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Quaternary climate change and the evolution of the mid-latitude western North  
Atlantic Ocean: palynological, foraminiferal, sedimentological, and stable isotope  
evidence from DSDP Sites 604, 607 and 612

by

Francine Marie Gisele McCarthy

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Submitted in partial fulfillment of the requirements  
for the degree of Doctor of Philosophy

at

Dalhousie University  
Halifax, Nova Scotia

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ISBN 0-315-76732-4

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## ABSTRACT

The fossil record of marine plankton (foraminifera and dinoflagellate cysts) and terrestrial spores and pollen along a transect across the western North Atlantic Ocean at approximately 40°N shows that climatic change through the Quaternary was accompanied by oceanographic changes that affected sedimentation, especially along the U.S. margin. The Gulf Stream was deflected from the margin in response to climatic deterioration ca 1.4 Ma, when polar surface water first penetrated between the western boundary of the subtropical gyre and the margin. The "Paleo-Slope Water" mass developed from a mixture of surface water of polar origin, fluvial influx, and water from the Sargasso Sea. The existence of this water mass allowed large-scale instabilities (meanders and warm core rings) to develop in the north wall of the Gulf Stream, and the topographic Rossby waves generated by eddies began to erode the slope, redepositing Quaternary and Neogene sediments onto the rise. When ice sheets reached the eastern U.S. around 0.3 Ma, ice-marginal conditions were gradually established over the New Jersey margin, and the influence of the Gulf Stream above the margin (and accompanying erosion of intercanyon areas of the mid-slope) was suppressed and the Slope Water mass developed ca. 0.2 Ma. The progressive seaward deflection of the Gulf Stream through the Pleistocene resulted in positive-feedback, allowing North American climates to become progressively more continental, forcing ice centres to migrate south and east within range of moist air masses, therefore further deflecting warm ocean currents from the east coast of North America. Because erosion of the New Jersey slope depends primarily on the oceanography of surface waters and not directly on sea level, the Vail depositional model may not be applicable to the Pleistocene sequence of this margin, where glacioeustatic sea-level highstands are marked by erosion of the mid- to lower slope by abyssal currents generated by instabilities of the Gulf Stream.

## ACKNOWLEDGEMENTS

I would like to acknowledge the assistance and support of everyone who made the production of this thesis possible; the following list will inevitably be incomplete, but will highlight those to whom I owe the deepest gratitude. I would like to thank David Scott who inspired me to pursue micropaleontology as a career, and for giving me the opportunity to do this project, and for sharing his benthic foraminiferal data. I would like to thank Peta Mudie for her help and guidance with dinoflagellate taxonomy and ecology and with paleoceanography, despite her hectic schedule. Thanks are also due to Pete Ogden for the use of his laboratory and for discussions of palynology, and to Franco Medioli whose fascinating anecdotes were my first introduction to micropaleontology. Veysi Baki performed stable isotopic analyses and assisted with computing and plotting; I want to thank him for his almost infinite patience in teaching an often difficult and demanding student. Eric Collins assisted in preparing photoplates, and with numerous (often tedious) tasks, for which I am truly grateful. Wolfgang Kuhnt also assisted with preparing photoplates. John Dickie assisted with drafting, providing useful suggestions on visual presentation of data. Mike Lozon (Brock University) was instrumental in improving the quality of computer-generated plots for the final draft. Ralph Stea shared his knowledge of the terrestrial record of the Quaternary, providing helpful insight into atmosphere-hydrosphere-lithosphere-cryosphere interactions. Chloe Younger assisted with sample preparation and with picking foraminifera for isotopic studies, as did Linda Dobbin. Shelley Thibaudeau assisted in palynological sample preparation. Jean Dabros taught me the techniques of dinocyst sample preparation. Wylie Poag and James Robb of the USGS were very generous with their time and with materials, discussing both published and unpublished data. Ron Boyd provided helpful criticism following my thesis proposal. I would also like to acknowledge the financial support

of the Killam Trust which supported me while in the PhD program, and additional support for this project from NSERC operating and strategic grants to D.B. Scott.

The five years I spent in the PhD programme were the most emotionally taxing years of my life, for a number of reasons, the difficulty (and occasional tedium) of the thesis work being only a relatively minor factor. Because of this, I would especially like to thank my family and friends who provided emotional support when the going got rough, and most of all, put up with me- Mom, Dad, Melissa, Veysi Baki, Eric Collins, Jane Barrett, Shelley Thibaudeau, John Dickie, Anne Jennings, Ralph Stea, Wolfgang Kuhnt, Cecily Honig, Helen Kerr, Peta Mudie, Dave Scott, Kumiko Azetsu Scott, Steve Varga, the Kevins and the other guys in Toronto, and special thanks to everyone in geology at St. F.X., and to the guys in geology at Brock who gave me a really good incentive to finish!

## CHAPTER I

### GENERAL INTRODUCTION

#### 1.0 THIS STUDY

The primary purpose of this study was to determine the effect of global climatic change on water masses and circulation in the western North Atlantic Ocean, and ultimately on the geologic record of the U.S. continental margin off New Jersey. Continental margins are the "bridge" between the relatively well-studied and well-understood deep sea and terrestrial environments, but the Pleistocene record of continental margins has seldom been studied in detail. Despite the difficulty of interpreting microfossil records in a depositional environment characterised by pervasive mass wasting, strong bottom currents, and very dynamic surface water characteristics, the Quaternary record of the margin is critical to improve correlation between the North American terrestrial record and the marine record of the western North Atlantic, and to improve our understanding of hydrosphere-atmosphere-cryosphere interactions.

A high resolution micropaleontological study was performed on hydraulic piston cores from three sites at mid-latitudes in the western North Atlantic. Deep Sea Drilling Project (DSDP) Sites 604 (Leg 93), 607 (Leg 94), and 612 (Leg 95) form a transect at about 40°N latitude from the mid-slope off New Jersey to the western flank of the Mid-Atlantic Ridge (Figure 1). Quaternary sediments from these sites were analysed for pollen and terrestrial spores, dinoflagellate cysts, and planktonic foraminifera. Oxygen isotopes were analysed using planktonic foraminiferal tests from two sites, DSDP Sites 607 and 612.

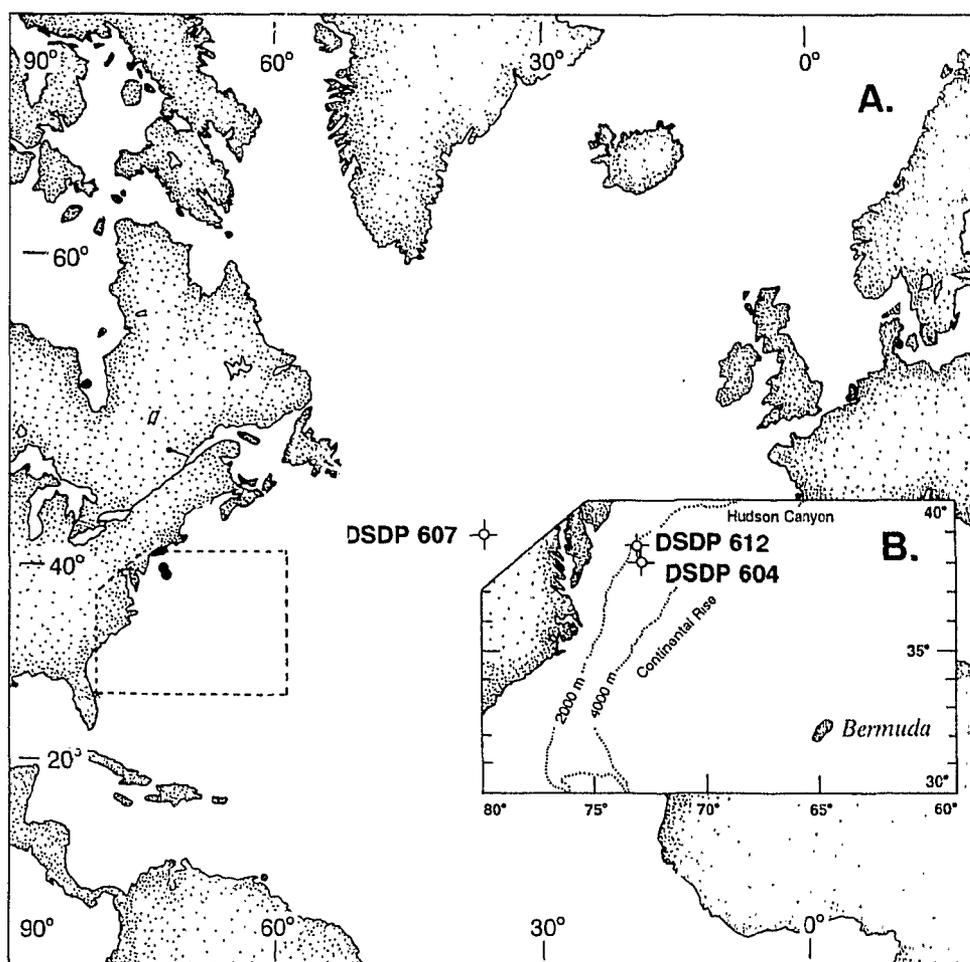


Figure 1. Map of the North Atlantic Ocean showing the location of the sites studied. Sites 604 and 612 are on the continental margin off New Jersey, and Site 607 is on the western flank of the Mid-Atlantic Ridge. The inset shows the location of the sites on the continental margin in relation to the bathymetry; the 2000m and 4000m contours are shown in the inset.

This study concentrated on the late Quaternary record (late Illinoian -Holocene) because global climate is best documented for the last 2 glacial-interglacial cycles, facilitating paleoceanographic interpretations and permitting a higher degree of confidence in climatostratigraphic correlation between late Quaternary sediments from sites on the slope, rise and the mid-ocean ridge. Complex depositional patterns on the continental margin, with frequent hiatuses and reworking, prohibited definite climatostratigraphic assignments in pre-late Quaternary sediments at Sites 612 and 604 on the mid-slope and upper rise; in the early-mid Quaternary, therefore, the relationship between climate change and the geologic record was examined without good chronostratigraphic control.

Comparing the planktonic foraminiferal, dinocyst, and pollen records at each site provided insights into the distribution and transport of these microfossils in the western North Atlantic, and into the degree of reworking of pre-Quaternary sediments from the continental slope into the Quaternary sequence of the continental rise in response to climatic and accompanying oceanographic changes. Cross-correlation also allowed paleoecological/ climatostratigraphic interpretations based on any one group of microfossils to be substantiated by data from other groups of microfossils. Dinoflagellate cysts are potentially valuable paleoecological tools, especially in neritic environments where calcareous microfossils are relatively rare; however, their modern distribution and ecological significance is less well understood than that of planktonic foraminifera. Comparison of the dinocyst records with planktonic foraminiferal and pollen records permitted testing of the ecological significance of these taxa compared to other reports in the literature (Mudie and Aksu, 1984; Aksu and Mudie, 1985; Scott *et al.*, 1984; Scott *et al.* 1989a; Aksu *et al.* 1989).

The records from the mid-slope, upper rise, and mid-ocean ridge reflect the way in which global change affected these three regions. While the similarities in the records permit correlation between sites and with the established Quaternary climatostratigraphy, the differences between the sites provide the key to understanding the effect of changes in global climate on water masses and oceanic circulation. These oceanographic changes affect the biota in the surface waters and on the seafloor, and sedimentary processes in the deep ocean and on the continental margin. The following questions were addressed by this study:

1) Can a climatostratigraphic definition of the Pleistocene established on the basis of faunal changes in low-latitude, deep-sea cores be applied to the continental margin? Is the identification of stratigraphic datums, such as the Plio-Pleistocene boundary, possible in continental margin sites subjected to erosion and intensive reworking?

2) Continental margins geographically link the terrestrial and deep sea environments. Can studies of continental margin sediments improve correlations between the terrestrial and oceanic record of the Quaternary?

3) How did progressive climatic deterioration through the Pleistocene affect surface water masses and circulation in the North Atlantic Ocean? How long have the present water mass configuration and circulation pattern existed? How and when did the Slope Water mass originate?

4) Can micropaleontological, sedimentological, and isotopic changes in the Quaternary record of the margin provide insight into the relationship between oceans and ice sheets? What was the effect of mid-latitude ice sheets on surface water masses and circulation in the temperate western North Atlantic? What is the relationship between oceanographic changes in the temperate western North Atlantic Ocean and global climate?

5) Unlike planktonic foraminifera, dinocysts are relatively common in neritic

environments. Can dinocyst records increase the accuracy of paleoclimatological/ paleoecological interpretations and facilitate correlation between deep-sea and continental margin sites?

6) Do pollen, dinocyst, and planktonic foraminiferal records from the continental margin closely match those from the mid-ocean? If not, what does this tell us about the ecology and transportation of these microfossils in marine environments? Do dinocysts show the same patterns of distribution as planktonic foraminifera (whose living environment they share) or do they correlate more closely with benthic foraminifera to confirm hypotheses regarding bottom water transport of dinocysts in pelagic environments?

7) Why do middle Eocene strata outcrop on the lower continental slope off New Jersey? What are the processes denuding the lower slope, and how long have they been active? How can the evidence for gravity-driven mass wasting processes and for along-slope bottom currents be reconciled?

8) Is the Vail depositional model applicable to the Quaternary sequence on the U.S. Atlantic continental margin? Have the mechanisms of erosion of the New Jersey continental margin remained constant, or have they varied through the Quaternary?

## **1.1 GEOLOGY OF THE U.S. ATLANTIC CONTINENTAL MARGIN NORTH OF CAPE HATTERAS**

The geology and physical oceanography of the temperate western North Atlantic Ocean between Cape Hatteras and Georges Bank are briefly summarised in this introductory chapter to provide the background information necessary for interpretation of the data in this thesis. For more detailed references the reader is directed to excellent summary volumes such as Emery and Uchupi (1972, 1984), Poag (1985a), and Milliman and Wright (1987).

The continental slope is a complex sedimentary environment (Doyle and Pilkey, 1979). The continental slope of the eastern United States is divided morphologically into 3 major segments (Shor and McClennen, 1988): 1) Cape Lookout to Georges Bank, 2) the Florida-Hatteras slope, and 3) the slope seaward of the Blake Plateau. DSDP sites 604 and 612 lie in the segment from Cape Lookout to Georges Bank (Figure 1). The depth of the shelfbreak off New Jersey occurs at ca. 120 m. The base of the slope is characterised by a marked decrease in gradient, typically between 2000 and 2200 m water depth. The average width of the slope between Cape Cod and Cape Lookout is 40 km. Average gradients on the upper slope are 1-2 degrees, on the lower slope, 3-8 degrees (Knebel, 1984). Submersible observations (*e.g.* Ryan *et al.*, 1978) showed that most of the upper slope north of Cape Lookout is incised by canyons whose wall gradients are commonly 30 degrees and are occasionally nearly vertical (Twichell and Roberts, 1982). The "average" gradients, therefore, apply to intercanyon regions of the slope on a regional scale, and thus can be misleading. The slope between Cape Lookout and Georges Bank is cut by many large canyons, smaller canyons, and gullies; this contrasts with the small number of canyons on the Florida-Hatteras slope. GLORIA surveys identified 51 canyons cutting the upper slope between Hudson and Baltimore canyons (Twichell and Roberts, 1982); the longest undissected slope segment is 40 km, located immediately southwest of Hudson Canyon (Shor and McClennen, 1988).

A standard stratigraphic reference section was established during DSDP Legs 93 and 95 along USGS seismic reflection line 25 (Figures 2 and 3). The "New Jersey transect" cuts

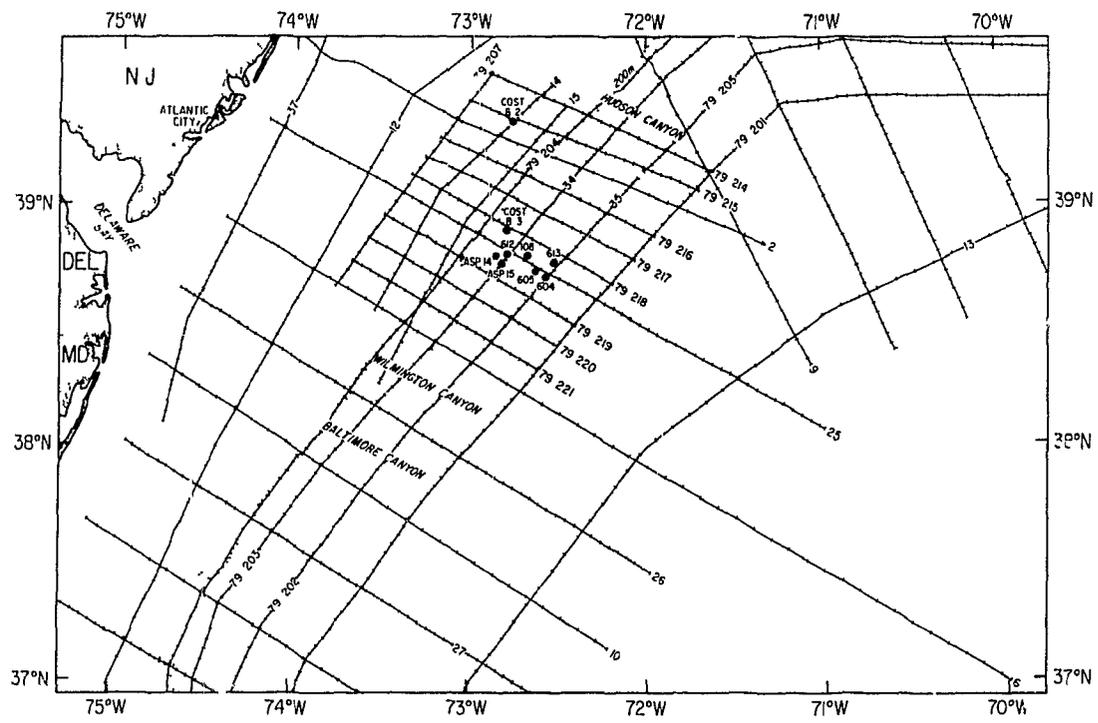


Figure 2. The location of Sites 604 and 612 along USGS seismic line 25, in relation to the grid of seismic lines run in the area of the Baltimore Canyon Trough (reprinted by permission from Poag, 1985).

