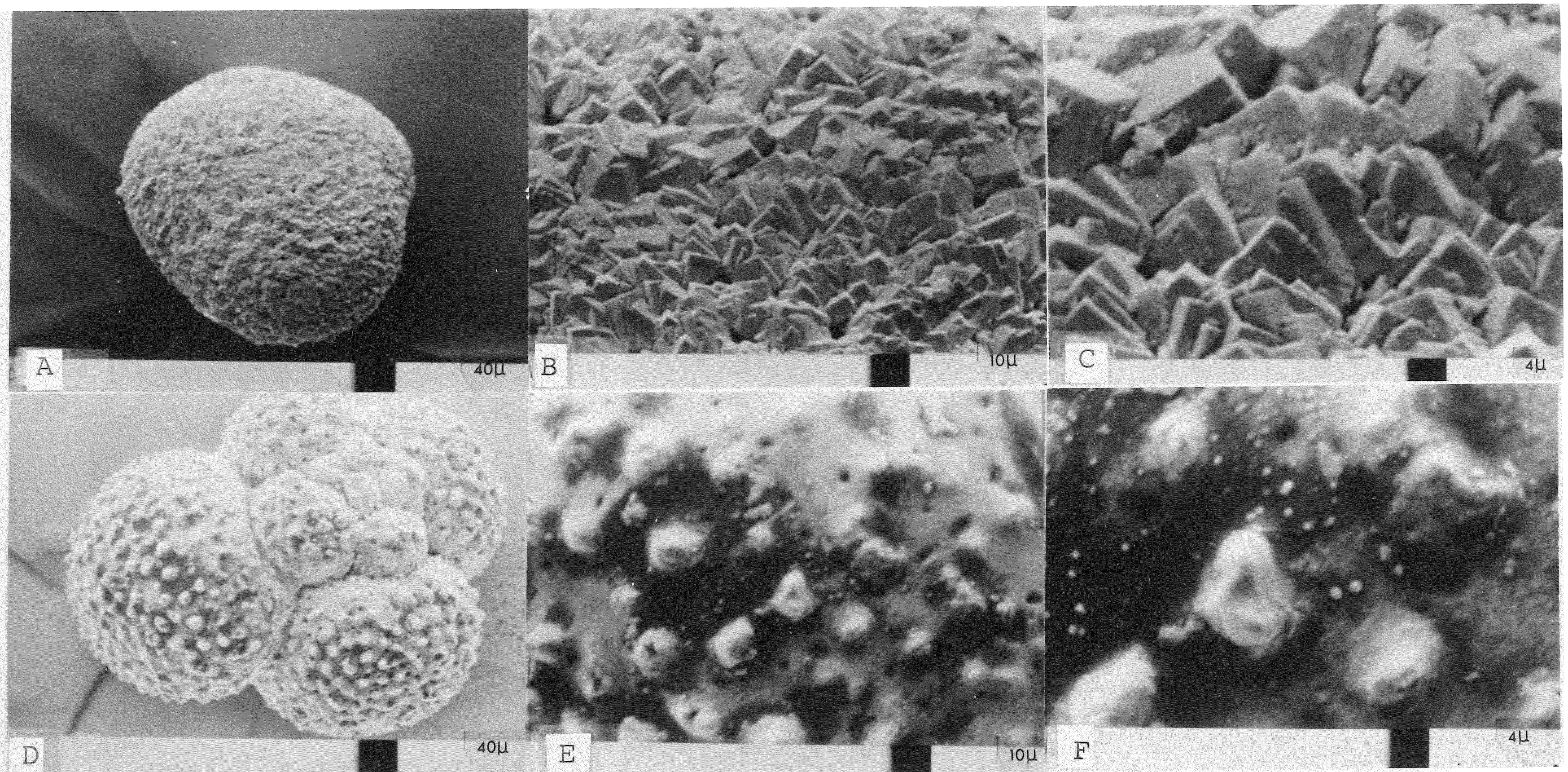


PLATE 6: Effect of carbonate dissolution on foraminiferal tests from red mud (core 74-021-29, depth 50 cm, (unit D). A-D. Globigerina pachyderma test, note very little affect of dissolution. D-F. Globigerina bulloides tests, subrhombic layer only partially effected. Total dissolution less than 10%.

Scale: A and D = 40 μ, B and E = 10 μ; C and F = 4 μ

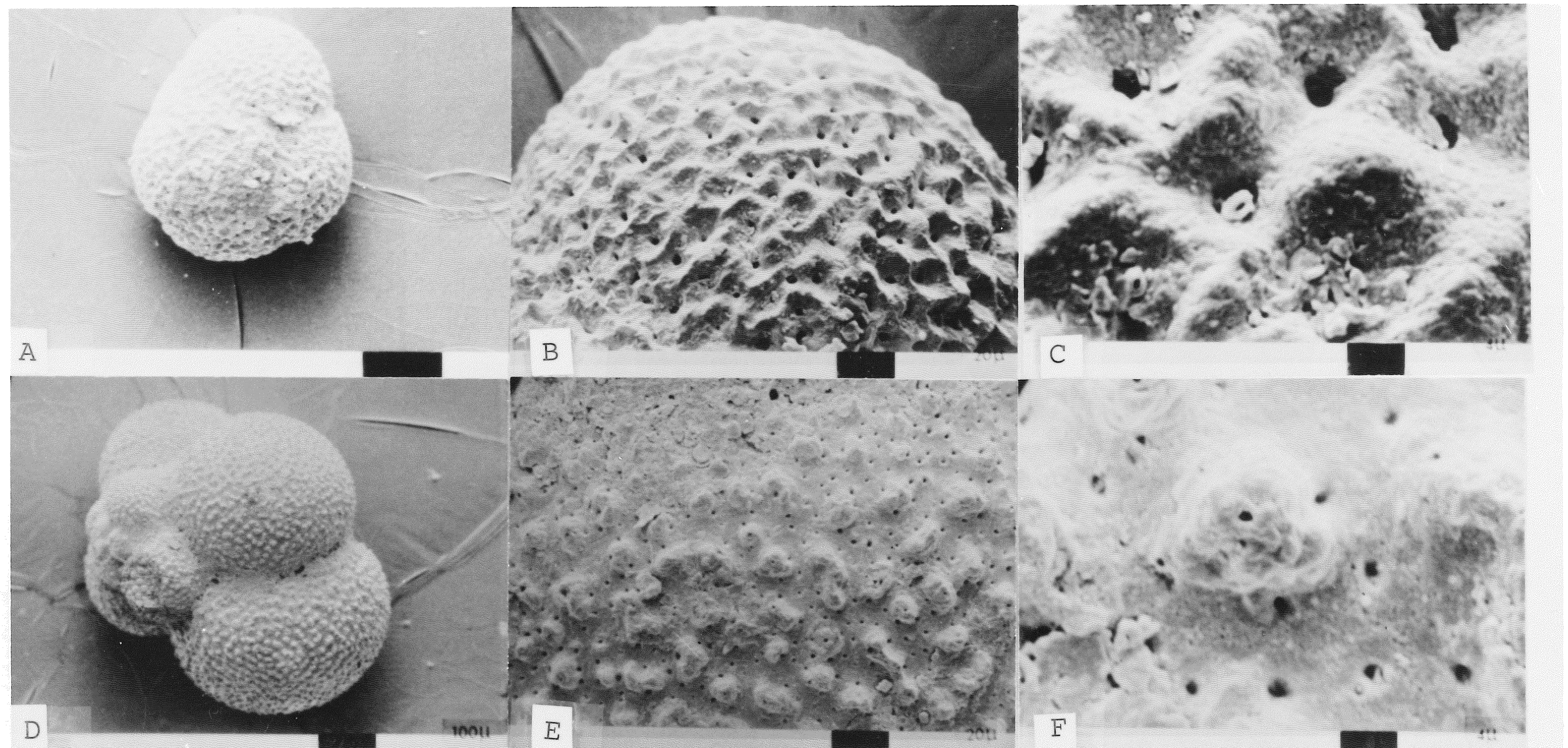


PLATE 7: Foraminiferal tests from unit G (core 74-021-29 depth 800 cm). A-C. *Globigerina pachyderma*, note the outer layer slightly affected by dissolution. D-F. *Globigerina bulloides* with etched subrhombic layer. Total carbonate dissolution about 20%.

Scale: A = 40 μ, B and E = 20 μ; C and F = 4 μ; D = 100 μ

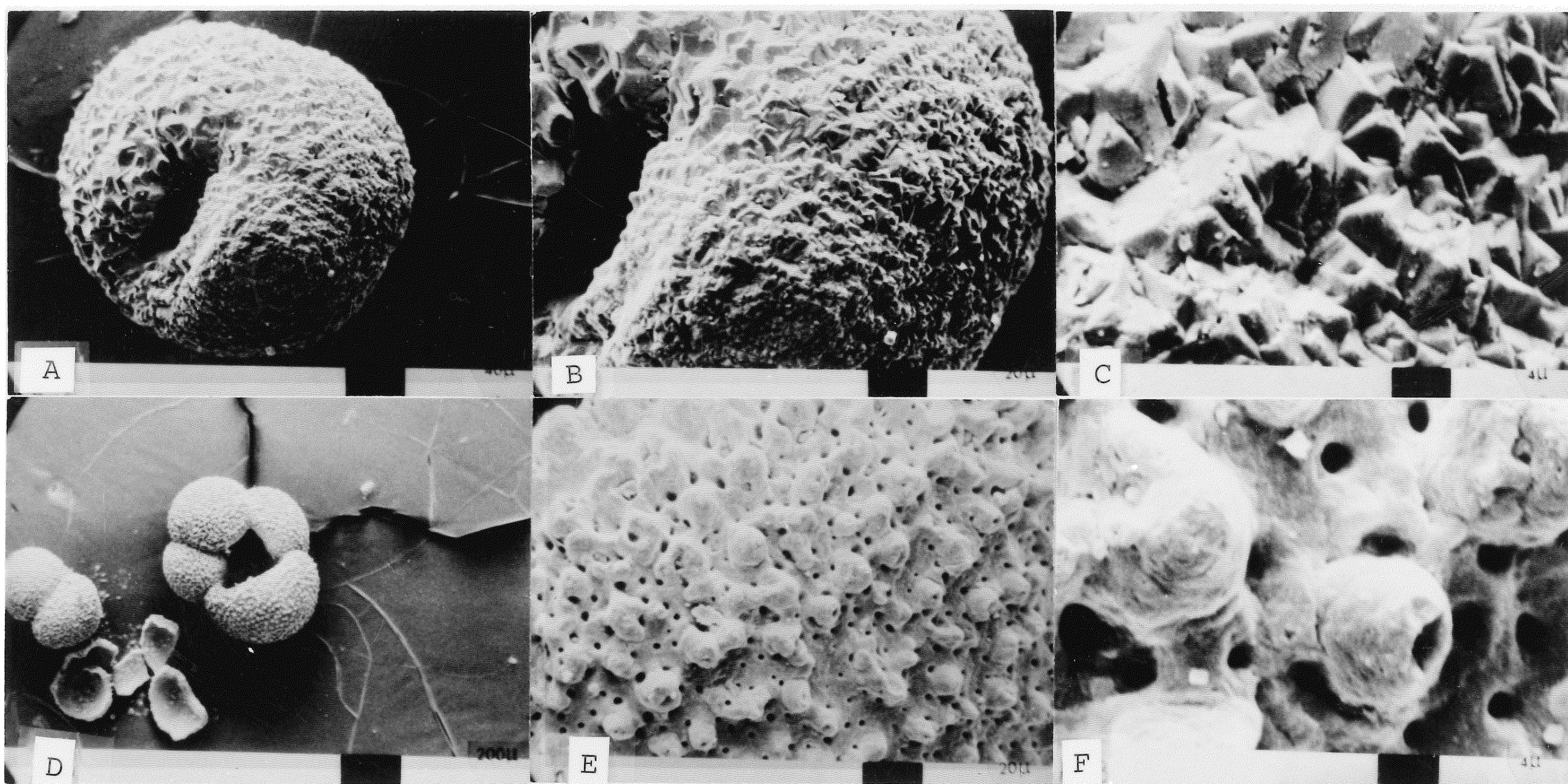


PLATE 8: Dissolution of foraminiferal tests from the lower gray mud (core 74-021-29, depth 540, subunit F3). A-C. *Globigerina pachyderma* tests only partially affected by dissolution. D-F. *Globigerina bulloides* tests, have undergone slight dissolution. Total dissolution approximately about 10%.

Scale: A = 40 µ ; B and E = 20 µ , C and F = 4 µ , D = 200 µ

CHAPTER IV

SEDIMENTOLOGY

GRAIN SIZE DISTRIBUTION

Grain size distribution of samples from all the cores was determined by standard pipette and sieve analysis techniques, with determination at 1 phi intervals. Since the predominant component of the sand and silt size fractions is biogenic carbonate, some of the samples were treated with hydrochloric acid before the size analysis and the results were compared with the untreated samples.

The results are presented in three different ways (1) The percentages of sand plus gravel, silt and clay are plotted against the sample position within the cores (Figure 10). These percentages are also plotted for all the samples on a ternary sand-silt-clay diagram (Figure 11). (2) Typical size analyses of untreated samples are plotted as cumulative curves to show grain-size distribution on logarithmic probability paper (Figure 12). (3) The ratio of silt to clay is plotted against depth in the cores to show the change in the fine fraction independent of the coarse fraction (Figure 13).

Some of the analyses were repeated to check the accuracy of analytical techniques. Duplicate analyses show good

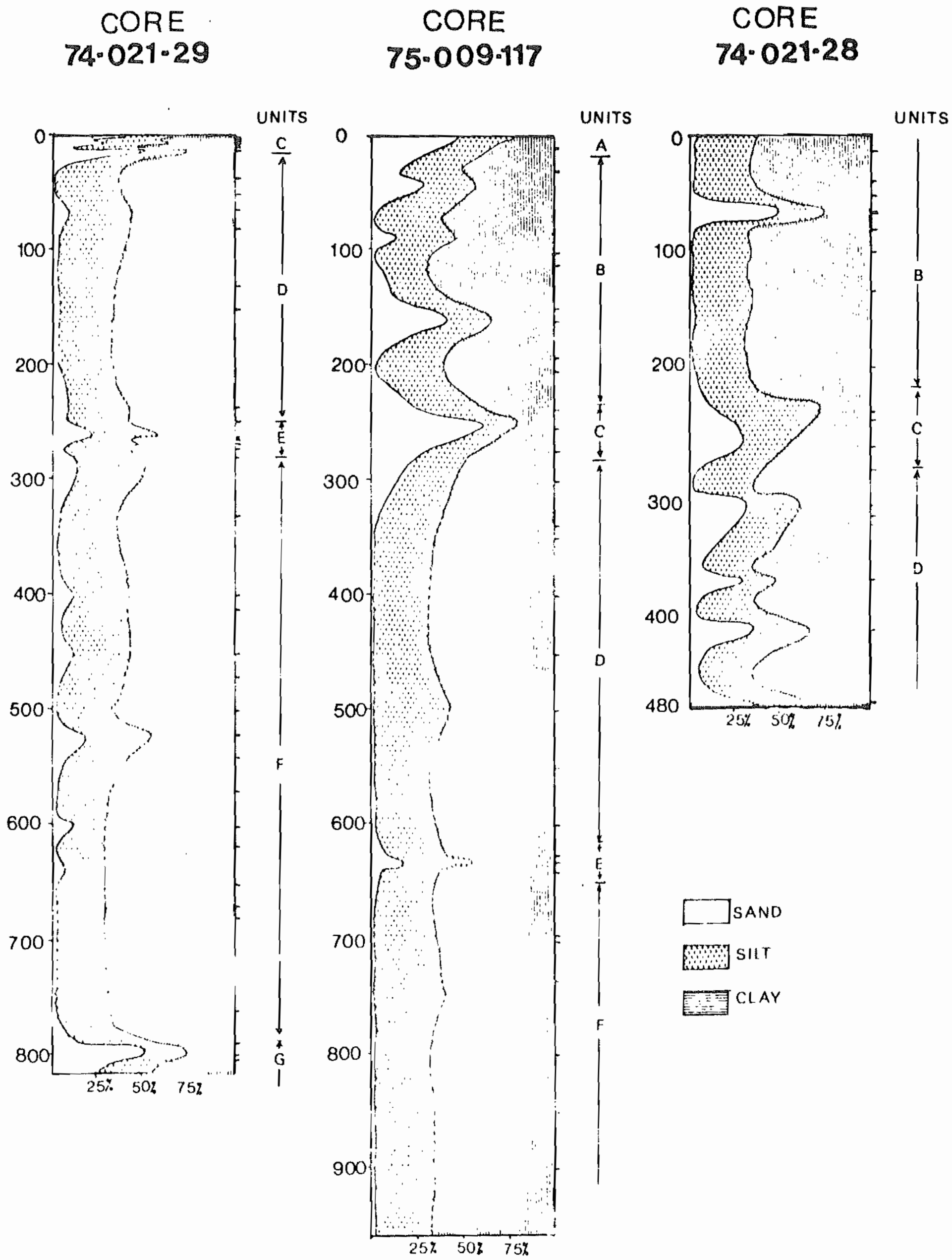


Figure 10: Distribution of sand, silt and clay in cores 74-021-29, 74-021-28, and 75-009-117. Smear slides have been used to interpolate between analyses.

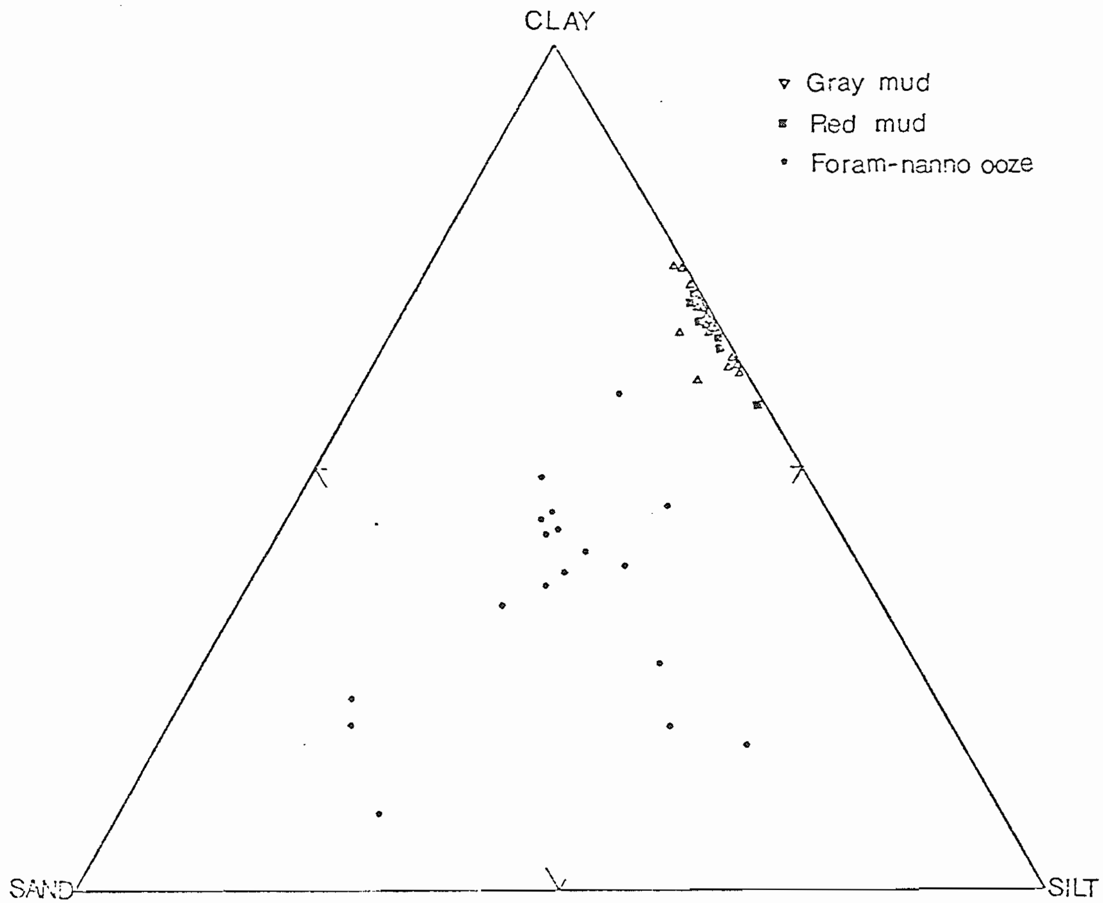


Figure 11: Size distribution of typical foram-nanno ooze red mud and gray mud.

reproducibility.

In general both the red and gray mud contain very small amounts of sand and coarse silt. The sand fraction is much greater in the foram-nanno ooze. The red and gray mud have an almost homogenous distribution of the sand, silt and clay. No apparent rhythmic or cyclic patterns which could be correlated with sedimentation processes were detected. It seems that the grain size distribution in the cores reflects the first order effect of climatic changes, and masks any second order effect which may be related with slight changes in the processes of sedimentation or in the source area.

Acid treated samples show sharp increases in the clay content, suggesting that the silt and sand component of the samples contain large amounts of carbonate (Figure 8). The typical red mud contains slightly higher amounts of sand and silt than the gray mud, but in the acid treated samples they do not show any difference. Silt and clay ratio increases in the foram-nanno ooze.

Water content of the cores is shown in Figure 14. In general water content gradually decreases with the depth in the core.

Pebbles

Pebbles are present mostly in the foram-nanno ooze, occasionally in the red mud and very rarely in the gray mud. The size and shape of these pebbles are quite variable; the maximum size observed was 2.5 cm. Their shapes vary from near spherical to almost blade-shaped. Pebble concentration is generally higher near the base of beds of foram-nanno ooze, and gradually decreases towards the top of the bed. In mud beds the pebble distributions do not show any systematic pattern.

PRIMARY SEDIMENTARY STRUCTURES

Sedimentary structures were studied on the split face of the cores, in x-radiographs and in thin sections. The clay is almost structureless, except for a few silt laminae, some faint bands (Plate 9,10) and a few mottles. The foram-nanno ooze shows only ice rafted pebbles and a few mottles.

Silt Laminae

Silt laminae are present in both red and gray mud, but are generally more common in the gray mud. In general these silt laminae can be divided into two major types on the basis of their mode of occurrence. Type A silt laminae are fine and occur in groups. Generally they are less than 1 mm thick but alternate with about 2 mm thick silt laminae. Type B consists of solitary silt laminae. Most of these are lenticular, having a maximum thickness of about 2 mm.

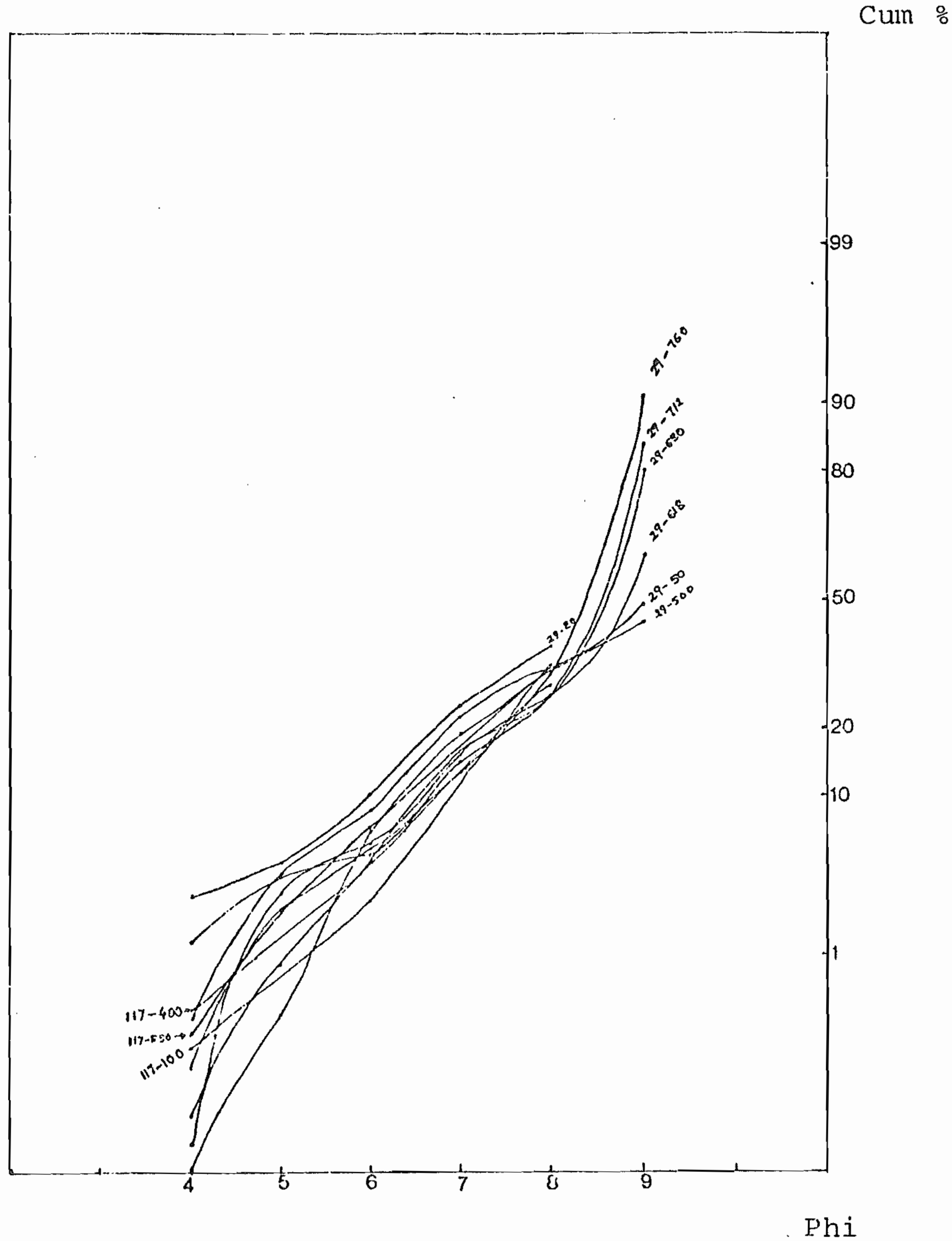


Figure 12: Cumulative curves of typical red mud and gray mud (not acid treated).

Type A silt laminae are common in the middle part of the cores, especially immediately above and below unit E. This type of silt laminae is present also in the gray mud of unit B, but they are visible only in x-radiographs (Plate 9). Most of these laminae have sharp bases and tops, but about 15% of them have a gradational base and sharp top and about 25% of them have a sharp base and gradational top. The thickness varies from 0.2 mm to 1 mm, but most of them fall in the lower part of this range. The majority of these silt laminae are fairly continuous and maintain a uniform thickness, but some of them are quite discontinuous and lenticular. They consist of mostly medium size (6 to 7ϕ) silt particles and are fairly well sorted and clean. Mineralogically the composition of these silt laminae is similar to the rest of the silt size particles in the clay, with the dominant components being carbonate and quartz with small amounts of feldspar and mafics.

Type B silt laminae are present mostly in the lower and middle part of the core (unit F) and are very rare in the upper part. This type of silt laminae may be either fairly continuous or in a well developed lenticular form. The thickness varies from 1.5 mm to 2.5 mm. Mineralogy and the contact nature of these laminae are same as the type A.

Bands

There are very faint bands in the apparently homogenous

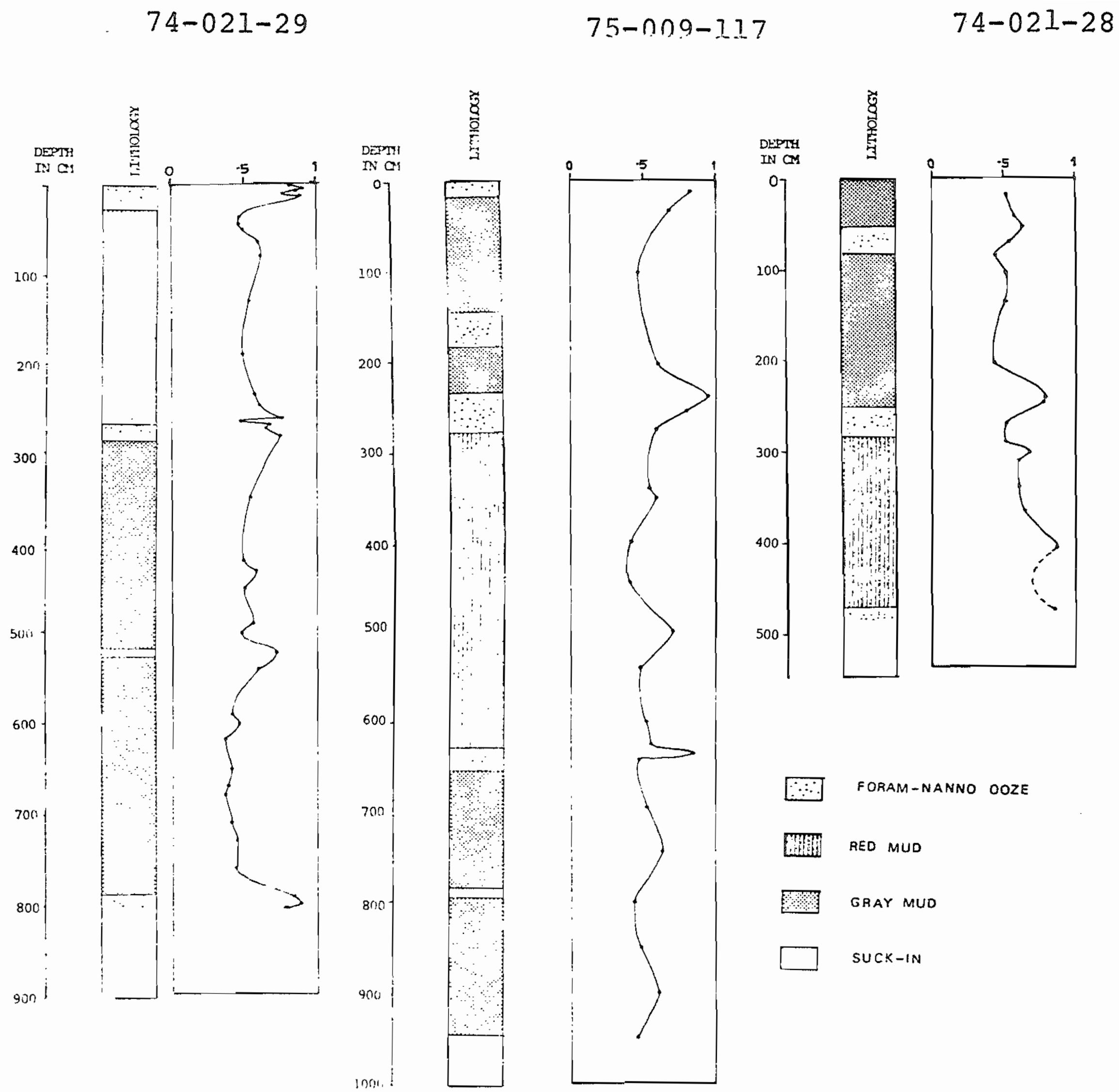


Figure 13: Silt and clay ratios in the cores, 74-021-29, 75-009-117 and 74-021-28

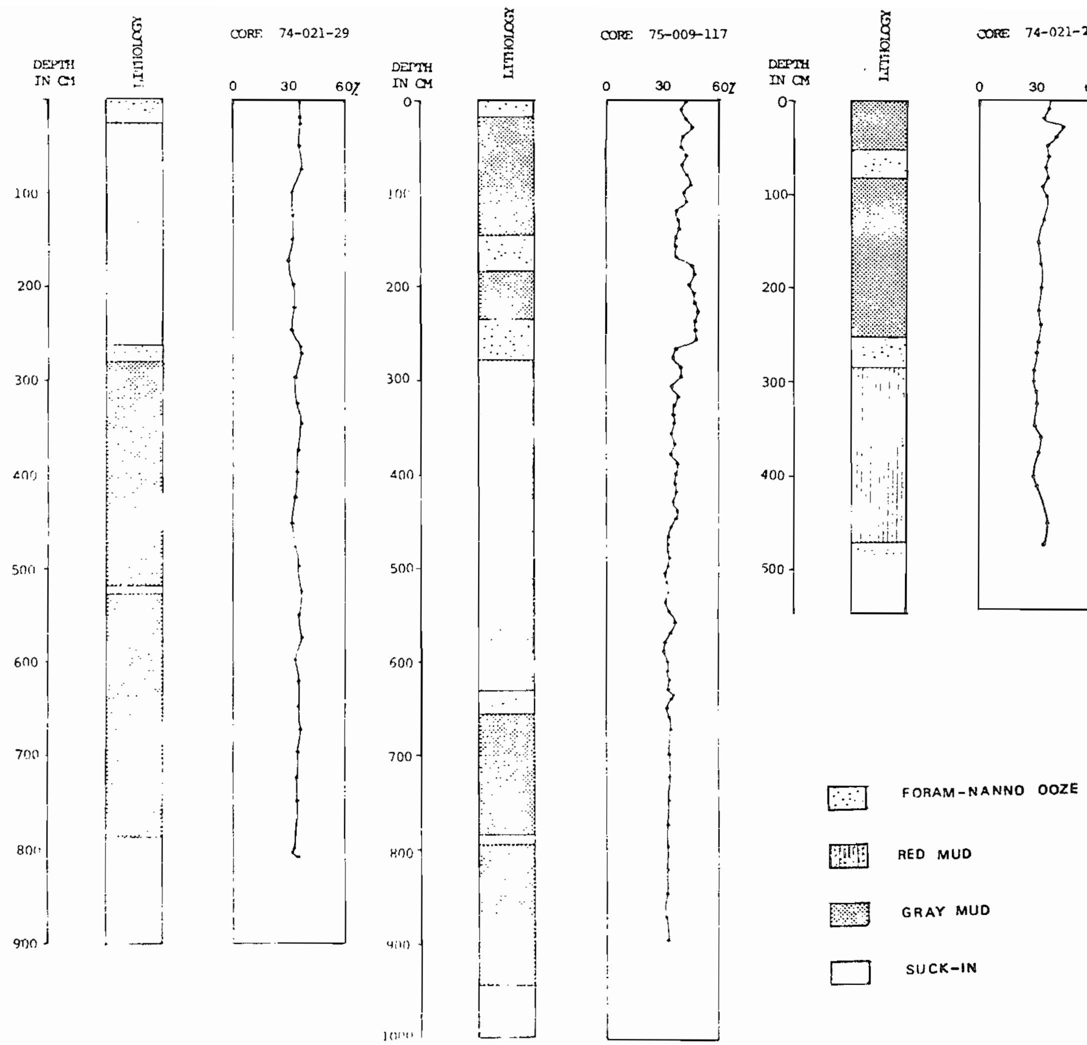


Figure 14: Plots of water content variations in cores

clay, which are only visible in the better quality X-radiographs (Plate 9,10) and in impregnated thin sections. They are quite continuous and uniform in thickness but sometimes disturbed near the edge of the core liner, presumably due to coring. There is slight grading and silt concentration near the base. These bands possibly represent short periods of nondeposition followed by sedimentation of apparently hemipelagic clay. Slight grading is found in these bands, possibly due to the size of the clay flocs and the presence of silt size particles.

Scouring

Evidence of scouring is rare in the cores. This may be either due to the homogenous nature of the sediments or to the rarity of scouring events. The major evidence of scouring appears at the tops of units E and G. Generally the upper and lower contacts of the foram-nanno ooze are gradational. It is shown in Chapter V that the contact nature of the foram-nanno ooze depends on climatic change. In general as the climate becomes warmer the foraminifera and coccolith concentration increases, and as it becomes cooler the concentration decreases. As a result these beds almost always have a gradational top and base. The most obvious reason for having a sharp upper contact of the foram-nanno ooze beds in units E and G is erosion. But it is believed that this erosion is very minor, most probably with removal of not more than a few centimeters of sediments, as indicated by the lack of pebble concentrations at the tops of the beds. The foram-nanno ooze always contains pebbles. As the finer fraction is eroded away the pebbles are left behind. Any large amount of removal of foram-nanno ooze is likely to leave a well defined pebble bed near the top of the units, but neither the clean core face nor the X-radiographs show any evidence of higher pebble concentration near the top of foram-nanno ooze beds. Scouring in the clay is generally suggested by the presence of lenticular silt laminae. One exception is near the top of the core 74-021-29, where the lower contact between the foram-nanno ooze (unit C) and the red mud appears to be rather erosional. However, since the top part of this core is missing, it is also possible that this contact may be due to coring disturbances (Plate 10A).

SECONDARY STRUCTURES

Although the rate of sedimentation is very slow, there is almost no burrowing visible in the sediments. Most of the silt laminae are undisturbed, except near the core liner. It appears from the x-radiographs that most of the observed mottles are due to disturbance of the sediments by coring or to diagenesis rather than burrowing. Normally these mottles appear where there is a change in the lithology. No bio-

PLATE 9

A. (Core 117, depth 70-80 cm): Granule bed and sharp silt laminae. Note the dragging effect on the granules.

B. (Core 117, depth 164-174 cm): Foram-nanno ooze with scattered pebbles.

C. (Core 117, depth 275-285 cm): A few sharp silt laminae with granules. Note the orientation of granules is possibly due to current.

D. (Core 117, depth 197-187 cm): Sharp silt laminae.

E. (Core 117, depth 540-550 cm): A few bands in the homogeneous clay.

F. (Core 29, depth 773-783 cm.): Microfault developed due to rotation of the core in the liner.

PLATE 10

A. (Core 29, depth 13-23 cm): Sharp contact between the red mud (unit D) and foram-nanno ooze (unit C).

B-E. Silt laminae with clay bands. Note absence of pebbles. (B = core 29, depth 452-462, C = core 29, depth 514-524, D = core 29, depth 556-566 cm , E = core 29, depth 580-590).

F. (Core 29, depth 795-805): Foram-nanno ooze with scattered pebbles.

PLATE 9

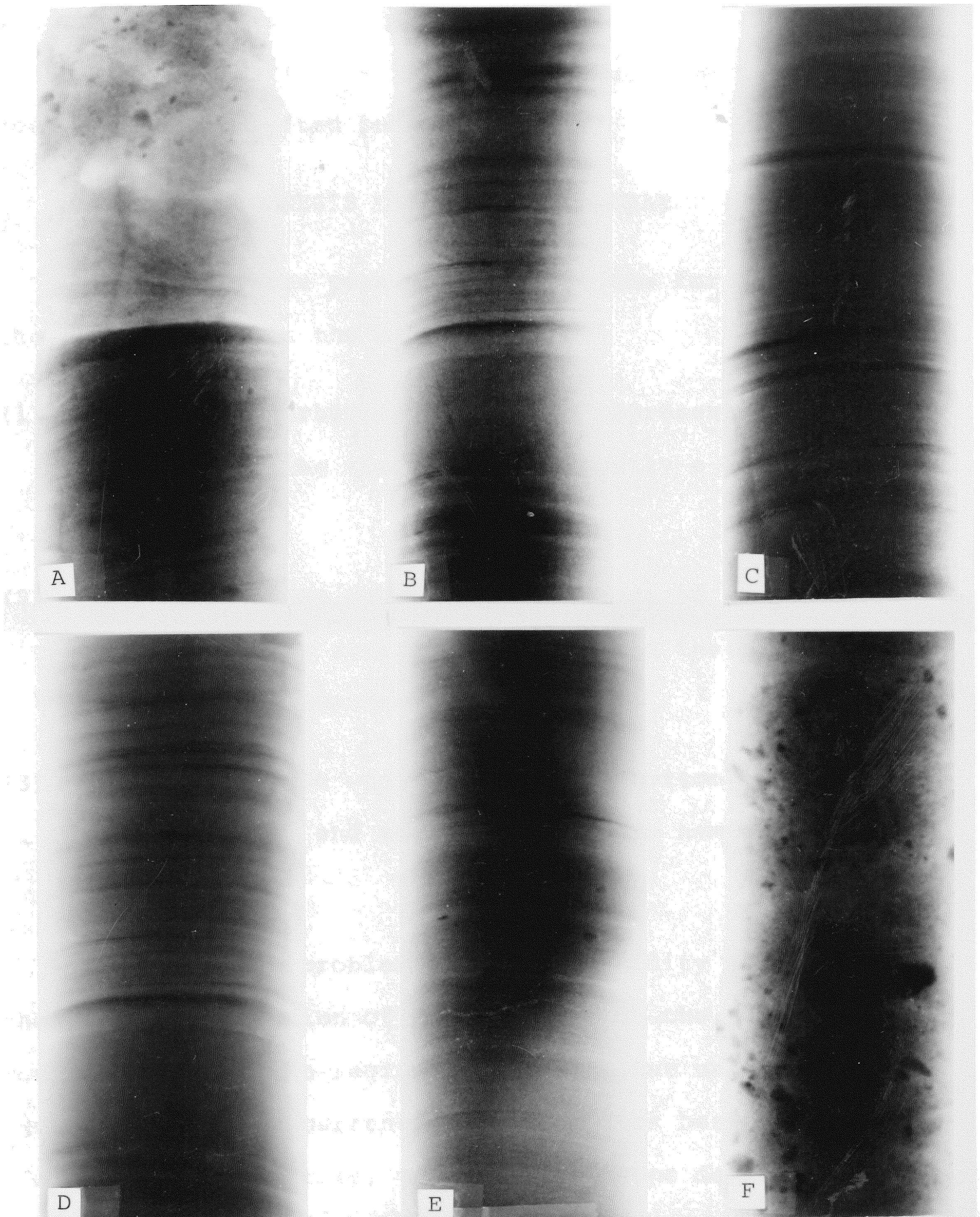


PLATE 10

turbation was detected in the X-radiographs. Pyrite mottles are present in two different forms, (1) as thin bands along the silt laminae, (2) as isolated patches, sometimes surrounding an ice rafted pebble.

GENESIS OF THE SILT LAMINAE

There are three possible explanations for the origin of the silt laminae in the clay.

(1) They are of turbidite origin and represent the d division (?) and the clay bands are mostly e division (?) (Bouma, 1962).

(2) The silt laminae formed from fall out of suspended material during period of storms, and the clay accumulated in fair weather.

(3) The silt laminae were deposited by bottom currents (contour currents) and the clay bands are hemipelagic sediments.

The principle problem with the turbidity current hypothesis is the location of the sediment on the top of a seamount 1 km above the regional sea bed. Let us assume that when the turbidity current meets the break between the continental slope and rise, the angle of slope decreases rapidly. The velocity of the tail of the current will still be relatively higher than the head and the tail will tend to build up over the body (Komar, 1972) and form a huge cloud. It is quite possible that at least a part of the fine grained sediments from the cloud will be trapped along a density interface within the ocean. These sediments, being dilute enough, should be able to travel for a considerable distance before the sediments concentrate just above the interface and move through it. Now, if the silt laminae formed by the density interface turbidity currents, we might expect to find in these muds characteristics found in normal muddy turbidites from the same source (Stow and Benteau, personal communication). This would include in general:

(a) Some change in carbonate and organic carbon content than the hemipelagic mud.

(b) Some near shore microfossils, especially diatoms, in the silt laminae, if they are derived from the Grand Banks.

(c) Gradual decrease in grain size towards the top, due to the decrease in the velocity of the fluid.

But our carbonate analyses do not suggest any remarkable change in carbonate content between the laminated mud and the fairly homogenous clay. The water content is also fairly uniform. Thin sections show that these silt laminae have the same composition as the rest of silt size particles in the clay and no evidence of any near shore microfossils or gradual change in grain size has been seen.

The second possibility is that these laminae has formed during storms when the sediments from the Grand Banks were swiftly carried in suspension by waves out to the deep ocean. In this case we can expect:

(a) Most of the laminae with coarse silt at the base and finer at the top. Since the silt would be deposited immediately after the storm, the coarse size particles would fall first and the finer later.

(b) Mineralogical composition of these laminae should be different from the silt associated with hemipelagic mud, but should be similar to the source area. In this case consisting mostly of quartz (Slatt, 1973 and personal communication, 1976).

We have already discussed that the contact nature of the silt laminae is quite variable, some with sharp tops and gradational bases and some with sharp tops and bases. These laminae have exactly the same mineralogical composition as the rest of the silt size particles in the mud, containing a lot of biogenic carbonate. In some places the Grand Banks sand is rich in carbonate, but they are mostly coarse sand size shell fragment, whereas these carbonates are mostly broken foram tests.

The third possibility is that they have formed by winnowing by bottom currents and hence could be so called contourite. In that case there should be:

(a) Direct evidence of bottom currents.

(b) At least some of the silt laminae should be lenticular and discontinuous.

(c) They should have the same mineralogical composition as the rest of the silt in the mud, since the current would mostly redistribute them.

(d) The nature of the contacts should be variable.

(e) They should be very thin.

The silt laminae are most common above the erosional surface of the foram-nanno ooze horizon (Figure 5), or when the coarse sands and granules suggest strong bottom traction (Plate 9). The lenticular, discontinuous nature of many silt laminae, their variable contacts and their mineralogical composition all support a bottom current hypothesis. Field and Pilkey (1971) and Bouma and Hollister (1972) proposed criteria to differentiate between the sand turbidites and contourites. Most of their criteria can be applied when there is full a succession of structures especially if cross lamination is present. However, if we consider their criterion that the microfossils are well preserved in turbidites, our data do not suggest that these laminae are of turbidite origin. Piper and Brisco (1975) suggested criteria to distinguish turbidite silt laminae from contourite laminae. Their first six critical criteria to distinguish the contourite silt laminae can be perfectly applied to these laminae. However, it is evident that more work is needed to prove beyond a doubt that these laminae are really contourites.

CHAPTER V

MICROPALEONTOLOGY AND QUATERNARY STRATIGRAPHY

Micropaleontology has been used in the study of marine Quaternary sediments with two different purposes. Firstly, to determine the temperatures or salinities that are recorded by particular sections of cores and secondly, to determine their age. Foraminifera are most extensively used for this purpose, but recently coccoliths are also being used. Quaternary biostratigraphy is more effective when both the foraminifera and coccoliths are used together. As extinction of micropaleontological forms in the Quaternary is rare, it is almost essential that the biostratigraphic data be controlled by radiochemical and paleomagnetic dating.

Planktonic foraminifera have long been known to be good paleoclimate indicators (Schott, 1935). In general it is assumed that the Plio-Pleistocene foraminifera have similar ecologic requirements to recent ones, and the latitudinal variation of different species in recent sediments can be used to make semiquantitative interpretations of past climatic changes. This is particularly true for the Pleistocene because the faunas are almost identical to recent ones. Pliocene and Miocene assemblages are slightly different and the interpretation of paleoclimate is somewhat hampered.

There are several methods for inferring paleoclimates by using foraminifera, but in general they can be divided into two types, one based on the total faunal assemblage, and the other on the relative abundance of a few temperature sensitive species and their coiling directions. Oxygen isotope analyses can also be used but requires special equipment.

Total faunal analysis has been used by several authors, including Ericson and Wollin (1956), Ericson et al. (1961), Ruddiman (1971), Barash (1971). This method involves dividing the fauna into three groups (i) species characteristic of cold and cold-temperate water, (ii) species characterizing warm and warm-temperate water, (iii) cosmopolitan species. The last group is excluded from counting and subsequent calculation. The ratio of the warm to cold species is calculated for each sample. The major advantage of this method is that it takes into consideration all the major species, thus increasing the reliability of the result. Total faunal analysis is time consuming. It is often difficult to evaluate properly the paleoclimatic significance of each of the species counted and overlapping ranges produce complex results. Some of the warm water species may be affected more by deep-sea carbonate dissolution, since most of them have comparatively thinner tests. Another disadvantage of total faunal analysis is that it can be used only in the temperate zones, where both the warm and cold water foraminifera are present. Imbrie and Kipp (1971) employed factor analysis to interpret the total faunal analyses, but their method fails when there is large scale variation in the faunal assemblages in the cores.

The most commonly used method based on specific foraminiferan species utilizes the relative abundance of _Globorotalia menardii_ and was first developed by Schott (1935). This method has been widely used by Ericson and his associates (1961, 1964, 1968, 1970). They prefer to use the whole _G. menardii_ complex, including _G. menardii tumida_ and _G. Menardii flexuosa_ and determine their relative percentage in the sample to draw the climatic curve. The advantage of this method is that it takes relatively little time. The climatic curves so produced appear quite reliable although, as Emiliani (1964) and Morin _et al_. (1970) point out, it is risky to draw paleoclimatic conclusions from the study of only one species which could be accidentally or for some unknown reason absent from the fauna. Lidz (1966) objects to the use of whole _G. menardii_ complex because different members of this group have different temperature ranges. Boltovskoy (1973) advocates use of the ratio of _G. menardii_ complex and _Globigerina inflata_, but the major problem in this case is that some authors (Kennett, 1970) report that _G. inflata_ is not recognized below the Brunhes-

Matuyama magnetic boundary. As a result, using this ratio will give consistantly high values for the Lower Pleistocene and earlier times.

Coiling directions of Globigerina pachyderma have been shown to be a very good paleoclimate indicator (Ericson, 1959; Parker, 1971), especially for the higher latitudes. In lower latitudes this species may be absent, Sinistral (left coiling) forms of this species prefer cold water whereas dextral (right coiling) ones prefer temperate water. The main disadvantage of this technique is that it can be used only in cold and cold-temperate zones. Sometimes this species is difficult to identify and can be confused with other species.

Both the abundance and the coiling directions of Globorotalia truncatulinoides can be used to reconstruct the paleoclimate. There is still some disagreement concerning the type of water mass that this species indicates. Bé (1969) and McIntyre et al. (1972) consider it as an indicator of subtropical water, whereas Boltovskoy (1966, 1969) and Ruddiman (1971) prefer to interpret it as a representative of cold-temperate water. According to Parker (1971) and Bé and Tolderlund (1971), in the South Atlantic, the left coiling (sinistral) form dominates at higher latitude and the right coiling (dextral) form at lower latitudes, but in the North Atlantic no such trend is found. Wollin *et al.* (1971) think that coiling direction of *G. trunctulinoides* in the North Atlantic has responded to the climatic changes of the past 75,000 years, and that the left coiling (Sinistral) form dominates during the time of warm climate and the right coiling (dextral) during cold climate. This contradicts the present distribution in the South Atlantic. Cifelli (1971) and Boltovsky (1973) believe that the coiling direction of *G. truncatulinoides* cannot be used for determining paleoclimates, at least in the areas that they studied.

Other techniques that have been used to reconstruct paleoclimates include the relative abundance of *Globigeri noides saculifer* (Herman, 1968) and *Globigerina rubescens* complex (Frerichs, 1968). Morphological variation of some species may also have temperature significance (Boltovskoy, 1969; Kennett, 1968). None of these techniques is widely used.

Porosity of Foraminiferan Tests

Pore concentrations on the test walls of foraminifera were first studied systematically by Wiles (1967). He counted the number of pores per unit area of *Globoquadrina* dutertrei collected from deep-sea cores and was able to demonstrate that the lower number correlate with core horizons corresponding to colder paleoclimates. Bé (1968) measured the pore diameter and pore concentrations of 22 species of planktonic foraminifera in an area of 25 μ x 25 μ and concluded that the shell porosities of planktonic foraminifera are useful indices for interpreting Cenozoic climates. Frerichs et al. (1972) measured pore densities and porosities on a number of specimens collected from surface sediments of the Indian Ocean and found a decrease in the test porosities with increasing distance from the equator.

Coccolith Assemblages

Several authors including McIntyre, Bé and Preikstas (1967); Bolli et al. (1968); McIntyre (1967); McIntyre and McIntyre (1971); McIntyre et al. (1972) used the coccoliths from deep-sea cores as a paleoclimate indicator. Although there has been sufficient work on the modern surface water species, the major problem with coccoliths is that fewer than 25% of them are preserved in oceanic sediments, which contain mostly robust forms. Most species produce fragile coccoliths prone to mechanical and chemical destruction in the water column and on the bottom. As a result studies dealing with modern surface water distribution of coccoliths are not perfectly applicable to fossil coccoliths. Recent work by Roch et al. (1974) and Ruddiman and McIntyre (in press) suggest that, with certain limitations, coccoliths can be used to infer paleoclimates.

METHOD USED IN THIS STUDY

Paleoclimatic curves based on a few temperature sensitive species exaggerate the effect of large scale temperature oscillations. This exaggeration makes the separation of glacial, interglacial and interstadial stages much easier. In this study emphasis was placed on a few temperature sensitive species, especially Globigerina pachyderma, Globorotalia menardii complex and G. truncatulinoides (Plates 11,12). Samples were split with a microsplitter and the relative abundance of all the major species present in the sample was estimated. The percentages of G. menardii, G. trunctulinoides and benthonics were then determined by a count of 300 individuals. Weight percentage of these species against the total weight of the sample was not used as the samples contain variable amounts of detrital particles. It seemed logical not to attempt to separate these detrital particles and hence disturb the sample. The ratio of benthonic to planktonic foraminifera was determined in order to provide a control over the variation in the input of the planktonic foraminifera, since it is unlikely that changes in the surface water temperature would affect the benthonic productivity at such a great depth.

Coiling direction of G. pachyderma was determined by counting 100 individuals, whenever possible. However, in

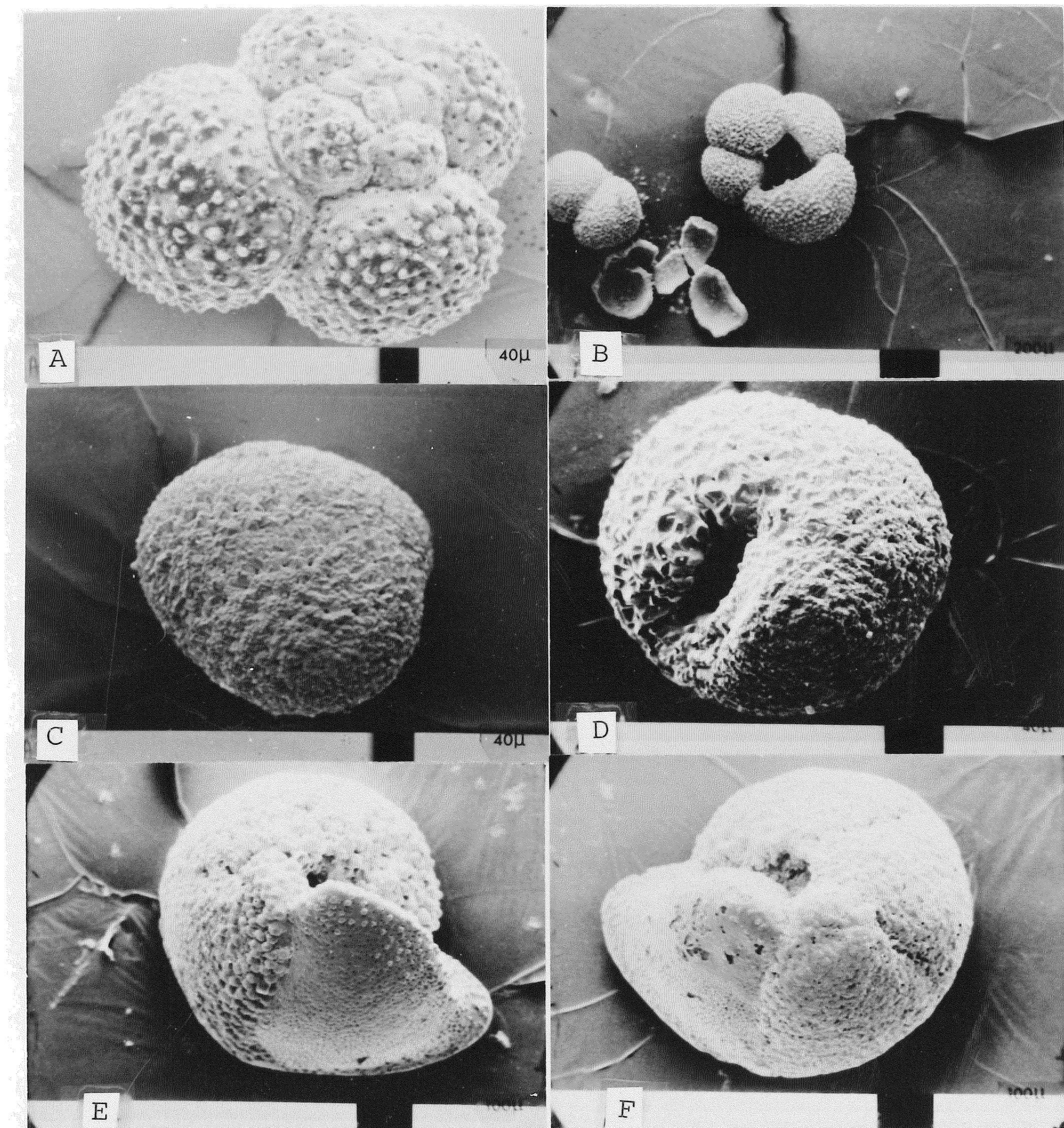


PLATE 11: A-B. Globigerina bulloides (d'Orbigny)
C-D. Globigerina pachyderma (d'Orbigny)
E-F. Globorotalia truncatulinoides (d'Orbigny)

Scale: A,C and D = 40µ , B = 200µ , E and F = 100µ

PLATE 12

Globorotolia menardii complex

A-B. *Globorotalia menardii menardii* (d'Orbigny).

C. *Globorotalia menardii tumida* (H. B. Brady).

D. *Globorotalia menardii fimbriata?*

E-H. *Globorotalia menardii flexuosa* (Koch).

Scale: the black bar at the bottom of each photograph is 200 μ.

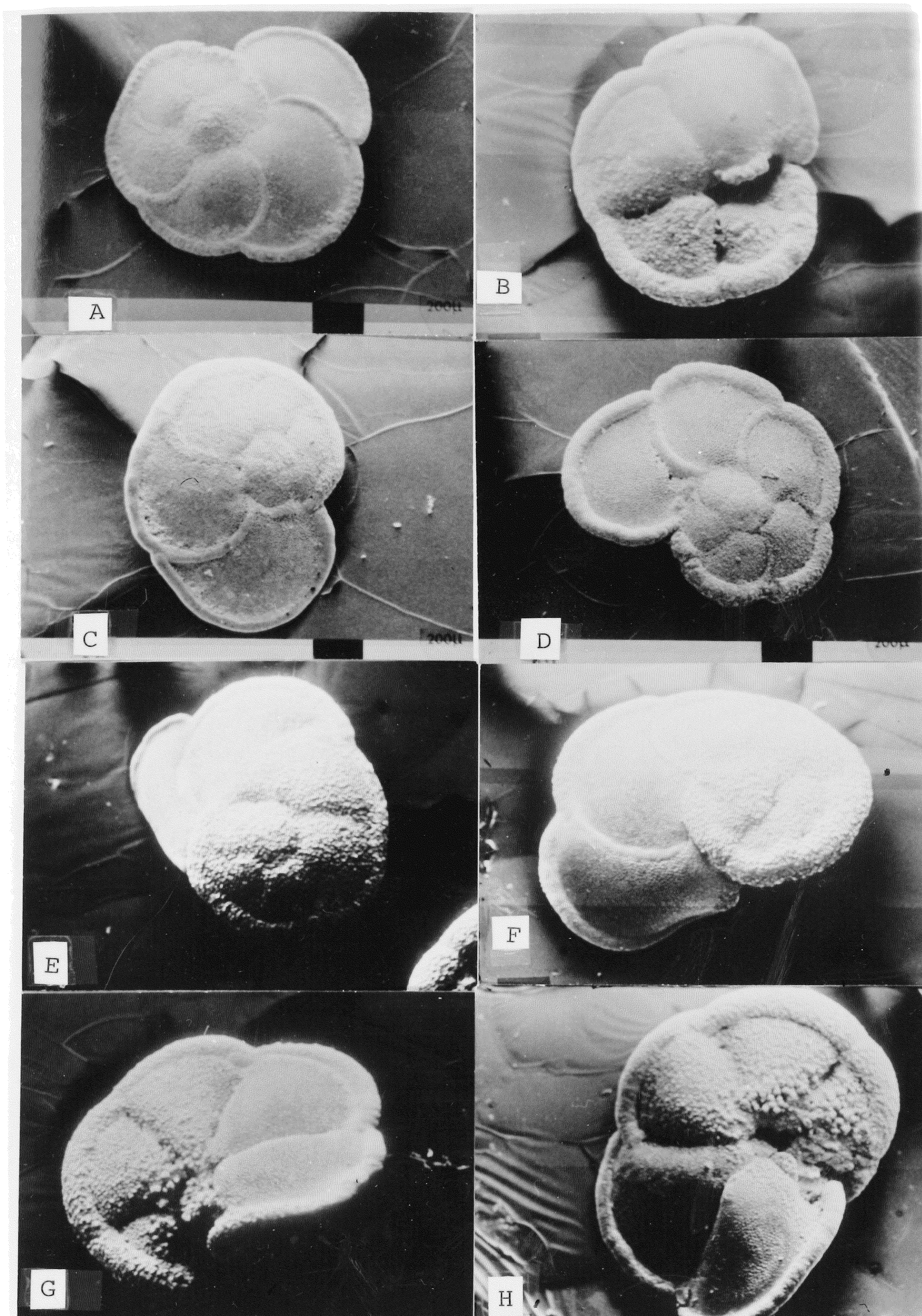


PLATE 12

some samples with low foram numbers (Figure 15) it was impossible to obtain 100 individuals. Coiling directions of G. truncatulinoides were also determined. It was later found that the coiling direction of G. truncatulinoides is rather unreliable as a climate indicator as suggested by Boltovskoy (1973). It is difficult to conclude from the study of a small area whether the coiling of this species is unreliable on a global scale or merely at the study area, which is close to a transitional zone. In the later phase of this work relative abundance of this species was determined.

Porosities of Gobigerina pachyderma and Globigerina bulloides were measured with a Cambridge 600 SEM to determine the relative temperature difference between two warm or cold horizons. For this purpose G. pachyderma specimens having the size range 200 μ to 300 μ and G. bulloides from 400 μ to 500 μ were mounted on an SEM stub with two-sided scotch tape; the tests were then broken with a needle and cleaned with a brush. The samples were coated and the pore concentration and diameter were determined. The measurements were carried out on the SEM screen and on photographs taken at X4000 magnification. At this magnification the displayed image corresponds to an area of 32 μ X 40 μ. All the pores within this area were counted and measured.

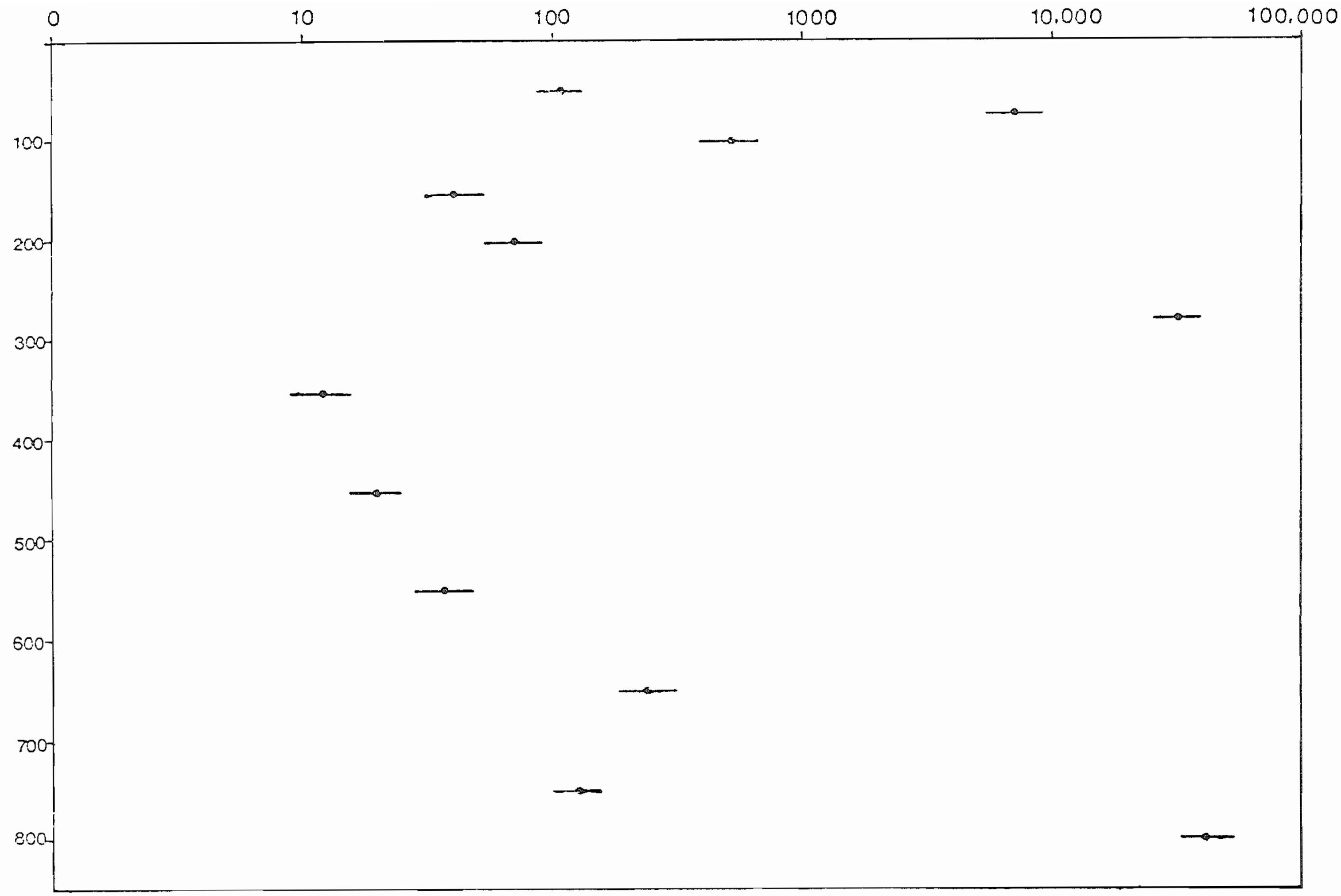


Figure 15: Number of foraminifera per gram of dry sediment (Foram Number) in core 74-021-29.