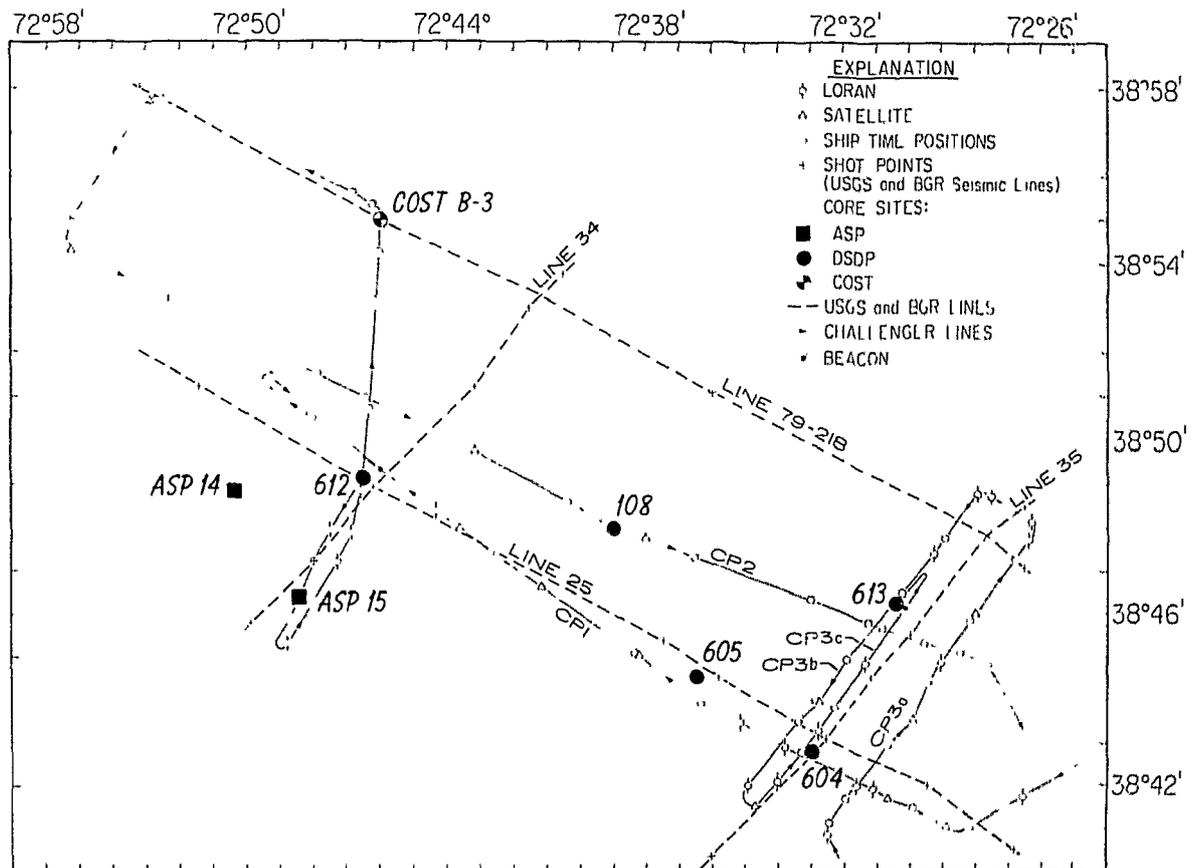


Figure 3. Location of the "New Jersey Transect", and location of DSDP Sites 604 and 612 relative to seismic lines 25, 34, and 35 (reprinted by permission from Poag, 1985).

across the Baltimore Canyon Trough, a major sedimentary basin, which has been intensely studied (Figure 2), due largely to its proximity to major oceanographic research institutions. There are two accretionary prisms on the margin, one on the upper slope-shelf and one on the upper rise-abyss, separated by an erosional outcrop of middle Eocene strata on the lower to mid-slope (Poag and Mountain, 1987). Thick accumulations of Neogene-Quaternary sediments on the shelfbreak and upper slope thin rapidly and downlap onto the Eocene sediments on the lower slope, and thicken again seaward on the upper rise (Figure 4). The lower slope is an area of extremely diverse morphology, with canyons, valleys and troughs formed in outcrops of Eocene shales, marls, and chalk (Robb *et al.*, 1983).

Pre-Quaternary sediments in the study area reflect an early rifting stage, then a carbonate-platform stage, followed by a clastic infill stage (Schlee *et al.*, 1988). The rifted Atlantic margin was filled by a Mesozoic rift-drift sequence; Triassic syn-rift sediments were deposited, followed by buildup of a carbonate platform. A conspicuous unconformity within the rise prism separates the lower sequence of Paleogene hemipelagic ooze and claystone from a complex association of Neogene slump deposits, turbidites, hemipelagic clays and channel fill that thickens seaward to 2.2 km under the mid-rise. Large, prograding deltas enriched in organic carbon and siliceous microfossils developed during the middle Miocene, and terrigenous sedimentation has dominated in the area to the present. Neogene slope/rise sediments consist of silty clays, glauconite sands and sandy pebbly conglomerates (Poag and Valentine, 1988). Neogene rocks are particularly thick in the Baltimore Canyon Trough, more than 1000 m thick in the depocenter (Poag, 1980, 1985b).

The distribution of surficial sediments on the continental margin in the area of the

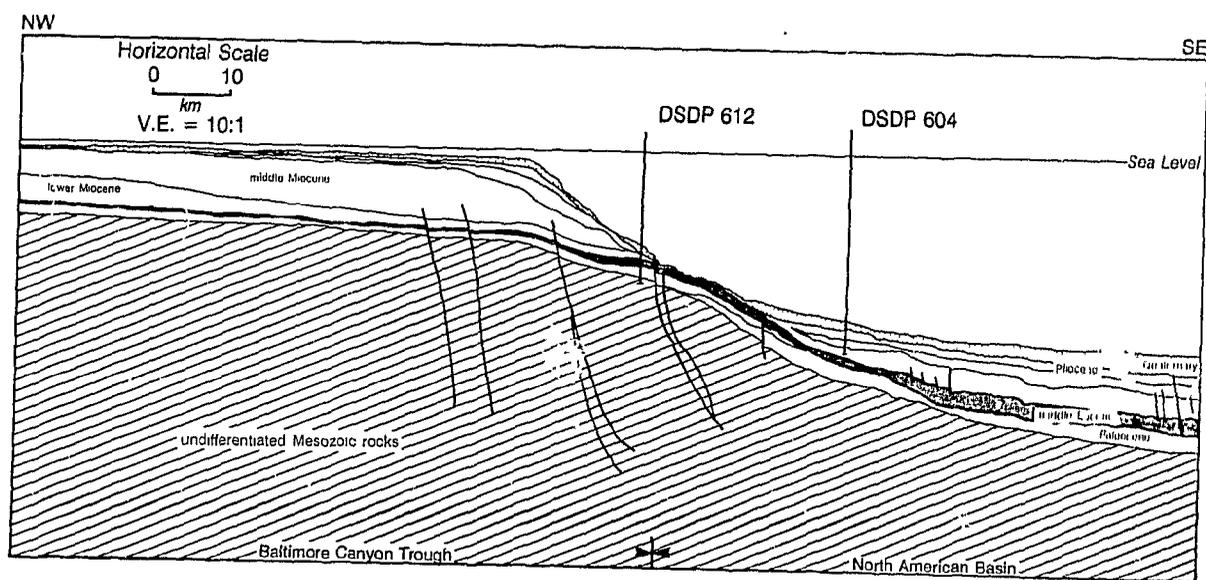


Figure 4. Stratigraphy of the continental margin off New Jersey, interpreted from seismic reflection profiles along USGS Line 25, with groundtruthing from numerous boreholes. This figure is simplified from Poag (1985a), showing only the Cenozoic stratigraphy. The Quaternary sediments (stipple), and the middle Eocene chinks (black) which outcrop on the lower slope, are highlighted, and the location of the sites studied, DSDP Sites 604 and 612, is shown.

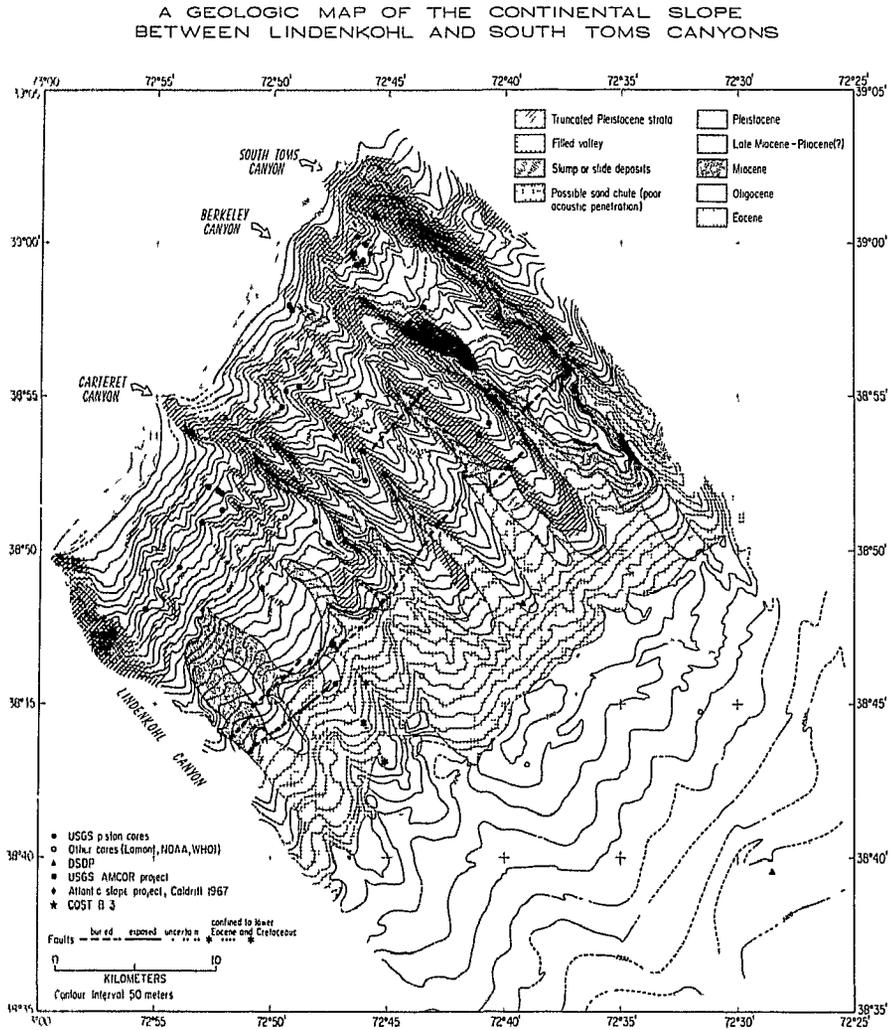


Figure 5. Surficial geology of the continental slope between Linden Kohl and South Toms Canyons, reproduced by permission from Poag (1985c).

New Jersey transect is illustrated in Figure 5. Pleistocene sediments exist on the mid-slope as fingerlike ridges trending downslope, separated by outcrops of Eocene to Miocene sediments.

The Pleistocene section consists of about 400 m of sand and silty clay beneath the outer shelf and upper slope, and it thins rapidly landward and seaward (Figure 4). Pleistocene sediments unconformably overlie sediments of Oligocene to Miocene age on the upper slope. The relief on the upper slope is generally subdued by thick drape of Pleistocene sediments which has buried the Tertiary surfaces (Hampson and Robb, 1984); Pleistocene deposits have extensively overlapped the mid-Eocene exposure on the lower slope, so that the Eocene outcrop belt is now much narrower than it was in the past. Widespread fluctuations in coastal onlap and deltaic outbuilding on the shelf resulted in extensive erosion of the slope landward during the Pleistocene, and the accumulation of a broad onlapping continental rise wedge (Schlee *et al.*, 1979). The Pleistocene sediments forming this pronounced seaward-thickening wedge on the upper rise are described as a complex association of slump deposits, turbidites, hemipelagic clays, and channel fill (Hampson and Robb, 1984).

## 1.2 PHYSICAL OCEANOGRAPHY OF THE WESTERN NORTH ATLANTIC OCEAN

Circulation of surface water in the North Atlantic is illustrated in Figure 6. Prevailing winds separate the surface waters of the North Atlantic into gyres. The subtropical gyre is a clockwise gyre driven by the overlying anticyclonic wind systems that blow around the subtropical high pressure regions; westerlies blow off North America at 45°N latitude and trade winds blow westward from Africa at 15°N latitude.

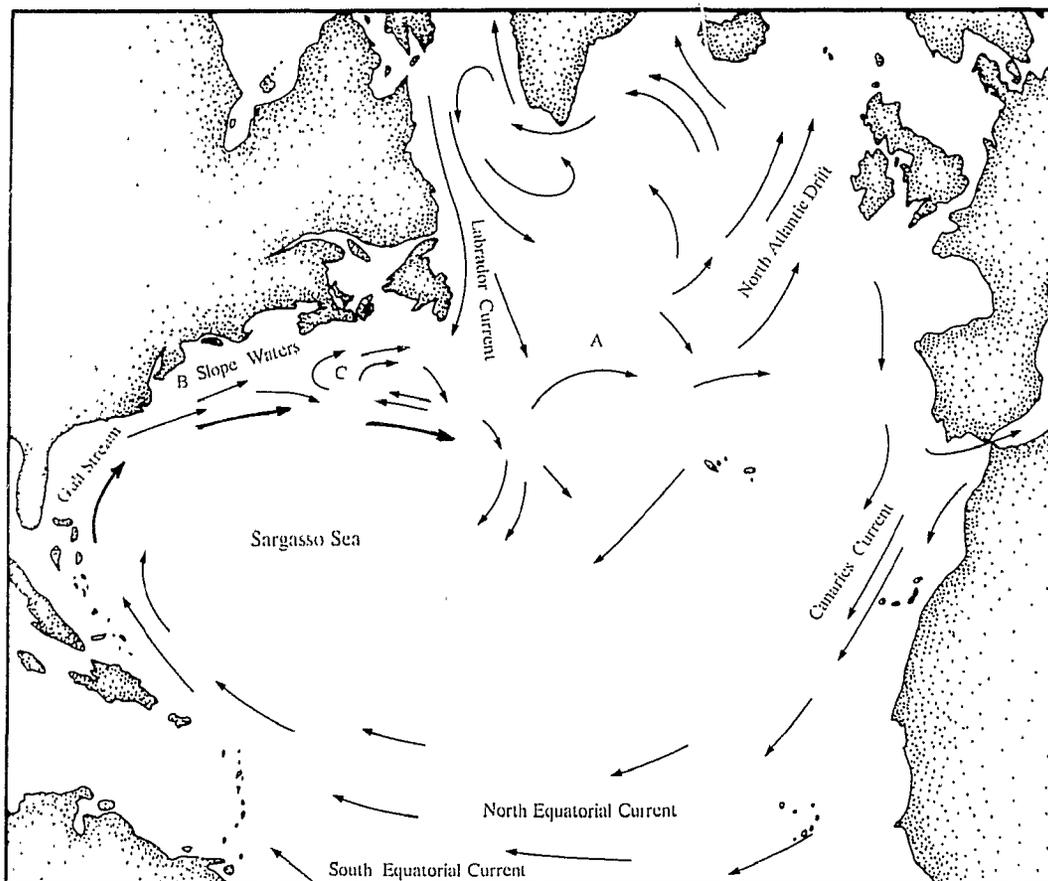


Figure 6. Surface water masses and current in the North Atlantic Ocean, modified from Open University Course Team (1989). The path of the Gulf Stream is indicated by bold arrows. The location of the transitional water mass is indicated by the letter A, and the Western and Eastern Slope Water masses are denoted by the letters B and C, respectively.

However, the centres of the atmospheric gyres tend to be displaced toward the eastern side of the oceans, while the centres of the oceanic gyres are displaced toward the western side, especially in the northern hemisphere, resulting in fast, intense, deep, and narrow western boundary currents, such as the Gulf Stream and the Kuroshio Current (Open University Course Team, 1989).

The Gulf Stream extends from the Straits of Florida to the Grand Banks, and carries warm water north along the east coast of North America. Much of the water carried by the South Equatorial Current is deflected north of the equator by the Brazilian coastline, and joins the North Equatorial Current; much of this combined flow enters the Caribbean Sea and passes through the Yucatan Channel to the Gulf of Mexico, and eventually flows through the Straits of Florida as the Gulf Stream. Between the Straits of Florida and Cape Hatteras, the Gulf Stream flows along the continental slope; in this region the flow remains narrow and well-defined, and meanders are restricted to a maximum of 55 km amplitudes. Beyond Cape Hatteras, the Gulf Stream is deflected seaward and moves into water 4000-5000 m deep, and the flow becomes more complex. As pointed out by Fuglister (1955), however, the Gulf Stream is not a single, tortuous current extending from Cape Hatteras to the Grand Banks, but rather is spatially and temporally variable, comprised of relatively short-lived filaments, meanders, and eddies. Because there are no topographic constraints on the Stream, meanders with amplitudes exceeding 350 km are common, often giving rise to Gulf Stream rings; amplitudes of over 500 km were measured south of Nova Scotia (Robinson *et al.*, 1974). This short-term variability has only recently begun to be understood by oceanographers, who have long been familiar with the mean motion of large-scale ocean currents (Open University Course Team, 1989).

Beyond the Grand Banks, the broader, more diffuse current is referred to as the North

Atlantic Drift (NAD). Much of the NAD flows south-east and recirculates in the subtropical gyre; most of the remainder continues north-eastward between Britain and Iceland, with major northward branches forming the Irminger Current flowing into the Labrador Sea. A mixed water body known as the transitional water mass occupies a triangular-shaped region of the North Atlantic with its apex at the Grand Banks and its base along the coast of northern Europe (McIntyre *et al.*, 1976). It is formed by the mixing of warm North Atlantic Drift waters with the cold waters occupying the counterclockwise subpolar gyre.

The central water mass in the subtropical gyre of the North Atlantic is known as the Sargasso Sea, a 700-800 m deep pool of warm water which forms the upper layer of most of the North Atlantic Ocean (Pickard and Emery, 1982). This water mass is characterised by relatively warm water (25-28°C), with little terrigenous input and low biological productivity.

In the space between the Gulf Stream and the North American coast there is a southwestward-flowing coastal current with an elongated counter-clockwise gyre between it and the Stream (Pickard and Emery, 1982). This constitutes the Slope Water, which is slightly fresher and cooler than the Gulf Stream or the Sargasso Sea because of the influx of Labrador Sea Water, as well as the drainage of several major rivers (*e.g.* Hudson, St. Lawrence). In the summer, the Slope Water rarely exceeds 23°C and 36‰, while the Gulf Stream reaches 28°C and 36.9‰ (Milliman and Wright, 1987). The Slope Water, sometimes geographically referred to as the Slope Sea, is comprised of two distinct water masses (Figure 6): the Western Slope Water occurs principally west of 66°W, and is relatively warm and saline, while the Eastern Slope Water (referred to as the Labrador Slope Water by Gatién, 1976), is cooler and less saline, and occurs mostly east of 66°W. Sea surface temperatures in the Slope Water show much greater seasonal

variability than do the waters above the slope southwest of Cape Hatteras or those of the Sargasso Sea. According to Csanady and Hamilton (1988), some of the key features of circulation in the upper 500 m of the Slope Water are inflow of Coastal Labrador Sea Water across the Grand Banks and isopycnal advection from the Gulf Stream thermocline, the total draining eastward, and a closed cyclonic gyre in the western Slope Sea, transporting ca  $3 \times 10^6 \text{ m}^3/\text{s}$  along the New Jersey coast southward. Deeper layers move southwestward through the Slope Water. Fluctuations in the Slope Water are generated by instabilities of the Gulf Stream (e.g. propagating eddies, or rings, and more wavelike disturbances). These eddies are in turn greatly modified by the underlying topography (Schmitz *et al.*, 1987), and can be much stronger than the mean. Deep eddies at the base of the Gulf Stream are thought to be responsible for north to northeast flowing currents measured in places in the Slope Water north of Cape Hatteras (Tucholke, 1987).

Circulation at mid-latitudes in the western North Atlantic basically consists of southward flow below 1400 m (*i.e.* beneath the northward-flowing "smoothed Gulf Stream" of Roemmich and Wunsch, 1985). Water masses are divided vertically by temperature into 3 main classes (Worthington, 1976): 1) deep water, colder than  $4^\circ\text{C}$ , which consists almost entirely of North Atlantic Deep Water (NADW). This southward thermohaline flow of cold, dense water of North Atlantic origin extends across the entire width of the ocean, though it is deepest and strongest in the west near the Gulf Stream where the Western Boundary Undercurrent (WBUC) flows along the lower rise to the southwest along the entire U.S. margin north of Cape Hatteras; 2) thermocline, water  $4\text{-}17^\circ\text{C}$ , which consists almost entirely of Western North Atlantic Water (WNAW, Wright and Worthington, 1970); and 3) warm water (warmer than  $17^\circ\text{C}$ ); warm water is a seasonal phenomenon north of Cape Hatteras, except in warm core rings. The depth of seasonal warming is approximately 200 m in the Slope Water. The depths at which the different

water masses occur varies spatially in the North Atlantic, especially between the Sargasso Sea and the Slope Water; the boundary between these water masses, the Gulf Stream, is a region of strongly sloping temperature/salinity gradients. Within the Sargasso Sea, for instance, the 6°C surface lies at about 1100 m, while at the inshore edge of the Gulf Stream, this isotherm lies at 800 m.