Coccoliths

Coccoliths were used in this study in order to provide a biostratigraphic control and to refine the paleoclimatic data obtained from the foraminiferan analysis. For paleoclimatic purposes the relative abundances of the coccoliths were estimated from smear slides and were compared with the foraminiferan cureves. Samples from different levels in the core were examined under the SEM to learn their biostratigraphic zonation.

Sample preparation was slightly different from the standard procedure (McIntyre et al., 1965) because of the presence of high amounts of clay. About 1.5 gm dried sample was soaked in 20 ml water, 10 ml 6% hydrogen perioxide and 10 ml calgon 1% for about 6 hours and then ultrasonically agitated for approximately 1 minute.

The solution was then sieved through a 45 μ sieve and the fine fraction was collected in a beaker. It was then centrifuged by an ultracentrifuge Sorvli RC2-B with SS-34 rotor in 50 ml tubes at 1400 rpm for 4 minutes. Samples were then transferred into small 20 ml vials and, after the addition of a sufficient amount of water and calgon, the material was put into suspension. It was allowed to settle for about 5 hr and the liquid from the top was removed by pipet. The process was repeated until the sample became completely clean. It was found that long use of a centrifuge to clean the sample results in breaking down of all the coccoliths. Once the sample was clean it was mounted on the SEM stub with a drop of acetone.

CLIMATIC CURVES FROM THE FORAMINIFERAN ANALYSIS

Figure 16 shows the relative distribution of the major planktonic foraminifera in the core 74-021-29. Transitional, subtropical and subarctic species, Globorotalia inflata, Globoquadrina dutertrie, Globigerinoides ruber, Globorotalia truncatulinoides, Globigerinoides sacculifer, and Globorolatia menardii dominate the upper (unit C), middle (unit E) and lower (unit G) foram-nanno ooze horizons. The reddish mud (unit D) and grayish mud (unit F) are dominated by subarctic and arctic species Globigerina pachyderma, Globigerina quinqueloba and Globigerina bulloides. This suggests comparatively warm climate during the deposition of foram-nanno ooze and cold climate during the deposition of both red and gray mud.

The coiling direction of G. pachyderma and the relative abundance of the G. menardii complex and G. truncatulinoides are shown in Figure 17. The percentage of the right coiling G. pachyderma is always higher in the foram-nanno ooze than in the muds. The unit C of this core shows some oscillation from cold to warm, but in general indicates a comparitively warm climate. This is supported also by the increase in the percentage of _G. menardii_ and _G. truncatulinoides_. In the underlying reddish mud, the left coiling variety of _G. pachyderma_ is dominant; this, as well as the absence or very low percentage of _G. menardii_ and _G. truncatulinoides_, suggests cold climatic conditions. The unit E separating the red mud from the gray mud shows an increase in the right-coiling _G. pachyderma_ accompanied by increase in the percentage of _G. menardii_ and _G. truncatulinoides_, suggesting a warmer climate, but possibly less warm than that of unit C. Below this horizon coiling directions suggest severe cold, but towards the base of this gray mud the climate was slightly warmer and in one case at 520 cm it shows substantially warmer conditions. The lower foram-nanno horizon (unit G) represents the warmest climatic condition documented in the whole core.

Calcium carbonate is higher where ever the micropaleontological evidence suggest warmer climatic conditions. One interesting aspect of this core is that both the lower and middle foram-nanno horizons (units G and E) show gradual increase in the temperature inferred from indicator species, but just near the top it shows a sudden decrease in the temperature. The percentage of calcium carbonate also shows the same pattern. Gravity core 74-021-29 G shows gradual increase of the temperature (Figure 18) and the upper part

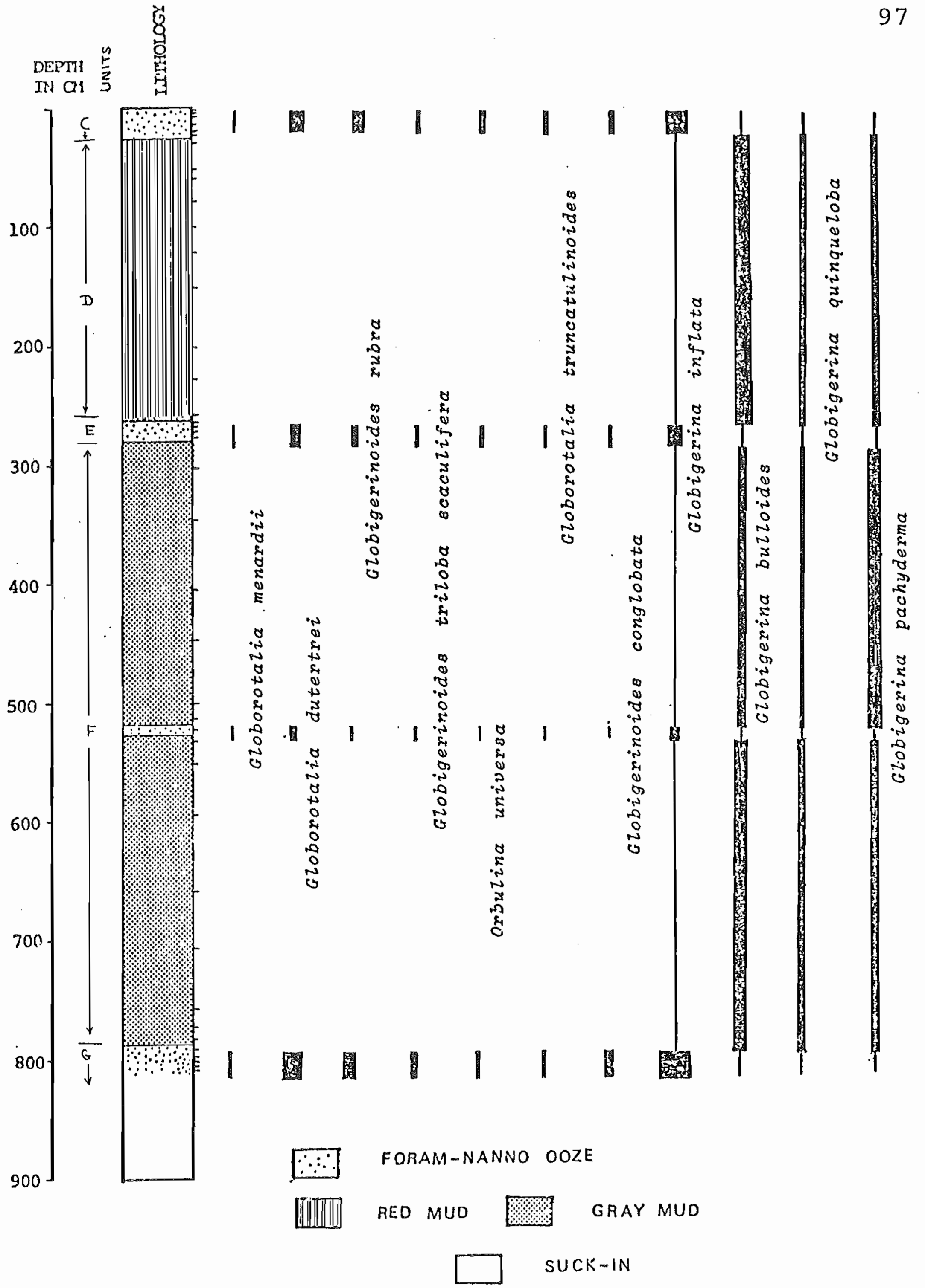


Figure 16: Relative abundance of major planktonic foraminifera in core 74-021-29.

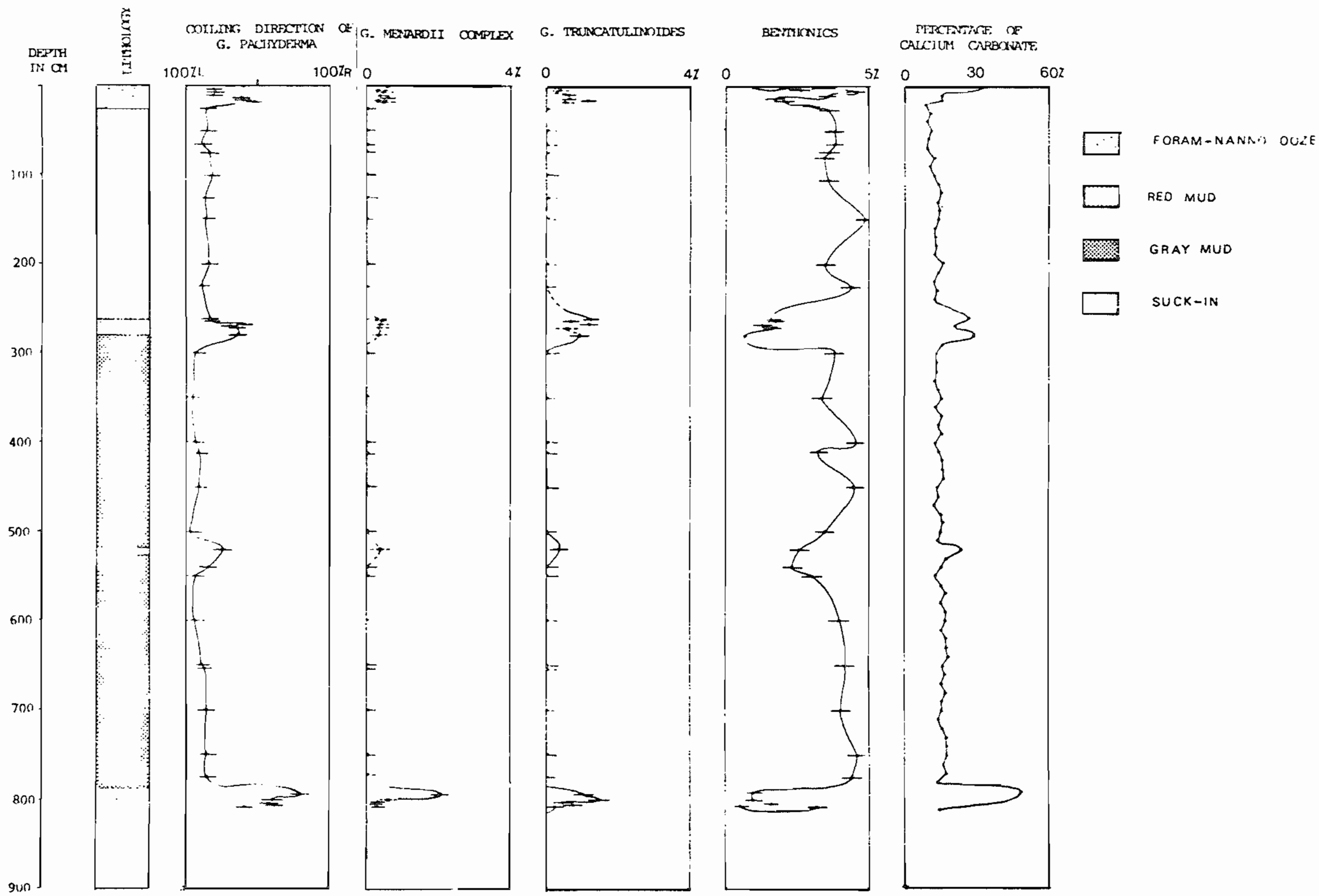


Figure 17: Climatic curve and carbonate content of core 74-021-29.

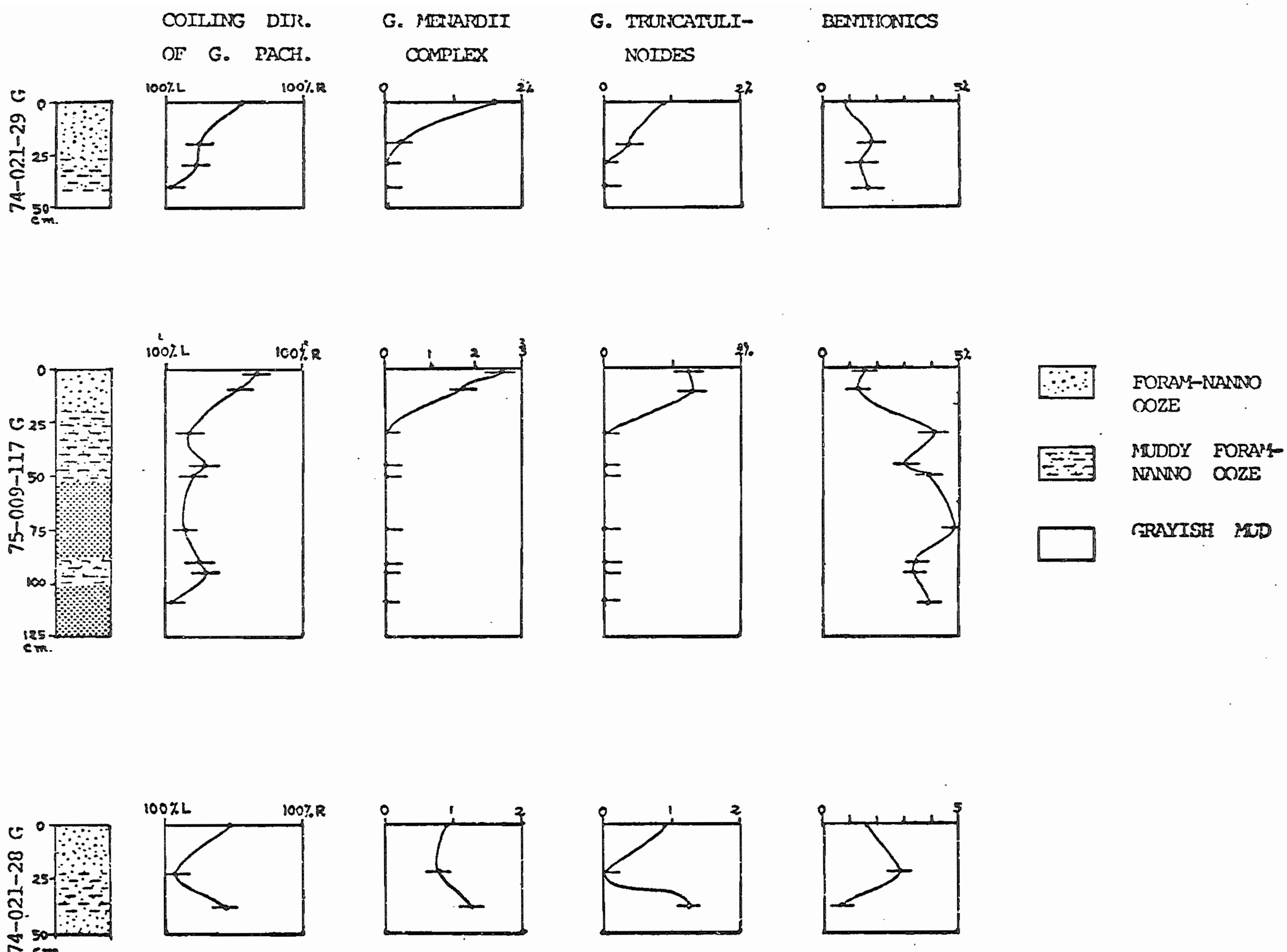


Figure 18. Lithology and paleoclimatic curve of the gravity cores.

of the core represent a climate equal to or warmer than, the lowest foram-nanno ooze horizon (unit G) of the main core. The percentage of benthonics normally decreases with warmer climatic conditions.

The overall faunal distribution in core 75-009-117 is shown in Figure 19. As in the previous core, in this core also, transitional, subtropic and subarctic species of foraminifera dominate the foram-nanno ooze horizon, whereas the mud is mostly inhabited by the arctic and subarctic species. The coiling direction of *G. pachyderma* shows evidence of a fairly warm climate just at the top of the core, and in the foram-nanno ooze (unit C), just above the red mud. In between these two horizons the upper gray mud (unit B), with muddy foram-nanno ooze, shows the presence of three comparatively warm periods (Figure 20). This increase of the right-coiling *G. pachyderma* is in most cases accompanied by a relative increase in the percentage of *G. menardii* and *G. truncatulinoides*. The red mud shows a cold climate with some variation in the intensity. The foram-nanno horizon (unit E), separating the red mud and the lower gray mud, shows an increase of right coiling *G. pachyderma* and of *G. truncatulinoides*. The coiling direction of *G. pachyderma* indicates severe cold in the lower gray mud with a slight warm stage at 790 cm depth.

Paleoclimatic analysis of core 74-021-28 shows one warm

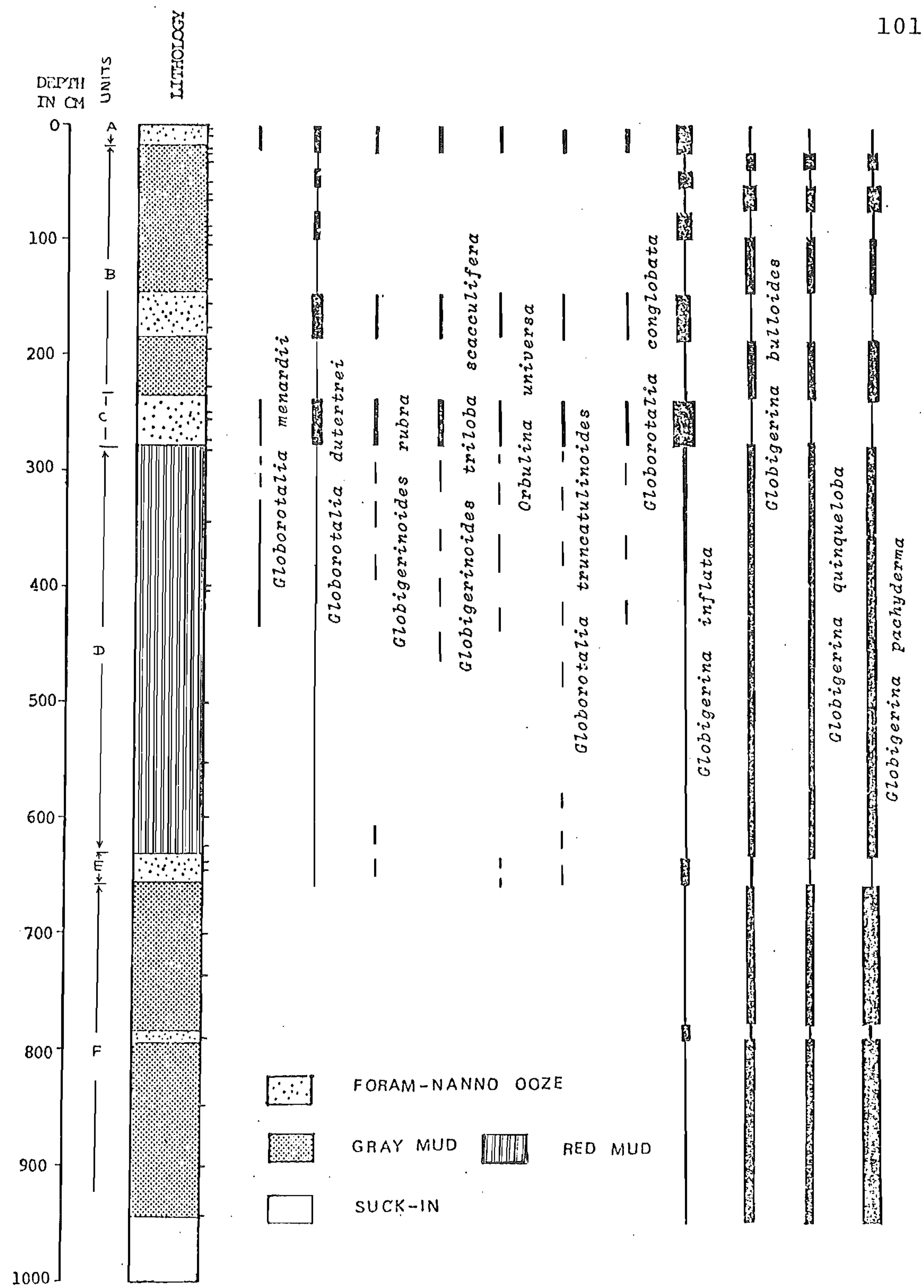


Figure 19: Relative abundance of major planktonic foraminifera in core 75-009-117.

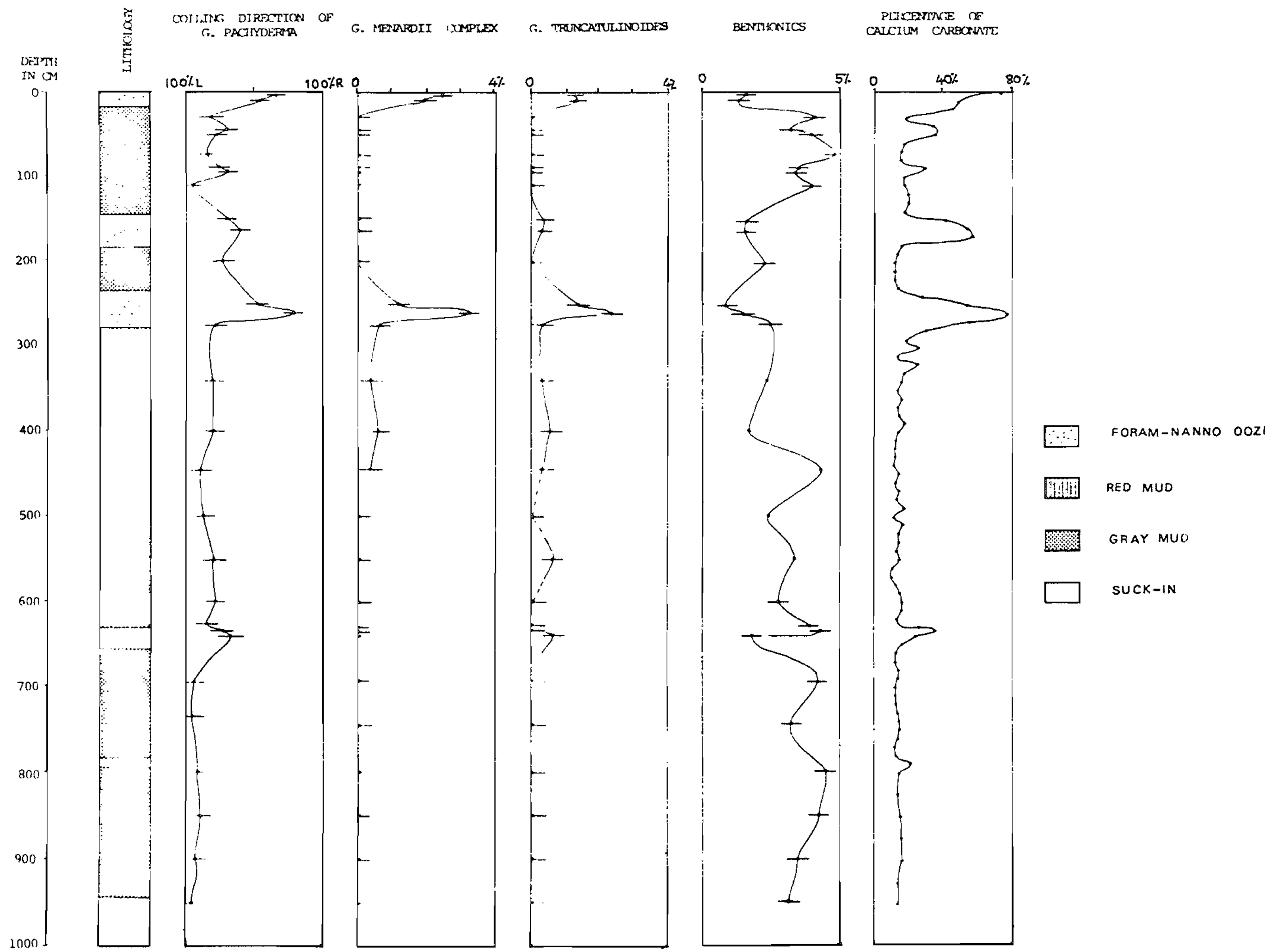


Figure 20: Climatic curve and carbonate content of core 75-009-117.

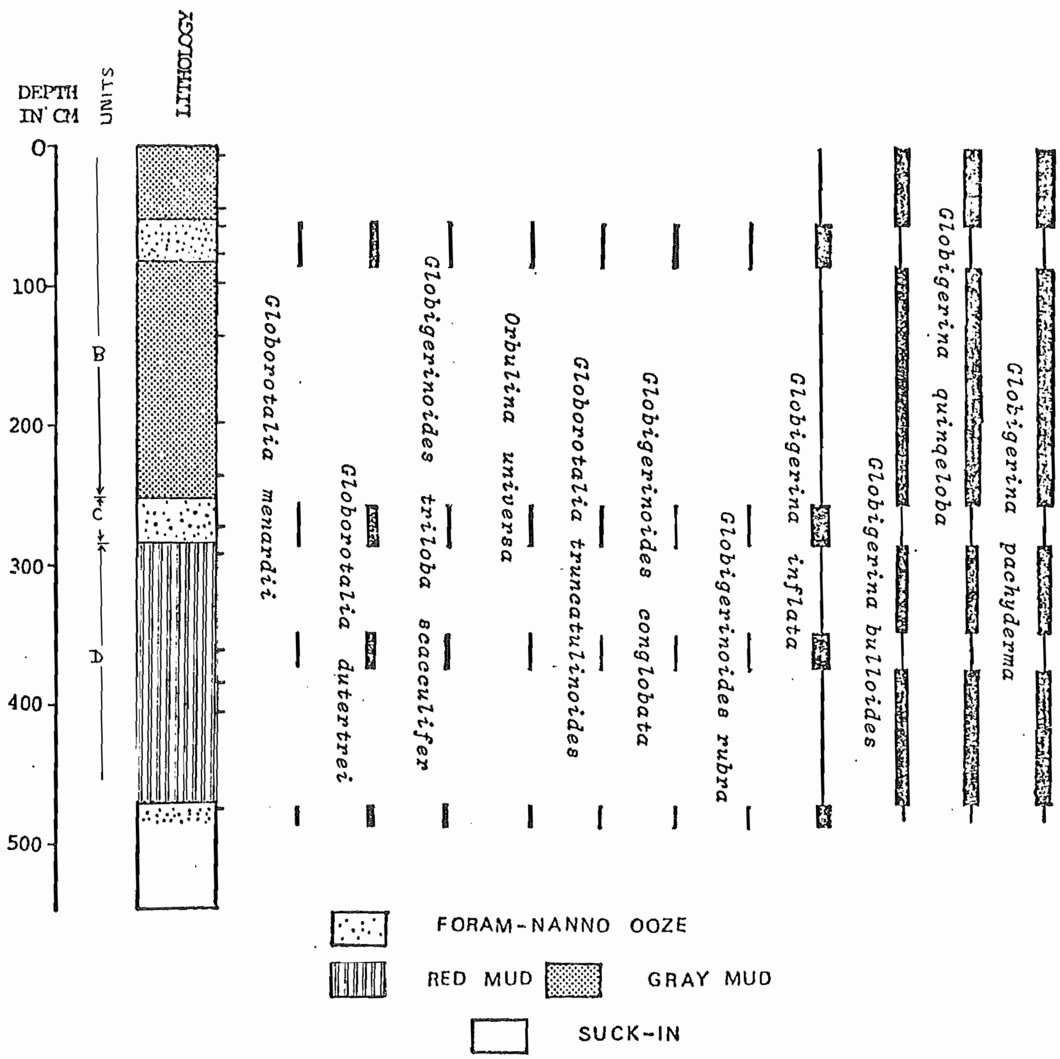


Figure 21: Relative abundance of major planktonic foraminifera in core 74-021-28.

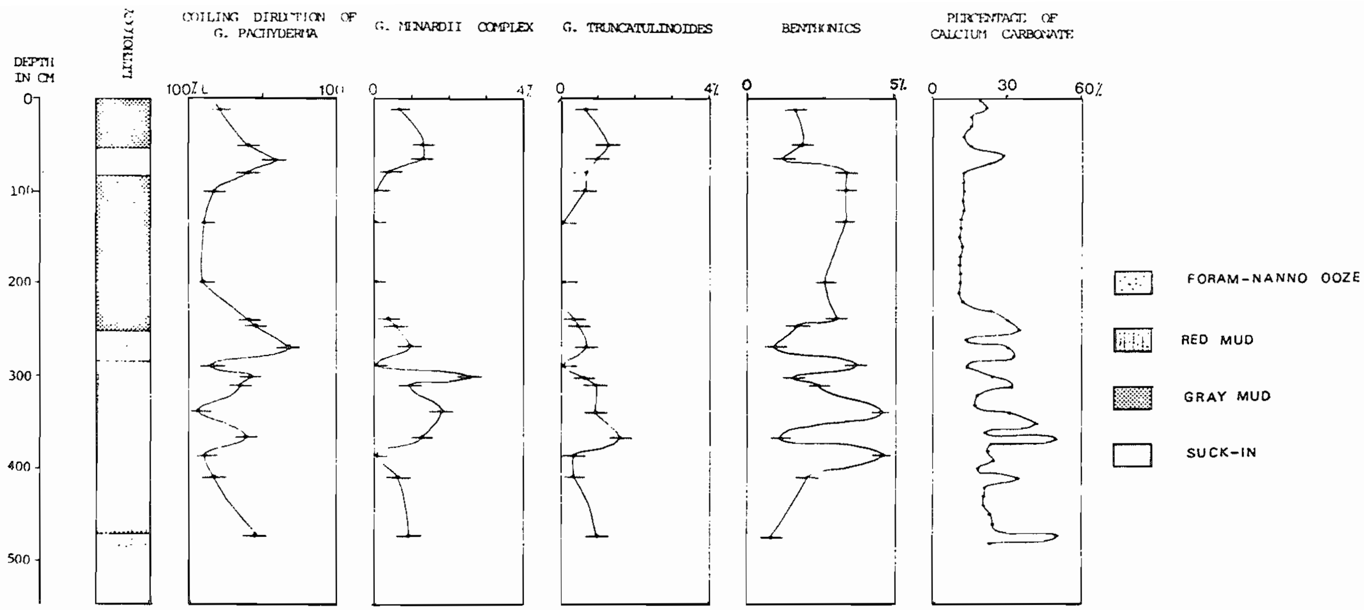


Figure 22: Climatic curve and carbonate content of core 74-021-28.

stage in the upper gray mud (unit B) (Figures 21, 22). The foram-nanno ooze (unit C), which separates the upper gray mud from the red mud, shows warm climatic conditions but with several oscillations. The red mud shows relatively cold conditions, again with some oscillation. As the core was collected from the flank of a seamount where slumping was possible, interpretation of the detail of this core must be limited. It is possible that many of the minor oscillations recorded in this core are due to spillover of the sediments from the top of the seamount. However, the major features recorded in this core are fairly consistent with the other cores, i.e. a remarkably warm stage in the upper gray mud (unit B) and a very warm stage (unit C) separating the red mud from the upper gray mud.

Porosity of the Foraminiferan Tests

Pore concentration and pore density of at least three specimens from each level in the core were randomly selected for measurements. Average pore diameter and pore number were calculated from these measurements (Appendix H). The outlines of the pores are often serrated and are seldom a perfect circle. Moreover, due to the hemispherical shape of the foraminiferal tests, not all pores could be viewed at right angles to the surface within the measured area. To minimize error, only maximum diameters were measured, even if this caused a slight over-estimation of the true pore area. Porosity of G. pachyderma and G. bulloides from core 74-021-29 are plotted in Figure 23. They suggest the result obtained from the foraminiferan count, that the climatic conditions during the deposition of mud were much cooler than during the deposition of the foram-nanno ooze. In comparison, the gray mud was deposited under colder climatic conditions than the red mud, and the unit G represents the warmest stage in this core.

COCCOLITH FLORA

Abundance of coccoliths (Figure 24) in the cores varies drastically. Maximum coccolith concentration is found in the units A, C and G. Unit B is marked by three coccolith-barren zones as well as by three coccolith-high concentrations. Coccolith concentration is very low in the red mud (unit D) and in the lower gray mud (unit F). These two units are occasionally marked by coccolith-barren zones. In general the coccolith concentration is very low whenever the foram data suggest a cold climate. This may itself be a useful piece of information.

RECONSTRUCTION OF THE PALEOCLIMATE OF THE AREA STUDIED

Figure 27 shows the changes in the climatic condition and the carbonate content plotted against the depth for the

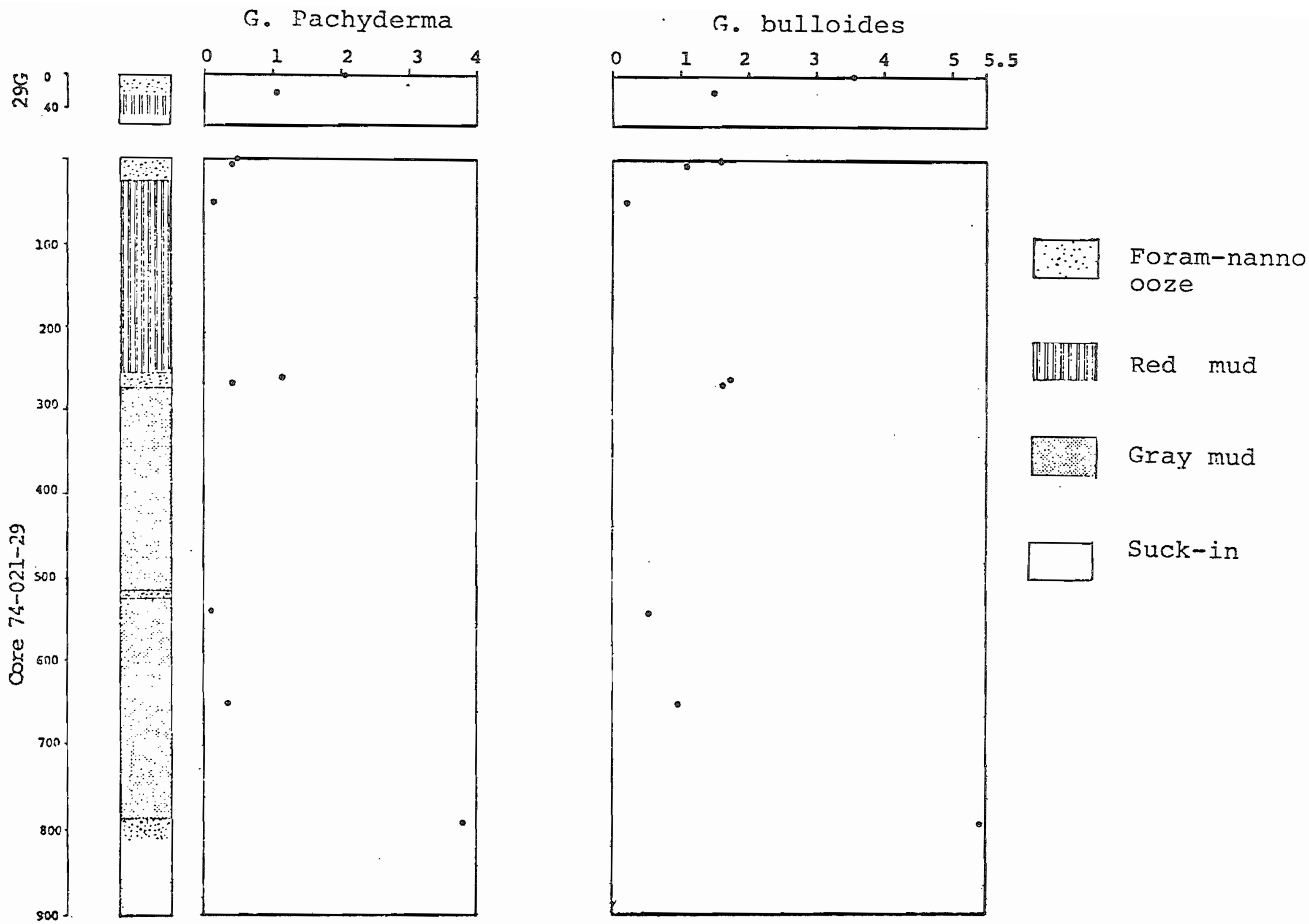


Figure 23: Porosity of G. Pachyderma and G. bulloides tests.

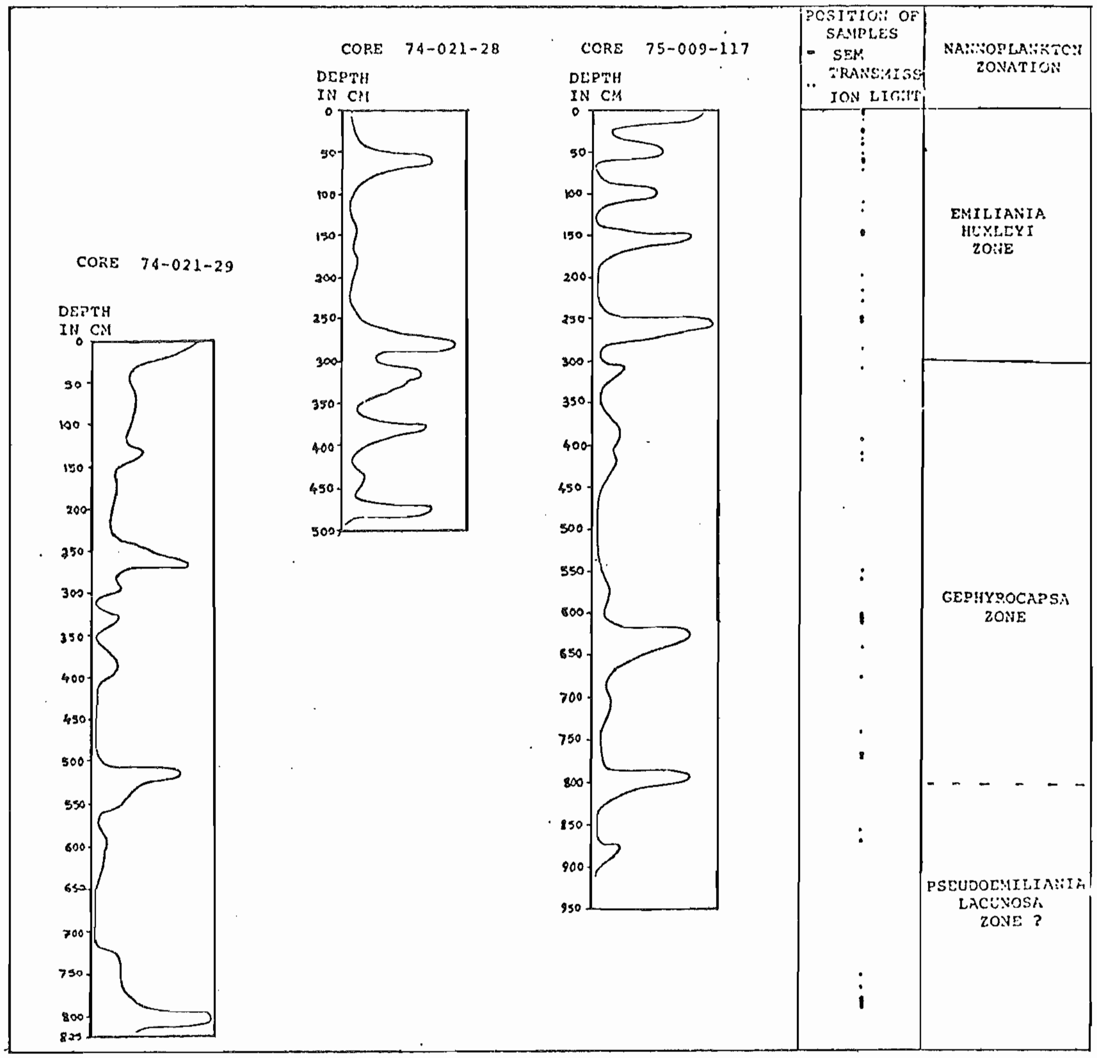


Figure 24: Relative abundance of coccoliths in different cores and their biostratigraphic zonation.

cores 74-021-29 and 75-009-117. As can be seen, the time span under consideration shows eight carbonate high and eight relatively warm periods. Three very remarkable warm periods are recorded in these cores. The first of these is just at the top of the cores (unit A); the second occurs between the upper gray mud and red mud (unit C); and the third is just below the lower gray mud (unit G). The upper one obviously represents the present climatic condition. If this is the case, then the lower horizons should represent periods when the studied area experienced temperatures equal to or warmer than the present.

BIOSTRATIGRAPHIC ZONATION

The presence of Globrotalia truncatulinoides throughout the core suggests that the cores are entirely in the *G. truncatulinoides* zone and therefore Quaternary in age.

Globorotalia menardii flexuosa is present in the lower part of the cores but it disappears above unit C in all the cores. Schott (1935) first reported the extinction of *G. m. flexuosa* and correlated this event with the last interglacial stage. Ericson *et al*. (1961, 1968) concluded from their study of equatorial Atlantic cores that this extinction occurs in their zone X, which they believe to represent an interstadial of the Wisconsin glaciation. Their zone X is separated from the Sangamon Interglacial (zone V) by the Early Wisconsin glaciation (zone W).

Pastouret *et al.* (1975) suggested from their study of the South East Newfoundland Ridge, that this extinction occurred in the Sangamon Interglacial. In the present case, Unit C, which is at least as warm or warmer than the present, is separated from another warm horizon (Unit B6) by a relatively severe cold period (Unit 7). The extinction of *G. m. flexuosa* in the Unit C in the present study is considered to represent the last interglacial as originally proposed by Schott (1935), and hence, as equivalent to the Sangamon Interglacial.

Nannoplankton Zonation

Garter (1969) suggested the following nonnoplankton zonation for the Central Atlantic.

Emiliania huxleyi zone

Gephyrocapsa zone

Pseudoemiliania lacunosa zone

The *Pseudoemiliania lacunosa* zone, extends from the first occurrence of Gephyrocapsa to the last occurrence of *Pseudoemiliania lacunosa* (Geitzenauer, 1972). Ruddiman and McIntyre (in press) draw the extinction of *P. lacunosa* at 400,000 years B.P., where as Geitzenauer places it between 520,000 and 580,000 years B.P.

The *Gephyrocapsa* zone, extends from the last occurrence of P. lacunosa to the first occurrence of Emiliania huxleyi.

The Emiliania huxleyi zone, extends from the first occurrence of E. huxleyi to the present. The base of the E. huxleyi zone is about 150,000 years B.P. (Geitzenauer, 1972).

Figure 24 shows the relative abundance of total nannoplankton and their zonation in different cores. Above unit D, the cores are entirely within the E. huxleyi zone. Positive identification of P. lacunosa was not possible and hence the boundary between the Gephyrocapsa and Pseudoemiliania lacunosa zones is uncertain in these cores.

CHAPTER VI

STRATIGRAPHY

MAGNETIC STRATIGRAPHY

Since the pioneer work of Cox et al. (1963), paleomagnetic measurements have been considered to be a very useful tool for dating marine sediments. The method is primarily based on the phenomena that the earth's magnetic field experiences changes in direction with time and certain rocks accurately record magnetic field directions existing at the time of their formation by means of a remanent magnetization. Two types of geomagnetic field changes during Quaternary time are known. One involves sudden complete reversals of geomagnetic field direction and the other (secular variation) involves uniform changes of a few tens of degrees with periods of hundreds of years.

The paleomagnetic time scale of Cox (1969) is used in this study. Different magnetic events in his time scale are dated by K/Ar methods. With increasing age the error in age determination became larger than the length of the shorter events. As a result, even the longer events of middle and early Pliocene reversal time scale are difficult to resolve. There also exists considerable inconsistencies and nomenclature conflicts between different groups (Watkins, 1972; Opdyke, 1972; Olausson and Svenoniu, 1975).

Core 74-021-29 was used for detailed paleomagnetic study. Samples were taken at 50 cm intervals from the upper few meters of the core but from the lower part they were taken at 10 cm intervals. Magnetic polarity was measured on a DSM-1 Digital Spinner Magnetometer before and after demagnetizing at 25, 50, 75, 100, 150, 250, 300 and 400 oersted. The field was found to stablize at about 100 oersted in most samples. Most of the sediments have fairly high magnetic moments, sometimes upto 0.15 $\times 10^{-2}$ gauss. No special precautions were taken to handle the core during removal from the piston corer and subsequent cutting. As a result declination has not been used in this study. The inclination has been plotted against the depth (Figure 25). Though the inclination shows secular changes, it fails to record any magnetic reversal. Sampling was at sufficiently close intervals that it is unlikely that any reversal epoch could have been missed. The time covered by the core thus appears to be entirely within the Brunhes Normal Epoch: that is the sediments are younger than 700,000 years B.P.

ABSOLUTE DATING

Carbon 14 dates are presented in Table 3. All these dates are on bulk samples from the foram-nanno ooze. No attempt was made to separate the forams, since the amount of sample was insufficient. The bulk samples thus include a small amount of inorganic carbonate (estimated at 0.5%) and

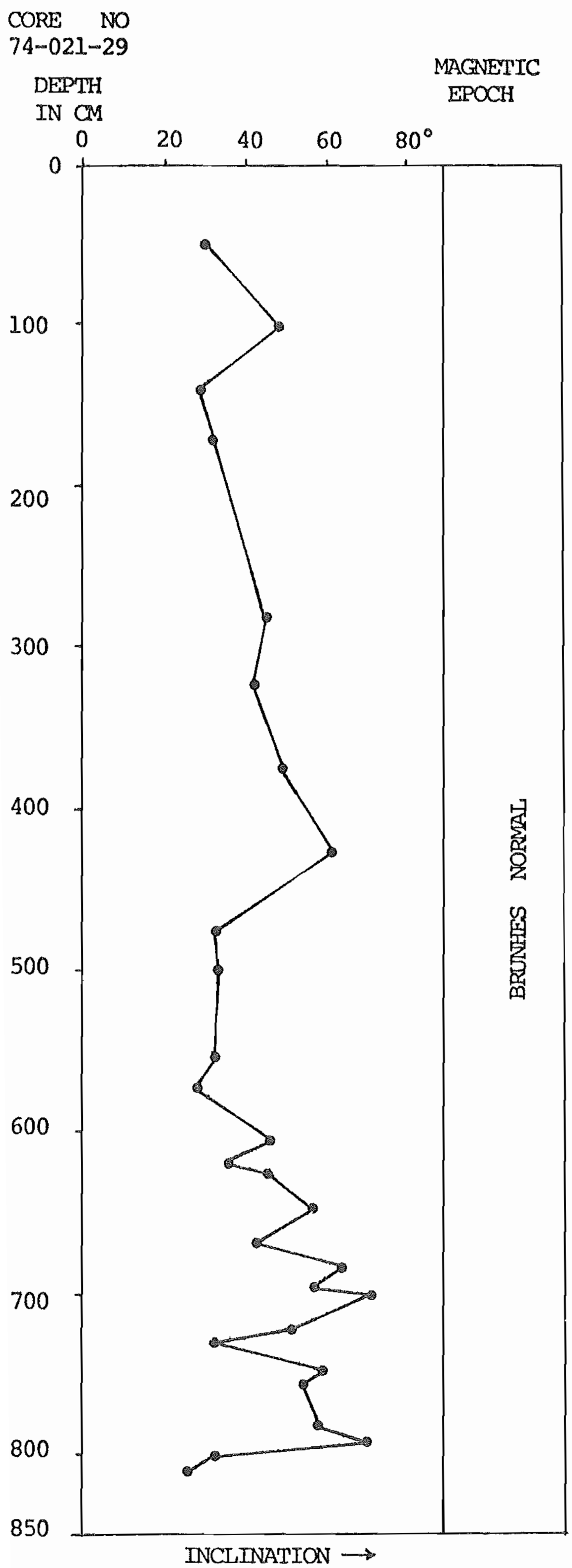


Figure 25: Magnetic Stratigraphy

may also contain occasional reworked older carbonate fossils (estimated at 0.1%).

The date from the upper part of the subunit B2, at 40-45 cm suggests that it is about 20,000 years B.P. and the age of units C and E are beyond the range of the carbon 14 method (see Table 3).

Pa^{231}/Th^{232} dates were not attempted. This method would have been useful to confirm the extrapolated dates in the lower part of the cores. Th^{230}/Th^{232} dates were not attempted since its reliability is questionable especially in the area which has been directly effected by glaciation (Broecker, 1965).

BIOSTRATIGRAPHIC CONTROL

The extinction of _G. m. flexuosa_ in unit C is here correlated with the Sangamon Interglacial. The first appearance of _E. huxleyi_ at the base of unit C also confirms this correlation.

RATE OF SEDIMENTATION

The rate of sedimentation southwest of the Grand Banks is controlled by three factors (1) terrigenous detrital input by turbidites, bottom currents, ice-rafting and hemipelagic processes, (2) biogenic input which is a function of

TABLE 3

CARBON 14 DATES

Core No.	Depth in cm.	Laboratory	Isotope Sample #	Method	$-\delta\ C^{14}$	Age in Years B.P.
74-021-29	258-263	Teledyne Isotope	1-8557	Total Carbonate	>993	>40,000
74-021-29	12-18	Teledyne Isotope	1-8700	Total Carbonate	>991	>37,500
75-009-117	40-45 cm	Teledyne Isotope	1-8986	Total Carbonate	920 ± 4	20,290±410

surface-water productivity (3) removal of biogenic carbonate by dissolution.

All the cores used in this study came from a depth less than 4000 m. Measurements of dissolution on the outer ultra-structure of the foram tests suggest that the effects of dissolution were minor and fairly uniform throughout the entire period. So we can isolate this factor from the other two and ignore its effect on the variation of sediment input.

The biogenic material in our cores is almost entirely calcium carbonate with a very small amount of silicious matter, such as sponge spicules, radiolaria and diatoms, all of which are usually less than 1% of the total biogenic input.

It should be obvious from the previous discussion that the predominate sediment type changes with change in climate. Here we are more interested how much change in the rate of sedimentation might have taken place. Although popular concepts favour a pronounced change in the rate of sedimentation on the continental margin from the glacial Pleistocene to the present, recent studies on all the continental margins around the world suggest that, while the change in some places may be considerable, it was never astronomical. One major reason is that the sediments on the continental shelves are still in disequilibrium with the present environment and the input from shelves is still high in comparison to pre-glacial conditions. It is true that with the rise of Holocene sea-level continental source areas have been pushed back, but the sediments left on the shelf during the last glaciation have much easier access to the outer continental margin. Cores collected from the Laurentian Fan suggest that the rate of sedimentation during the Holocene was about 10cm/1000 years and during the Wisconsin glaciation from 12 to 17 cm/1000 years (Stow, 1975). It can be assumed that the change in the rate of sedimentation on the top of seamounts which are beyond the range of turbidity currents, was much less than on the Laurentian Fan.

Broecker et al. (1958) suggested that in order to determine the rate of sedimentation, the rate of accumulation of each individual component of the sediment has to be determined. If the rate of accumulation of components such as clay, carbonate and sand remains constant, then the overall rate of sedimentation will remain constant. However, it is often possible that the decrease of one of these components is accompanied by an equal increase of another component and the overall rate of sedimentation remains fairly constant.

In order to determine the variation in the input of different components of the sediments each of them were examined throughout the cores.

The first noticable change is the carbonate content of the sediments. While warm stages are marked by high carbonate, the cold periods show much lower carbonate content. There are three possible explanations for this change:

1. Biogenic input was the same both during glacial and interglacial stages, but the effect of dissolution was much higher during the glacial stages. We can reject this explanation since we have already seen that there was not enough change in the intensity of dissolution.

2. Carbonate input was uniform but it has been diluted by high influx of terrigenous material. In order to check this hypothesis, we can relate the input of planktonic foraminifera to the benthonics, assuming that the rate of productivity of benthonics was uniform both during glacial and interglacial stages and that the change in the ratio of these two indicates the change in the input from the surface. Our results suggest (Figures 17, 20, 22) three-hundred to four-hundred percent change in the input of planktonic foraminifera which is quite sufficient to produce low carbonate in the glacial stages. We have already discussed that the terrigenous input even on the Laurentian Fan did not change more than one hundred percent, so it is almost impossible that a change

in terrigenous input can produce such large scale dilution of carbonate.

3. The third possibility is the change in the surface productivity is mainly responsible for this carbonate low. The major faunal assemblage present during the warm stages changed abruptly in the cold stages. Transitional and subtropical foram species were completely replaced by arctic and subarctic species. It is very rational to assume that the productivity was much less when this area was under Arctic and Subarctic water mass. Our benthonic and planktonic ratio also suggest that the surface productivity was low during the cold stages.

Size analysis of the sediments, both acid treated (to remove $CaCO_3$) and untreated samples, suggests that the input of the coarse grained particles was always higher in the warm than the cold stages. In order to determine whether the decrease of coarse size terrigenous particles and biogenic sediments were compensated by an equal rise in the influx of fine grained sediments, the rate of influx of sediments was determined at different level in the core (Figure 26).

Warming up of the climate in the upper part of the cores can possibly be considered to be related with the last termination of ice about 11,000 to 10,500 years B.P.

Since it can be used as a reasonable check of the reliability of Carbon 14 dates, an attempt was made to get a date from the immediately lower warm horizon (unit B2). This date is about 20,000 years B.P. and the middle point of sampling is 42.5 cm in core 75-009-117, which suggests an average rate of sedimentation of about 2.12 cm/1000 years. Again considering the warming up event at 20-22 cm level related with Holocene termination of ice we get an average rate of sedimentation 1.9 cm to 2.0/1000 years. Unit C has been assigned a Sangamon Interglacial age; the highest warm peak of this interglacial stage occurred about 127,000 years B.P. (Broecker and Van Donk, 1970; McIntyre et al., 1972; Ericson et al., 1964), which gives an average rate of sedimentation of 2.03 cm/1000 years for the Post-Sangamon section of the core.

Figure 26 shows the depth in the core having time control. Average rate of sedimentation for the first climatic cycle is about 2.12 cm/1000 years and that during the Interglacial stage 2 cm/1000 years and glacial stage 2.27 cm/1000 years. It seems that though the input of different components of the sediments such as the carbonate, clay and detrital sand varies with time, this variation was approximately reciprocal. When the input of the carbonate and sand decreases, the input of the clay increases. An average rate of sedimentation of 2 cm/1000 years is used.

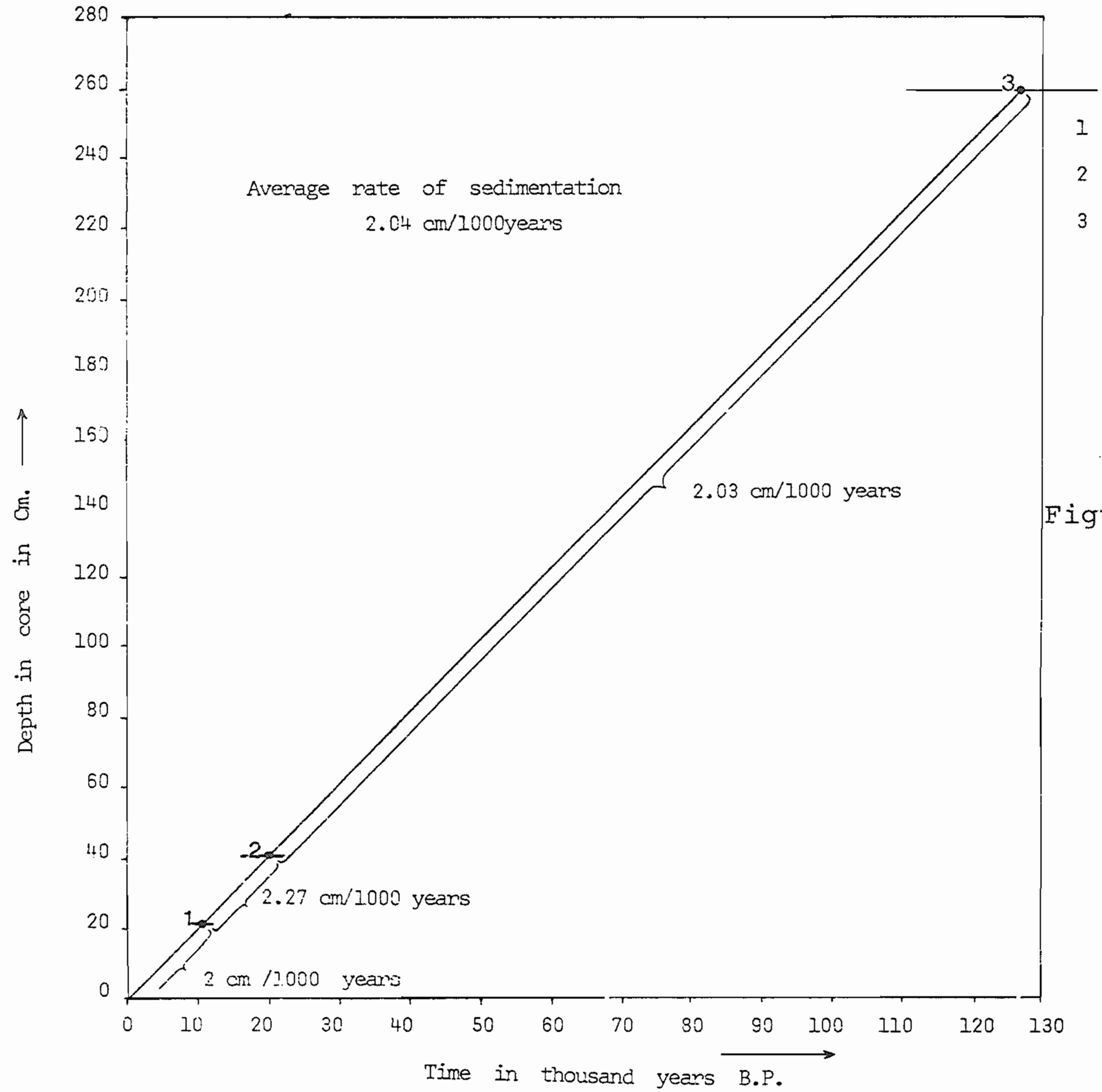
DATING THE CLIMATIC HIGHS

Three very warm intervals are represented in the cores by units A, C, and G. Figure 27 gives the tentative dates of different warm and cold stages based on carbon 14 dates and micropaleontological extinction and controlled by the paleomagnetic time scale as well as the data available from the land and sea.

The last warm event at the top of the cores (unit A) gradually became cooler to a depth 20 to 22 cm in all the cores and is separated from another warm stage unit B2 by a relatively cooler horizon. The carbon 14 date obtained from the second warm horizon (unit B2) has been used to confirm the date at the base of unit A. Considering the other work in the North Western Atlantic (Wollin et al., 1971; Broecker and Van Donk, 1970; McIntyre and Ruddiman, 1972) this date is quite reasonable (see Figure 27).

The absolute date of unit B2 suggests that it is equivalent of the Plum Point Interstadial and is separated from the Holocene by a cooler stage presumably representing the Late Wisconsin glaciation.

Subunit B4 is again a warmer horizon which extends from



1 Holocene warming up

2 Carbon 14 Date

3 Mid-Sangamon, micropale-ontological evidence

Figure 26: Rate of Sedimentation of core 75-009-117.

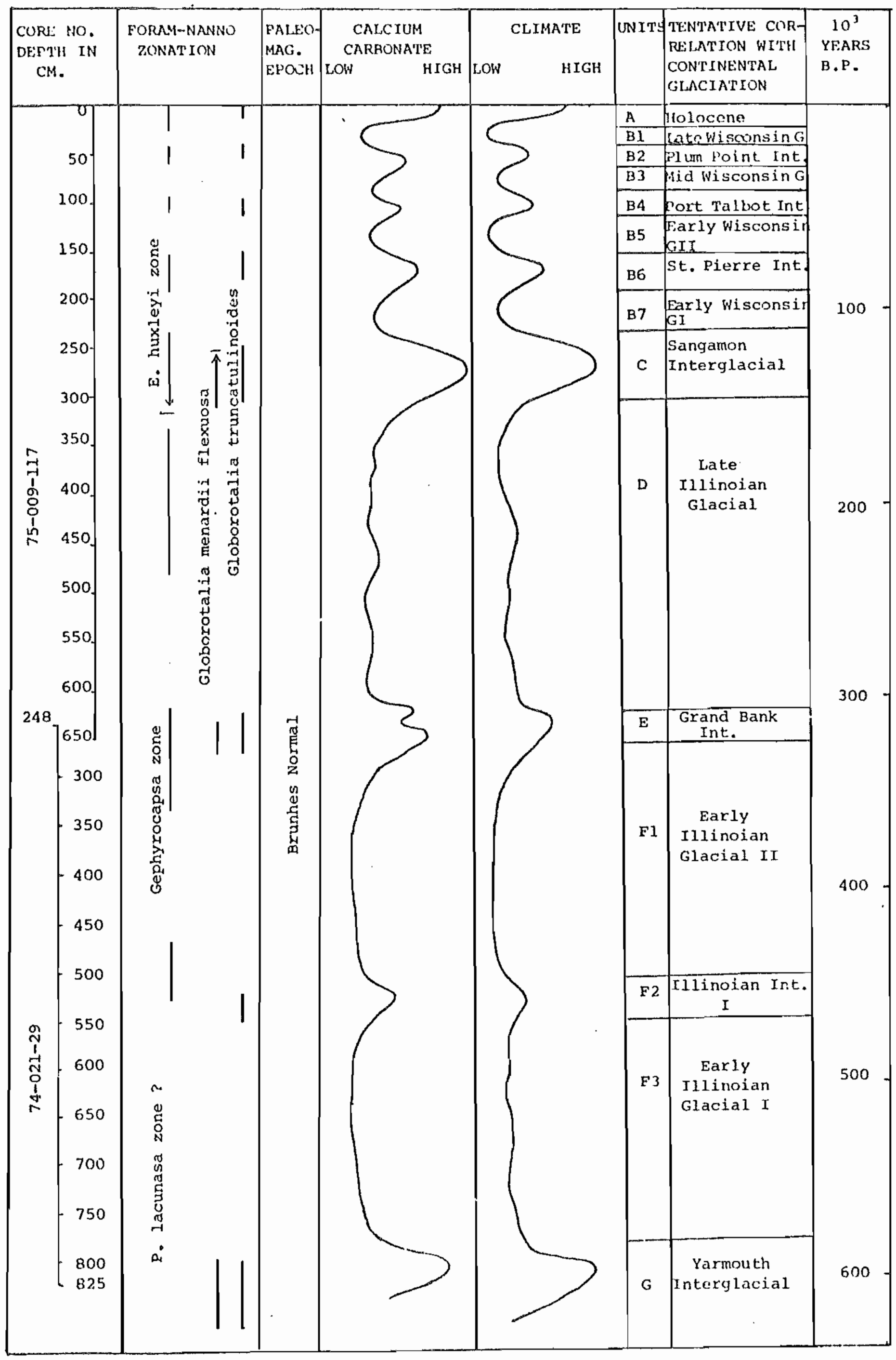


Figure 27: Generalized climatic curve for the studied area and tentative correlation with terrestrial events.

40,000 years B.P. to 50,000 years B.P. and coincides with the Port Talbot Interstadial, separating the Mid Wisconsin from the Late Wisconsin glaciation. Another fairly warm stage in the cores is represented by the unit B6 which is relatively much warmer than any of the other two interstadials and is easily recognizable. An extrapolated date suggests that the range of this interstadial was from 70,000 years B.P. to 90,000 years B.P. The well known St. Pierre Interstadial also falls within the same range and the correlation seems to be quite reasonable. The stratigraphic position of this interstadial is such that is divides the Early Wisconsin into two phases, Early Wisconsin I,which appears to be longer and cooler than the Early Wisconsin II.

Unit C has been correlated with the Sangamon Interglacial by its stratigraphic position and by micropaleontology. Extrapolated dates suggest that this interglacial stage extended from 115,000 to 145,000 years B.P.

The rate of sedimentation for the last 130,000 years was more or less uniform, but in order to extrapolate dates for the lower part of the cores we must be more cautious. The presence of *Globorotalia truncatulinoides* suggests that we are still within the Pleistocene, so the age must be less than 1.8 m.y. Paleomagnetic stratigraphy suggests that the entire section of the core is within the Brunhes Normal Epoch, so our extrapolated dates have to be less than 700,000 years. In order to check these requirements, I will first attempt to date the lower foram-nanno ooze horizon, unit G, which the foram assemblage shows is at least as warm as the present. An extrapolated date assuming a constant rate of sedimentation suggests that this warm event has a 600,000 years B.P. which is equal to the age of zone T of Ericson and Wollin (1968), which they correlated with the Uarmouth Interglacial. Since the data from the land suggest that the Kansan glaciation ended at about 700,000 years B.P. (Richmond, 1970) and was followed by the Yarmouth Interglacial, I have tentatively correlated the unit G with the Yarmouth Interglacial. Ericson and Wollin (1968) found a magnetic reversal towards the base of their zone T, but our cores did not penetrate below the foram-nanno ooze horizon. The second important consideration is whether this reversal should have occurred within this foram-nanno horizon. If it is equivalent to zone T, then it is reasonable to expect a magnetic reversal. But it may not be there simply because the phase of the glacials at lower latitudes will be felt at a later date than in higher latitudes. Even for the Holocene-Pleistocene boundary, Ruddiman and McIntyre

(1973) found that the relation is time transgressive between the equatorial Atlantic and higher North Atlantic.

Olausson and Svernonius (1975) suggested that reversal of the earth's magnetic field may be triggered by rapid glaciation and deglaciation. If this is true then Brunhes-Matuyama boundary which has been dated as approximately within the Yarmouth Interglacial (Ericson and Willon, 1968) would have occured either at the beginning or at the end of the Yarmouth Interglacial.

Moreover, the boundary drawn by Ericson and his associates between zone T and zone S is slightly arbitrary, since they do not have sharp changes in the climatic conditions in the Equatorial Atlantic which could be used to pick up this boundary confidently and the recent dates from land (Richmond, 1970; Cooke, 1973) conflict with this boundary between the Kansan and the Yarmouthian.