## CHAPTER II

### STUDY APPROACH AND LABORATORY METHODS

#### 2.0 INTRODUCTION

Despite the fact that the Baltimore Canyon Trough is one of the most intensively studied sedimentary basins on the U.S. continental margin, relatively few high resolution micropaleontological studies of the Quaternary sediments have been done in this region, due to the complexity of the depositional environment. This study examined hydraulic piston cores from three sites at mid-latitudes in the western North Atlantic. The record of two sites on the continental margin in the Baltimore Canyon Trough (Sites 604 and 612) were compared with the record from Site 607 in the deep sea. These sites form a transect from the mid-slope off New Jersey to the western flank of the Mid-Atlantic Ridge (Figure 1). Cores were collected during DSDP Legs 93 (Site 604: Shipboard Scientific Party, 1987a), 94 (Site 607: Shipboard Scientific Party, 1987b), and 95 (Site 612: Shipboard Scientific Party, 1987c).

Quaternary sediments from these sites were analysed for pollen, dinoflagellate cysts, and planktonic foraminifera. Isotopic analyses of ratios of  $^{18}\text{O}:^{16}\text{O}$  in planktonic foraminiferal tests (*Neogloboquadrina pachyderma*) were obtained from two sites, DSDP Leg 94 Site 607, and DSDP Leg 95 Site 612. No isotopic analysis was done on the sediments from Site 604 because core recovery was poor, foraminiferal concentrations are very low through most of the Upper Pleistocene section, and because reworking is common at this site. However, Ganssen (1987) published isotopic data for this site, which is referred to in the discussion of the microfossil data. Radiocarbon dates were obtained from intervals of particular significance in late Wisconsinan

sediments from Sites 607 and 612.

The Quaternary sediments from Sites 607 and 612 are relatively homogeneous, so that the core descriptions in the initial reports volumes suffice (Shipboard Scientific Party, Leg 94, 1987b; Shipboard Scientific Party, Leg 95, 1987c). Because the sedimentology of Site 604 is highly variable (Shipboard Scientific Party, Leg 93, 1987a), however, the volume of sediments > 0.063 mm was measured for each 10cc sample, and a visual analysis of the composition of sand-sized fraction of the sample was made during foraminiferal analysis; percentages of 4 major sediment types: terrigenous sand, glauconite, biogenic CaCO<sub>3</sub> (basically planktonic foraminiferal tests) and biogenic silica (reworked Tertiary radiolarians and diatoms), were estimated.

## **2.1 PALYNOLOGICAL ANALYSIS**

Palynomorphs are microfossils with walls made of complex organic material which is acid insoluble; these include dinoflagellate cysts, pollen, and terrestrial spores.

### **2.1.1 Dinoflagellate cysts**

Dinoflagellates are single-celled eukaryotes belonging to the kingdom Protista. Because about half of the extant dinoflagellate species are photosynthetic autotrophs possessing chloroplasts, and the remainder are nonphotosynthetic heterotrophs, they have been classified under both the International Code of Botanical Nomenclature and the International Code of Zoological Nomenclature; most paleontologists follow the botanical classification.

The life cycle of a dinoflagellate consists of a motile stage (theca) and an encysted stage. Biologists classify living dinoflagellates according to the theca, cell characteristics, and the life history of the organism, while paleontologists classify dinoflagellates according to the characteristics of the fossilised cyst (*e.g.* paratabulation, archeopyle, and ornamentation); a dual taxonomy has therefore developed (Loeblich, 1970; Sarjeant, 1974). Several workers have attempted to unify the taxonomy of fossil and living dinoflagellates (*e.g.* Harland, 1982; Bujak and Davies, 1983), but only about a third of modern cysts have been correlated with their motile stages, and it is difficult to infer anything about the motile stage of many dinoflagellate from their resting cysts. Only about 10% of modern genera produce geologically preservable (acid-resistant) cysts, and the probability of fossilisation varies widely among the taxa which produce acid-resistant cysts (*e.g.* Dale, 1976). There are geologically-preserved cysts produced by dinoflagellates in a wide variety of environments, from freshwater to open marine, however, and dinoflagellates are important components of the marine phytoplankton (Marshall, 1984), thus forming an important part of the marine food chain.

#### **2.1.1.1 Dinocysts as paleoecological indicators**

Because the fossil record of dinoflagellates represents one stage in the life cycle of the relatively small percentage of dinoflagellate species that produce cysts with acid-resistant walls, the fossil record is selective and incomplete. Observation of modern plankton in surface waters and comparison with resting cysts in bottom sediments is the ideal method for determining how environmental conditions (*e.g.* sea-surface temperature, salinity, etc.) affect dinoflagellate distribution, and how these are reflected in the geologic record; few studies of this nature have been done, however, so that an accurate data base does not exist. Relatively little information is

available even on the distribution of dinocysts in surface sediments, although the database is continuously growing (e.g. Harland, 1977; Wall *et al.*, 1977; Mudie, 1980; Harland, 1983; Turon, 1984; Mudie, 1984; Mudie and Short, 1985; Dale, 1986; Mudie *et al.* 1990).

Mudie *et al.* (1990), Mudie (in press), and Edwards *et al.* (1991) have recently reviewed the distribution of dinocysts in the northern North Atlantic and the application to paleoclimatic reconstruction. They employed the Imbrie-Kipp CABFAC and multiple regression programs to analyse the database. They found that the slope ecofacies was represented by *Operculodinium centrocarpum* and *Spiniferites* spp., *Multispinula? minuta* + *Brigantedinium* spp. comprised the Labrador Current-NW Greenland shelf ecofacies, and the ice margin was represented by *Leiosphaeridia*, *Brigantedinium* and *Polykrikos*; *Polycystodinium zoharyi* and *Impagidinium aculeatum* characterised the Gulf Stream and *Nematosphaeropsis labyrinthus*, *Bitectatodinium tepikiense* and *Impagidinium sphaeridium* marked the NAD. *O. centrocarpum* has also been associated with the Gulf Stream (Harland, 1983).

*Nematosphaeropsis labyrinthus* and the genus *Impagidinium* characterise oceanic environments, such as the Sargasso Sea (Wall *et al.*, 1977; Mudie, 1987). For temperate, oceanic environments, in samples with >2% *Impagidinium* spp., Edwards *et al.* (1991) developed an *Impagidinium* Index:  $100 (I. aculeatum + I. striatum) / \text{total } Impagidinium \text{ specimens}$ . Values of the *Impagidinium* Index greater than 40 are equal to winter sea surface temperatures (WSST) of 10-13°C, values between 0 and 40 correspond to WSST of 6-10.5°C, and values of 0 correspond to WSST of 0-10°C.

The ratio of autotrophic gonyaulacoid dinocysts versus protoperidinoid dinocysts, which

lack chlorophyll and which eat diatoms, is generally employed as a crude paleoecological tool; the G:P ratio decreases with decreasing available light, such as beneath perennial pack ice (Mudie and Short, 1985), and the ratio tends to increase with distance from shore (Harland, 1973). The increase in heterotrophic protoperidinioid cyst species during the Pleistocene in the Norwegian Sea noted by Mudie (1989), for example, is consistent with other data suggesting an increase in pack ice in the Arctic Ocean (*e.g.* Gilbert and Clark, 1983; Mudie, 1985; Aksu and Mudie, 1986).

The approach taken in this study was to compare downcore variations in dinocyst assemblages with variations in planktonic foraminifera, pollen and terrestrial spores in the same samples. Because the ecology of these two groups of microfossils is fairly well understood, paleoecological interpretations can be made on the basis of the planktonic foraminiferal and pollen record, and projected onto dinocyst assemblages, increasing our understanding of the paleoecological significance of dinocyst assemblages.

### **2.1.2 Pollen and terrestrial spores**

Pollen grains and terrestrial spores are the male gametophytes of vascular plants (Knox, 1979). Their outer cell walls (exines) are composed of a highly resistant complex organic substance called sporopollenin, susceptible only to strong oxidation. Fossil pollen grains and terrestrial spores are therefore commonly preserved in sediments in a wide variety of sedimentary environments. They are small (ca. 10- 300  $\mu\text{m}$ ), and therefore only sediment samples of ca. 5 cc are sufficient for analysis.

Because they reflect terrestrial vegetation, pollen and terrestrial spores are indicators of

continental climates. The abundance of fossil pollen, and its application in vegetational reconstruction, climatic interpretation, and Quaternary stratigraphy has been appreciated since the study of von Post (1916). The pollen assemblage is a complex product of several factors, including pollen production of a given taxon, its resistance to corrosion, the mode of pollen dispersion and transport, and sedimentary environment (Birks and Birks, 1980). Only about 30 of more than 300 families of flowering plants are adapted for pollen dispersal in air currents (Knox, 1979), but wind-dispersed pollen strongly dominates fossil pollen assemblages.

When interpreting paleovegetation and paleoclimates from pollen assemblages, it is important to account for the fact that pollen assemblages are not direct reflections of vegetation. The amount of pollen produced is taxon-specific, and the probability of preservation of different pollen types depends on the composition of the exine, which again is species-specific; as a result, some taxa are over-represented in the fossil record (*e.g. Pinus*) while others (*e.g. Acer*) are under-represented (Delcourt *et al.*, 1984; Webb and McAndrews, 1976).

#### 2.1.2.1 Pollen in the marine realm

The ultimate aim of palynology is often to reconstruct vegetational succession and interpret paleoclimates from the vegetational reconstruction by comparison with modern analogs. The climatic reconstructions are also used for stratigraphic correlation, since the Quaternary is defined and subdivided climatostratigraphically. Because pollen comes from the continent, and pollen assemblages are proxy climatic indicators, pollen records in marine sediments provide a mechanism for correlating marine and terrestrial sediments. Although these attributes suggest that palynomorphs are potentially ideal paleoclimatic indicators, in practice palynological records are

difficult to interpret paleoecologically.

There are also special considerations when attempting to correlate marine pollen records with terrestrial records (Groot and Groot, 1966). Interpreting marine pollen records requires an understanding of the role of ocean and air currents in transporting terrestrial palynomorphs, and the hydrodynamic properties of different types of palynomorphs resulting in varying distances of transport, both of which affect the resulting assemblages in different parts of the ocean. Previous palynological studies in the western North Atlantic, however, have illustrated that the distribution of pollen on the continental margin is similar to the distribution onshore in eastern North America, and therefore can provide meaningful interpretations of paleovegetation and paleoclimates (e.g. Mudie, 1982; Heusser, 1983; de Vernal, 1986). Some things are common to all studies of pollen in the marine realm; those pollen types which transport long distances (e.g. bisaccate pollen grains, especially *Pinus*) tend to be overrepresented in marine sediments and to increase in abundance away from the continent relative to those pollen types which do not tend to be transported long distances.

### **2.1.3 Laboratory methods**

Sediment samples of 5 cc volume were treated using standard processing techniques to digest silicates and carbonates using acid, thus concentrating the acid-resistant palynomorphs for analysis. Samples were first disaggregated using a 0.02% Calgon solution; one hour to overnight in a warm water bath, depending on clay and moisture content. Samples were "spiked" with a known number of marker grains; two different types of marker particles were used: microspheres (Ogden, 1986), and *Lycopodium* spores (Stockmarr, 1971). Although theoretically, the use of

different marker particles should not affect resulting estimated palynomorph concentrations, some samples were spiked with both types of marker grains, and resulting concentrations did differ significantly. The effect of using different marker grains in processing for palynomorphs is addressed in Appendix A. Calcium carbonate was then dissolved using HCl; dilute HCl was first added to digest most of the rapidly reacting carbonates. This treatment was followed by a treatment of full-strength HCl of at least two hours duration in a hot water bath; this step was repeated until all carbonates were digested. Silicates were then dissolved using concentrated (48-52%) HF; samples were left in HF in a hot water bath at least 12 hours. As with treating for carbonates, this step was repeated until all silicates were dissolved. In samples rich in terrigenous silicates, the coarser clastics were carefully removed by swirling after the first HF treatment, and the coarse residues retained to check for palynomorphs lost by swirling; because swirling was done carefully and conservatively, with only the coarsest clastics separated by this process, no palynomorphs were lost in this process (as verified by spot-checking the decant), which reduced the time and amount of acid required to treat each sample. Samples were stained using safranin-O, and mounted in glycerine gel.

Identifications were made using a Zeiss Ultraphot microscope; a magnification of 250X sufficed for most palynomorphs, although magnifications of 400X and sometimes 1000X were sometimes necessary to distinguish small features. Dinocyst percentages are based on a minimum of 100 dinoflagellate cysts and minimum of 100 upland pollen grains comprise the pollen sum, except at Site 607, where very low palynomorph concentrations existed; the number of palynomorphs constituting the sum at this site are indicated in the data table. Pollen identifications follow the key of McAndrews *et al.* (1973). An abbreviated taxonomy of dinocyst taxa identified in this study is found in Appendix B, along with photoplates (Plates 4-7)

several of the most common dinocysts.

## 2.2 FORAMINIFERAL STUDIES

### 2.2.1 Planktonic foraminifera

Foraminifera are protozoans assigned to the class Sarcodina because they possess pseudopodia (Boersma, 1978). There are relatively few genera within the order Foraminiferida that are planktonic; all of these have calcareous tests. Their distribution is controlled by water masses (Bradshaw, 1959; Be and Tolderlund, 1971; Kipp, 1976; Be, 1977). Morphologic variation within species is also climatically controlled, e.g. sinistral forms of *Neogloboquadrina pachyderma* with a crystalline, heavily calcified test are most abundant when surface water temperatures are less than 8°C, while less calcified, reticulate, predominantly dextrally coiled tests are most common in surface waters warmer than 8°C (Be and Tolderlund, 1971; Reynolds and Thunell, 1986). Coiling ratios are therefore commonly employed as indices of sea surface temperatures. Ratios of sinistral:dextral tests of *N. pachyderma* were used as such in this study; "coiling ratios" refer hereafter to ratios of sinistral:dextral *N. pachyderma*.

The climatic/water mass affinities defined by Kipp (1976) were primarily followed in this work, since it concentrated on the North Atlantic; Table 1 shows the distribution of common species of planktonic foraminifera on the North Atlantic seabed defined by Kipp.

Table 1. Distribution of common planktonic foraminiferal species on the North Atlantic seabed (from Kipp, 1976).

<i>POLAR</i>	<i>SUBPOLAR</i>	<i>TRANSITIONAL</i>	<i>SUBTROPICAL</i>	<i>TROPICAL</i>	<i>GYRE-MARGIN</i>
<i>Neogloboquadrina pachyderma (sin)</i>	<i>Neogloboquadrina pachyderma (dex)</i>	<i>Globorotalia inflata</i>	<i>Globigerina falconensis</i>	<i>Globigerinoides ruber</i>	<i>Neogloboquadrina dutertrei</i>
<i>Globigerina quinqueloba</i>	<i>Globigerina bulloides</i>		<i>Globorotalia scitula</i>	<i>Globigerinoides sacculifer</i>	<i>Globorotalia menardii</i>
	<i>Globigerinita glutinata</i>		<i>Globorotalia truncatulinoides</i>		<i>Globorotalia tumida</i>
			<i>Globorotalia hirsuta</i>		<i>Pulleniatina obliquiloculata</i>

### 2.2.2 Laboratory methods

Sediment samples of 10cc volume were washed through a #230 mesh (0.063 mm) sieve for foraminiferal studies, retaining the 0.063 mm fraction. Only the greater than 0.150 mm fraction was examined for planktonic foraminifera, to eliminate juveniles which are difficult to identify (Kellogg, 1983). Because micropaleontologists commonly examine this size fraction for planktonic foraminifera, the results of this study can easily be compared with their work. The samples were dried at low temperatures (not exceeding 40°C). Foraminiferal tests were concentrated from samples rich in clastic sediments by floating using carbon tetrachloride, which has a specific gravity of 1.58. An Otto microsplitter was used to separate samples rich in foraminifera into fractions of a minimum of 300 individuals, and the entire fraction was picked and identified. Identifications basically follow Saito *et al.* (1981); an abbreviated taxonomy follows in Appendix C, together with photoplates (Plate 1-3) illustrating the most common planktonic foraminiferal taxa.

### 2.3 STABLE ISOTOPES

Dextral tests of *Neogloboquadrina pachyderma* and *Globigerina bulloides* were picked from DSDP Site 612, and sinistral and dextral tests of *N. pachyderma* were picked from DSDP Site 607 for stable isotope analysis. A minimum of 80 specimens from a narrow size range (0.212-0.250 mm) were picked, since inherent ratios of  $^{18}\text{O}:^{16}\text{O}$  differ not only between species, but also change throughout the lifetime of the foraminifer, and hence the size of the test (Williams *et al.*, 1981; Deuser *et al.*, 1981). Samples were ultrasonically cleaned in methanol, dried at 50°C, and roasted at high vacuum for 1 hour at 400°C. The isotopic composition of the tests was

measured using a VG-602D mass spectrometer with on-line carbonate preparation system (Shackleton and Opdyke, 1973; Duplessy, 1978); the carbonate was reacted with 100% orthophosphoric acid at 50°C. Values are expressed in parts per mil relative to the laboratory standard, Carrera marble calibrated to universal PDB standard. All values are Craig-corrected (Craig, 1957).

## 2.4 ESTIMATES OF SEA SURFACE TEMPERATURES (SST)

Sea surface temperatures (SST) were calculated from dinocyst assemblages at Site 607 using the method of Imbrie and Kipp (1971), with the transfer functions of Mudie *et al.* (1990). The standard errors associated with estimates of summer sea surface temperature are +/- 2.6°C, while those associated with estimates of winter sea sea surface temperatures are +/- 2.4°C.

## 2.5 QUATERNARY CHRONOLOGY

The Quaternary has been defined and subdivided according to geologic-climate, or climatostratigraphic units (see chapter 5). Because there is still no common consensus regarding the subdivision of the Pleistocene (or even the base of the Pleistocene) the chronostratigraphic framework for this study is outlined in this section. The basis of subdividing the cores in this study following geologic-climate or climatostratigraphic units and comparison with the established chronostratigraphy is discussed more fully in a later section (Chapter 5).