The next warm stage recorded in our cores is subunit F2 which appears at approximately 440,000 B.P. The intensity of this warm stage is more or less like the Plum Point and Port Talbot Interstadial. I will refer to this interstadial as Illinoian Interstadial I. It divides a cold period which I believe is Early Illinoian, into two parts--Early

Illinoian I and Early Illinoian II. Climatic data suggests that the Early Illinoian I was severe for longer duration than the Early Illinoian II. The next warm event we encounter is Unit E. It is very well defined datum between the lower gray mud and the red mud. Climatic data suggests that this horizon is fairly warm but slightly less warm than the St. Pierre Interstadial. An extrapolated date suggests that this horizon has an average age of about 310,000 years B.P. Since this interstadial played a very important role in the sedimentation South of Grand Banks, I will refer to this interstadial as the Grand Banks Interstadial. The next cold stage is represented by unit D, and according to its stratigraphic position and age I would correlate it with the Late Illinoian glaciation. Unit D is fairly cold but less so than subunits F1 and F3. Although it shows certain variation in the climatic conditions, it was never warm enough to record an interstadial condition.

COMPARISON WITH OTHER OCEANIC DATA

The sediments on the continental margin are believed to document the climatic change of the continent more accurately than those in the mid-ocean areas. Mid-ocean climatic signature is complicated by several factors, including

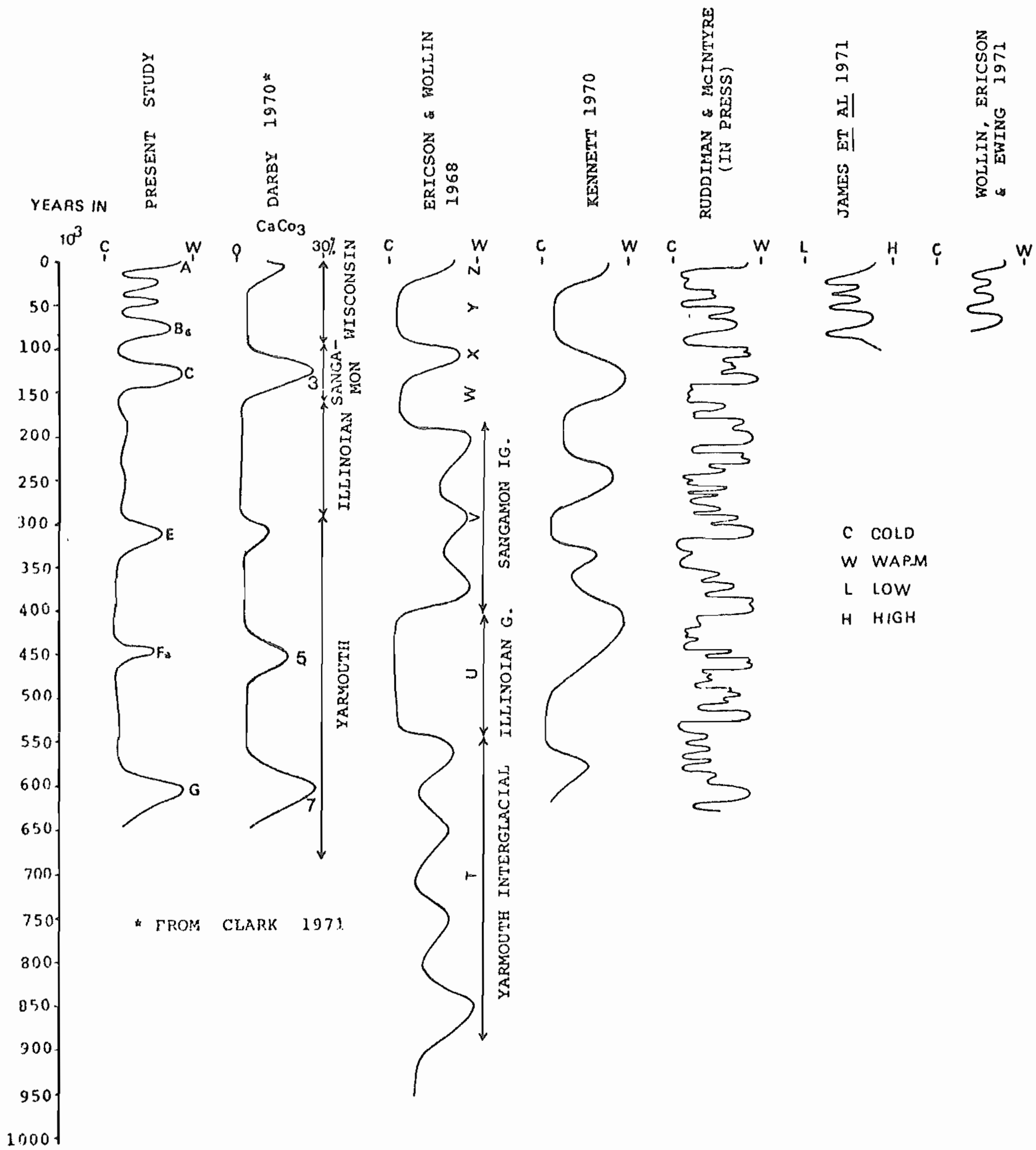


Figure 28: Comparison and proposed correlation with other oceanic curves.

the shifting of the Arctic convergence and the Equatorial divergence. With one glacial advance and retreat they will pass through the same latitude twice. Ocean circulation during glacial and interglacial stages changed quite considerably and with our present knowledge of mid-ocean sediments, it is difficult to quantify the effect of such changes. From this point of view continental margin sediments are less complicated but again they are subjected to considerable change in the process of deposition and in the influx of sediments.

The generalized paleoclimatic curves of South of Grand Banks have been compared with other deep-sea climatic curves (Figure 28). The comparison is based more on the time-stratigraphy, but due to lack of confirmed time horizon at the base of our cores, some of these correlations are open to discussion. Most of the curves span through the last 600,000 years B.P. They are redrawn and in some cases have been stretched in length. The curve by Darby (1970) is based on $CaCO_3$ in the Arctic Ocean. Ericson et al. (1968, 1971), and Ruddiman and McIntyre (1976) used foraminifera to reconstruct the paleoclimate in the Equatorial Atlantic and in the Polar Front area of the North Atlantic respectively. James et al. (1972) data are from the sea-level in Barbados.

The most obvious correlation of the upper part of our paleoclimatic curves is with those of Ericson *et al*. (1970) and James *et al*. (1972). Three Wisconsin Interstadials documented by them match with our climatic highs. They suggested slightly different ages for these interstadials, but approximately their dates are 25,000, 40,000 and 65,000 years B.P. for the Plum Point, Port Talbot and the St. Pierre interstadials. Our dates are slightly different from theirs --22,000, 45,000 and 75,000 years B.P.

There are two possible interpretations of the lower part of our cores. Assuming that a constant sedimentation rate was maintained, as shown in Figure 26, then our climatic curves shows an impressive correlation with Darby's (1970) curve which is dated by magnetic stratigraphy. His high $CaCO_3$ at 127,000 years B.P. almost perfectly correlates with our unit C. The three climatic highs and his carbonate highs appears almost at the same time horizons. However, Darby (1970) correlates the whole section below the 300,000 years B.P. datum with the Yarmouth Interglacial whereas we consider that only unit G represent the Yarmouth Interglacial.

If the constant sedimentation rate does not hold for the lower part of the core then other correlations are possible. The extinction of *G. m. flexuosa* in zone X and unit C, the apparent lack of *P. lacunosa* in the lower part of the cores, and the absence of magnetic reversal in unit G, all possibly demand an alternative correlation. These indicators however, may all be latitude dependant, as discussed earlier.

If we consider that average rate of sedimentation was slightly different for the glacial and interglacial stages, we can work out a correction factor. Our sedimentation curve (Figure 25) suggests a difference from 2 cm/1000 years to 2.27 cm/1000 years for the first glacial cycle--an increase by about 13% for the glacial stage. To be more realistic we can consider this variation to be about 25%. Applying a correction factor of ± 25% for the base of the core we get an approximate date of 400,000 years for unit G. This suggests a correlation of units E, F and G with three warm peaks of zone V and unit C with zone X. However, this correlation results in a poor match of interstadials and interglacials between our cores and Ericson _et al's_ curve. Moreover, if we consider the water content of our cores (Figure 14) we need to apply a higher compaction factor for the lower part of the cores which may well balance small variations in sediment input. However, confirmation of our dates will require examination of at least a few more cores.

The upper part of our climatic curve can be correlated with the curves of Ruddiman and McIntyre (in press), but the correlation for the lower part is rather poor.

In conclusion, our climatic curve shows interesting similarity with work in the Arctic Ocean, and with work on land of Eastern Canada, as well as with the work in the Mid-Continent (Richmond, 1970), but there is considerable amount of discrepencies with work in mid-ocean areas, especially in the Equatorial Atlantic and in the Polar Front area of the North Atlantic. It is not clear from this study whether some of these differences are due to preferential penetration of the Polar water mass towards low latitudes, or due to the fact that the climatic signal from the continent gets distorted as one moves too far away from the continents. Additional cores are required to resolve this problem as well as to complete the whole climatic record of the Quaternary on the continental margin of Eastern Canada.

CHAPTER VII

DISCUSSION

The ocean's response to climatic change on the continent may be more complicated that in usually assumed. The ocean has a vast heat storage capacity, as a result there may be a lag in its response to change of continental climate. With climatic change, there is a change of the ocean circulation, a change of the carbonate compensation depth and change in the position of the biogenic productive zones. All these factors complicate the data from the open ocean and its bearing on the climate of the continents.

On the continental margin the pulses of ice advance and retreat are recorded much more directly than in the open ocean. The area south of the Grand Banks is especially interesting because it is in the climatically sensitive mid-latitudes. The change in ocean circulation of the area can also be reasonably reconstructed. During warm periods it is dominated by the Gulf Stream and during cold periods by the Labrador Current. Furthermore, there has been a considerable amount of work in the Mid-Atlantic, at the same latitude, which can be used for comparison.

Sediment records from the seamounts from the outer continental margin off the coast of Newfoundland suggest that the continental ice sheet terminated three times during the last 600,000 years. The term "termination" of an ice sheet is being used here to indicate major deglaciation, having an intensity at least equal to the present. Major deglaciation in Eastern Canada took place at about 11,000 years B.P., 130,000 years B.P. and 600,000 years B.P.

In between these terminations, the ice sheet partially retreated several times. The most recent partial retreat of the ice sheet is recorded in our sediments by an interstadial about about 22,000 years B.P., which we have correlated with the Plum Point Interstadial.

It was preceded by an intense glaciation and then by another interstadial at about 45,000 B.P., which has been correlated with the Port Talbot Interstadial. The climatic conditions during both Plum Point and Port Talbot Interstadial were much cooler than at the present. Although the ice partially retreated, it is believed that it still covered large parts of the continent, and the ocean circulation was not completely restored to a non-glacial pattern. The Early Wisconsin stage is divided into two parts by a very prominent interstadial which peaked at about 70,000 B.P. and has been correlated with the St. Pierre Interstadial. This interstadial is much warmer than the two younger interstadials, being only slightly cooler than the present.

The next major deglaciation took place at 130,000 B.P. and has been correlated with the Sangamon Interglacial. The climatic record for the last 130,000 years B.P. can be easily correlated with the known stratigraphic record on land and, with some adjustments, with the open ocean. Before the Sangamon Interglacial, our data suggests a relatively cold climate until the Grand Banks Interstadial, at 320,000 B.P. This was preceded by an intensely cold climate interrupted by a interstadial at about 430,000 B.P.

The earliest evidence of deglaciation that we found was at around 600,000 B.P., and we consider that event equivalent to the Yarmouthian Interglacial.

Our data suggests close correlation with the known glacial history of Eastern Canada and with work in the Arctic (Darby, 1970): in other words with the areas close to our's and receiving the direct influence of glaciation. Our data shows significant differences from some of the well known open ocean work, even at the same latitude (Ruddiman and McIntyre, in press). We did not find convincing evidence of a very warm climate at 200,000 and 400,000 years B.P. as suggested from the study of some open ocean cores. It is possible that some of the differences are due to the dating technique, and future work may require readjustment of our time scale, especially for the lower part of the cores. There was no evidence of large scale variation in the rate of sedimentation and it seems that only small adjustment of the extrapolated dates may be required.

The type of sediment changes with the change in climate. Two sedimentological parameters supplement our climatic interpretation: the carbonate content of the sediments, and the presence of ice rafted pebbles. Both of these parameters are quite variable and cannot be applied on a world wide scale unless their regional significance is properly evaluated.

It has been known since the work of Arrhenius (1952) that the input of calcium carbonate to deep-sea sediments is influenced by climatic change. However, the variation shows different patterns in different oceans. Arrhenius found that high calcium carbonate in the Equatorial Pacific represented the glacial stages. Olausson (1967) reported that in glacial times calcium carbonate was higher in the Pacific but lower in the Atlantic ocean, which he ascribed to a massive net transport of carbonate from one ocean to the other ocean. Ruddiman (1971) suggested that the net carbonate deposition probably increased in both equatorial Atlantic and Pacific during glacial stages, but that the input of lutite from Africa increased during the glacial stages which diluted and overwhelmed the increase of carbonate deposition in the Atlantic but failed to do so in the Pacific.

McIntyre (1967) reported high carbonate in interglacial stages at high latitudes. McIntyre *et al*. (1972) suggested that a glacial stage in high latitudes is characterized by a coccolith carbonate minimum. In intermediate latitudes the maximum productivity zones move through the area twice during migration in a single glacial cycle and leave two maxima. Low latitudes are marked by a single carbonate maximum (Figure 27).

In order to evaluate how the carbonate cycle of the outer continental margin off the coast of Newfoundland changed with the climate, the percentages of calcium carbonate in the cores have been compared with foraminiferal data (Figure 17, 20, 22). There is a marked correlation between the climatic highs and carbonate highs, with an increase in the amount of calcium carbonate, with increase of temperature. None of the cores penetrated pre-glacial sediments; as a result a direct comparison with pre-glacial sediments is impossible. The pronounced decrease in carbonate during glacial stages may result from any of three causes: (1) change in the biogenic input, especially due to change in the productivity of microfossils (2) increase in the detrital input (3) increased carbonate dissolution.

It has been shown that changes in productivity and detrital input are the most important reasons for this carbonate variation. With cold climate and lower sea level the detrital input increased whereas with warm climate and higher sea level the biogenic productivity was higher.

Although the Newfoundland continental margin is in mid-latitudes, the suggestion of McIntyre et al. (1972) that in the mid-latitude a single glacial cycle leaves two carbonate maxima (Figure 29) is not valid for this area. The continental margin itself is a highly productive belt and this belt moves towards the deep sea with the advance of a glacial and returns to its original position with the termination of the continental ice sheet. As a result at the continental margin glacial advance is recorded by a carbonate low, but in the deep sea moving of this belt during a single glacial advance may produce two carbonate highs. However, when considering a case like this as well as the one illustrated by McIntyre et al. (1972) one has to be careful whether the productive zone while passing through an area will produce a significant difference in carbonate sedimentation. In order to produce any remarkable difference the productive zone has to be stable over an area for a considerable amount of time to overcome the effect of dissolution and the dilution by detrital materila both of which may vary with the change in climate.

The quantity of ice-rafted pebbles was also used as a possible climate indicator. The relationship between the

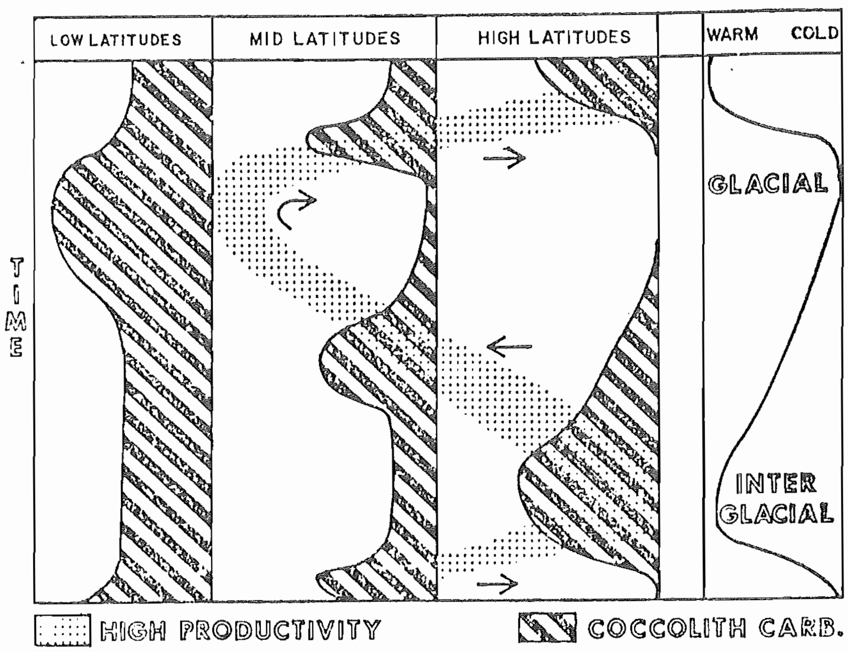


Figure 29: Theoretical migration sequence of the Coccolithophorida high productivity zone through an interglacial-glacial climatic cycle with attendant effects on the concentrations of the less than 74μ carbonate in cores underlying the zone of migration. (McIntyre *et al*., 1972).

rate of ice-rafted sedimentation and climate is complex, as discussed by Kent et al. (1971), Piper (1976), Fillon (personal communication). For example, whether bergs are calved from ice shelves or from glaciers will control the amount of sediment that they carry. Most authors will, however, agree that position of the Arctic (or Antarctic) convergence and the number of icebergs calved play a critical roled in ice rafted sedimentation in the higher latitudes. If we consider the other factors are less important, then the ice rafted sedimentation on the south of the Grand Banks are mostly controlled by the position of the "cold wall", (Hachey, 1961), especially the contact between the Labrador Current and the Gulf Stream and the number of icebergs reaching the Tail of the Bank. The "cold wall" in the neighbourhood of the Tail of the Bank is subjected to considerable variation in position due to relative strenghts of the Labrador Current and Gulf Stream. In the winter months with the increased flood of cold Arctic water the "cold wall" is pushed towards the south (Hachey, 1961).

Recent work by Clark (1971), Herman (1970), and Hunkins et al. (1971) suggest that the amount of ice rafted sediments in the higher northern latitudes decreased during the glacial stages. Most of these workers attribute this to the increase of the extent of ice sheet during glaciation and closing of open water passages in the Arctic region. The condition is to some extent the reverse of conditions in lower latitudes, where the amount of ice rafted material increased during the glacial stages. We do not have any data whether the Davis Strait or Labrador Sea was covered by ice during the glacial stages. Under present climatic conditions the stream of ice from the Arctic region is most voluminous and extends farthest southward during spring and early summer than the winter or late summer (Hachey, 1961).

The changes in position of the "cold wall" from summer to winter and increase in the frequency of ice in the spring and early summer provides a guide to changes between glacials and interglacials. We can assume that with a decrease of temperature, there was a southward extension of the cold water mass. As a result icebergs which now melt near the Tail of the Banks would melt further south of the area studied during cold periods. Moreover, the frequency of icebergs was possibly also less, the result was decrease in the total amount of ice rafted sediment on the south of the Grand Banks during the glaical stages. With the warming up of the climate, as the ice sheet would start to melt there would be a sudden increase in the number of icebergs on the south of the Grand Banks. As temperature increased again the Gulf Stream would move closer to the Grand Banks so that melting of these icebergs would be accelerated. As a result one can expect to find large numbers of ice rafted pebbles documenting the first phase of warming up, but as the climate became warmer and warmer, the number of icebergs will decrease and so will the number of ice rafted pebbles.

All three of the climatic criteria we have used--micropaleontology, carbonate cycles and ice rafted materials--show very good correlation with one other. The sequence of climatic change appears to be reliable. The time frame-work for the lower part of the cores is less well established and may have to be revised as more data (radiometric and paleomagnetic) becomes available.

With the change of climate at least five variables changed drastically, which in turn influenced the type and amount of sediments on the outer continental margin south of the Grand Banks. A model is developed considering the effect of these variables on the sedimentation. These are

(1) Lowering of the sea-level
(2) Advance and retreat of ice and sheet on the continent
(3) Opening and closing of the Avalon Channel
(4) Southward penetration of the cold water front
(5) Change in the intensity of the Labrador Current and the Gulf Stream.

(1) During glacial maxima there was a considerable drop in sea-level, sometimes about 130 m (Emery, 1969), which ex-

posed vast parts of the continental shelf above sea-level. Again during most of the Wisconsin Interstadial sea-level was around 50 m lower than the present (Fairbridge, 1966), with a possible exception being the St. Pierre Interstadial. During the last Sangamon Interglacial sea-level was possibly 10 m higher than the present (Shepard and Curray, 1967). We have to limit our discussion about differential effect on sedimentation, of having different sea-level, since we do not have almost any data on the absoluate sea level from the Atlantic Canada. Again the Quaternary sea level is complicated by several other factors except climatic change. Bloom (1971) discussed the difficulties of evaluating the Quaternary sea level without completely understanding the sea floor spreading, plate tectonics, isostasy and regional tectonics.

A lowering of sea level by 100 m or more would expose almost all of the Grand Banks to the atmosphere. Part of this area could have been covered by glacial ice. Lowered sea level would have influenced the sedimentation on the outer continental margin in two different ways. The ultimate source of most of the deep-sea clastic sediments is land, and the coast forms essentially a line source of terrigenous sediments. As a result shifting of the shore line plays a very important role in the input of sediments on outer continental margin. In the case of the Grand Banks a drop of sea level by 100 m would shift the coast line a few hundred kolometers seawards. It would cut down the cross shelf dispersal time for fine-grained sediments. With the coast line near the shelf break, sedimentation would be rapid and direct onto the outer continental margin. Glacial melt water streams flowing across the shelf would debourch large amounts of sediments directly into the ocean basin, through narrow valleys incised into the outer shelf and upper slope. These features are preserved today as numerous submarine canyons around the Grand Banks. Since the breaker zone of the waves would have been very close to the head of the canyons, large amounts of sediments could have been transported directly into the deep sea by rip-currents. Considerable amounts of fine-grained sediments could have been put into suspension near the shelf break by the waves. A drop in sea level would make the sediments at the shelf edge and on the continental slope hydrostatically more unstable, so that slumping and resulting turbidity currents would be more common. Possibly the rise of sea level would also promote slumping because the pore pressure of the sediments is not in equilibrium with the new hydrostatic pressure. Lower sea level would act as a barrier to icebergs from the Labrador Sea. A lowering of 100 m would almost close the Avalon Channel, diverting bergs towards the Tail of the Banks, mostly through Flemish Pass.

A 50 m drop in sea level would leave most of the Grand Banks still under water and the Avalon Channel open. This would accelerate the input from the shelf since the effect of storms on the shelf sediments would increase. Moreover, sucn an intermediate sea level would partially prevent the ice sheet from extending from Newfoundland over the Grand Banks.

(2) The change in climate controlled the advance and retreat of ice sheets on the land. The direction from which ice sheet advanced determined the type of sediments whereas the extent and nature of the ice sheet controlled the amount of sediment input. The lowering of sea level was also controlled by the extent and thickness of the ice. It has been suggested that the major lowering of sea level is mostly related to the development of the Laurentide ice sheet because of its large size compared with other Pleistocene ice sheets except the polar ice sheets. However, such generalizations can only be approximate.

Whether the ice sheet from Labrador ever over rode the Atlantic provinces is not completely known. Flint (1951) and Murray (1955) envisioned an extension of Labrador ice onto Newfoundland prior to the development of a discrete island ice cap. However, recent workers (Grant, 1969, 1972, and Brookes, 1970, 1974) feel that Labrador ice may have only invaded the extreme northwest tip of the island. Well developed local ice caps in Newfoundland and Nova Scotia effectively prevented the growth of Laurentide ice in the Wisconsin (Prest and Grant, 1967; Grant 1973). They had further control over the isostatic lowering of the sea level and possibly redistributed the sediments already present in Nova Scotia and Newfoundland, in most cases pushing further towards the sea as indicated by the moraine complex on the Scotian Shelf (King, 1969).

(3) Changes in ice distribution and sea level would affect the opening and closing of the Avalon Channel as mentioned earlier. This channel carries one branch of the Labrador Current, and probably plays an important role in the sedimentation on the south of the Grand Banks. The deep water prevents seaward transport of coarse sediments and the Labrador Current flowing through this channel advects fine sediments to the continental margin Haddock Channel, restricting fine sediment input to the Grand Banks.

(4) During glacial stages the water mass off the coast of Newfoundland would become intensely cold throughout the year, because of the presence of an ice sheet on the continent and the southward penetration of cold water front from the Arctic. This would drastically change the ecologic conditions of this area. At present this continental margin is a very highly productive zone and biogenic material is a

major sediment component, in some cases predominating. This biogenic component is mostly foraminifera and coccoliths. During the Pleistocene glacial stages, biogenic productivity was greatly reduced and the detrital input from the land increased. As a result the pelagic biogenic sediments were almost completely replaced by detrital sediments.

(5) As the climate changed the ocean circulation also changed. As the climate became cooler, the Norwegian Sea Overflow intensified and consequently the intensity of the Western Boundary Undercurrent increased. At present considerable amounts of fine-grained suspended sediments are being transported by the Western Boundary Undercurrent (Heezen and Hollister, 1972). During the glacial stages both the erosional and depositional ability of the Western Boundary Undercurrent would have increased.

The intensity of the Labrador Current would also have increased during glacial stages and it would have pushed the Gulf Stream further towards the southeast, away from the Grand Banks (Pastoret *et al.*, 1975). As a result, any sediment input from the Gulf Stream on the south of the Grand Banks would be greatly reduced. The sediment contribution from the Labrador Current would also change. At present the major sediment contribution from the Labrador Current is mostly ice rafted material and some fine-grained suspended sediments. During the glacial stages the number of icebergs passing over the Newfoundland Ridge was possibly greatly reduced; moreover, due to southward penetration of the cold water front the icebergs were melting further south. So the amount of ice rafted sediments on the south of the Grand Banks were substantially reduced.

The predominent sediment on the top of these seamounts is clay and understanding the processes of sedimentation of clay plays an important role in evaluating the Quaternary history of this area. It appears that the clay was mostly deposited from suspension. Presumably most of these fine-grained sediments came with glacial melt water and a part is wave suspended sediment from the shelf break; and formed a plume at the outer continental margin (Heezen and Hollister 1972). This plume could have been fed by several other processes. It is possible that considerable amounts of sediment were trapped from the tails of turbidity currents at the density interface in the ocean especially as the currents approach the break between the slope and rise (Komar 1972). Further because of higher suspended sediment concentration near the shelf edge, a very low density current flows along the continental slope and rise. This current is frequently called lutite flow (McCave, 1972) and as it reaches a density interface of the ocean, a part of the sediment is trapped and moves along the interface. Cascading cold water

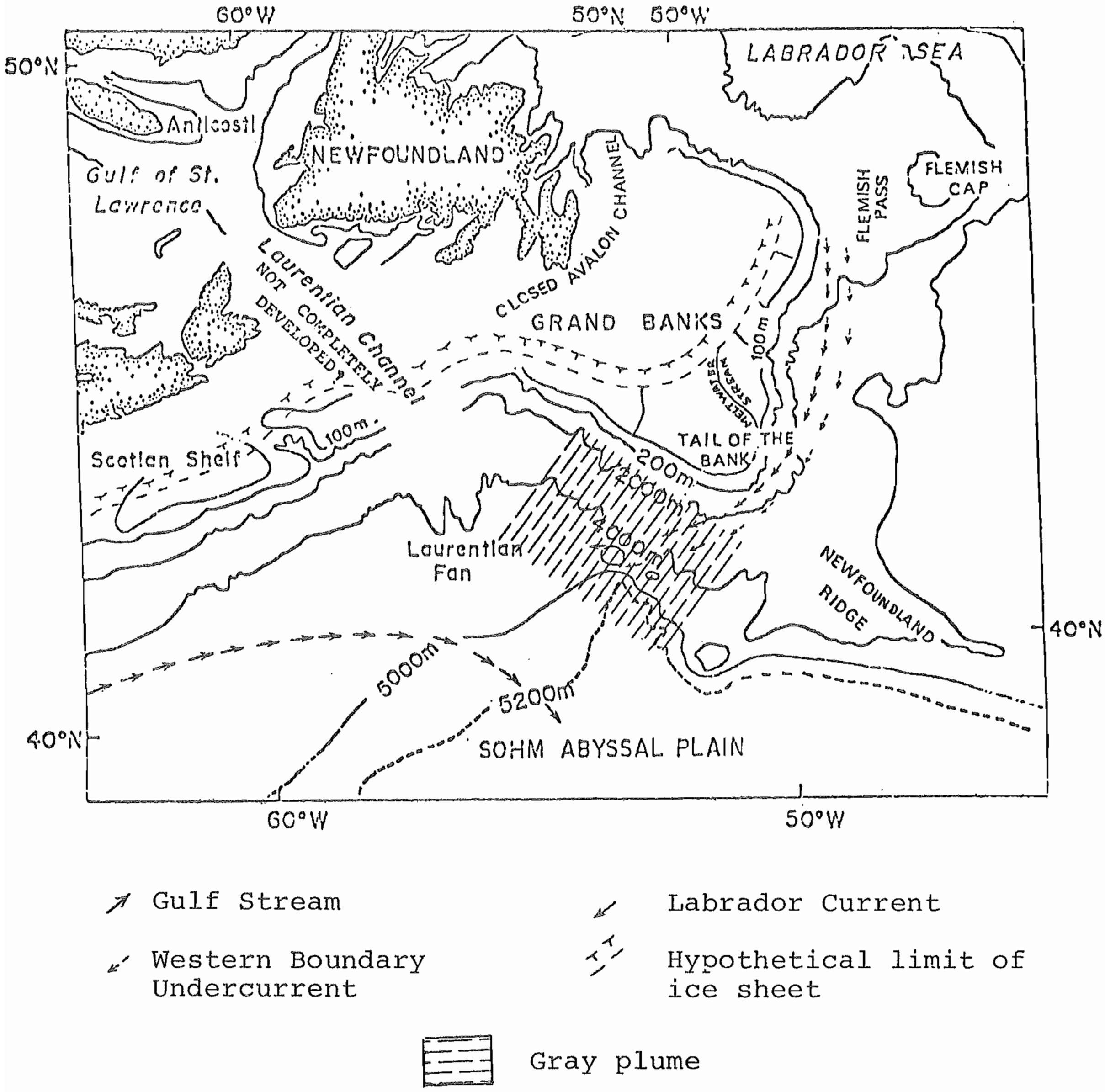


Figure 30: Hypothetical sedimentation pattern of Early Illinoian (unit F3) with closed Avalon Channel and ice sheets covering large part of Grand Banks and Scotian Shelf. Presumably the Laurentian Channel was not well developed and consequently discharge of water and sediments through it was low.

at the shelf edge is also likely to feed the system in the same way. It is possible that the bands in the clay formed due to settling of sediments from different density interface.

This study does not have direct data on the advance and retreat of glaciers or on their areal extent. Since the paleoclimatic data correlate with sediment data indicating ice advance and lowering of sea level, we can assume a correlation of cold/warm in our cores with ice advance/retreat on land. We can indirectly measure the intensity of glaciation by the intensity of cold documented in our cores and by the duration of cold episodes. The relative extent of the glaciers can be inferred from our cores by the types and amount of sediments contributed from the ice.

The paleoclimatic data suggests that during the Early Illinoian glaciation (unit F), the climate was very cold and it lasted much longer than any subsequent glaciation. Clay mineralogy and the mineralogy of the sand and silt indicate that sedimentation on the south of the Grand Banks at that time was controlled by the advance of the Labrador-Newfoundland ice sheet. "Newfoundland type" illite-chlorite rich gray clay was deposited during this stage. Surface textures of the quartz grains, as well as the feldspar/quartz ratio and the ratio of angular/rounded quartz grains suggest that the source of the sediments was very close, and the sediments had undergone very little reworking before final deposition. It appears that the ice sheet at that time covered more than half of the Grand Banks and had completely closed the Avalon Channel (Figure 30). The increased intensity of the Western Boundary Undercurrent is reflected by the scouring at the tops of unit G and by the dragging effect on some granules in unit F.

After the Grand Bank Interstadial (unit E) the ice sheet readvanced in the Late Illinoian (unit D), but this time it appears that the Laurentide ice sheet was controlling the pattern of sedimentation. The clay mineralogy and the petrology of the pebbles suggests that most of the sediments came through the Laurentian Channel. The discharge through the Laurentian Channel during this period was presumably very high. It appears from the presence of "Laurentian Channel type" red clay and occasional ice rafted pebbles in our cores, that the glacial melt water, rich in suspended sediments and sometimes with icebergs, covered a large part of the continental margin off Atlantic Canada. The Avalon Channel was possibly still closed because otherwise this "red plume" (Heezen and Hollister, 1972) would have been diverted towards the southwest by the Labrador Current flowing through this channel, and the area south of

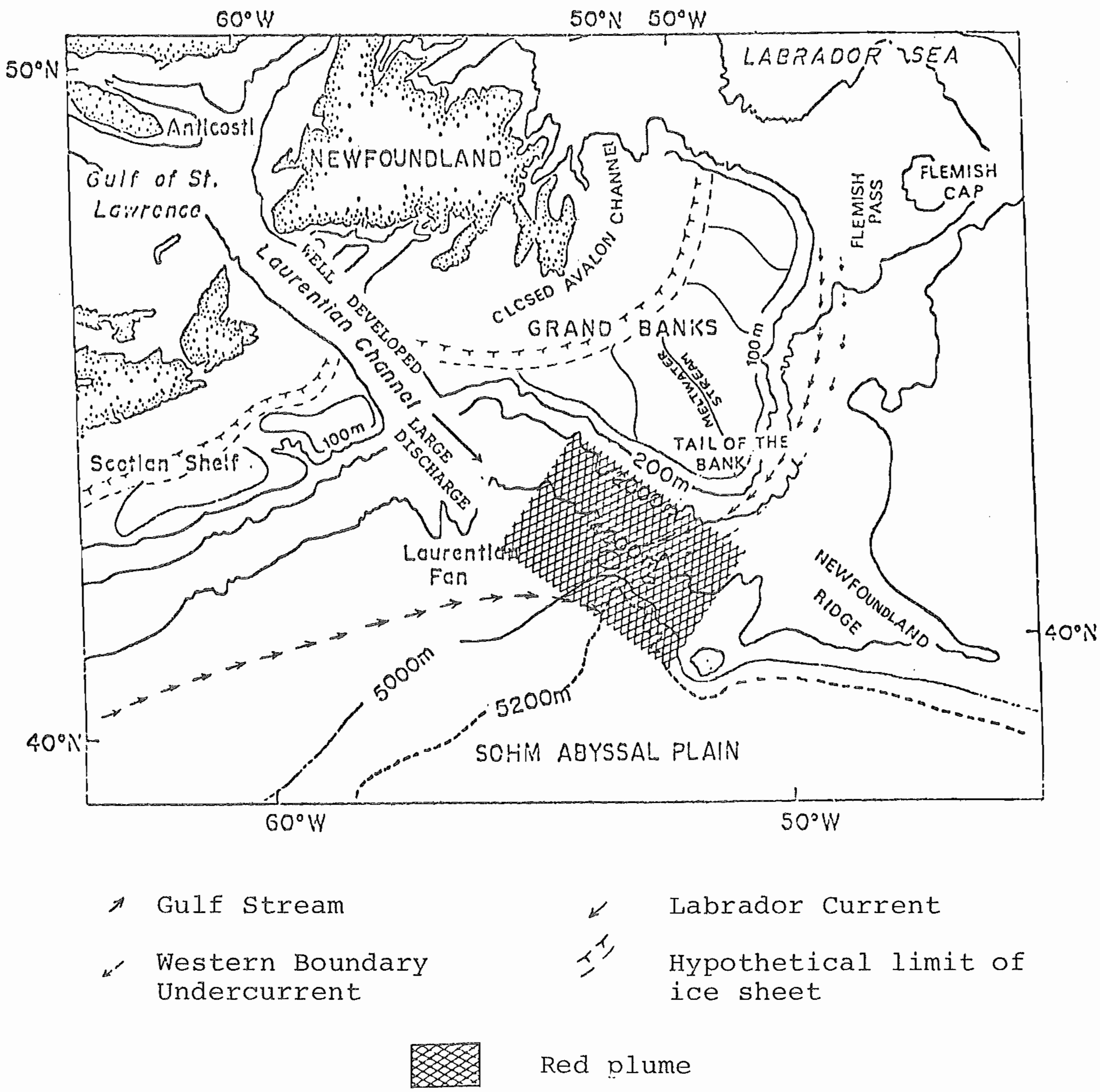
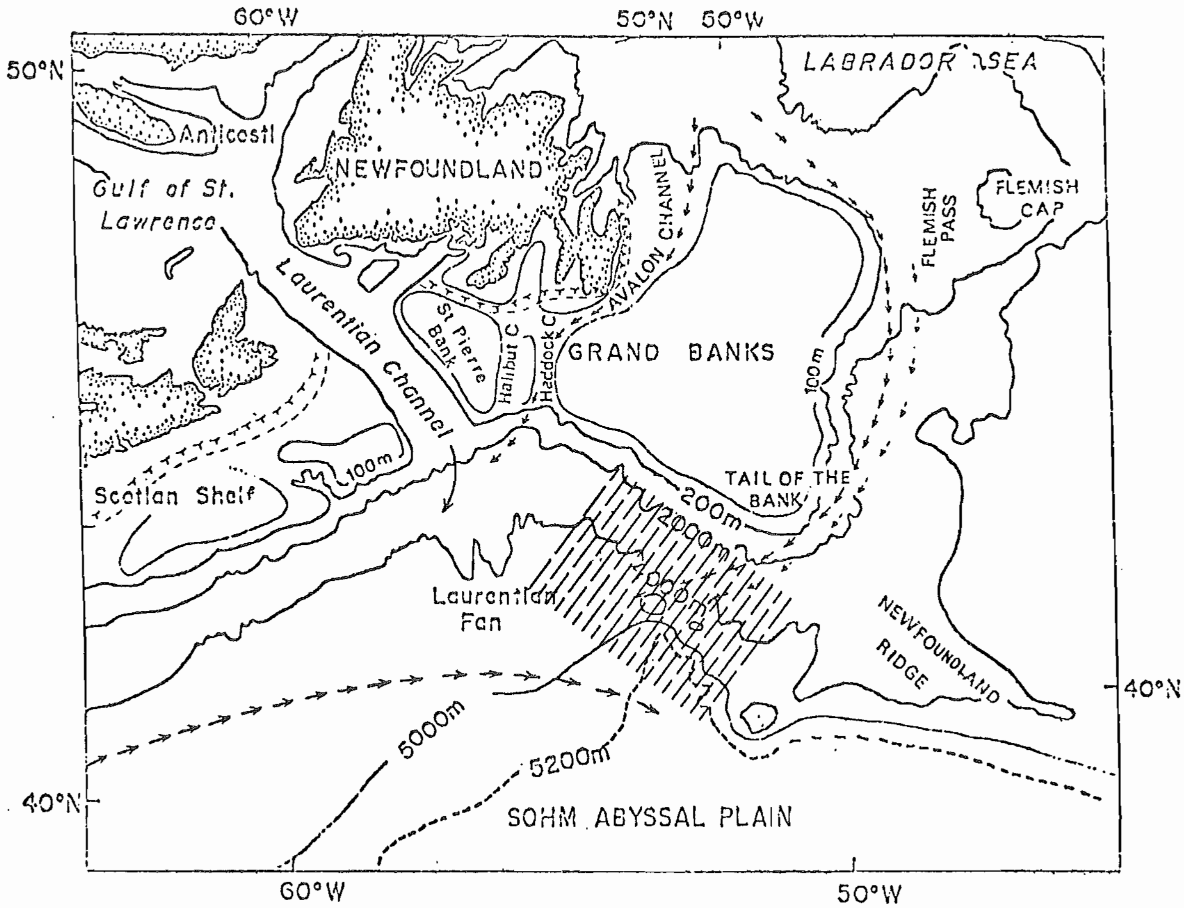


Figure 31: Sedimentation pattern in Late Illinoian (unit F1) after Grand Banks Interstadial, with very active Laurentian Channel, but closed Avalon Channel and ice sheets still covering large part of Grand Banks.

the Grand Banks would remain free from its influence (Figure 31). Furthermore, the direct input of sediment from the Newfoundland ice was possibly still substantial, as indicated by mixed clay mineralogy and the fresh appearance of the silt and sand fraction. An open Avalon Channel would reduce sediment supply from Newfoundland. Our inference that the Labrador Current was not very strong during this period is supported by the fact that a strong Labrador Current coming through the Flemish Pass would have kept the area around the Tail of the Bank free from the Laurentian Channel icebergs and from the red sediments.

Eastern Canada was possibly completely deglaciated in the Sangamon Interglacial (unit C); our paleoclimatic data suggest that during this interglacial Eastern Canada was at least as warm as the present. After that, when the continental glacial reappeared in the Wisconsin they seem to be mostly in the form of local ice caps. Our data do not suggest the presence of a large ice sheet on the continent during this period; at least we do not have large contributions of sediments from such an ice mass. Slatt (1974) and Müller and Milliman (1973) concluded that there was a general lack of evidence of Late Wisconsin glaciation on the Grand Banks. If small ice caps were present on the outer shelf, they would have redistributed the sediments already present on the shelf, and accelerated the sedimentation on the outer continental margin. Several lines of evidence suggest that the Grand Banks was free from the continental ice sheet and the Avalon Channel was open throughout the Wisconsin glaciation. First, the surface texture of sand grains from the Wisconsin sediments suggest that most of them were reworked before deposition. This is possible only when the direct input of Newfoundland ice was blocked by the Avalon Channel. Second, the discharge through the Laurentian Channel was a very important source of sediment to the Laurentian Fan in the Wisconsin (Piper, 1975b; Stow, 1975). It appears that the suspended sediments coming throught the Laurentian Channel were effectively diverted toward the southwest by the Labrador Current flowing through the Avalon Channel. Third, the closing of the Avalon Channel will possibly be reflected in the increased intensity of the Labrador Current and possibly also in the Western Boundary Undercurrent, since physical oceanographic evidence suggests that they are interdependent (Emery and Uchupi, 1972). This increased intensity near the Tail of the Bank generally leaves some small scale erosion on the top of these seamounts,(Piper, 1975) also found such evidence . But since the Sangamon Interglacial, there is no evidence of erosion in our cores. It perhaps indicates that the intensity of Arctic Water Overflow was decreased and also that a part of the Labrador



Gulf Stream

Labrador Current

Western Boundary Boundary

Hypothetical limit of ice sheet

Grey plume

Figure 32: Wisconsin sedimentation with open Avalon Channel, and ice sheets retreated from most part of the shelves. Suspended sediments coming through the Laurentian Channel being diverted by the Labrador Current flowing through the Avalon Channel.

Current was flowing through the Avalon Channel. However, admittedly this is a weak argument.

With the Avalon Channel open and the extent of ice on the continent being less than in the Illinoian stage, the overall sedimentation pattern again changed (Figure 32). The Avalon Channel during most of Wisconsin times blocked the input of sediments from Newfoundland and fine-grained suspended sediments coming through the Laurentian Channel. Although the mineralogy of the Wisconsin clay (unit B) and silt is very similar to the Early Illinoian clay and silt, the sand fraction is quite different, being more mature. Most of the Wisconsin fine-grained sediments on the tops of the seamounts south of the Grand Banks are probably re-worked sediments from the Grand Banks and the upper slope. Several Wisconsin transgressions and regressions across the Grand Banks and possible intermediate lowering (50 to 70 m) of the sea level helped to rework the Illinoian glacial sediments there. Some may have been eroded from the Meso-zoic and Cenozoic strata at the head of submarine canyons, but the volume of these sediments appears to be small. Cores collected from the Grand Banks slope shows that it was covered with thick Quaternary sediments suggesting lack of large-scale erosion. Moreover, most of such eroded sediments would have been transported to continental rise and abyssal plain with the turbidity currents.

With the Holocene termination of the ice sheet, the sedimentation pattern again changed. Due to rise of sea level, the coastline retreated a few hundred kilometers. As a result the contribution of detrital sediments from the land was substantially reduced. There was sudden increase in the number of icebergs on the south of the Grand Banks. As temperature increased, the Gulf Stream moved closer to the Grand Banks and the number of icebergs melting near the Tail of the Bank increased. So the first phase of warming up is documented by high pebble concentrations in the Holocene sediments. But as the climate became warmer and warmer, the number of icebergs was reduced and so was the number of ice rafted pebbles.

With the warming climate, biogenic productivity increased and biogenic sediments became the dominant sediment on the tops of seamounts. The intensity of the Labrador Current and Western Boundary Undercurrent decreased and they become less important in the sedimentation in this area with the exception of the ice rafted sediments. The northward movement of the Gulf Stream resulted in the presence of Subtropical foraminifera and higher montmorillonite in the clay.

The sedimentation pattern during the Sangamon and Yarmouthian Interglacial stages was similar to that during the Holocene sedimentation.

During the interstadial stages the sedimentation pattern was slightly different from that in the Holocene. The climate being warmer than glacial time, productivity was higher resulting in a higher biogenic carbonate input. As the sea level was about 50 to 70 m lower than the present, input of clastic sediments from the Grand Banks was higher than the Holocene.

Some of the sedimentation processes discussed above will remain speculative until more data are available. In particular other models are possible to explain the variation in sediment supply. For example the gray mud may be mostly transported by the Western Boundary Undercurrent. Only during the Late Illinoian glaciation was there a sufficiently large discharge through the Laurentian Channel introducing large amounts of red sediments into the area.

The examination of more core material will resolve problems such as these.

CONCLUSION

1. Eastern Canada was deglaciated three times during the last 600,000 year B.P. Major deglaciations occurred at 11,000 B.P., 130,000 B.P., and 600,000 B.P. (although the 600,000 B.P. date is open to question, and could be younger). In between these interglacials the ice sheet partially retreated five times. Three of these interstadials occurred in the Wisconsin and two in the Illinoian stage.

2. The ecologic condition of this area changed with the advance of glacial ice and southward penetration of the cold water front from the Arctic. Under the present climatic condition, the area studied receives mostly biogenic sediments. During glacial stages, the biogenic component was greatly reduced and the area was dominated by fine-grained clastic sediments.

3. In comparison to the Illinoian, Wisconsin glaciation appears to be much less extensive. Early and Late Illinoian ice sheets appear to have covered large parts of the shelf, but the Wisconsin ice advance on the Shelf was less extensive and played a less important role in the sedimentation on the outer continental margin off Eastern Canada.

4. With the advance of the continental ice the Gulf Stream moved towards southeast, and colder northern waters were more important in the Grand Banks area.

5. The stratigraphic record in the cores back to the Sangamon Interglacial can easily be correlated with the record from the land and with some adjustment with that from the open ocean. The record of the Pre-Sangamon section varies slightly from many other workers. Nevertheless, it tends to support the glacial history as proposed from studies on land rather than from studies of the open ocean.

6. The response of surface water temperatures of the ocean to climatic change on the continent is probably indirect. Because of the uncertainties in dating the lower parts of our cores, this suggestion cannot be proved.

7. There was no large scale change in carbonate dissolution during the glacial and interglacial stages in the studied area.

8. Change in the colour of the clay is mostly due to change in the source area.

9. The provenance of the clays deposited during the glacial stages can be determined from their mineralogy. The

Early Illinoian sedimentation pattern was controlled by the advance of an ice sheet across Newfoundland, the Laurentian ice sheet eroded the Gulf of St. Lawrence and Laurentian Channel in the Late Illinoian. Sediment contribution from the Wisconsin ice sheet was relatively small.

10. There was a significant change in the process of sedimentation of fine-grained sediments from glacials to interglacials. Sedimentation from clay plumes off outwash streams appears to be a very important process during the glacial stages.

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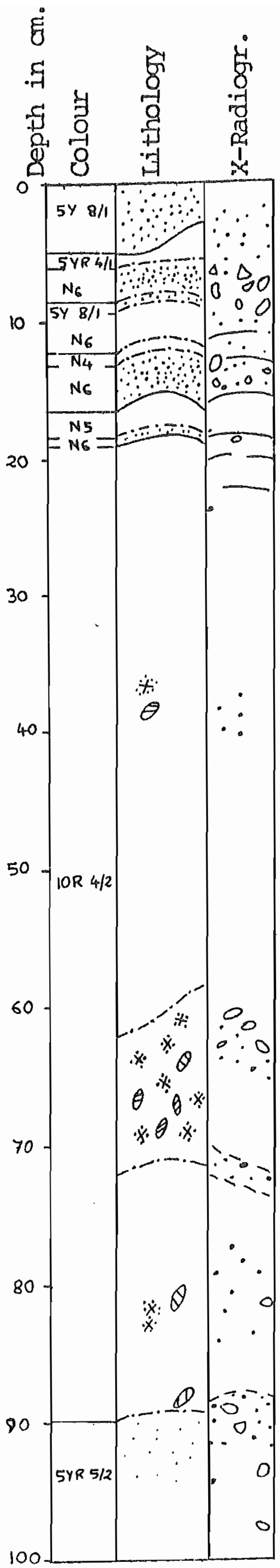
APPENDIX A

DESCRIPTION OF THE CORES

APPENDIX A

Core 74-021-29



Yellowish gray 5Y 8/1 foram-nanno ooze with coarse sand and pebbles.

Alternating brownish gray 5YR 4/1, medium gray N5 mud and foram-nanno ooze. Mud contains about 40% silt and fine sand. Fairly sharp lower contact.

Grayish red 10R 4/2 mud with about 30% silt size particles.

Grayish mud with mottles of foram-nanno ooze and medium gray mud.

Grayish red mud with no apparent structure.

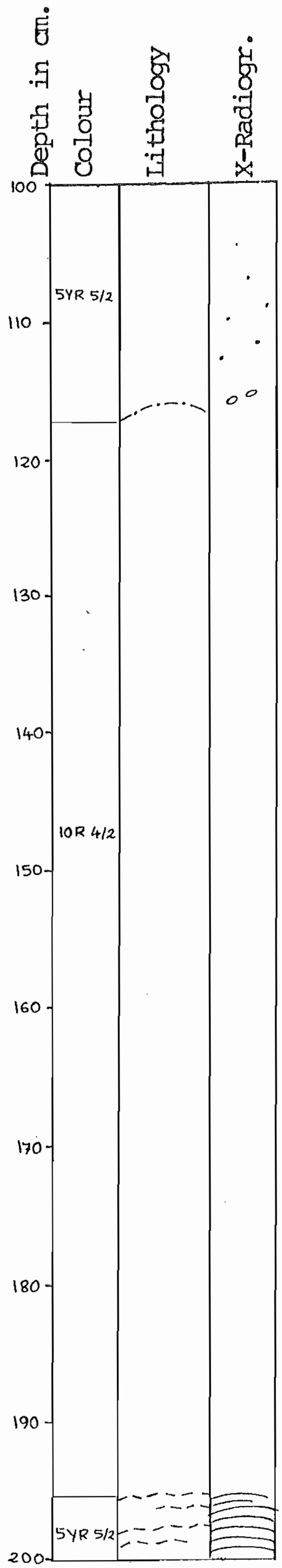
Mottles of light gray N6 foram-nanno ooze and grayish red 10R 4/2 mud.

Grayish red mud with a few mottles of foram-nanno ooze and gray mud.

Mottle of pale brown 5YR 5/2 mud.

Pale brown 5YR 5/2 mud with about 45% silt and fine sand size particles including broken foram tests. Slight high foram concentration near the top of the bed.

Core 74-021-29

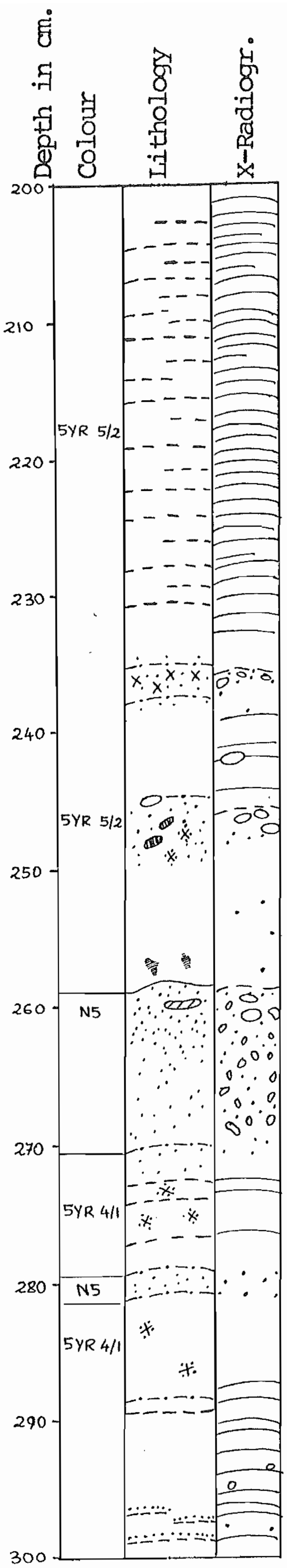


Same as above.

Pale brown 5YR 5/2 homogenous structureless mud with about 40% silt sized particles, mostly broken foram tests, coccoliths, sponge spicules, quartz, carbonate clasts and a few feldspar and ilmenite grains. No silt laminae are visable.

Pale brown 5YR 5/2 mud with a few disturbed silt laminae. 0.5 mm to 1 mm thick.

Core 74-021-29



Frequent silt laminae of generally less than 1 mm, but a few are up to 2 mm thick. Upper and lower contacts of most of the laminae are sharp. Some laminae are continuous, others discontinuous.

Mottles of coarse silt and sand size particles, including foram and quartz. Very faint bedding.

Pale brown 5YR 5/2 mud.

Pale brown 5YR 5/2 mud with about 45% fine sand and silt size particles, with a few pebble and black mottles. Gradational upper and lower contact.

Pale brown mud with black mottles.

Medium gray N5 foram-nanno ooze with scattered pebbles and granules. 3 cm long mottle of grayish red clay. Sharp upper and gradational lower contact. About 30% clay.

Brownish gray 5YR 411 mud with mottles of foram-nanno ooze and a few silt laminae.

Medium gray N5 foram-nanno ooze.

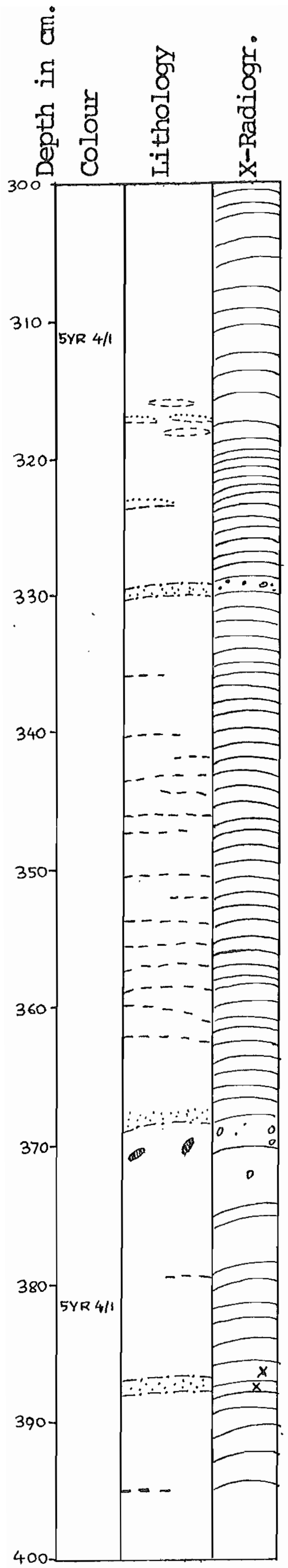
Brownish gray 5YR 4/1 mud with some mottles of foram-nanno ooze.

Sharp silt lamina.

Brownish gray 5YR 4/1 mud.

A few gradational silt laminae. Thickness less than 3 mm,contain biogenic and detrital particles.

Core 74-021-29



Brownish gray 5YR 4/1 mud with about 40% silt sized particles. Frequent fine silt laminae and faint bands visable in x-radiograph.

A few discontinuous and disturbed silt laminae, rich in quartz, broken foram tests, coccoliths and opaque minerals. Most of the laminae have sharp bases. A few have gradational tops, others have sharp tops.

1 cm zone rich in coarse silt and fine sand sized carbonates and quartz.

Brownish gray mud with frequent silt laminae, most of the laminae are disturbed and comparitively rich in quartz.

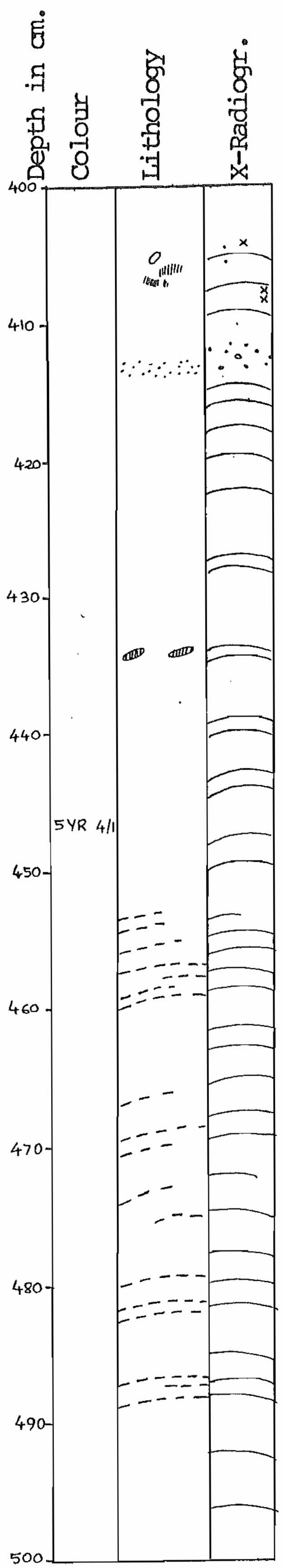
Brownish gray 5YR 4/1 mud. Slightly more silty.

Discontinuous silt lamina.

1 cm thick zone rich in foram. Pyrite in x-radiograph.

Discontinuous silt lamina.

Core 74-021-29



Brownish gray 5YR 4/1 mud with pebble and black mottles.

Pyrite mottes.

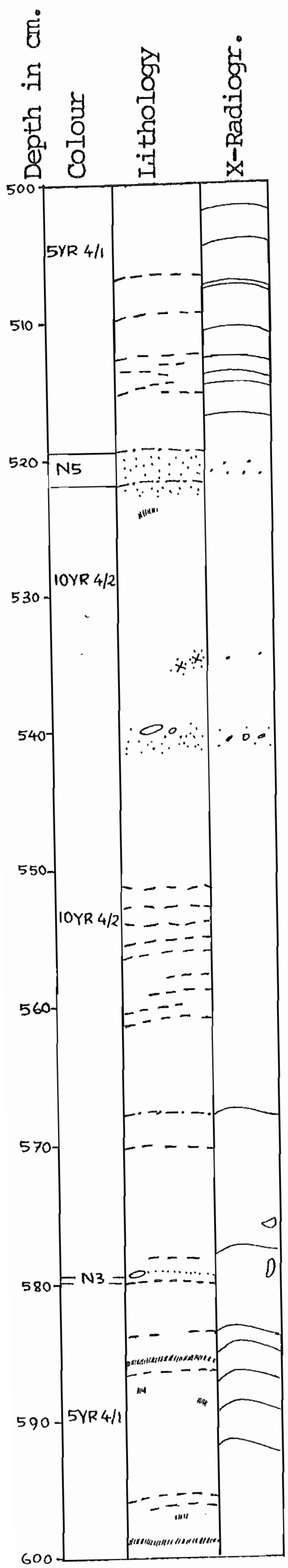
Slightly more silty, relatively rich in foram and quartz.

Brownish gray mud with silt laminae and bands.

Some back mottles.

Brownish gray 5YR 4/1 mud with frequent silt laminae; some continuous others discontinuous or disturbed. Most of the laminae are less than 1 mm thick. Only one or two laminae are more than 1 mm thick but rarely exceed 2 mm. The laminae contain about 50% silt size broken foram test, quartz, coccoliths, carbonate clasts, etc. but are relatively richer in quartz than the mud.

Core 74-021-29



Brownish gray 5YR 4/1 mud with two sharp 1 mm thick silt laminae.

Brownish gray mud with frequent disturbed silt laminae.

Medium gray N5 foram-nanno ooze. Relatively sharp upper contact, gradational lower contact.

Dark yellowish brown 10YR 4/2 mud with some mottles of foram-nanno ooze.

Slightly higher concentration of foram, coccoliths and quartz grains.

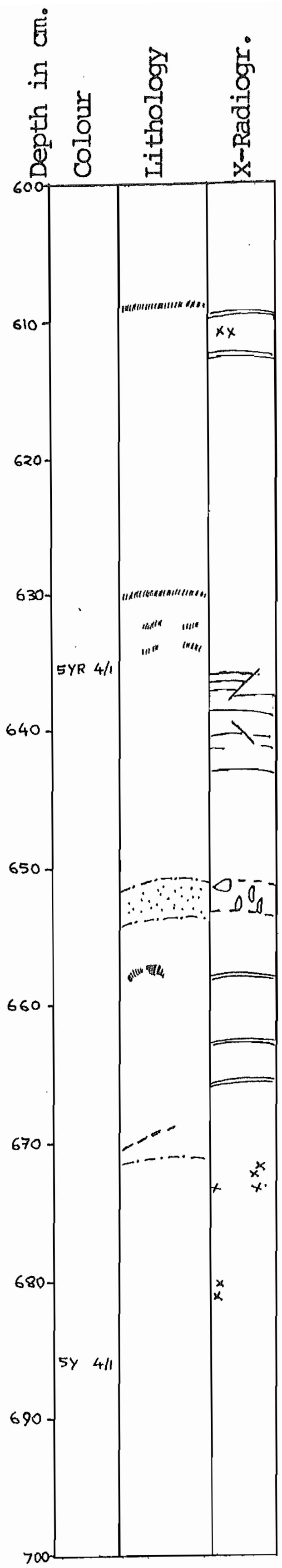
Dark yellowish brown mud with frequent very fine silt laminae.

Sharp 2 mm thick silt lamina with concentration of black opaque minerals (pyrite ?).

4 mm thick dark gray N3 silt laminae with sharp base and gradational top.

Brownish gray 5YR 4/1 mud with a few discontinuous silt laminae, and a few black mottles and bands.

Core 74-021-29



Slight change in colour but still brownish gray 5YR 4/1 mud with about 40% silt sized particles.

Pyrite

Black mottles and band.

Microfault

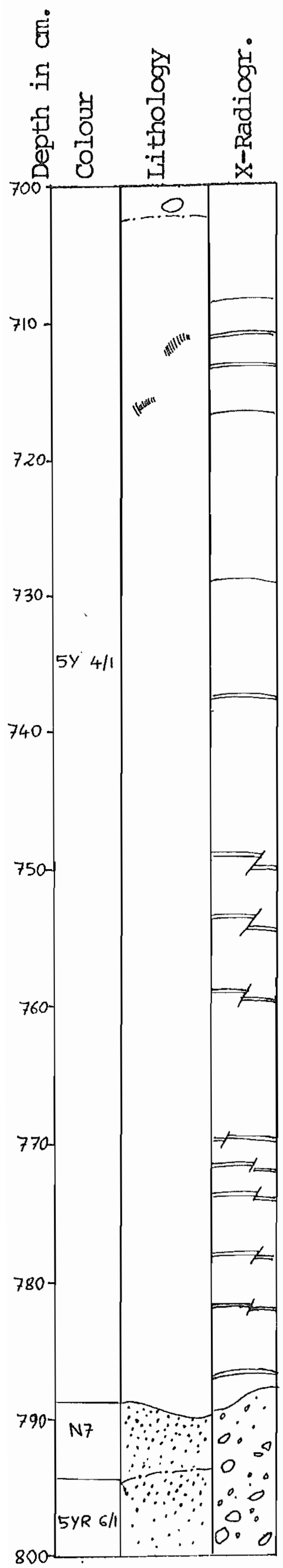
Brownish gray 5YR 4/1 foram-nanno ooze with about 50% clay, both upper and lower contacts are relatively sharp.

Olive gray 5Y 4/1 mud with about 40% silt and sand sized particles.

Olive gray 5Y 4/1 mud, slightly more silty.

Olive gray 5Y 4/1 mud.

Core 74-021-29



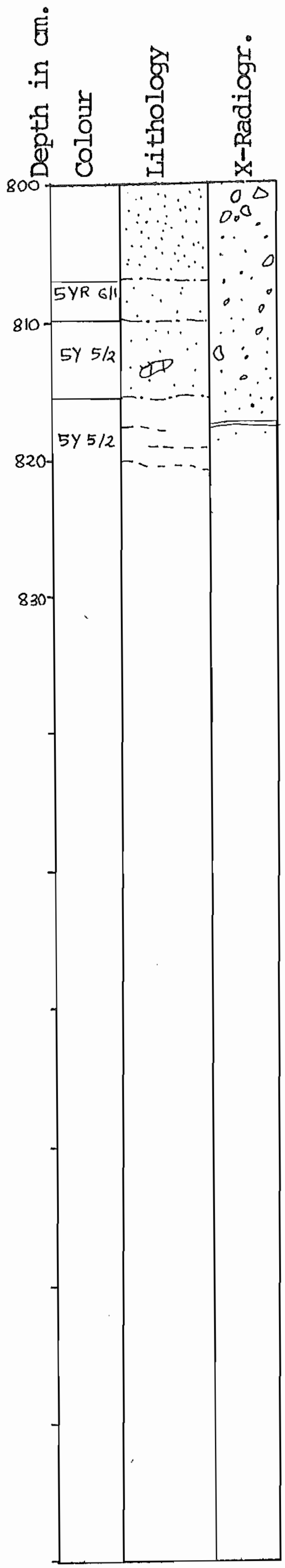
1 cm pebble

Olive gray 5Y 4/1 mud with a few silt laminae and black mottles.