The chronostratigraphic framework used in this study is illustrated in Figure 7. There are relatively few biostratigraphic datums useful in defining and subdividing the Quaternary. The

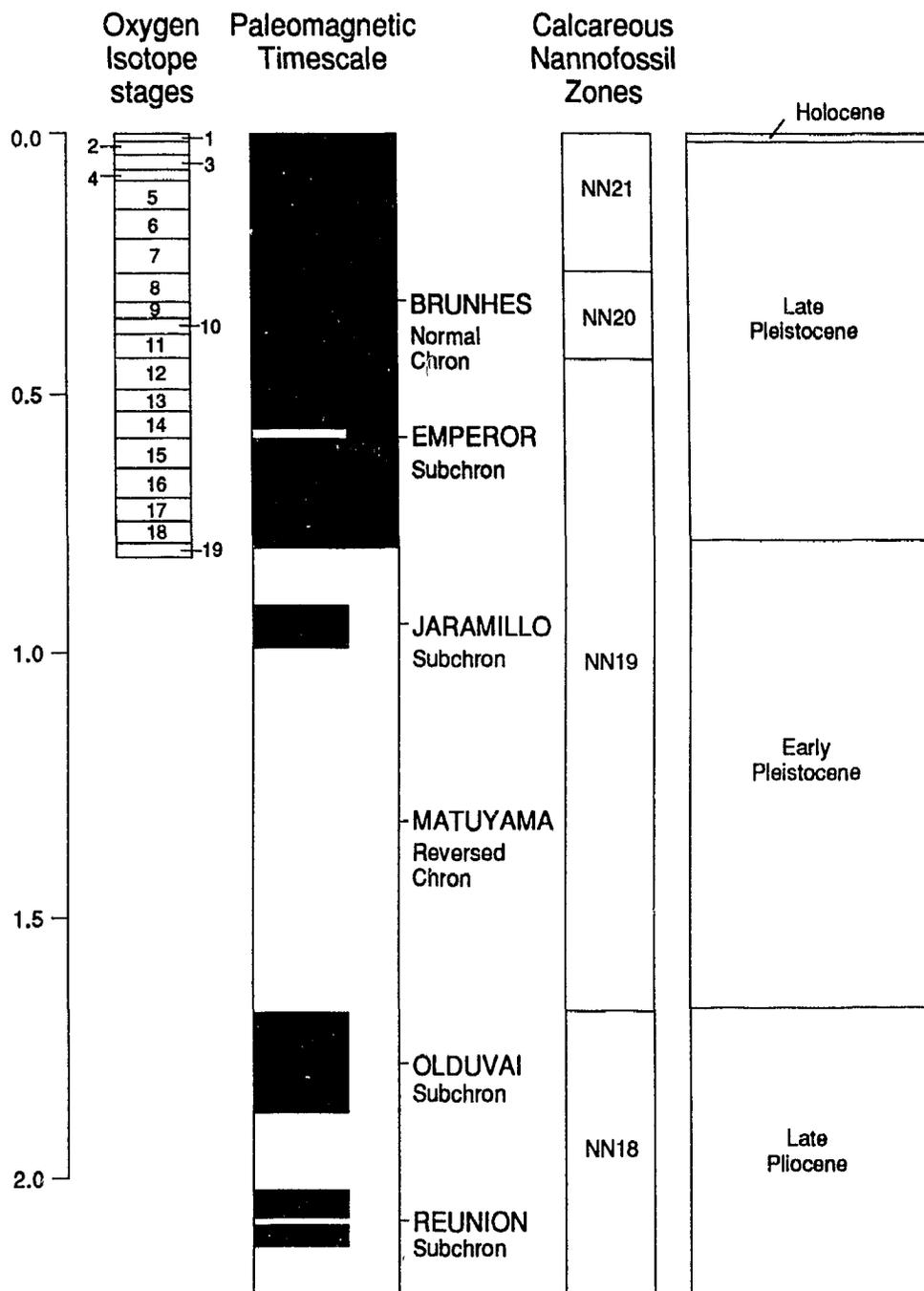


Figure 7. Chronostratigraphic framework used in this study. Calcareous nannofossil zones are from Perch - Nielsen (1985); absolute ages of oxygen isotope stages and magnetochrons follow Richmond & Fullerton (1986).

Quaternary contains only 1 planktonic foraminiferal zone, the *Globorotalia truncatulinoides* zone, and 3 calcareous nannofossil zones: NN19 *Pseudoemiliana lacunosa*, NN20 *Gephyrocapsa oceanica* zone, and NN21 *Emiliana huxleyi*.

Paleomagnetically, the base of the Pleistocene is equated with the top of the Olduvai event; the absolute age of this datum is approximately 1.7 Ma. The boundary between the Matuyama Reversed and Brunhes Normal magnetochrons, dated at 0.79 Ma, is equated by most workers with the boundary between the early and late Pleistocene. The Jaramillo normal polarity subchron occurred near the end of the Matuyama Reversed polarity chron, spanning 0.97-0.9 Ma, and the Empereor reversed polarity subchron has been dated at ca. 0.59 Ma.

Because the resolution available using biostratigraphic and magnetostratigraphic datums is much less than the high level of precision required to examine the Quaternary record at the level of individual glacial-interglacial cycles, these "normal" chronostratigraphic tools can only serve as a means of verifying specific levels within the climatostratigraphy of such high-resolution studies.

Numerous glacial-interglacial cycles have occurred over the last 2.4 Ma at northern latitudes, and the age of a sedimentary unit can be determined by "counting from the top" and comparing with the global climatic signal if a complete stratigraphic record exists. Glacial-interglacial sediments can sometimes be easily identified: for example tills and paleosols or glaciomarine sediments containing ice-rafted debris vs. homogeneous, fine-grained interglacial marine sediments. Glacial-interglacial cycles can be determined by examining alternating microfossil assemblages, and simplified zonations have been developed for some regions (*e.g.* the

Pleistocene planktonic foraminiferal zones of Ericson and Wollin, 1968). Simplified ratios, such as variations in coiling ratios of sinistral:dextral tests of the planktonic foraminifer *N. pachyderma* or gonyaulacoid vs. protoperidinioid dinocysts can provide crude paleoenvironmental interpretations.

Biogeochemical changes, such as stable isotope variations in calcareous tests, also reflect paleoclimates. Since the pioneering work of Emiliani (*e.g.* 1955, 1966), Shackleton (1967), Shackleton and Opdyke, (1973, 1977), and Ruddiman and McIntyre (1981, 1984), Quaternary climatostratigraphy has been based primarily on continuous oxygen isotope records, which reflect global ice volumes and are theoretically correlatable world-wide; the ratio of the "heavy" isotope of oxygen (containing 10 neutrons) to the most common isotope (with 8 neutrons) increases in sea water during glacials because the lighter isotope is preferentially evaporated, falls onto the continents, and is retained in ice sheets, leaving the ocean enriched in the heavy isotope.

Because they primarily reflect global ice volumes, oxygen isotope ratios should be easier to correlate globally than faunal changes which vary with latitude, environment, etc. (see chapter 5). A stable isotope stratigraphy has been established, with absolute ages assigned to stage boundaries (Shackleton and Opdyke, 1973). The boundary between the early and late Pleistocene (Brunhes/Matuyama paleomagnetic boundary) occurs in isotope stage 19, and the Plio-Pleistocene boundary (top of the Olduvai paleomagnetic event) occurs during isotope stage 37. I equate the Illinoian with stages 8-6 (302-132 ka), the Sangamon with stage 5 (132-79 ka), the early Wisconsinan with stage 4 (79-65 ka), the middle Wisconsinan with stage 3 (65-35 ka), the late Wisconsinan with Stage 2 and the lower part of stage 1 (35-10 ka) following Richmond and Fullerton (1986). The Holocene spans 10,000 years B.P. to the present. The pre-Illinoian is

referred to only as early or late Pleistocene, due to the lack of chronostratigraphic resolution in pre-Illinoian sediments in this study.

However, caution must be used when interpreting oxygen isotope signals because other factors besides global ice volume can affect the isotopic signal (*e.g.* salinity, sea surface temperature). These effects are especially prevalent in planktonic foraminifera, which were used in this study. The factors which affect isotopic ratios are especially variable along continental margins, and in close proximity to the polar front; these factors were therefore estimated based on microfossil assemblages at the sites in this study, to correctly interpret the isotopic signal. Carbon isotopes in tests are thought to reflect the metabolic activity of the foraminifer (Berger *et al.*, 1978; Kahn, 1979); this may at times be positively correlated with temperature, but may also be positively correlated with nutrient availability, which peaks during deglaciation, when nutrients are liberated from melting ice.

All of these climatostratigraphic interpretations, however, whether based on the sedimentology, micropaleontology, or stable isotopes, depend upon continuous sedimentation; hiatuses spanning more than a few millenia will result in the absence of one glacial or interglacial stage, invalidating any chronostratigraphic interpretation below the hiatus. It is therefore very difficult to interpret the chronostratigraphy of sites on the continental margin due to the complexity of this depositional system.

## CHAPTER III

### RESULTS

#### 3.1 DSDP LEG 93 SITE 604

##### 3.1.1 Location and core description

DSDP Site 604 was drilled on the upper continental rise off New Jersey in 2361 m water at latitude 38° 42.79 N, longitude 72° 32.95 W. The Pleistocene sediments were lithologically divided into two units and four subunits (Shipboard Scientific Party, 1987a). Acoustically, the Pleistocene sediments are assigned to one alloformation, the Hudson Canyon Alloformation; two distinct reflectors occur within this alloformation, and correspond fairly well with the lithological units (Figure 8). A sample from 8.04 m at this site which was submitted for radiocarbon dating, yielded insufficient carbon for analysis; this is thought to be due to the presence of small amounts of pyrite in the sample, which can sometimes affect carbon dating (IsoTrace Laboratories, pers. comm.). Recovery was generally poor in lithological unit I, but somewhat better in unit II. Lithological unit I spans the upper 84 m, and consists of grey to dark greenish grey alternations of clay and silt. These sediments were assigned a late Pleistocene age based on calcareous nannofossil content (Lang and Wise, 1987), while Moullade (1987a, 1987b) places the boundary between the early and late Pleistocene slightly higher at this site, (ca. 77 m). The work of Lang and Wise (1987) on calcareous nannofossils suggests that rates of sedimentation during the late Pleistocene-Holocene were extremely high, 201 m/Ma. Calcareous nannofossils are relatively common and well-preserved in this unit (Lang and Wise, 1987); planktonic foraminifera are rare through most of the upper part of this unit (subunit IA), (Moullade, 1987a; this study). The upper 35.3 m of this unit consist of interbedded clay and silt layers, referred to as subunit IA (Shipboard Scientific Party 1987a), and are characterised by redeposited middle Eocene

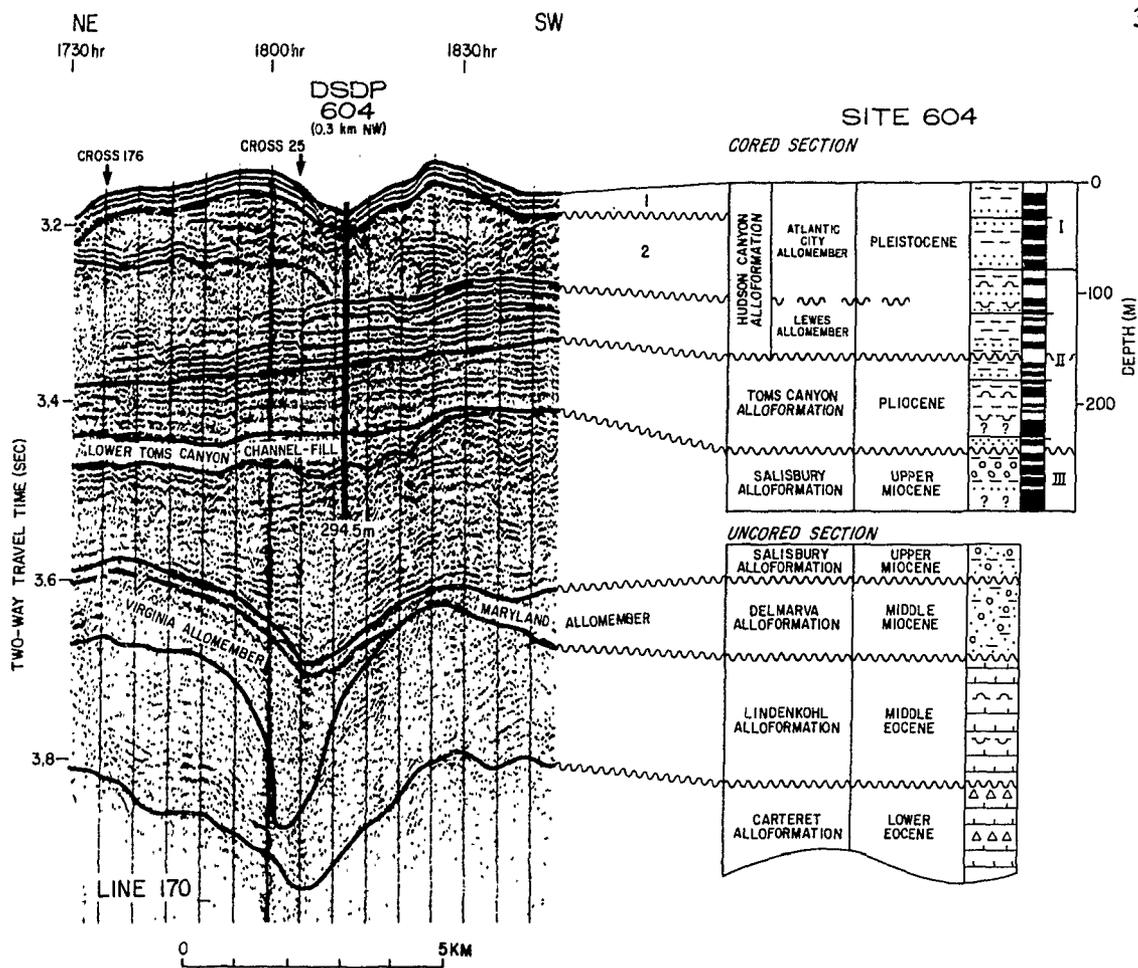


Figure 8. Correlation of seismic stratigraphy along USGS Line 170 and lithostratigraphy at DSDP Site 604 (reprinted by permission from Poag and Mountain, 1987). The boundary between lithologic units IA and IB correlates with a strong seismic reflector, as does the boundary between lithological units I and II.

biosiliceous nannofossil chalk; clasts of chalk containing Eocene nannofossils occur in cores 1 as well as in core 5 (Lang and Wise, 1987), and Eocene foraminifera were identified by Moullade (1987a) in core 2. Below this, to 84 m, the sediments of subunit IB consist of grey and dark greenish grey clay and silt enriched in glauconite, pyrite and heavy minerals, with highly contorted and convolute patterns interpreted as slump structures. Although glauconite is not mentioned as an important component of this unit in the core descriptions in the Initial Report (Shipboard Scientific Party, 1987a), visual analysis of the sand-sized fraction (see below) has shown that glauconite is common to abundant through much of this unit.

Lithological unit II spans the interval between 84 and 238.9 m (Figure 9); the sediments are greenish grey clay with glauconite-rich intervals and variable amounts of biogenic silica. The upper 2 subunits contain sediments of Pleistocene age. Subunit IIA, between 84 and 121.7 m, consists of greenish grey clay with glauconite-rich sand and biogenic silica. Subunit IIB, from 121.7 to 179.3 m, consists of greenish-grey clay with glauconite-rich sand. The Plio-Pleistocene boundary was placed at 162.1 m by Lang and Wise (1987), the base of the *Calcidiscus macintyreii* calcareous nannofossil zone, and between sections 2 and 3 of core 17 (154 m), based on planktonic foraminifera (Moullade, 1987a). The biostratigraphy of this site is discussed further in chapter 5.

The seismic stratigraphy of the upper rise in the vicinity of Site 604 (Figure 8) indicates that the depositional history at this site has been complex. Three strong reflectors at ca. 35, 102 and 162 m record potential erosional surfaces. The reflector at 162 m is thought to represent the Plio-Pleistocene boundary (Poag and Mountain, 1987); this interpretation is supported by the calcareous nannofossil data, which shows a boundary between the *Discoaster brouweri* and

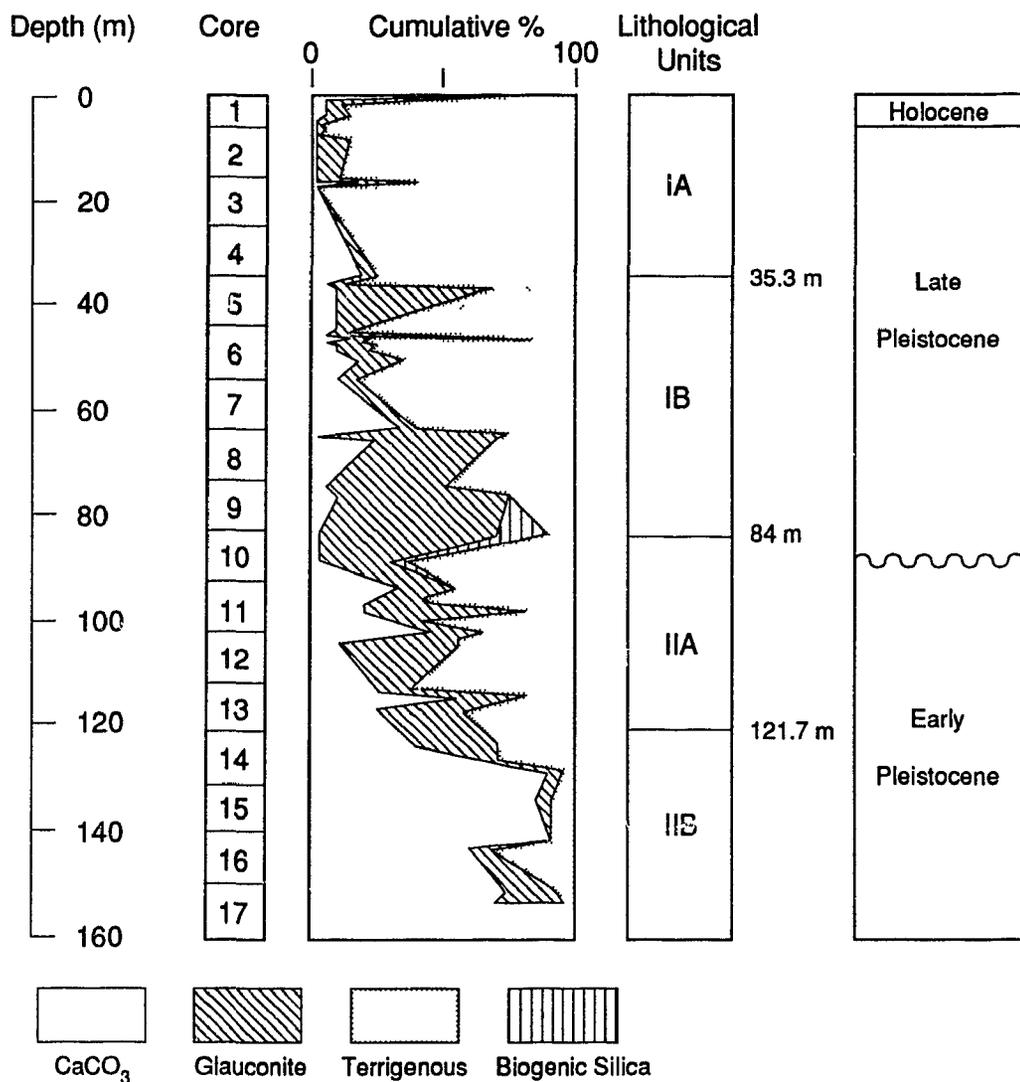


Figure 9. Composition of the sand-sized (>0.063 mm) fraction at Site 604 based on visual estimate.

*Calcidiscus macintyreii* zones at this depth (Lang and Wise, 1987). The reflector at 102 m can be traced all along the diagram in Figure 8, but a sedimentary unit to the left of Site 604 on the diagram appears to pinch out just to the northeast of this site, indicating the existence of an unconformity. Lang and Wise (1987) identified a hiatus spanning at least 1.1 to 0.44 Ma from the calcareous nannofossil data within core 10, which they associate with reflector P1, which they place at ca. 90m, approximately 10 m below that documented by Poag and Mountain (1987). The strong reflector at 35 m identified by Poag and Mountain (1987) separates lithologic units IA and IB; this reflector may also record an unconformity, but there are no biostratigraphic data allowing this to be confirmed.