Olive gray 5Y 4/1 mud with faint bands and silt laminae. Sometimes cut by microfault.

Light gray N7 foram-nanno ooze with very sharp upper contact and gradational base.

Light brownish 5YR 6/1 foram-nanno ooze, both the grain size and foram concentration increases towards 797 cm.



Core 74-021-29

Light brownish gray 5YR 6/1 foram-nanno ooze with pebbles and granules.

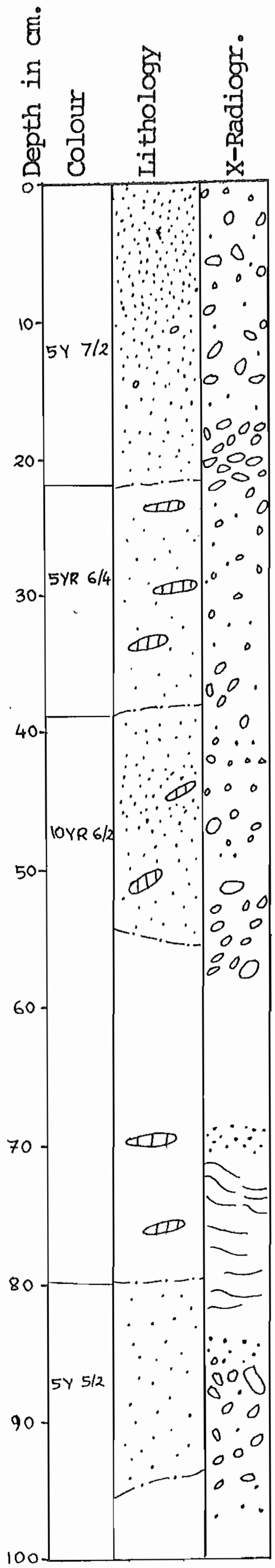
Light brownish gray muddy foram-nanno ooze.

Light olive gray 5Y 5/2 muddy foram-nanno ooze with some light grayish mottles.

Light olive gray 5Y 5/2 mud with irregular and disturbed fine silt laminae.

Suck in.

Core 75-009-117



Yellowish gray 5Y 7/2 foram-nanno ooze with pebble and granules.

Higher pebble concentration.

Light brown 5YR 6/4 muddy foram-nanno ooze with some gray mottles.

Pale yellowish brown 10YR 6/2 muddy foram-nanno ooze with some light brown and gray mottles.

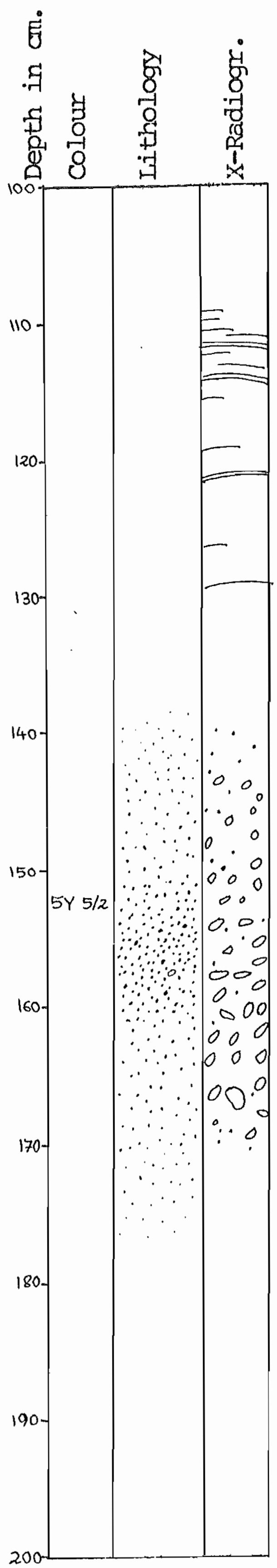
Higher pebble concentration.

Pale yellowish brown 10YR 6/2 mud with light brown mottles.

A few disturbed silt laminae.

Light olive gray 5Y 5/2 muddy foram-nanno ooze. Higher foram concentration around 90-95 cm.

Core 75-009-117

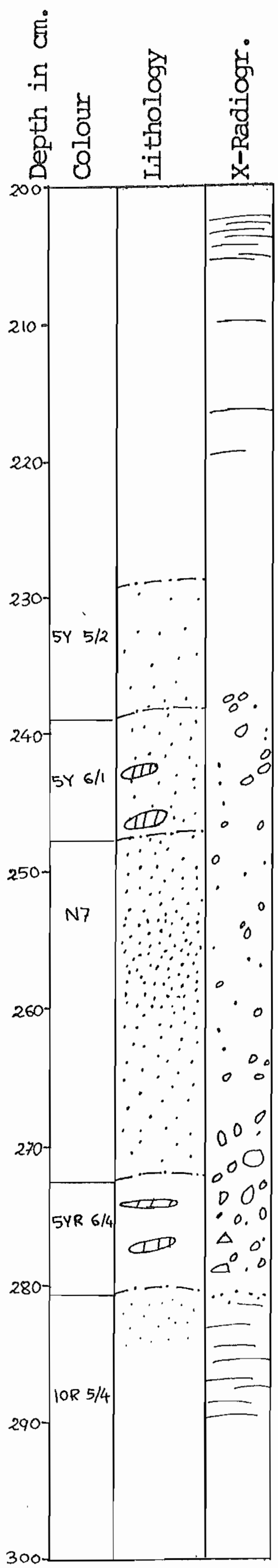


Light olive gray 5Y 5/2 mud with about 40% silt and fine sand sized particles. A few fine silt laminae and faint bands visable only in x-radiograph.

Light olive gray 5Y 5/2 foram-nanno ooze with pebbles and granules. Foram and coccolith concentration is highest between 150 to 160 cm, and the pebble concentration between 160 to 165 cm. Both the upper and lower contact is gradational.

Light olive gray 5Y 5/2 mud with about 35% silt sized particles. No apparent structure.

Core 75-009-117



Light olive gray 5Y 5/2 mud with a few silt laminae and faint bands.

Almost homogenous mud.

Light olive gray 5Y 5/2 muddy foram-nanno ooze.

Light olive gray 5Y 6/1 foram-nanno ooze with mottles of 5Y 5/2 mud and ooze.

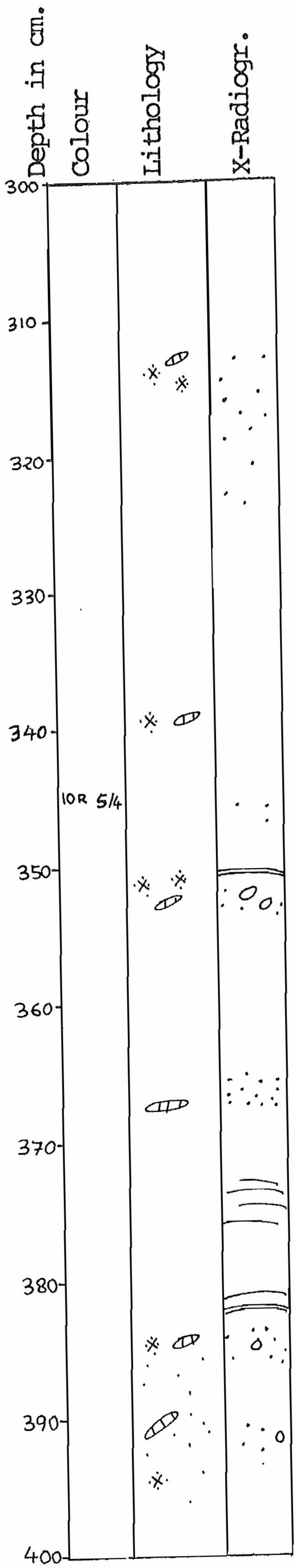
Light gray N7 foram-nanno ooze with pebbles and granules. Both the upper and lower contacts are gradational. High pebble concentration near the lower contact.

Light brown 5YR 6/4 muddy foram-nanno ooze with some light gray mottles. High pebble concentration.

Pale reddish brown 10R 5/4 mud with some foram and frequent silt laminae.

Pale reddish brown 10R 5/4 mud with about 40% silt sized particles. No apparent structure.

Core 75-009-117



Pale reddish brown 10R 4/2 mud with no apparent structure.

Mottles of foram-nanno ooze and grayish mud with some granules.

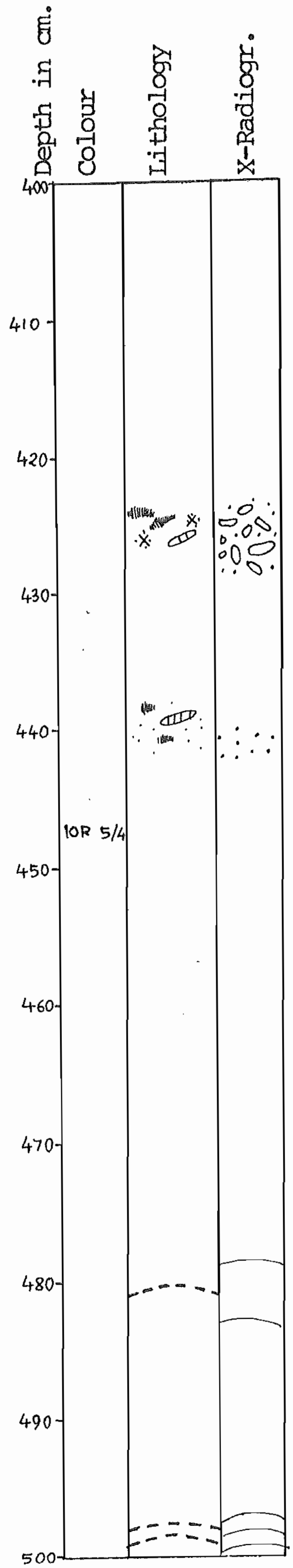
Pale reddish homogenous mud.

Mottles of foram-nanno ooze and gray mud.

Pale reddish brown 10R 4/2 mud with occasional mottles of foram-nanno ooze and gray mud. Very rare pebble and granules. Sometimes faint bands in the clay.

Pale reddish brown mud. Slightly more silty. Percentage of silt and sand sized particles about 45% most of which are foram and quartz. Some grayish mottles and rare pebbles.

Core 75-009-117



Pale reddish brown 10R 5/4 mud with about 40% silt and very fine sand sized particles. No apparent structure.

Black and grayish mottles with pebbles.

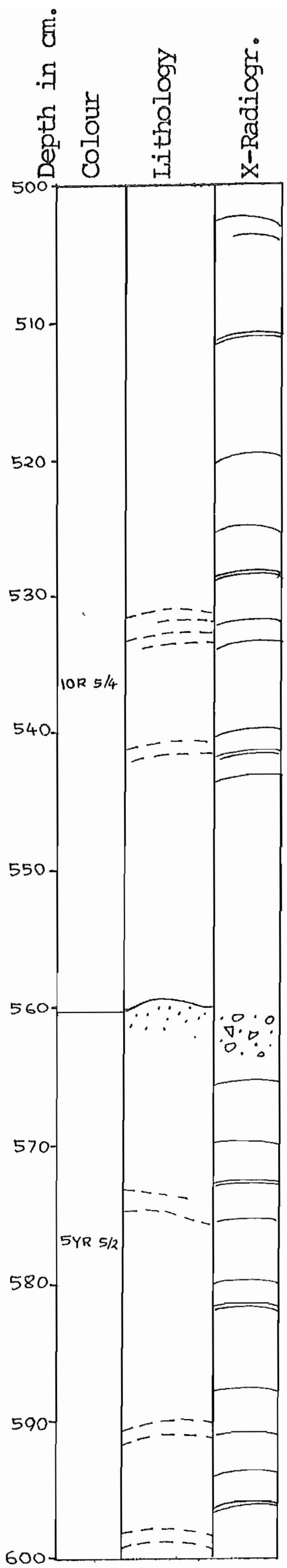
Pale reddish mud with black and grayish mottles.

Pale reddish brown mud without any apparent structure.

Sharp 1 mm thick silt laminae.

Pale reddish mud with rare silt laminae.

Core 75-009-117



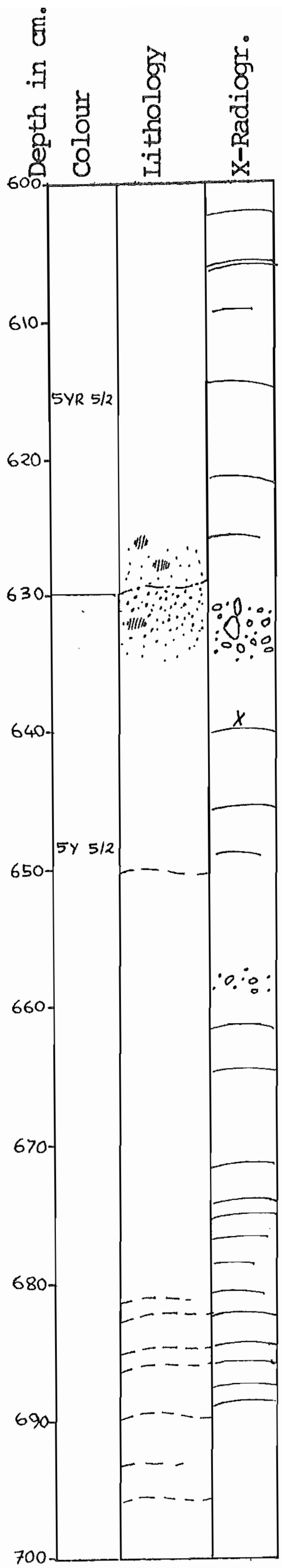
Pale reddish brown 10R 5/4 mud with a few silt laminae and faint bands.

Pale reddish brown mud with a few silt laminae and faint bands visable only in the x-radiograph. The silt laminae are generally less than .5 mm thick and most of them have sharp upper and lower contacts. A few are however, with gradational contacts.

Pale brown 5YR 5/2 muddy foram-nanno ooze with some pebble and granules.

Pale brown 5YR 5/2 mud with a few silt laminae and bands. The silt laminae are less than .5 mm thick, consist of mostly broken foram tests and quartz.

Core 75-009-117



Pale brown 5YR 5/2 mud with a few fine silt laminae and a few faint bands.

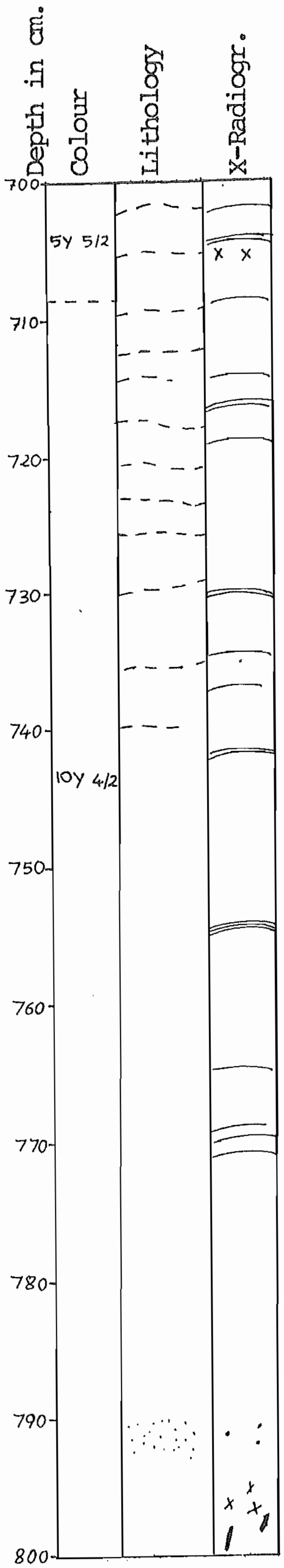
Light olive gray 5Y 5/2 foram-nanno ooze with pebbles and granules and some black mottles. Concentration of both foram and coccoliths gradually increases towards 630 cm and 626 cm.

Sharp silt lamina.

Light olive gray 5Y 5/2 mud with a few faint silt laminae and bands.

Light olive gray 5Y 5/2 mud with frequent fine silt laminae and faint bands. Thicknesses of the silt laminae are generally less than 0.5 mm, very rarely they exceed 1 mm. Most of them are continuous and have sharp upper and lower contacts.

Core 75-009-117



Pyrite

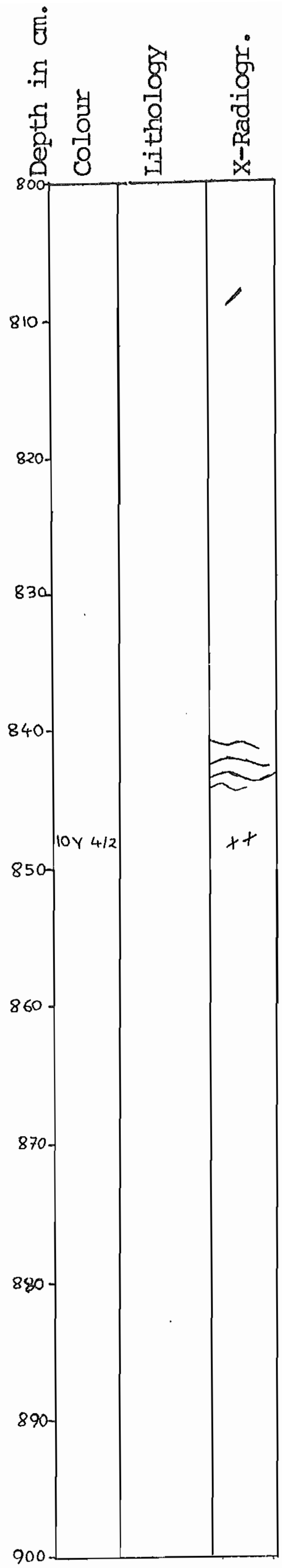
Grayish olive 10Y 4/2 mud, sometimes with fine silt laminae.

Almost homogenous grayish olive 10Y 4/2 mud with occasional silt laminae and bands, visable only in x-radiograph.

Slightly higher foram concentration.

Pyrite mottes and tubes.

Core 75-009-117



Pyrite (?) tube.

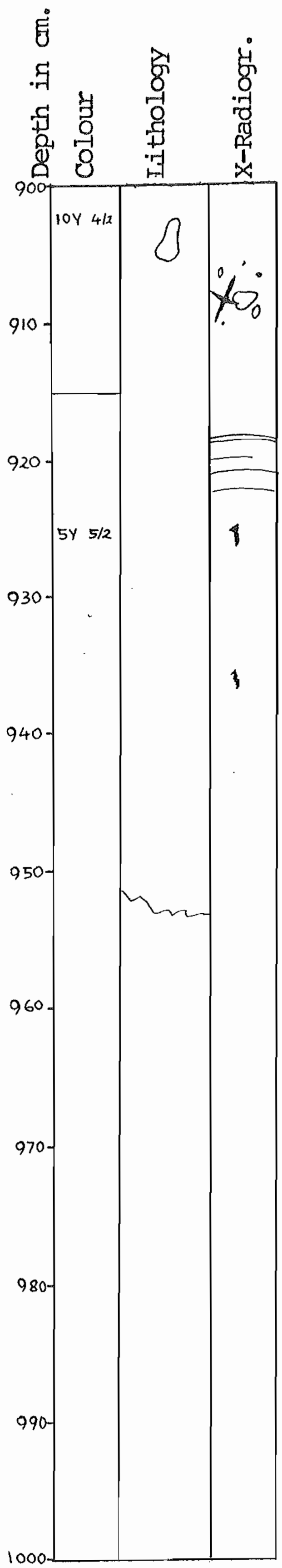
Grayish olive 10Y 4/2 mud with about 35% silt sized particles most of which are broken foram tests and quartz. No apparent structure.

A few disturbed silt laminae.

Pyrite mottles.

Homogeneous grayish olive 10Y 4/2 mud. No apparent structure.

Core 75-009-117



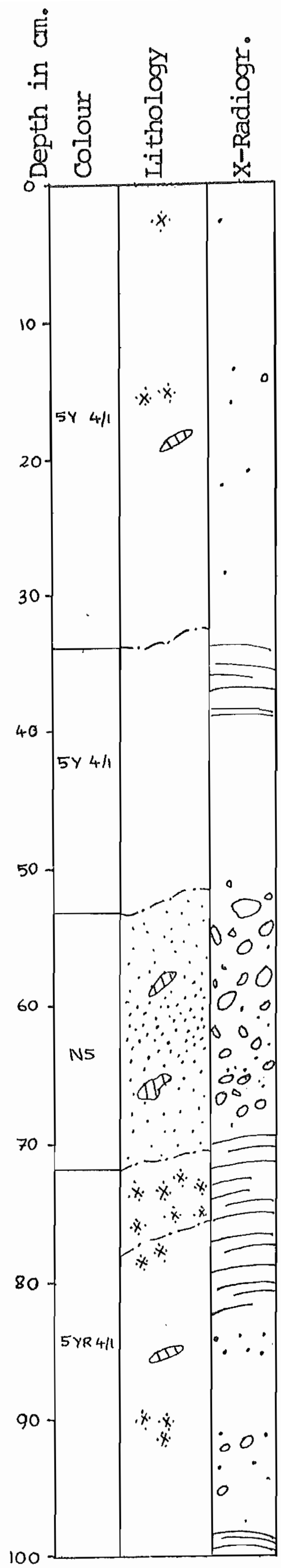
Mud clast

Pebble and pyrite tubes.

Light olive gray 5Y 5/2 mud with rare silt laminae and some pyrite mottles.

Suck in.

Core 74-021-28



Olive gray 5Y 4/1 mud with mottles of medium gray foram-nanno ooze and reddish mud. Quite disturbed.

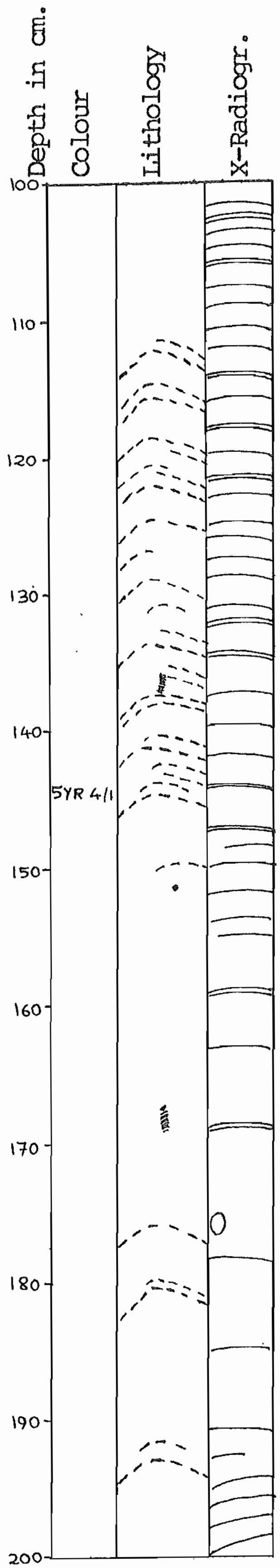
Olive gray 5Y 4/1 mud with about 40% silt sized particles. A few silt laminae on the upper part of the bed.

Medium gray N5 foram-nanno ooze with some brownish gray 5YR 4/1 mottles. The ooze is coarser at the middle part and relatively finer towards base and top of the bed.

Brownish gray 5YR 4/1 mud with mottles of foram-nanno ooze and a few silt laminae.

Brownish gray 5YR 4/1 mud with mottles of medium gray foram-nanno ooze and olive gray mud.

Core 74-021-28



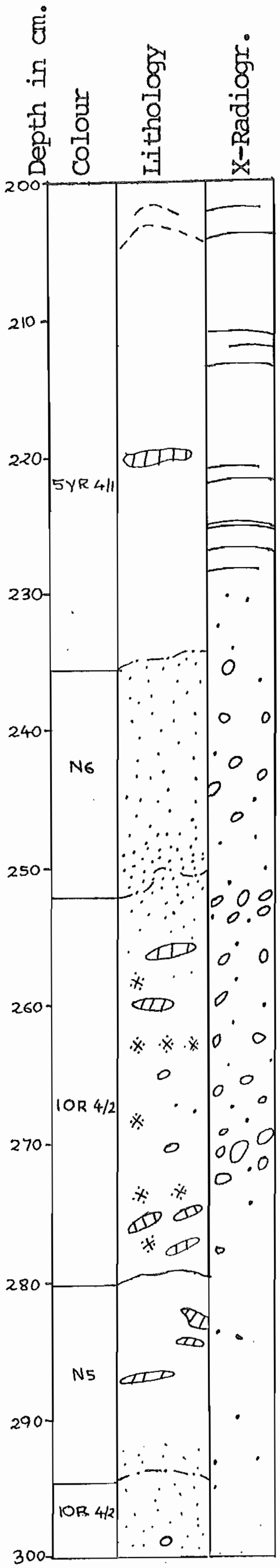
Brownish gray 5YR 4/1 mud with about 40% silt sized particles including broken foram tests, quartz, coccoliths, opaque minerals etc. Frequent silt laminae and bands in x-radiographs.

Brownish gray 5YR 4/1 mud with frequent silt laminae. Thickness of these laminae are generally less than 0.5 mm but a few are up to 2 mm. Almost all the laminae are disturbed especially on one side of the core. Composition of the laminae are more or less like the mud, with exception that they contain about 60% silt sized particles, mostly broken foram tests and quartz.

Black mottle

Brownish gray 5YR 4/1 mud with occasional silt laminae.

Core 74-021-28



Brownish gray 5YR 4/1 mud with a few silt laminae and reddish mottles.

Reddish mottle.

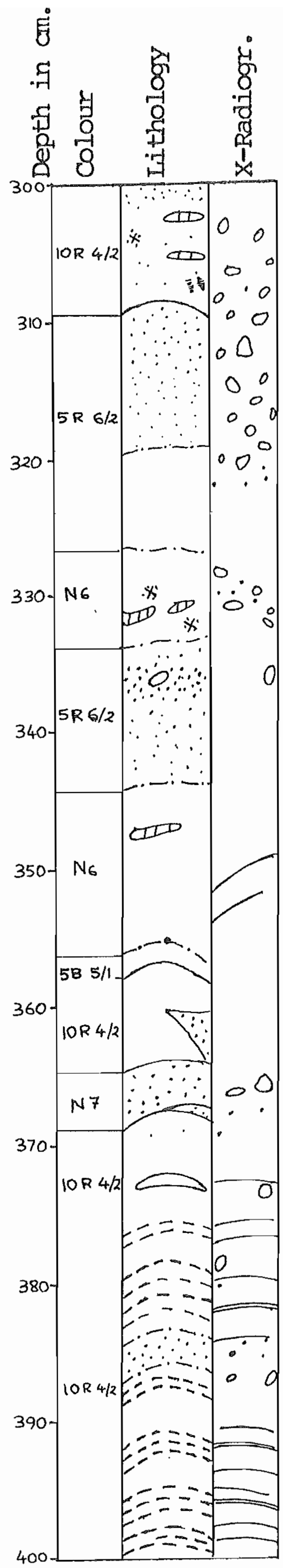
Medium gray N6 foram-nanno ooze with pebbles and granules. Foram-nanno concentration is highest near 250 cm. Gradational top and base.

Grayish red 10R 4/2 mud with mottles of foram-nanno ooze and grayish mud. Higher foram concentration near the top and base of the bed. Pebble and granules throughout the bed.

Medium gray N5 mud with some grayish red mottes. Upper contact is fairly sharp but the lower contact is gradational.

Grayish red 10R 4/2 mud with foram.

Core 74-021-28



Grayish red 10R 4/2 mud with mottles of foram-nanno ooze, gray mud and some black pyrite. Foram and coccolith concentration is highest between 301 to 297.5 cm.

Pale red 5R 6/2 muddy foram-nanno ooze. Fairly sharp upper contact, gradational lower contact.

Pale red 5R 6/2 mud.

Medium gray N6 mud with mottles of foram-nanno ooze and reddish mud.

Pale red 5R 6/2 muddy foram-nanno ooze. High foram and coccolith concentration around 336 cm.

Medium gray N6 mud with some reddish mottles and rare silt laminae.

Bluish gray 5B 5/1 mud. Sharp base.

Light gray N7 foram-nanno ooze with some pebbles and granules. Sharp base, with a bluish gray mud clast.

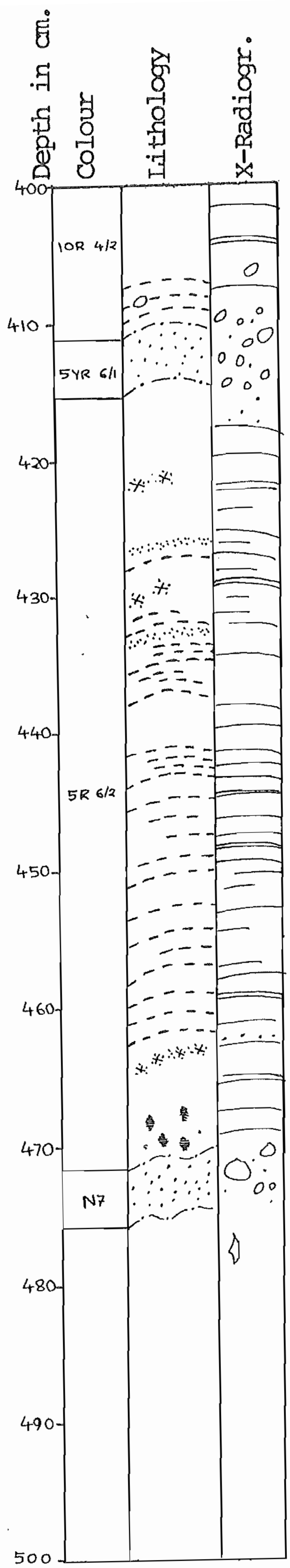
Lenticular silt bed (?)

Grayish red 10R 4/2 mud with frequent silt laminae. Slight foram concentration near top.

Thin bed of foram-nanno ooze.

Grayish red 10R 4/2 mud with frequent silt laminae and faint bands. Thickness of the silt laminae varies from .5 to 3 mm. Most of the silt laminae have sharp bases and gradational tops other with sharp tops and bases.

Core 74-021-28



Grayish red 10R 4/2 mud with a few silt laminae.

Light brownish gray 5YR 6/1 foram-nanno ooze with pebbles and granules.

Pale red 5R 6/2 mud with frequent silt laminae and a few bands of foram-nanno ooze. Most of the silt laminae are disturbed. Almost all the thick laminae have sharp base and gradation top.

A few black mottles

Light gray N7 foram-nanno ooze with pebbles and granules.

Suck in.

APPENDIX B-I

RESULT OF ANALYSES

APPENDIX B

Wet water content

Core 74-021-29

Depth in cm.	Wet water content %	Depth in cm.	Wet water content %
0	34.37	425	33.97
18	35.21	450	32.35
25	35.93	475	33.068
50	35.30	500	35.167
75	36.85	525	36.32
100	31.84	550	35.28
125	31.80	575	36.32
150	31.84	600	33.95
175	29.87	625	34.597
200	32.27	650	35.39
225	33.53	675	36.12
250	31.42	700	35.20
265	36.98	725	34.47
275	36.59	750	34.91
300	33.87	775	33.79
325	34.88	800	33.31
350	37.99	803	32.00
375	34.99	807	36.93
400	34.561		

APPENDIX B

Wet water content

Core 75-009-117

Depth in cm.	Wet water content %	Depth in cm.	Wet water content %
0	42.72	200	47.24
10	40.49	210	44.39
20	42.09	220	46.89
30	46.27	230	48.95
40	41.48	240	47.51
50	40.21	250	46.80
60	43.11	260	48.17
70	40.06	270	37.41
80	43.06	280	36.30
90	45.70	290	40.95
100	41.31	300	40.92
110	43.11	310	34.89
120	37.46	320	38.86
130	38.16	330	36.10
140	39.15	340	36.33
150	37.14	350	36.46
160	37.139	360	34.35
170	36.99	370	36.04
180	46.24	380	35.53
190	47.24	390	38.35

APPENDIX B

Wet water content (continued)

Core 75-009-117

Depth in cm.	Wet water content %	Depth in cm.	Wet water content %
400	37.61	590	31.10
410	37.03	600	33.46
420	37.66	610	32.60
430	36.198	620	34.09
440	38.327	630	33.87
450	37.75	635	36.23
460	34.08	640	35.91
470	33.99	650	33.55
480	33.27	660	34.58
490	33.23	675	34.21
500	33.86	700	33.29
510	31.97	725	33.37
520	32.48	750	33.89
530	33.02	775	33.21
540	32.08	800	32.28
550	34.93	825	33.31
560	37.28	850	32.89
570	34.11	875	31.81
580	31.07	900	33.21

APPENDIX B

Wet water content

Core 74-021-28

Depth in cm.	Wet water content %	Depth in cm.	Wet water content %
0	38.08	225	31.57
10	36.95	240	32.21
20	35.41	260	31.35
30	44.81	270	30.96
40	41.80	290	29.51
50	36.16	300	29.51
60	37.07	312	30.30
70	36.407	325	30.67
80	37.37	350	29.33
90	33.91	367	32.70
100	36.13	375	31.29
125	34.21	400	28.53
150	31.47	410	30.21
175	32.67	450	36.553
200	33.00	473	35.540

APPENDIX C

Carbonate Content

Core 74-021-29

Depth in cm.	% of $CaCO_3$	Depth in cm.	% of $CaCO_3$
0	33.41	210	14.55
10	15.53	220	12.35
20	10.21	230	14.86
30	11.03	240	13.76
40	10.72	250	21.59
50	12.43	260	30.28
60	10.25	270	22.44
70	9.83	280	30.83
80	12.60	290	15.97
90	10.67	300	13.64
100	13.24	310	14.40
110	14.71	320	14.59
120	15.27	330	13.25
130	14.31	340	14.96
140	14.09	350	15.40
150	14.95	360	13.98
160	13.05	370	15.15
170	13.15	380	14.73
180	13.88	390	15.91
190	12.94	400	13.63
200	15.85	410	14.73

Carbonate Content (continued)

Core 74-021-29

Depth in cm.	% of $CaCO_3$	Depth in cm.	% of $CaCO_3$
420	15.52	620	17.26
430	15.72	630	17.93
440	16.65	640	17.20
450	14.77	650	16.02
460	14.75	660	16.94
470	13.28	670	15.62
480	15.59	680	16.93
490	16.05	690	15.70
500	15.21	700	15.54
510	14.58	710	14.70
520	25.07	720	15.70
530	17.14	730	17.90
540	15.42	740	16.70
550	13.79	750	16.21
560	15.37	760	16.20
570	16.19	770	17.72
580	15.51	780	14.40
590	17.20	790	57.33
600	16.20	800	50.33
610	15.90	810	20.89

APPENDIX C

Carbonate Content

Core 75-009-117

Depth in cm.	% of $CaCO_3$	Depth in cm.	% of $CaCO_3$
0	73.63	210	11.14
10	48.51	220	10.41
20	45.83	230	13.01
30	18.67	240	27.68
40	33.21	250	63.75
50	35.54	260	78.80
60	16.12	270	54.87
70	15.41	280	29.56
80	14.19	290	18.52
90	28.26	300	25.36
100	16.41	310	12.22
110	17.70	320	25.14
120	19.54	330	16.01
130	19.87	340	15.31
140	17.32	350	13.97
150	40.25	360	15.12
160	53.10	370	13.41
170	57.13	380	14.48
180	15.36	390	18.21
190	12.08	400	14.11
200	11.54	410	13.86

APPENDIX C

Carbonate Content

Core 75-009-117

Depth in cm.	% of $CaCO_3$	Depth in cm.	% of $CaCO_3$
420	13.06	650	17.695
430	12.26	660	13.98
440	11.97	670	13.20
450	14.93	680	14.29
460	12.86	690	15.21
470	15.58	700	13.89
480	14.99	710	13.27
490	17.30	720	13.80
500	11.88	730	14.95
510	17.34	740	15.21
520	15.21	750	15.89
530	14.23	760	14.20
540	13.75	770	13.21
550	15.67	780	13.76
560	11.10	790	19.94
570	10.99	800	15.95
580	13.87	825	14.70
590	15.37	850	16.66
600	16.52	875	16.14
610	16.78	900	16.28
620	13.70	925	14.94
630	27.76	950	15.97
640	24.45	635	36.55

APPENDIX C

Carbonate Content

Core 74-021-28

Depth in cm.	% of $CaCo_3$	Depth in cm.	% of $CaCo_3$
0	20.52	250	35.36
10	22.08	260	14.00
20	15.96	270	31.25
30	16.61	280	33.21
40	13.26	290	21.58
50	15.86	300	23.61
60	29.55	312	32.05
70	23.73	320	18.20
80	13.23	330	17.80
90	14.30	340	31.87
100	13.24	350	43.49
110	12.87	360	21.81
120	13.20	367	51.25
130	12.17	370	23.51
140	11.89	380	22.69
150	11.45	390	24.76
160	12.10	400	18.90
170	11.76	410	38.74
180	11.50	420	21.61
190	11.97	430	20.91
200	12.77	440	21.81
210	12.31	450	23.34
220	13.85	460	24.34
230	23.76	473	50.39
240	30.75	480	23.81

APPENDIX D

Mineraloy of sand

1. Angular quartz grain
2. Rounded quartz grain
3. Iron stained quartz grain
4. Altered feldspar
5. Microcline feldspar
6. Plagioclase feldspar
7. Feldspar undifferentiated
8. Muscovite
9. Rock fragments
10. Amphibole
11. Ilmenite
12. Chert
13. Opaque
14. Biotite
15. Garnet
16. Epidote
17. Sphene
18. Glauconite
19. Unknown
20. Total number of grains counted

All expressed in percentage

APPENDIX D

Mineralogy of Sand

Core # and depth cm.		1	2	3	4	5	6	7	8	9	10
75-009-117	No.	103	116	2	4	2	1	--	1	2	3
100	%	42.04	47.34	.82	1.65	.82	.41		.41	.82	1.23
75-009-117	No.	111	113	1	3	2	3	1	--	3	2
220	%	43.88	44.66	.40	1.20	.80	1.20	.40		1.20	.80
74-021-29	No.	131	129	5	3	6	6	1	--	2	3
100	%	42.67	42.02	1.63	.98	1.95	1.95	.33		.65	.98
74-021-29	No.	163	115	3	6	2	5	3	--	10	1
129-131	%	50.94	35.94	.94	1.88	.63	1.56	.94		3.13	.31
74-021-29	No.	155	112	5	5	--	3	--	--	7	3
189-191	%	50.82	36.72	1.64	1.64		.98			2.30	.98
74-021-29	No.	156	121	2	1	4	3	3	--	6	2
225	%	49.21	38.17	.63	.32	1.26	.95	.95		1.89	.63
74-021-29	No.	225	57	1	2	7	2	4	--	7	2
300-302	%	68.81	17.43	.31	.61	2.14	.61	1.22		2.14	.61
74-021-29	No.	131	18	--	2	3	9	4	--	15	3
350-352	%	61.79	8.49		.94	1.42	4.25	1.89		7.08	1.42

Mineralogy of Sand (continued)

Core # and depth cm.		11	12	13	14	15	16	17	18	19	20
74-009-117	No.	1	5	1	2	--	--	--	2	--	245
100	%	.41	2.04	.41	.82				.82		
74-009-117	No.	--	6	2	3	--	--	--	3	--	253
220	%		2.37	.80	1.20				1.20		
74-021-29	No.	10	8	2	1	--	--	--	--	--	307
100	%	3.26	2.61	.65	.33						
74-021-29	No.	2	5	2	--	--	--	1	2	--	
129-131	%	.63	1.56	.63				.31	.63		320
74-021-29	No.	1	3	2	1	3	--	1	4	--	305
189-191	%	.33	.98	.66	.33	.98		.33	1.31		
74-021-29	No.	1	12	1	--	--	--	2	3	--	317
225	%	.32	3.79	.32				.63	.95		
74-021-29	No.	--	9	4	1	--	1	1	4	--	327
300-302	%		2.75	1.22	.31		.31	.31	1.22		
74-021-29	No.	1	9	4	1	--	1	5	1	2	212
350-353	%	.47	4.25	1.89	.47		.47	2.36	.49	.94	

Mineralogy of Sand (continued)

Core # and depth cm.		1	2	3	4	5	6	7	8	9	10
74-021-29	No.	167	40	1	2	5	3	3	--	13	2
400-402	%	66.27	15.87	.40	.79	1.98	1.19	1.19		5.16	.79
74-021-29	No.	160	29	--	2	6	18	6	--	49	1
450-452	%	51.12	9.29		.64	1.92	5.75	1.92		15.65	.32
74-021-29	No.	43	6	--	--	1	6	4	--	5	1
589-591	%	55.13	7.79			1.30	7.79	5.19		6.49	1.30
74-021-29	No.	198	42	--	--	5	10	13	--	36	4
600-602	%	60	12.73			1.52	3.03	4.94		10.91	1.21
74-021-29	No.	212	47	3	1	5	6	6	1	11	4
669-691	%	65.23	14.46	.92	.31	1.54	1.85	1.85	.31	3.38	1.23

Mineralogy of Sand (continued)

Core # and depth cm.		11	12	13	14	15	16	17	18	19	20
74-021-29	No.	1	3	4	1	--	--	2	5	--	252
400-402	%	.40	1.19	1.59	.40			.79	1.98		
74-021-29	No.	1	29	4	4	1	--	--	3	--	313
450-452	%	.32	9.27	1.28	1.28	.32			.96		
74-021-29	No.	--	6	1	--	1	--	2	--	1	77
589-591	%		8.22	1.30		1.37		2.61		1.30	
74-021-29	No.	4	7	1	2	1	1	--	6	--	330
600-602	%	1.21	2.12	.30	.61	.30	.30		1.82		
74-021-29	No.	--	15	3	5	1	--	--	3	2	325
669-691	%		4.62	.92	1.54	.31			.92	.62	

APPENDIX E

Peak Areas of Other Constituents in Less than 2μ Fraction

Core No.	Depth in cm	Quartz 4.26Å	Feldspar 3.18Å	Feldspar 3.22Å	Illite & Quartz 3.33Å	Chlorite & Kaolinite 3.53Å	Amphibole 8.4-8.5Å
75-009-117	30	14x5	28x7	26x7	145x5	54x7	--
75-009-117	100	20x3	41x5	32x8	192x4.5	86x5	8x6
75-009-117	205	11x4	30x5	20x5	83x4.5	25x6	10x6
75-009-117	239	17x4	40x5	27x5	120x4	30x6	9x5
74-021-28	100	16x4	45x6	35x5	142x5	50x7	9x4
74-021-28	290	15x4	42x5	32x5	152x5	55x5	13x4
74-021-29	3	8x4	28x5	24x5	80x4	35x6	--
74-021-29	6	8x5	24x5	28x5	109x5	59x6	--
74-021-29	10	13x3	31x5	31x6	118x5	66x5	--
74-021-29	37	13x5	28x4	--	135x5	69x5	--
74-021-29	44	17x2.5	36x5	--	192x4.5	110x5	13x6
74-021-29	50	14x3	34x4	39x5	161x5	79x6	--
74-021-29	80	12x3	30x5	40x6.5	152x5	69x6	--
74-021-29	129	18x2.5	39x5	36x6.5	170x5	105x5	--
74-021-29	189	15x2.5	40x3.5	36x6	172x4.5	108x4.5	--
74-021-29	236	14x3	36x5	39x5	160x5	98x5	--
74-021-29	248	13x2	34x5	34x5	114x5	62x5	11x4
74-021-29	262	14x3	38x4	32x4	126x5	64x4	--
74-021-29	489	20x15	45x4	37x5	132x4.5	70x4	14x3
74-021-29	500	11x2.5	46x5	38x6	122x5	66x4	20x4
74-021-29	589	12x3	39x5	34x5.5	142x5.5	80x4	14x8
74-021-29	617	12x5	42x4	28x5	90x5	43x4	12x7
74-021-29	669	16x2.5	50x4	44x6	158x6	89x4.5	20x5
74-021-29	678	10x5	31.5x4	22x3	52x4	18x6	9x7
74-021-29	791	14x3	51x4	40x5	102x6	46x6	14x5

APPENDIX F

Lithology of pebbles from Newfoundland Basin

Lithology	No. of Pebbles	Percentage
Limestone	4	18.18
Sandstone	2	9.09
Quartzite	1	4.54
Gresses	6	27.27
Granite	2	9.09
Gabbro	3	13.06
Basalt	4	18.18

APPENDIX G

Grain Size Analysis

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ	> 9ϕ
74-021-29 2	34.21	30.51	35.29	.86	65.79		60.37		35.29	
74-021-29 3	23.17	36.43	40.40	.90	76.83		67.89		40.40	
74-021-29 6	30.28	32.44	37.28	.87	69.72	62.13	55.35	46.15	37.28	
74-021-29 10	9.70	37.61	52.69	.71	9.30	82.05	76.48	65.80	52.69	
74-021-29 14	50.07	23.84	27.09	.88	49.93		41.38		27.09	
74-021-29 37	.24	31.95	67.87	.47	99.76	98.55	94.90	82.29	67.81	
74-021-29 45	.28	31.76	67.96	.47	99.72	98.88	96.06	84.01	67.96	55.04
74-021-29 50	.32	32.76	66.92	.49	99.68	96.10	91.19	77.71	66.92	50.36
74-021-29 65	9.89	33.57	56.54	.59	90.11	86.56	78.69	69.26	56.54	
74-021-29 80	2.24	37.02	60.73	.61	97.76	95.76	89.70	75.21	60.73	

Grain Size Analysis (continued)

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ	> 9ϕ
* 74-021-29 100	.35	29.70	69.95	.42	99.65	98.58	96.07	84.26	69.95	
74-021-29 130	.07	34.16	65.77	.52	99.93	99.71	91.35	80.83	65.77	54.10
74-021-29 190	.08	32.86	67.06	.49	99.92	99.17	92.95	80.37	67.06	54.36
74-021-29 236	7.79	33.60	58.61	.57	92.21		87.21		58.61	
74-021-29 248	5.67	35.35	58.97	.60	94.33	90.91	84.54	71.90	58.97	
* 74-021-29 255	.11	46.47	53.43	.87	99.89	96.41	85.08	66.49	53.43	
74-021-29 262	21.25	34.05	44.70	.76	78.75		70.37		44.70	
74-021-29 263	14.78	27.80	57.42	.48	85.22	81.78	79.83	68.85	57.42	
74-021-29 268	8.34	36.80	54.86	.67	91.46		85.27		54.86	
74-021-29 272	5.26	36.97	57.77	.64	94.74		85.71		57.77	
74-021-29 280	11.13	37.99	50.88	.75	88.87	86.41	81.37	64.90	50.88	

* acid treated

Grain Size Analysis (continued)

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ	> 9ϕ
74-021-29 300	13.42	29.69	56.89	.52	86.58	85.51	79.92	68.71	56.89	46.66
74-021-29 350	.01	34.51	65.48	.53	99.99	96.71	91.91	79.43	65.48	53.78
74-021-29 400	12.22	28.72	59.06	.49	87.78	87.91	82.59	71.47	59.06	48.63
74-021-29 412	5.56	34.49	59.95	.58	94.44	91.72	75.40	65.76	59.95	
74-021-29 450	10.81	29.75	59.45	.50	89.19	88.43	81.98	69.73	59.45	48.18
74-021-29 490	.03	35.83	64.13	.56	99.97	97.14	90.42	77.85	64.13	52.27
74-021-29 500	.01	32.23	67.76	.48	99.99	99.63	93.20	80.88	67.88	55.11
74-021-29 520	17.02	35.18	47.80	.73	82.98		74.21		47.80	
74-021-29 540	8.54	33.80	57.65	.59	91.46	87.15	81.18	65.39	57.65	
74-021-29 590	.01	29.91	70.08	.43	99.99	97.97	95.11	84.12	70.08	57.47

Grain Size Analysis (continued)

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ	> 9ϕ
74-021-29 600	11.55	28.00	60.45	.46	88.45	87.85	81.71	72.81	60.45	49.90
74-021-29 618	.02	26.97	73.01	.37	99.98	97.17	36.36	84.97	73.01	
74-021-29 650	1.69	28.67	69.64	.41	98.31	95.53	92.83	82.38	69.64	57.05
74-021-29 670	.10	27.31	72.59	.38	99.90	98.69	95.57	86.91	72.59	57.23
74-021-29 680	1.27	25.33	73.41	.35	98.73	96.45	95.30	85.08	73.41	
74-021-29 712	.04	28.71	71.25	.40	99.96	99.11	95.60	83.51	71.25	
74-021-29 730	.10	30.77	69.13	.45	99.90	88.30	86.85	84.85	69.13	
74-021-29 760	.08	30.30	69.62	.44	99.92	97.87	94.96	86.55	69.62	
74-021-29 791	42.34	26.85	31.82	.84	57.66		49.92		31.82	
74-021-29 800	50.01	23.84	26.15	.91	49.99		40.69		26.15	
74-021-29 805	31.11	30.12	38.77	.78	68.89		59.84		38.77	

APPENDIX G

Grain Size Analysis

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ	> 9ϕ
75-009-117 9	39.22	27.62	33.16	.83	60.78		47.06		33.16	
75-009-117 30	13.55	36.30	50.15	.72	86.45		71.62		50.15	
75-009-117 100	1.21	31.92	66.88	.48	98.79		97.57		66.88	
75-009-117 * 115	.03	26.47	73.50	.36	99.97	98.52	96.74	85.92	73.50	61.43
75-009-117 150	29.43	19.21	51.36	.37	70.57		58.21		51.36	
75-009-117 160	34.26	22.54	43.20	.52	65.75		57.28		43.20	
* 75-009-117 198	.18	28.19	71.62	.39	99.82	99.14	90.19	84.17	71.62	57.05
75-009-117 205	1.87	37.62	60.51	.62	98.13		90.27		60.51	
75-009-117 239	24.02	37.19	38.79	.96	75.98		63.44		38.79	
75-009-117 255	60.32	17.64	22.04	.80	39.68		30.86		22.04	

* acid treated

Grain Size Analysis (continued)

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ	> 9ϕ
75-009-117 275	28.14	26.95	44.91	.60	71.86		60.86		44.91	
75-009-117 300	12.30	32.10	55.60	.58	87.70		73.80		55.60	
74-009-117 340	1.16	35.19	63.66	.55	98.84		92.07		63.66	
* 75-009-117 350	.76	37.75	61.49	.61	99.24	96.89	89.79	74.73	61.49	51.46
75-009-117 400	.37	29.55	70.09	.42	99.63		95.50		70.09	
75-009-117 445	.87	29.98	69.15	.43	99.13		96.28		69.15	
* 75-009-117 450	.84	26.71	72.45	.37	99.16	98.12	95.71	86.02	72.45	60.11
75-009-117 500	.01	42.04	57.95	.73	99.99		90.61		57.95	
75-009-117 550	.26	32.75	67.00	.49	99.74		94.06		67.00	

* acid treated

Grain Size Analysis (continued)

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ	> 9ϕ
75-009-117 600	.06	34.28	65.66	.52	99.94		96.86		67.00	
75-009-117 625	6.12	33.26	60.61	.55	93.88		84.90		60.61	
75-009-117 635	14.68	39.14	46.18	.85	85.32		71.56		46.18	
75-009-117 640	4.88	30.79	64.34	.48	95.12		85.14		64.34	
75-009-117 695	.03	34.40	65.57	.52	99.97		87.22		65.57	
* 75-009-117 700	1.80	26.88	71.32	.38	98.20	95.75	92.24	83.83	71.32	59.37
75-009-117 745	.04	38.48	61.47	.63	99.96		94.01		61.47	
75-009-117 800	.22	30.29	69.49	.44	99.78		95.13		69.49	
* 75-009-117 800	.04	26.30	73.66	.36	99.96	98.62	96.25	87.08	73.66	62.92

* acid treated

Grain Size Analysis (continued)

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ	> 9ϕ
75-009-117 850	.16	32.96	66.88	.49	99.84		93.44		66.88	
75-009-117 900	.05	37.82	62.14	.61	99.95		94.68		62.14	
75-009-117 950	.03	32.01	67.97	.47	99.97		95.44		67.97	

APPENDIX G

Grain Size Analysis

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	≥ 6ϕ	> 7ϕ	> 8ϕ
74-021-28 15	1.52	33.42	65.06	.51	98.48		92.83		65.06
74-021-28 38	7.19	33.83	58.98	.57	92.81	88.62	82.15	70.57	58.98
74-021-28 50	.85	38.45	60.70	.63	99.15		91.02		60.70
74-021-28 65	49.53	19.78	30.69	.64	50.47		40.96		30.69
74-021-28 80	1.31	30.43	68.26	.45	98.69		92.63		68.26
74-021-28 100	.04	34.02	65.94	.52	99.96		93.92		65.94
74-021-28 135	.07	33.60	66.33	.51	99.93		92.74		66.33
74-021-28 200	.06	30.42	69.53	.44	99.94		93.59		69.53
74-021-28 240	14.64	37.28	48.08	.78	85.36		78.21		48.08

Grain Size Analysis (continued)

Core # and depth cm.	% of sand	% of silt	% of clay	silt/clay ratio	> 4ϕ	> 5ϕ	> 6ϕ	> 7ϕ	> 8ϕ
74-021-28 247	29.86	30.62	39.52	.77	70.14		62.79		39.52
74-021-28 270	27.59	24.63	47.77	.52	72.41		65.64		47.77
74-021-28 290	.75	33.50	65.76	.51	99.25		93.60		65.76
74-021-28 F301	30.13	28.84	41.03	.70	5.9	5.4	9.6	7.9	41.03
74-021-28 F312	29.73	26.33	43.95	.60	5.6	6.8	6.3	7.5	43.95
74-021-28 F339	13.43	33.13	53.44	.62	2.7	7.4	10.5	12.5	53.44
74-021-28 F367	28.27	28.39	43.35	.65	6.1	5.2	9.0	8.0	43.35
74-021-28 F410	33.47	30.87	35.66	.87	8.6	7.4	8.7	6.1	35.66
74-021-28 F473	27.27	33.61	39.12	.86	6.7	8.1	7.0	11.8	39.12

APPENDIX H

Foraminiferal Analysis

1. Percentage of benthonics
2. Percentage of *G. menardii* complex
3. Percentage of *G. truncatulinoides*
4. Number of *G. pachyderma* counted
5. Percentage of right coiling *G. pachyderma*
6. Percentage of left coiling *G. pachyderma*
7. Number of *G. truncatulinoides* counted
8. Percentage of right coiling *G. truncatulinoides*
9. Percentage of left coiling *G. truncatulinoides*

APPENDIX H

Foraminiferal Analysis

Core # and depth cm.	1	2	3	4	5	6	7	8	9	Total # counts
74-021-29 0	.88	.58	Present	100	37	63	--	--	--	340
74-021-29 3	2.6	Present	Present	101	18.83	83.16	--	--	--	308
74-021-29 6	4.66	Present	Present	101	21.78	78.21	6	83.33	16.67	300
74-021-29 10	3.42	.31	.62	100	19	81	12	58	42	321
74-021-29 14	1.62	.64	.64	100	37	63	21	61.9	39.1	308
74-021-29 16	2	Present	1.66	100	38	62	--	--	--	300
74-021-29 18	3	Present	Present	102	47.05	52.94	--	--	--	300
74-021-29 25	3.57	n.s.	n.s.	56	14.28	85.72	--	--	--	207
74-021-29 50	3.8	n.s*	n.s.	102	15.7	84.3	--	--	--	312

* Not seen

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	7	8	9	Total # counts
74-021-29 65	3.82	Present	.64	107	14.01	85.99	14	85.7	14.3	314
74-021-29 75	3.51	Present	Present	100	17	83	--	--	--	313
74-021-29 100	3.42	n.s.	n.s.	104	18.3	81.7	--	--	--	314
74-021-29 125	3.65	n.s.	n.s.	106	13.2	86.8	--	--	--	246
74-021-29 150	49	n.s.	n.s.	100	13	87	--	--	--	302
74-021-29 200	3.5	n.s.	n.s.	104	15.38	84.62	--	--	--	312
74-021-29 225	4.4	n.s.	n.s.	88	11.36	88.64	--	--	--	270
74-021-29 262	1.61	Present	1.61	105	17.14	82.86	25	96	4	310
74-021-29 263	1.76	Present	.83	102	18.37	81.37	20	100	?	317

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	7	8	9	Total # counts
74-021-29 268	1.29	Present	1.61	100	41	59	20	75	25	306
74-021-29 270	1.57	Present	.94	109	30.3	69.7	--	--	--	317
74-021-29 272	1.6	Present	Present	100	36	54	14	85	15	305
74-021-29 280	.65	Present	1.31	101	36.63	53.37	42	97.6	2.4	306
74-021-29 300	3.83	n.s.	n.s.	100	5	95	--	--	--	287
74-021-29 350	3.35	n.s.	n.s.	94	4.25	95.75	--	--	--	268
74-021-29 400	4.67	n.s.	n.s.	100	6	94	--	--	--	278
74-021-29 412	3.22	n.s.	n.s.	100	9	91	--	--	--	310
74-021-29 450	4.54	n.s.	n.s.	100	8	92	--	--	--	286

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	7	8	9	Total # counts
74-021-29 500	3.45	n.s.	n.s.	100	2	98	--	--	--	318
74-021-29 520	2.59	Present	.32	104	25	75	--	--	--	309
74-021-29 F540	2.3	n.s.	n.s.	100	15	85	--	--	--	305
74-021-29 550	3.03	n.s.	n.s.	100	6	94	--	--	--	288
74-021-29 600	4	n.s.	n.s.	100	6	94	--	--	--	300
74-021-29 650	4.19	n.s.	n.s.	100	11	89	--	--	--	310
74-021-29 652	5.9	n.s.	n.s.	101	12.87	87.12	--	--	--	319
74-021-29 700	4.04	n.s.	n.s.	100	14	86	--	--	--	308
74-021-29 750	4.67	n.s.	n.s.	100	13	87	--	--	--	300

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	7	8	9	Total # counts
74-021-29 775	4.38	n.s.	n.s.	107	13.1	86.9	--	--	--	319
74-021-29 791	.95	2.85	1.27	102	79.41	20.59	50	82	18	316
74-021-29 800	.97	Present	1.9	100	60	40	53	83	17	309
74-021-29 803	1.63	.32	.65	100	57	43	--	--	--	308
74-021-29 805	.60	.30	.90	103	60.19	39.8	29	100	?	306
74-021-29 807	3.20	.32	.32	100	40	60	--	--	--	312

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	Total # of counts
75-009-117G 45	3.17	n.s.	n.s.	104	28.85	71.15	315
75-009-117G 50	3.93	n.s.	n.s.	103	20.39	79.61	331
75-009-117G 75	4.85	n.s.	n.s.	61	14.75	85.25	314
75-009-117G 90	3.55	n.s.	n.s.	117	23.08	76.92	394
75-009-117G 95	3.40	n.s.	n.s.	109	28.44	71.56	324
75-009-117G 110	3.93	n.s.	n.s.	108	3.70	96.30	331
75-009-117 9	1.28	.64	1.28	103	47.57	52.42	312
75-009-117 30	4.06	n.s.	n.s.	106	10.4	89.6	345
75-009-117 100	3.42	.34	.34	95	26.31	73.68	292

Foraminiferal Analyis (continued)

Core # and depth cm.	1	2	3	4	5	6	Total # of counts
75-009-117 155	1.27	n.s.	.32	115	27.83	72.17	309
75-009-117 164	1.56	n.s.	Present	109	38.53	61.47	325
75-009-117 200	2.26	n.s.	n.s.	103	25.24	74.76	310
75-009-117 205	3.08	n.s.	n.s.	116	16.38	83.62	324
75-009-117 239	1.56	.94	1.25	109	30.27	69.72	320
75-009-117 250	.84	1.12	1.40	103	51.46	48.58	356
75-009-117 255	1.25	1.87	2.18	103	70.87	29.13	321
75-009-117 260	1.18	3.24	2.36	102	82.35	17.65	338
75-009-117 275	2.58	.66	.32	100	24	76	309

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	Total # of counts
75-009-117 340	2.33	.33	.33	82	20.71	79.29	301
75-009-117 400	1.61	.66	.66	106	21.70	78.30	311
75-009-117 445	4.31	.31	.31	107	12.15	87.85	325
75-009-117 500	2.32	n.s.	n.s.	73	13.69	86.30	259
75-009-117 550	3.29	n.s.	.63	103	21.15	78.84	315
75-009-117 600	2.77	n.s.	n.s.	106	22.24	77.75	289
75-009-117 625	4.13	n.s.	n.s.	104	15.73	84.27	315
75-009-117 635	4.56	n.s.	n.s.	114	27.19	72.80	329
75-009-117 640	1.61	n.s.	.64	104	32.69	67.30	311

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	Total # of counts
75-009-117 695	4.30	n.s.	n.s.	22	4.50	95.50	60
75-009-117 745	3.23	n.s.	1.62	28	3.57	96.43	62
75-009-117 800	4.55	n.s.	n.s.	108	7.55	92.45	308
75-009-117 850	4.19	n.s.	n.s.	55	10.90	89.09	143
75-009-117 900	3.49	n.s.	n.s.	83	7.23	92.77	143
75-009-117 950	2.91	n.s.	n.s.	101	3.96	96.03	369
75-009-117G 2	1.54	2.46	1.23	102	65	35	325
75-009-117G 10	1.27	1.90	1.22	107	52.34	47.66	316
75-009-117G 30	4.09	n.s.	n.s.	105	17.14	82.86	318

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	Total # of counts
74-021-28 15	1.61	.64	.64	104	20	80	311
74-021-28 50	1.85	1.23	1.23	123	39.02	6098	324
74-021-28 65	1.24	1.24	.93	103	56.31	43.69	322
74-021-28 80	3.34	Present	.61	105	40	60	329
74-021-28 100	3.36	n.s.	.66	106	16.04	83.96	327
74-021-28 135	3.35	n.s.	n.s.	42	9.52	90.47	179
74-021-28 200	2.60	n.s.	n.s.	100	8	92	308
74-021-28 240	3.00	Present	Present	109	40	69	315

Foraminiferal Analysis (continued)

Core # and depth cm.	1	2	3	4	5	6	Total # of counts
74-021-28 270	.97	.97	.65	104	66.35	33.65	310
74-021-28 290	3.91	n.s.	n.s.	112	14.29	85.71	295
74-021028 F367	.96	1.28	1.66	100	39	61	313
74-021-28 F387	4.73	n.s.	.32	108	10.19	89.81	317
74-021-28 F410	1.62	.65	.32	100	17	83	308
74-021-28 F473	.65	.97	.97	100	45	55	308

APPENDIX I

Pore diameters and porosities of *G. pachyderma*

Core No.	Depth in cm	Mean Pore Diameter (microns)	# of pores per 32 x 40 μ	Porosity	Mean Porosity
74-021-29G	0	1.45	12	1.55	
	0	1.55	13	1.92	2.030
	0	1.75	14	2.629	
74-021-29G	20	1.25	11	1.05	
	20	1.12	12	.923	1.080
	20	1.31	12	1.263	
74-021-29	0	.86	9	.408	
	0	.95	9	.498	.486
	0	.95	10	.553	
74-021-29	6	.90	11	.546	
	6	.78	10	.373	.471
	6	.82	11	.493	
74-021-29	50	.59	9	.175	
	50	.62	7	.165	.164
	50	.50	10	.153	

Pore diameters and porosities of <u>G. pachyderma</u>

Core No.	Depth in cm	Mean Pore Diameter (microns)	# of pores per 32 x 40 μ	Porosity	Mean Porosity
74-021-29	262	1.32	11	1.175	
	262	1.33	11	1.193	1.146
	262	1.32	10	1.069	
74-021-29	268	.76	10	.35	
	268	.75	11	.38	4.20
	268	.90	11	.55	
74-021-29	540	.86	4	.186	
	540	.69	4	.116	.135
	540	.59	5	.107	
74-021-29	650	1.00	5	.307	
	650	1.00	6	.368	.387
	650	1.15	6	.487	
74-021-29	791	2.25	10	3.105	
	791	2.65	12	5.168	3.830
	791	2.29	10	3.216	

APPENDIX I

Pore diameters and porosities of *G. bulloides*

Core No.	Depth in cm	Mean Pore Diameter (microns)	# of pores per 32 x 40 μ	Porosity	Mean Porosity
74-021-29G	0	1.39	26	3.081	
	0	1.45	28	3.610	3.535
	0	1.51	28	3.915	
74-021-29G	20	.90	30	1.49	
	20	.92	28	1.45	1.493
	20	.90	31	1.54	
74-021-29	0	.96	24	1.356	
	0	1.02	24	1.531	1.581
	0	1.10	25	1.855	
74-021-29	6	1.00	18	1.104	
	6	.92	19	.986	1.094
	6	1.04	18	1.194	
74-021-29	50	.59	9	.175	
	50	.62	7	.165	.164
	50	.5	10	.153	

Pore diameters and porosities of *G. bulloides*.

Core No.	Depth in cm	Mean Pore Diameter (microns)	# of pores per 32 x 40 μ	Porosity	Mean Porosity
74-021-29	262	1.18	25	2.135	
	262	.93	26	1.379	1.740
	262	1.10	23	1.707	
74-021-29	268	1.34	19	2.09	
	268	1.18	18	1.53	1.613
	268	1.00	20	1.22	
74-021-29	540	.694	20	.591	
	540	.614	19	.439	.518
	540	.688	18	.523	
74-021-29	650	.68	19	.539	
	650	.90	20	.994	.879
	650	1.00	18	1.104	
74-021-29	791	1.94	26	6.001	
	791	1.63	32	5.214	5.476
	791	1.63	32	5.214	