#### 3.1.1.1 Sediment analysis

A visual estimate of the composition of the sand-sized (> 0.063 mm) fraction was made; Figure 9 shows the contribution of biogenic  $\text{CaCO}_3$  (dominantly planktonic foraminifera), terrigenous sand, reworked biogenic silica (comprised of Eocene radiolaria) and reworked Neogene glauconite to the total sand fraction. The percentage of sand is highly variable through the core, and peaks in glauconitic samples. Similarly, the percentage of glauconite, biogenic carbonate and terrigenous components is highly variable; short term fluctuations occur, but there is an overall trend from dominantly planktonic-foraminifera in the lower ca. 35 m (lithological unit IIB), to glauconite (sporadically abundant in units IIA and IB, and pervasive through most of core 10), and then to terrigenous sand (mainly quartz) in the upper ca. 65 m of this site, especially in the upper 35 m. Reworked Tertiary radiolarians were noted in core 10.

### 3.1.2 Micropaleontology

#### 3.1.2.1 Planktonic Foraminifera

Percentage occurrences of Quaternary planktonic foraminifera are listed in Appendix Table 1, and downcore variations in planktonic foraminiferal abundance and percentages of the most common planktonic foraminiferal taxa are plotted in Figure 10. Percentage curves are arranged from left to right to reflect modern foraminiferal assemblages in North Atlantic water masses following Kipp (1976). Percentages of reworked foraminiferal tests are plotted as well, but will inevitably be an underestimate since most Tertiary taxa are still extant. Glauconite and biogenic silica (sedimentological indicators of Neogene and Paleogene reworking, respectively), correspond quite well with occurrences of reworked planktonic foraminifera. Stratigraphic zonations indicated on the diagrams are discussed in chapter 5, and paleoclimatic/ paleoceanographic interpretations are discussed in chapter 6.

The most common taxa in the Pleistocene record of Site 604 are "cold water" taxa, *i.e.* taxa whose distributions are transitional to polar waters, according to Kipp (1976): *Neogloboquadrina pachyderma* (dextral and sinistral varieties), *Globigerina bulloides*, *Globigerina quinqueloba*, *Globorotalia inflata*, and *Globigerinita glutinata*. The percentage of *Globigerina quinqueloba* rises upcore, and this taxon is common only in lithological unit I. "Warm water taxa", whose distributions are in subtropical, tropical and gyre margin environments, according to Kipp (1976), *e.g.* *Globigerinoides ruber*, *Globorotalia truncatulinoides*, *Globorotalia scitula*, *Globorotalia crassaformis*, *Globorotalia hirsuta*, *Globigerinella aequilateralis*, *Globigerinoides sacculifer*, *Globorotalia tumida*, *Globorotalia menardii*, *Pulleniatina obliquiloculata*, etc., are present in significant percentages in the surface sample, but common only around 45-48 m, 51

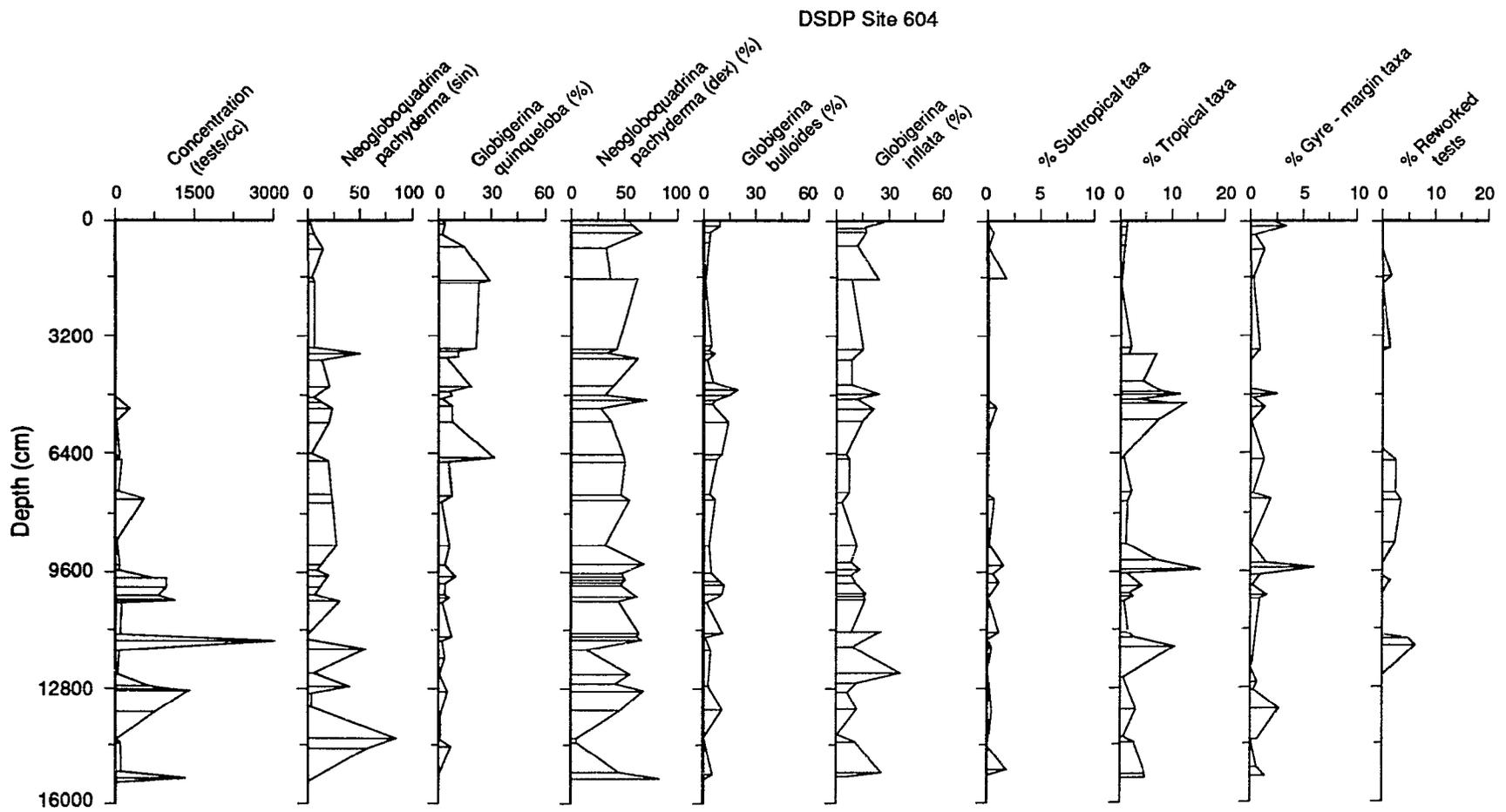


Figure 10. Downcore variation in percentages of common planktonic foraminifera at Site 604.

m, 94-100 m, 117 m, 134 m and 152 m. Subtropical taxa are largely absent from late Pleistocene sediments, although they are more or less consistently present in low numbers through the early Pleistocene record. Sums of subtropical, tropical, and gyre margin taxa in Figure 10 are based on total percentages of the taxa which Kipp (1976) found to be characteristic of these water masses; these taxa, and their water mass affinities, are listed in Table 1. The ratio of sinistral to dextral tests of *Neogloboquadrina pachyderma* (i.e. the "coiling ratio") is plotted in Figure 11. High coiling ratios are found around 8 m, 36.5-37.5 m, 45.4 m, 51 m, 54.5 m, 65.5 m, 75-77 m, 89 m, 97.3 m, and 103-105 m, but very high coiling ratios (>2) are restricted to samples at 117.5 m, and between 142 and 144 m. The very high sinistral:dextral ratios in these samples may reflect the higher resistance to dissolution of sinistral tests of *N. pachyderma* (c.f. Berger, 1968), given the independent evidence of dissolution in these samples. These generally correspond with low percentages of "warm water" taxa; exceptions, such as around 117 m, are generally associated with evidence of reworking, which could account for the "mixed" environmental signal.

Concentrations of planktonic foraminifera (tests/cc) in this study are compared with concentrations of benthic foraminifera found in the same samples by Scott (in prep.) as a P:B ratio in Figure 11. Concentrations of benthic foraminifera refer to specimens larger than 0.063 mm, however, while planktonic foraminiferal concentrations refer to specimens larger than 0.150 mm, so that the calculated ratio consistently underestimates concentrations of planktonic foraminifera. The ratio of planktonic:benthic foraminifera is commonly employed as a crude indicator of distance from paleoshorelines, since stenohaline planktonic foraminifera are intolerant to the lowered salinities found in coastal waters; the plotted P:B ratio is intended only as a crude indicator of the "oceanic" vs. "neritic" character of the surface water mass. High P:B ratios are found only in the surface sediments and around 77 m, between 97 and 99 m, around 104 m, 155

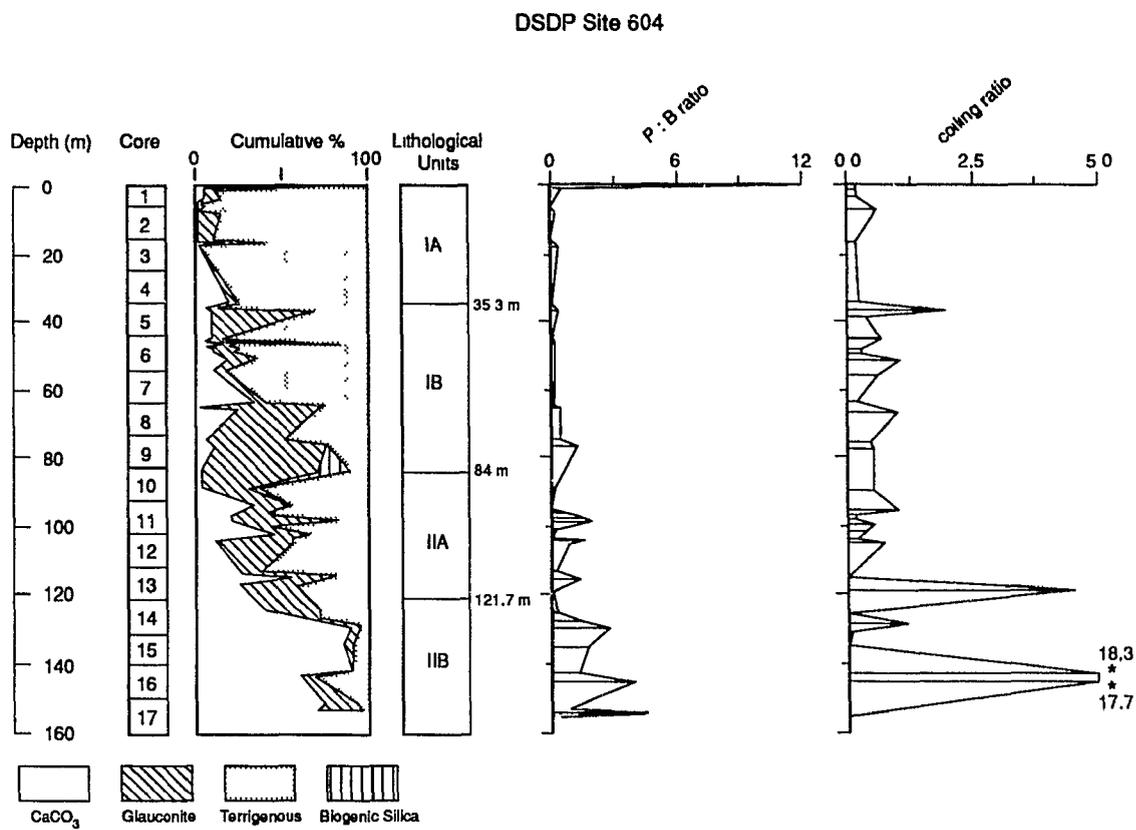


Figure 11. A comparison of coiling ratios (sinistral:dextral tests of *N. pachyderma*) and ratios of planktic:benthic foraminifera (P:B ratios) with Quaternary lithologic units at DSDP Site 604.

m, and below 127 m (*i.e* through lithological unit II).

Concentrations of planktonic foraminifera are highly variable through the Pleistocene section; high concentrations (>1000 tests/cc) occur only in the uppermost sample in late Pleistocene sediments (lithological unit I), but concentrations commonly exceed 1000 tests/cc in early Pleistocene sediments (lithological unit II). Concentrations are especially low in the upper 35 m (lithological unit IA) below the surface sample, often less than 10 tests per cubic centimetre.

One of the important factors restricting the distribution of planktonic foraminifera in deep sea sediments is the solubility of calcite.  $\text{CaCO}_3$  solubility increases with decreasing temperature and salinity; low concentrations of planktonic foraminifera in the sediment can result from low productivity, usually reflecting cold surface water temperatures or reduced salinity, or from high rates of clastic sedimentation diluting the influx of planktonic microfossils. Such samples would be characterised by foraminiferal taxa which tolerate cold water of slightly reduced salinity (e.g. *Neogloboquadrina pachyderma*, *Globigerina quinqueloba*). *N. pachyderma*, for instance, is a deep-dwelling foraminifer, living at or beneath the thermocline (Fairbanks and Wiebe, 1980); high percentages of this taxon could indicate that conditions in the upper layers of the thermocline were unfavourable to the growth of planktonic foraminifera which inhabit shallower depths. Percentages of *G. quinqueloba* presently increase away from the Gulf Stream in the Slope Water (Be and Tolderlund, 1971), so the relatively cold, relatively fresh conditions in the Slope Water appear to be favourable for this taxon. High percentages of these taxa suggest that these conditions are found through most of lithological unit I.

Alternatively, low concentrations of planktonic foraminifera can reflect post-mortem

effects such as dissolution. Low concentrations which are the result of  $\text{CaCO}_3$  dissolution can be identified by 1) low concentrations of calcareous benthic foraminifera, which do not share the same living environment, but which are subject to the same post-depositional conditions, and 2) altered assemblages which are depleted in taxa most susceptible to dissolution and enriched in solution-resistant taxa such as *N. pachyderma*, *G. inflata*, *G. truncatulinoides*, etc. (e.g. Thunell and Honjo, 1981; Berger, 1968). These conditions are present between 3 and 8 m, at 17.6 m, 46.2 m, 65.5 m, 88.9 m, and between 142 and 144 m. Low planktonic foraminiferal concentrations at Site 604 are generally associated with high percentages of terrigenous sand (Fig. 9). This is the case even in samples with relatively high concentrations of benthic foraminifera. The influx of material from the continent onto the rise may promote benthic foraminiferal growth, and simultaneously dilute the planktonic foraminifera.

### 3.1.2.2 Dinoflagellate cysts

Percentage occurrences of dinoflagellate cysts in the Quaternary sediments of Site 604 are listed in Appendix Table 2, and downcore variations in dinocyst concentrations (cysts/cc) and percentages of the most common taxa are illustrated in Figure 12. The ratio of gonyaulacoid: protoperidinoïd cysts (G:P ratio), a crude indicator of temperature and salinity, is plotted in Figure 13. High G:P ratios characterise relatively warm, saline surface waters, while low G:P ratios characterise cold, less saline surface waters. Very low G:P ratios are typically found beneath pack ice. Stratigraphic zonations indicated on the diagrams are discussed in chapter 5, and paleoclimatic/ paleoceanographic interpretations are discussed in chapter 6. Pre-Quaternary taxa indicate reworking, but as with the planktonic foraminifera, reworking will be underestimated because the biostratigraphic ranges of the majority of modern dinocyst taxa extend back through

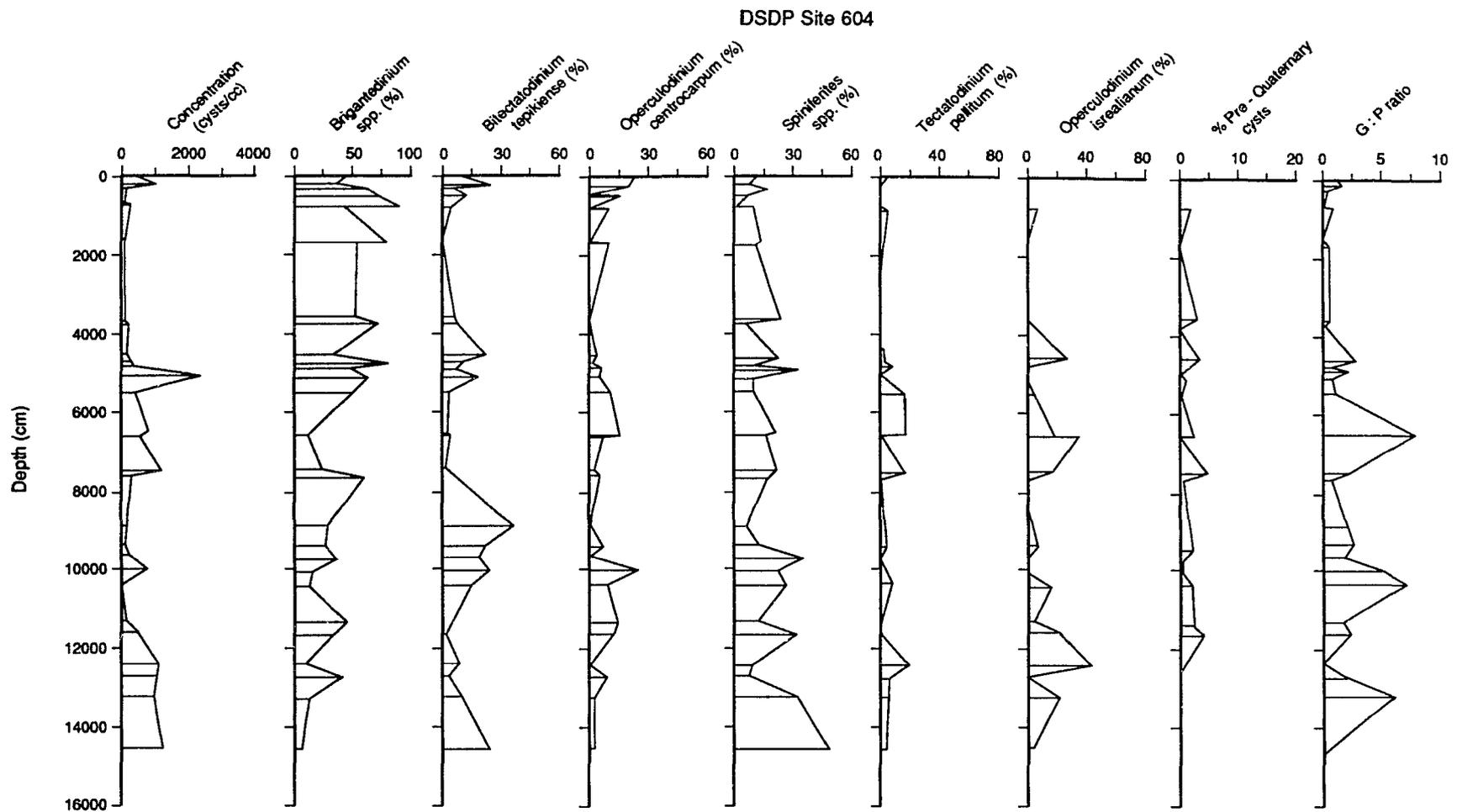


Figure 12. Downcore variation in percentages of common dinoflagellate cyst taxa at Site 604. Cyst concentration is low in the upper 50m below the surface.

the Tertiary.

The surface sediments are dominated by *Operculodinium centrocarpum* and *Brigantedinium* spp., together with lesser, but significant percentages of *Impagidinium* spp., *Spiniferites* spp. (mainly *S. ramosus*), and *Bitectatodinium tepikiense*. These are the most common dinocyst taxa in late Pleistocene sediments at Site 604. The percentage of protoperidinioid dinocysts generally increases up-core relative to temperate gonyaulacoid taxa; the G:P ratio is very high in lithological unit IIB (averaging around 11), falls to 3 in unit IIA, to 2.7 in unit IB, and is only 0.7 in unit IA. There is a correlation between protoperidinioid dinoflagellate cysts and terrigenous sand, and between gonyaulacoid dinoflagellate cysts and calcareous "sand" (dominantly planktonic foraminiferal tests). *Spiniferites* spp., *B. tepikiense*, *O. centrocarpum*, *Operculodinium israelianum* (= *O. crassum*), and *Tectatodinium pellitum* are common in the early Pleistocene.

Glaucinite-rich samples tend to be rich in *Operculodinium israelianum*, *Spiniferites* spp. (with high percentages of *S. pseudofurcatus*) and *Tectatodinium pellitum*. These species are not restricted to sediments rich in glauconite, however; they are also common in early Pleistocene samples with minor glauconite, as in sample 15-1 (140-143), at 132.72 m. Where they occur in late Pleistocene glauconitic sediments, especially in association with "cold-water" taxa, they probably do indicate Neogene reworking. Extinct (*i.e.* pre-Quaternary) taxa, such as *Corrudinium harlandii* occur in low percentages at 8.04 m, 36.57 m, 46.26 m, 65.5 m, 88.95 m, 103.9 m, 132.72 m, and 146.02 m.

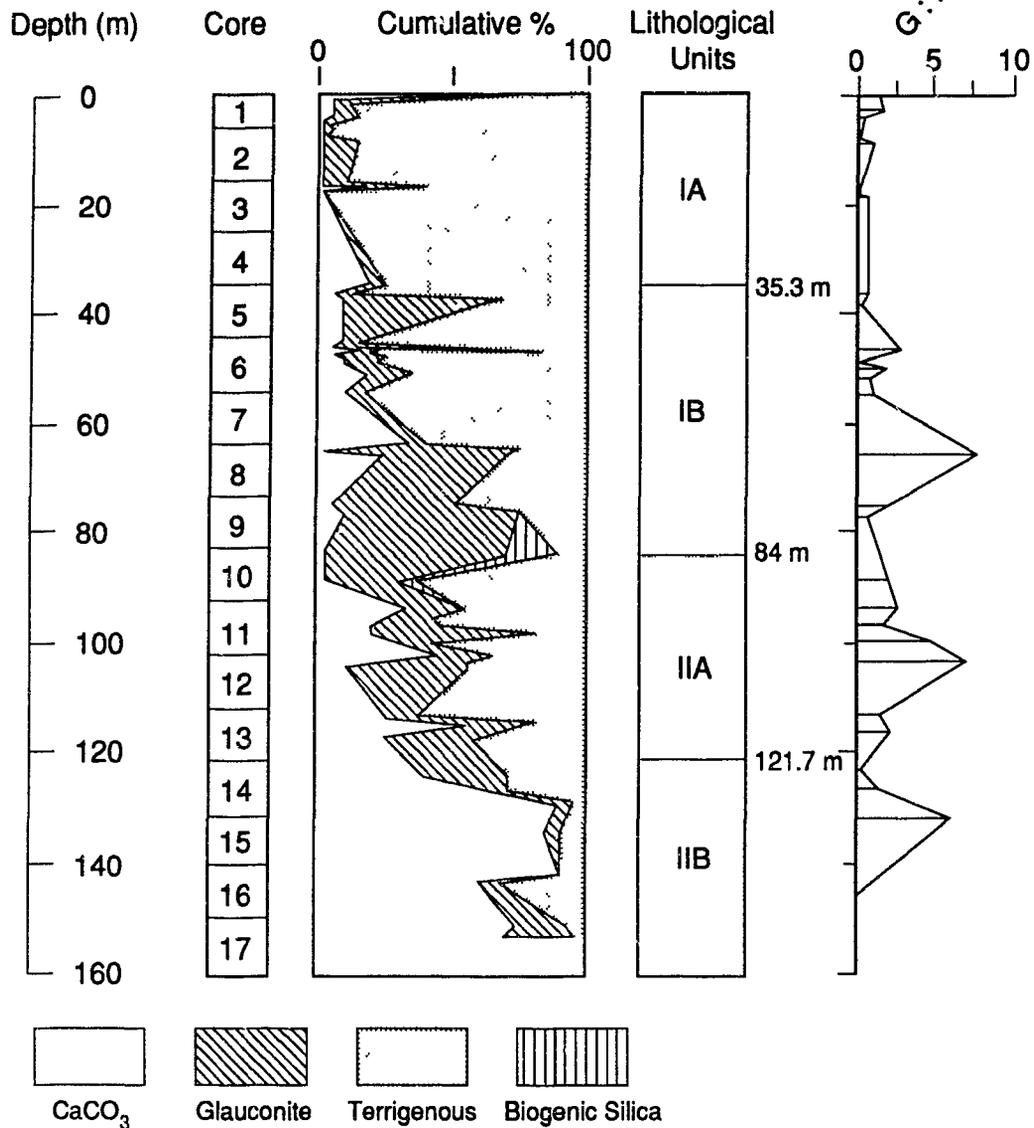


Figure 13. Ratio of gonyaulacoid vs. protoperidinioid dinocysts and lithology at Site 604. There is a correlation between protoperidinioid dinocysts and terrigenous sand. Ratios of gonyaulacoid: protoperidinioid cysts are highest in lithologic unit IIB and lowest in unit IA.

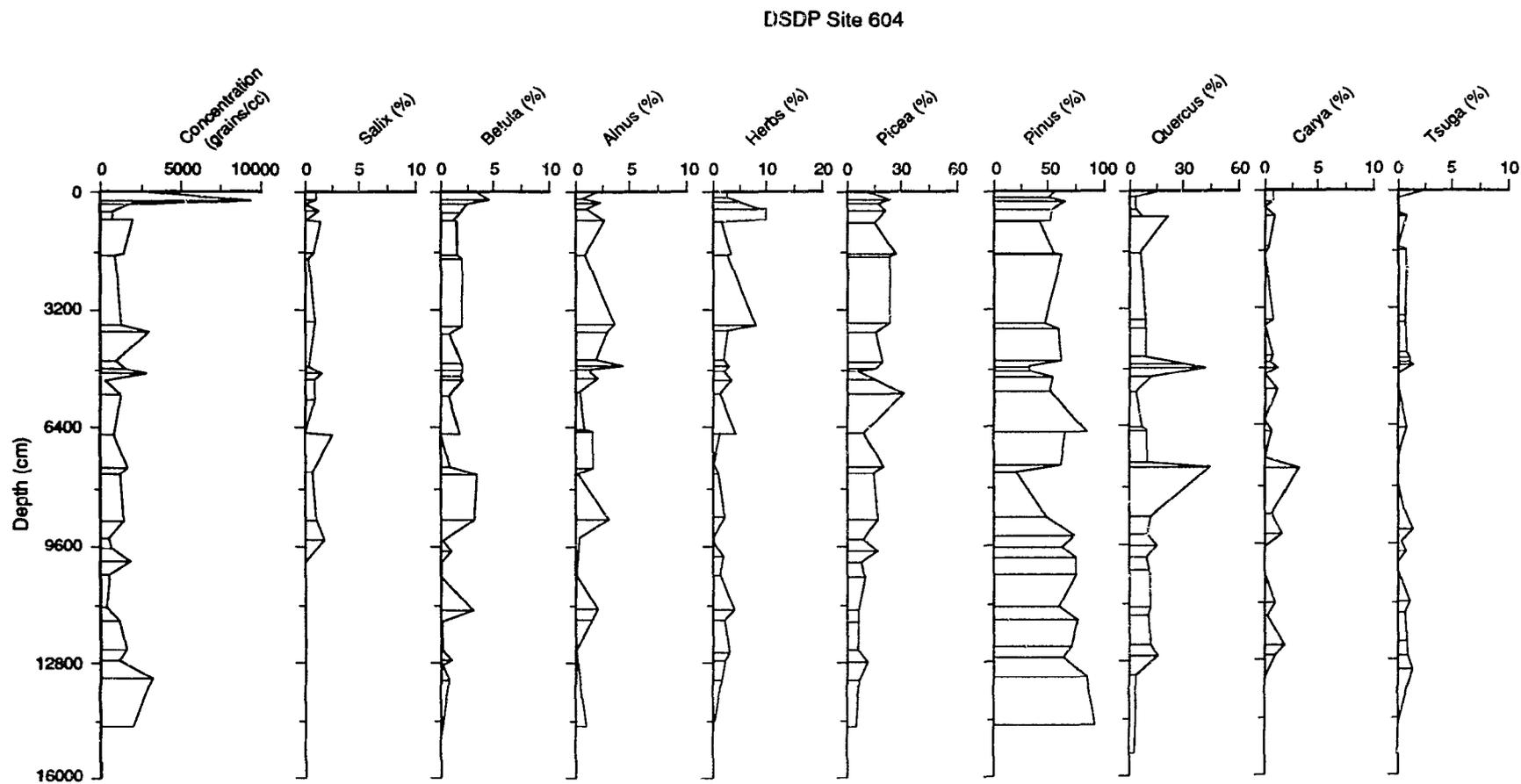


Figure 14. Downcore variation in percentages of common pollen and terrestrial spore taxa at Site 604.

### 3.1.2.3 Pollen and terrestrial spores

Percentage occurrences of pollen and terrestrial spores in Quaternary strata at DSDP Site 604 are listed in Appendix Table 3, and percentages of the most common pollen taxa are plotted in Figure 14. The surface sediments are characterised by high percentages of *Quercus* (19.4%) and *Pinus* (58.1%). The majority of samples in the upper 90 m (cores 1-10) contain high percentages of *Picea* (over 15%), samples between 90 and 145 m (cores 11-16) generally contain less than 10 % *Picea*. *Pinus* strongly dominates the assemblage in cores 11 to 16 (over 60 %), peaking at 90.5 % 146.1 m. Pollen diversity is also much higher in the upper 10 cores than in cores 11-16. *Quercus* and other temperate tree taxa (e.g. *Acer*, *Carya*, *Fagus*, *Ostrya*, *Tilia*, *Tsuga*, *Ulmus*, etc.) are abundant below the surface only sporadically, as at 8.07m, 47.8-49.24m, and 76.6m.

## 3.2 DSDP LEG 94 SITE 607

### 3.2.1 Location and core description

DSDP Site 607 is located on the western flank of the Mid-Atlantic Ridge at latitude 41° 00.07 N, longitude 32° 57.44 W in 3427 m water. Its location at mid-latitudes just south of the glacial-stage polar front, results in very large glacial-interglacial temperature fluctuations, estimated by CLIMAP (1981) for the late Pleistocene to be around 12°-15° C, and by Ruddiman *et al.* (1983) for the early Pleistocene at up to 20° C. This site could therefore be critical to understanding the causes of global climate change and glaciation.

Lithological unit 1 (Pleistocene-upper Pliocene) consists of homogenous planktonic foraminiferal-calcareous nannofossil ooze, the top 28 cm of which are yellow-gray in colour reflecting the circulation of oxygen-rich water through the sediments. The rest of the unit consists of interlayered dark-coloured sediments relatively rich in terrigenous material (probably reflecting increased erosion of the continents during glaciations) and lighter coloured ooze with a relatively small terrigenous component (Shipboard Scientific Party, DSDP Leg 94, 1987b). A  $^{14}\text{C}$  date of  $16,800 \pm 110$  years B.P. (16,390 y. B.P. corrected for a 410 year reservoir effect; IsoTrace Radiocarbon Laboratory Lab No. TO-2448) was obtained from foraminiferal tests at 62-65 cm.

### 3.2.2 Micropaleontology

#### 3.2.2.1 Planktonic Foraminifera

Percentage occurrences of planktonic foraminifera at Site 607 are listed in Appendix Table 4, and the most common foraminiferal taxa are plotted in Figure 15. The surface sediments are characterised by high percentages of taxa restricted to subtropical or warmer waters (36%). Percentages of these taxa decrease below the surface relative to taxa whose ecologic range extends to subpolar-polar waters, such as *N. pachyderma*, *G. quinqueloba* and *G. bulloides*. The coiling ratio of *N. pachyderma* also increases downcore to a maximum of 1.11 at 122 cm; this interval is characterised by low percentages of "warm water" foraminifera, which reach a minimum of 3.8% at 122 cm. The sample at 153 cm shows a lower coiling ratio and percentages of "warm water" taxa increase to ca. 10%, a trend that is continued at 184 cm, which has a coiling ratio of 0.73 and ca. 8.9% "warm water" taxa. At 212 cm, the coiling ratio falls to 0.16, the lowest level since 31-34 cm (early Holocene) and percentages of "warm" taxa rise to 9.9%, highest level since

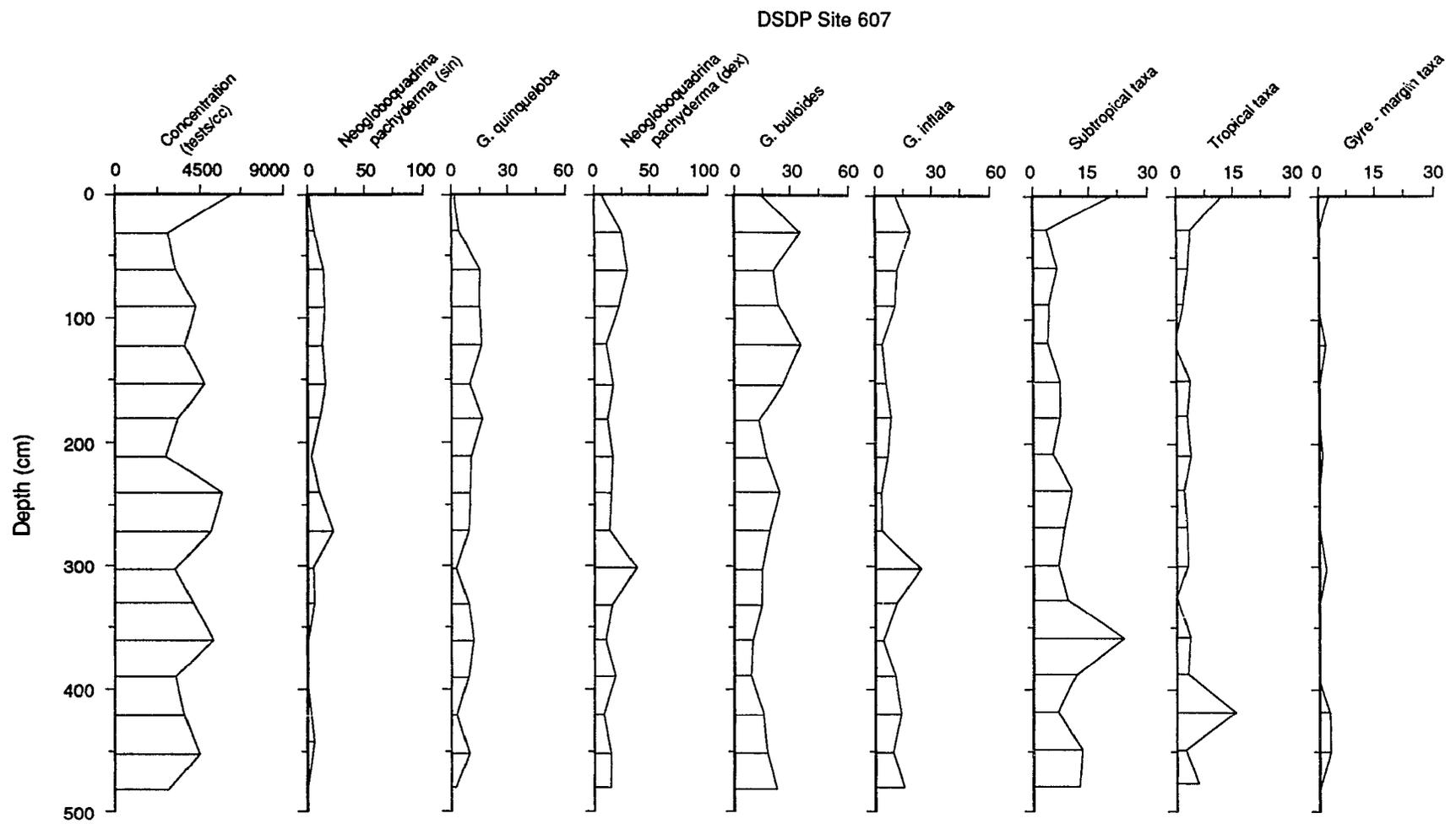


Figure 15. Downcore variation in percentages of common planktonic foraminiferal taxa at Site 607.

the surface. The coiling ratio rises again at 241 cm to 0.61, although percentages of "warm water taxa" are relatively high (12%). The coiling ratio peaks at 272 cm (ca. 1.42), but percentages of "warm water taxa" remain high (ca. 11%). Coiling ratios are low, and percentages of "warm water taxa" are generally high from 300 cm to the lowest sample at 480 cm, although slightly higher coiling ratios are found at 331 and 453 cm.

### 3.2.2.2 Dinoflagellate Cysts

The surface sample at DSDP Site 607 is almost barren, with fewer than 10 dinocysts/cc; only a few corroded cysts of *Impagidinium aculeatum* and *I. striatum* were noted. At 31 cm, dinocyst concentrations are ca. 300 cysts/cc; the assemblage is dominated by *Impagidinium aculeatum* and *Nematosphaeropsis labyrinthus* (Figure 16; Appendix Table 5). Dinocyst concentrations rise sharply to 700/cc (4475/cc based on *Lycopodium* spike) at 62 cm where *Brigantedinium* spp. dominate, making up nearly 90 % of the assemblage. The sample at 91 cm contains high percentages of *Operculodinium centrocarpum*, *N. labyrinthus*, *I. aculeatum* and *Impagidinium patulum*. Sediments at 122 cm are rich in *Brigantedinium*, *O. centrocarpum* and *I. patulum*. *Spiniferites ramosus* is also present in low quantities at 91 and 122 cm. At 153 cm, *O. centrocarpum* dominates, with cf. *Hystrichokolpoma* (?), *Micrhystridium* sp., *I. patulum*, and *N. labyrinthus*. Samples from 181 cm to 212 cm are characterised by low dinocyst concentrations, and are dominated by *I. aculeatum* and *N. labyrinthus*. At 241 cm dinocyst concentrations fall again, and the assemblage is dominated by *Impagidinium* spp. Dinocyst concentrations rise at 272 cm, where *Brigantedinium* spp. and *O. centrocarpum* dominate, comprising 24 and 22 % of the sample, respectively.

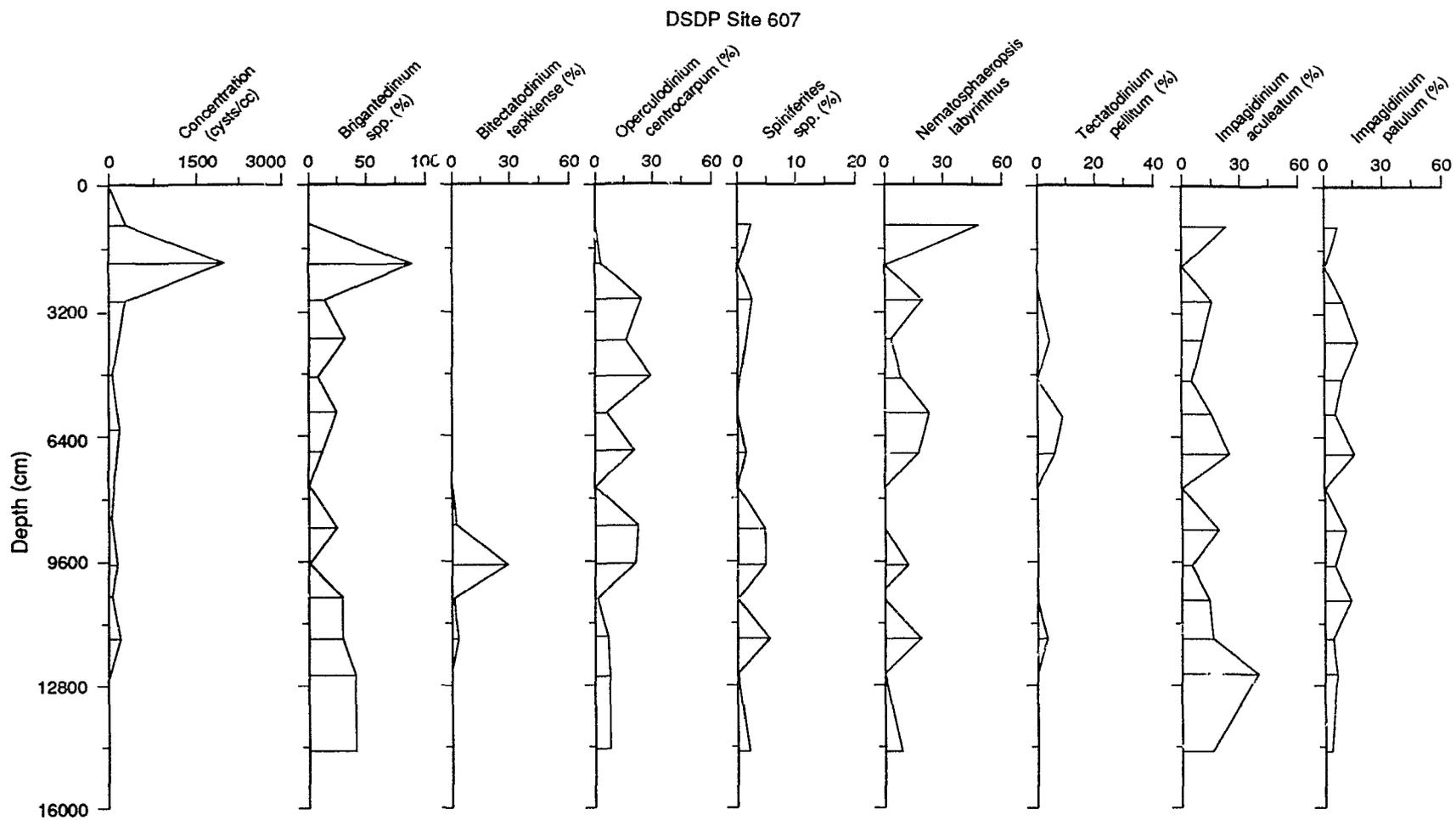


Figure 16. Downcore variation in percentages of common dinoflagellate cyst taxa at Site 607.

At 303 cm *Bitectatodinium tepikiense* and *O. centrocarpum* dominate, together with *N. labyrinthus* and *Impagidinium* spp. From 331 to 453 cm, dinocyst concentrations fall again; *N. labyrinthus* and *I. aculeatum* are important components, together with *Brigantedinium* spp. in the more productive samples.

### 3.2.2.3 Pollen and terrestrial spores

Concentrations of terrestrial palynomorphs are too low at this mid-ocean site to be statistically significant, but concentrations in dark-coloured glacial sediments (*i.e.* between 30 and 303 cm) are about an order of magnitude higher than in oxidised interglacial sediments (Appendix Table 6). The most common pollen taxa are *Pinus*, *Picea*, *Quercus*, *Ulmus*, *Carya*, and *Betula*.

### 3.2.3 Stable isotopes

Isotopic analyses were performed on tests of *Neogloboquadrina pachyderma*; dextral (right-coiling) tests were analysed throughout the core, to give ratios of  $^{18}\text{O}:^{16}\text{O}$ , allowing global climates to be estimated, and ratios of  $^{13}\text{C}:^{12}\text{C}$  were measured to provide insights into paleoproductivity in surface waters of the subtropical gyre above Site 607. Sinistral (left-coiling) tests were analysed where sufficient numbers were found, to give some indication of seasonality.

The lightest  $^{18}\text{O}$  value (0.963‰) occurs in the surface sample (Figure 17, Appendix Table 7). The change in lithology at 28 cm, with higher percentages of terrigenous material imparting a grey colour to the sediments, is associated with a sharp increase in the ratio of  $^{18}\text{O}:^{16}\text{O}$  to 2.305‰, and decrease in the ratio of  $^{13}\text{C}:^{12}\text{C}$  to -0.005‰. Relatively heavy oxygen isotope values in

## DSDP Site 607

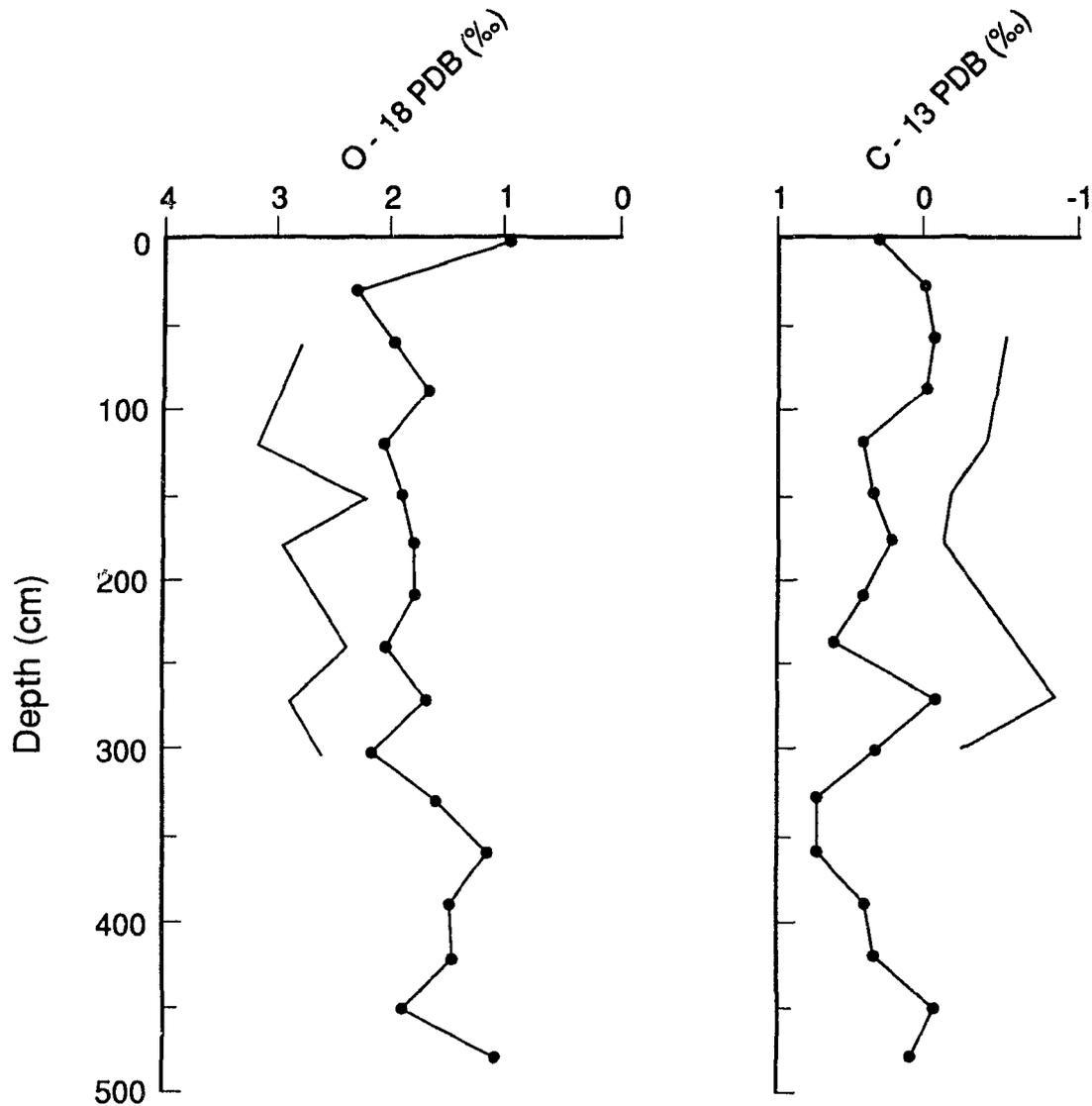


Figure 17. Stable isotope ratios ( $^{18}\text{O}:^{16}\text{O}$  and  $^{13}\text{C}:^{12}\text{C}$ ) measured in dextral and sinistral tests of *Neogloboquadrina pachyderma* at DSDP Site 607. Values measured on dextral tests are indicated by large dots, those measured on sinistral tests are indicated by small dots.

dextral *N. pachyderma* are found between 31 and 303 cm, with heaviest values ( $> 2\text{‰}$ ) at 31, 122, 241, and 303 cm.

Heavier  $^{18}\text{O}$  values were measured in sinistral tests; values range from 2.404 to 3.199 ‰. Heavy oxygen isotope values ( $>2.5\text{‰}$ ) were measured at 62, 122, 181, 272, and 303 cm. There is a large difference between  $^{18}\text{O}$  values measured in dextral and sinistral tests at 62, 122, 181, and 272 cm, but only a small difference at 153, 241, and 303 cm.

Positive values of  $^{13}\text{C}:^{12}\text{C}$  were found in dextral *N. pachyderma* through most of the core; negative values were measured from 31 to 91 cm and at 272 cm. Negative values were measured in sinistral tests throughout the site, with very low values ( $<-0.5\text{‰}$ ) measured at 62, 241 and 272 cm.

### 3.2.4 Sea Surface Temperatures

Sea surface temperatures, SST (summer) and SST (winter), estimated from dinocyst assemblages are given in Appendix Table 8. Summer SST values substantially below the modern value of 19-20°C occur only at 122 cm, 181 cm, 272 cm, and 453 cm. Values of winter SST below the modern value of 14°C occur throughout the core from 30 to 450 cm, and values below 10°C occur at 62, 122, 181, 272, 362, and 453 cm.

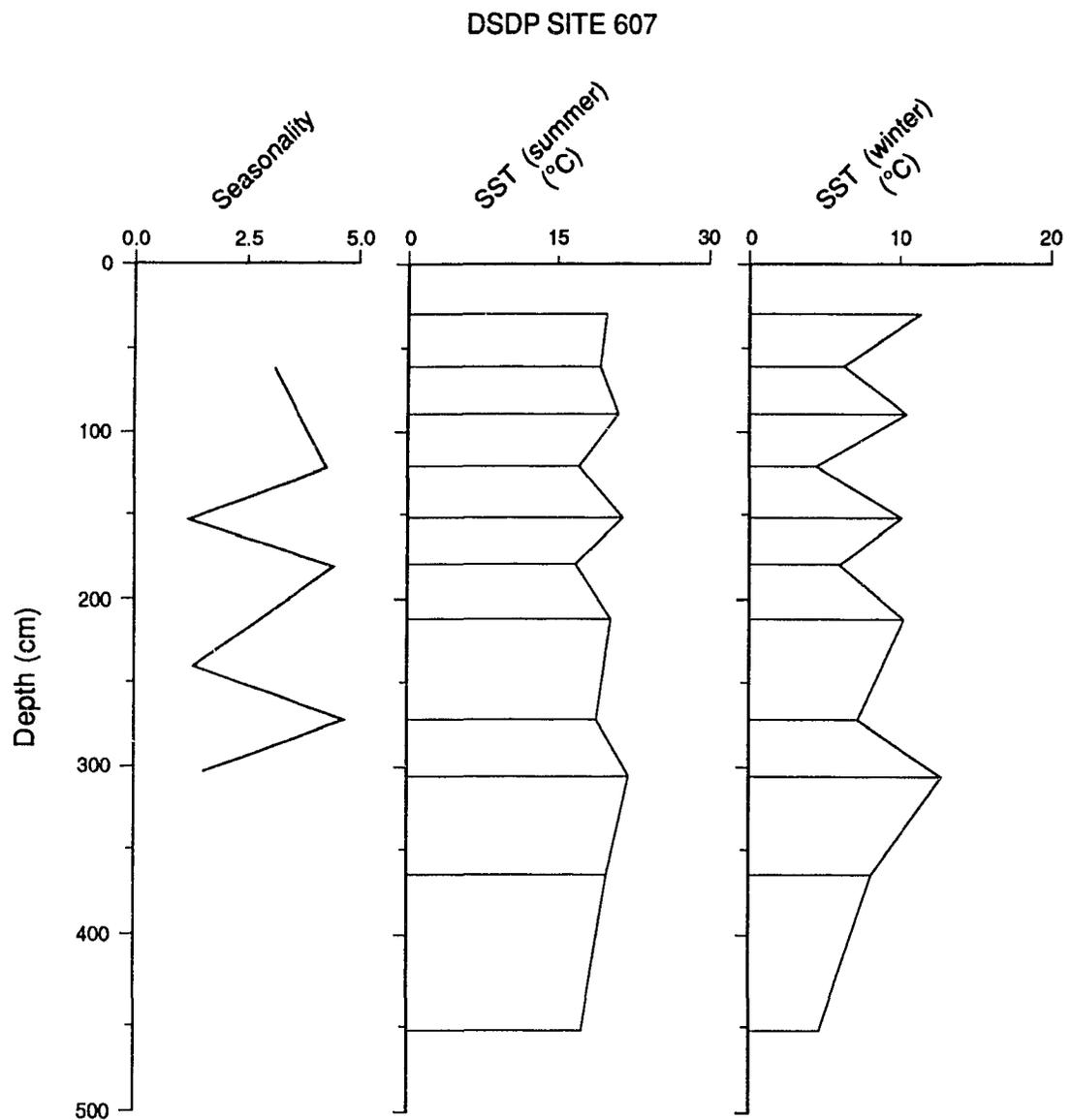


Figure 18. Sea surface temperatures (SST) calculated for summer and winter from dinocyst assemblages at Site 607. The curve on the left indicates differences in seasonality estimated from the difference in  $^{18}\text{O}$  values measured in dextral and sinistral tests of *N. pachyderma*.

### 3.3 DSDP LEG 95 SITE 612

#### 3.3.1 Location and core description

Site 612 is at the shoreward end of the New Jersey transect on the mid-slope in 1386 m of water at latitude 38° 49.21 N, longitude 72° 46.43 W. The upper 32.5 m consist of homogeneous dark grey mud with irregularly-distributed burrow mottles; these are most abundant in core 1 and in the upper half of core 3. There is a sharp colour change to reddish grey mud near the base of core 4, below which silty laminae are common, implying distorted bedding possibly resulting from a debris flow (Shipboard Scientific Party, 1987c). Within section 5-3, at 37 m, is a distinct eroded surface thought by most workers to correspond to the Plio-Pleistocene boundary (e.g. Poag and Low, 1987). There do not appear to be any other unconformities in the upper 37 m, but below 37 m, sedimentation is much more variable. The entire section from section 5-2 (110 cm), i.e 36.2 m, was assigned to the latest Pleistocene zone CN15 by the presence of the calcareous nannofossil *Emiliana huxleyi* (Shipboard Scientific Party, 1987c). A radiocarbon date of 11,730 +/- 100 years B.P. (11,320 +/- 100 y. B.P. corrected for a 410 year reservoir effect; IsoTrace Radiocarbon Laboratory Lab No. TO-2446) was obtained from the top of core 3-1, at about 15 m, suggesting extremely high sedimentation rates on the mid slope over the last 11.5-10 ka, especially between 10.9 and 10.5 ka.

### 3.3.2 Micropaleontology

#### 3.3.2.1 Planktonic foraminifera

The surface sediments are dominated by dextral *N. pachyderma*, with high percentages of *G. inflata* (Figure 19, Appendix Table 9); these sediments contain relatively high percentages of "warm water taxa" and very low percentages of polar indicators (3.7% sinistral *N. pachyderma* and 0.5% *G. quinqueloba*). Percentages of these "cold water" taxa and the coiling ratio rise sharply below the surface; at 230 and 330 cm, over 40 % sinistral *N. pachyderma* and over 5% *G. quinqueloba* are found, and the coiling ratio is greater than 1.

Very low concentrations of planktonic foraminifera are found between 4 and 14 m, (<< 10 tests/cc). The foraminiferal record is dominated by dextral *N. pachyderma* and by *G. quinqueloba*. Percentages of sinistral *N. pachyderma* are also fairly high, but lower than in the lower part of core 1, and coiling ratios are around 0.6. Around 15 m, concentrations rise to over 10 tests/cc; the percentage of sinistral *N. pachyderma* is still quite high, as are percentages of dextral *N. pachyderma*, *G. bulloides*, *G. quinqueloba*, and *G. glutinata*. Percentages of sinistral *N. pachyderma* fall sharply around 16 m. The coiling ratio falls between 16 and 26 m, percentages of sinistral *N. pachyderma* are fairly low, while percentages of the transitional species *G. inflata* and "warm water taxa" are significant. Concentrations average around 35 tests/cc. From 26 to 32 m, two very high peaks in sinistral *N. pachyderma* are separated by an interval characterised by very high percentages of *G. inflata*. Below 32 m, percentages of sinistral *N. pachyderma* fall, and percentages of "warm water taxa" rise slightly; very low percentages of

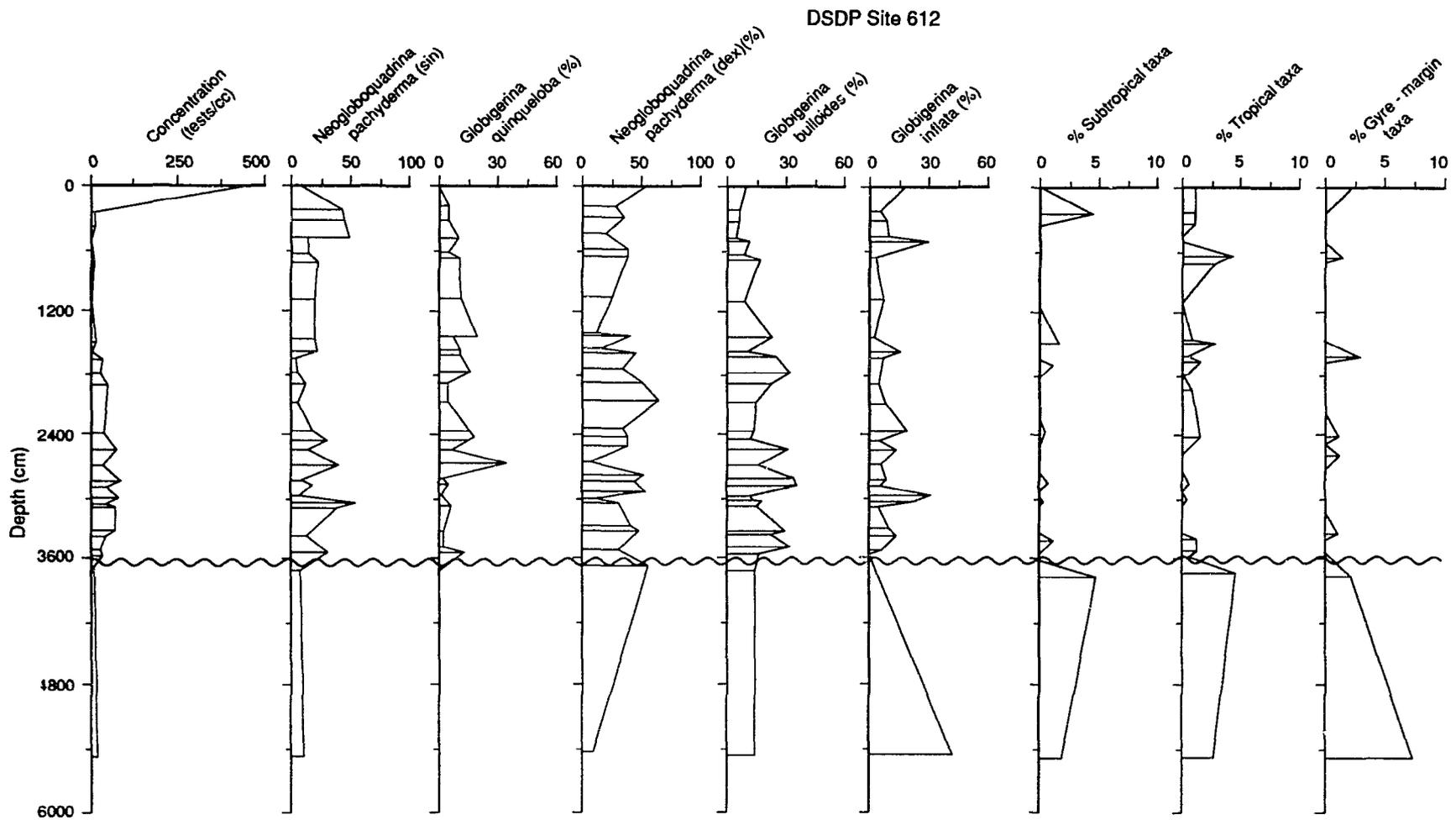


Figure 19. Downcore variation in percentages of common planktonic foraminiferal taxa at Site 612. At a depth of 15 m, a  $^{14}\text{C}$  age of 11,730 y. B.P. (corrected age 11,320 y. B.P.) was measured from picked foraminiferal tests using an accelerated mass spectrometer.

sinistral *N. pachyderma* (8.7 %) are found in sample 5-1 (47-53), but the sample is still dominated by "cold water taxa", dextral *N. pachyderma*, *G. bulloides*, and *G. inflata*, and "warm water taxa" comprise less than 3 % of the sample. The coiling ratio rises sharply below 35 m.

Below the unconformity at 37 m (the Plio-Pleistocene boundary as identified by the Shipboard Scientific Party, Leg 95, 1987c), foraminiferal assemblages are quite different from those above the unconformity; percentages of "warm water foraminifera", notably *G. crassaformis*, are considerably higher than in the upper 37 m. There are no diagnostic pre-Quaternary foraminifers in those samples below 37 m which do not contain abundant glauconite, however, making age determinations difficult.

### 3.3.2.2 Dinoflagellate cysts

The surface sample is dominated by *Operculodinium centrocarpum* (37.2%) and *Brigantedinium* spp. (24.4%), with significant percentages of *Spiniferites* spp. (7%) and *Impagidinium* spp. (10.9%) (Figure 20, Appendix Table 10). Between 2.5 and 14 m, the G:P ratio is very low, and *Brigantedinium* spp. heavily dominate the dinocyst assemblage, but high percentages of *Bitectatodinium tepikiense* occur around 7 m. Between 15 and 20 m percentages of *O. centrocarpum*, *Spiniferites* spp., and *B. tepikiense* are fairly high, and *Impagidinium* spp. are also present in low percentages. Percentages of *Brigantedinium* spp. are relatively high between 20 and 29 m and then fall relative to *O. centrocarpum* and *Spiniferites* spp. Percentages of *Brigantedinium* spp. rise sharply around 32 m, and then fall around 33 m; percentages of *Spiniferites* spp. rise to highest values at this site around 33 m, where *O. centrocarpum* is also abundant, and these taxa also characterise the next sample (35 m). One sample was examined

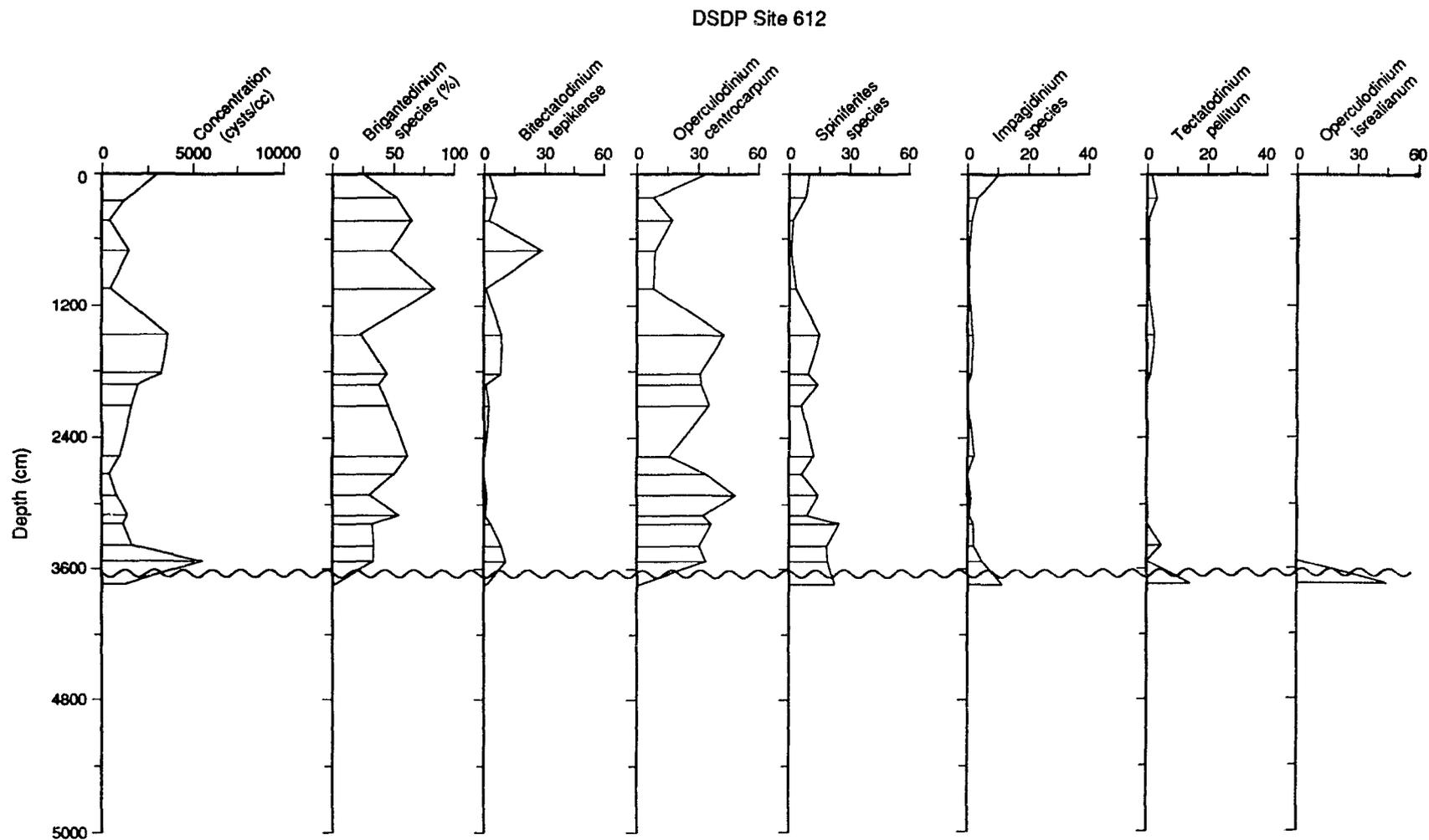


Figure 20. Downcore variation in percentages of common dinoflagellate cyst taxa from Site 612.

below the unconformity at 37m; this sample was relatively rich in glauconite, and was dominated by the warm-water dinocysts *Operculodinium israelianum* and *Tectatodinium pellitum*.

### 3.3.2.3 Pollen and terrestrial spores

Pollen concentrations are relatively high throughout the upper 37 m of Site 612, and contain diverse assemblages (Figure 21, Appendix Table 11). *Pinus* is the dominant taxon throughout the core, but *Picea* and *Quercus* are also abundant in some samples. Percentages of *Picea* are high throughout the core below the surface, but percentages of this taxon are relatively lower in cores 3 and in the upper part of core 5. *Quercus* is abundant in the surface sample as well as in sample 3-1 (130-133) and 4-3 (80-86).

### 3.3.3 Stable Isotopes

Isotopic analysis was performed on right-coiling tests of *Neogloboquadrina pachyderma*. Ratios of  $^{18}\text{O}:^{16}\text{O}$  at the surface are around 1.3 ‰ (Figure 22, Appendix Table 12). Such light isotopic values recur in the core only once, at the top of core 2 where a very light value of 0.87 ‰ was measured; this is thought to represent contamination by surface sediments (Scott, 1987). Fairly heavy isotopic values persist elsewhere throughout the Pleistocene sediments of Site 612, at the sampling resolution employed in this study, although relatively light values occur between 15 and 21 m, around 28-30 m and below 33 m. Foraminiferal concentrations were insufficient between 4 and 14 m to permit isotopic analysis, but samples just above and below this "barren" interval show heavy isotopic values (nearly 3‰). Heaviest values are found around 2.5 m, 25 m and around 31 m. Light isotopic values are also found below the unconformity at 37 m (ca. 1.8 ‰).

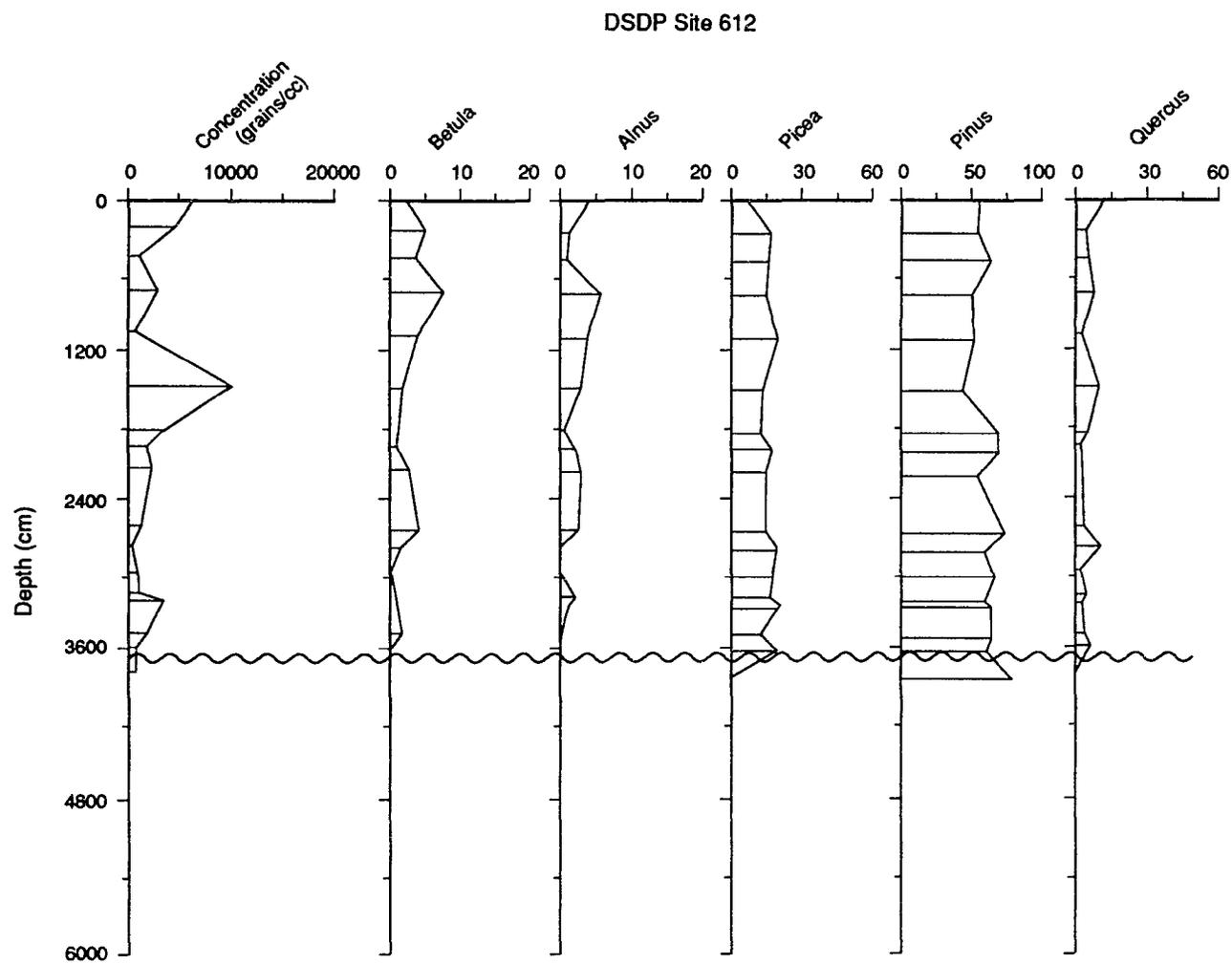


Figure 21. Downcore variation in percentages of common pollen and terrestrial spores at Site 612.

## DSDP Site 612

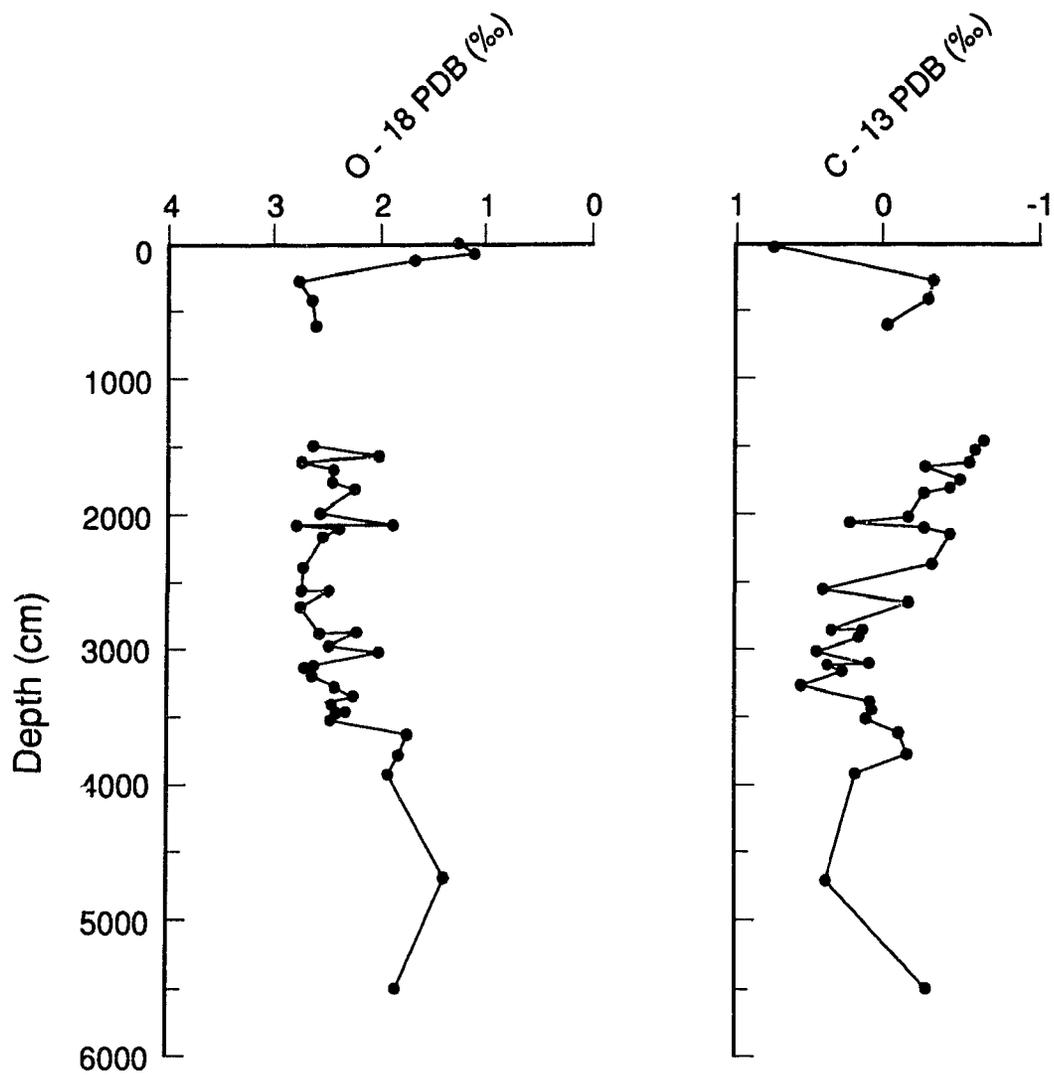


Figure 22. Stable isotope ratios ( $^{18}\text{O}:^{16}\text{O}$  and  $^{13}\text{C}:^{12}\text{C}$ ) measured in dextral tests of *Neogloboquadrina pachyderma* at Site 612. The interval 4-14m contained too few tests for analysis.

Zonations shown in Figure 22 are tentative, since the record differs from the classic signals obtained from mid-ocean sites. The frequent, fairly high-amplitude variations may be due to fluctuations in salinity. Climatostratigraphic correlation therefore cannot be based only on the isotopic signal; microfossil data must also be examined in order to correctly interpret the isotopic signal.

## CHAPTER IV

### COMPARISON OF DINOCYST, POLLEN, FORAMINIFERAL AND STABLE ISOTOPE DATA FROM THE THREE SITES STUDIED

#### 4.0 INTRODUCTION

Global climate changes are reflected differently by different groups of organisms. Climatic changes at mid-latitudes not resulting in polar glaciation will have little effect on benthic organisms, which respond to the climate at the poles where the dense bottom waters originate. Climatic change at high latitudes will rapidly affect the terrestrial biota in polar-subpolar regions, and lower surface water temperatures will cause rapid changes in plankton assemblages; scientists, however, are divided on the effect that high latitude glaciation will have on terrestrial and planktonic organisms at low latitudes. Some, such as Braun (1955), suggest that climatic changes associated with the Laurentide ice sheet were insufficient to displace the terrestrial biota except within about 100 km of the ice sheet. Others, such as Deevey (1949), feel that the climatic change was great, displacing boreal vegetation into the southern United States, virtually eliminating most deciduous forest elements from these areas.

The comparison of planktonic foraminiferal, dinocyst, and pollen and terrestrial spore, stable isotope and the benthic foraminiferal data of Scott (1987 and in prep.) permits more precise paleoclimatic and paleoceanographic interpretations, and allows comparison of the response to the

global climate signal in the surface and deep western North Atlantic Ocean and the North American continent. Paleoenvironmental and sedimentological data from each site are summarised in Figures 23-25. The climatostratigraphic zonation illustrated on the diagrams is discussed in chapter 5.

#### **4.1 ECOLOGICAL COMPARISON OF TWO GROUPS OF FOSSIL PLANKTON: FORAMINIFERAL TESTS AND DINOFLAGELLATE CYSTS**

Both dinoflagellates and planktonic foraminifera inhabit the surface waters, and hence reflect conditions at the ocean's surface. Despite their more ubiquitous distribution and their higher likelihood of preservation, dinoflagellate cysts are less commonly used to reconstruct Quaternary paleoceanography because the modern distribution of the cyst is not well known. Few high-resolution dinocyst studies have been done in Quaternary sediments from mid-latitudes in the western North Atlantic; palynologists have concentrated on "high latitudes", around 55°N, critical to ice sheet growth and global climate (Milankovitch, 1941). The modern distribution of planktonic foraminifera in the mid-latitude western North Atlantic, in contrast, is relatively well understood, and planktonic foraminifera have long been used to interpret the paleoecology of surface waters, and hence to reconstruct paleoceanography and paleoclimates (*e.g.* McIntyre *et al.*, 1976). Both groups of planktonic microfossils were examined from the sites on the slope, rise, and mid-ocean ridge for the following reasons:

- 1) Planktonic foraminifera and dinoflagellates occupy different microhabitats and niches—many dinoflagellates are autotrophs (primary producers), while all planktonic

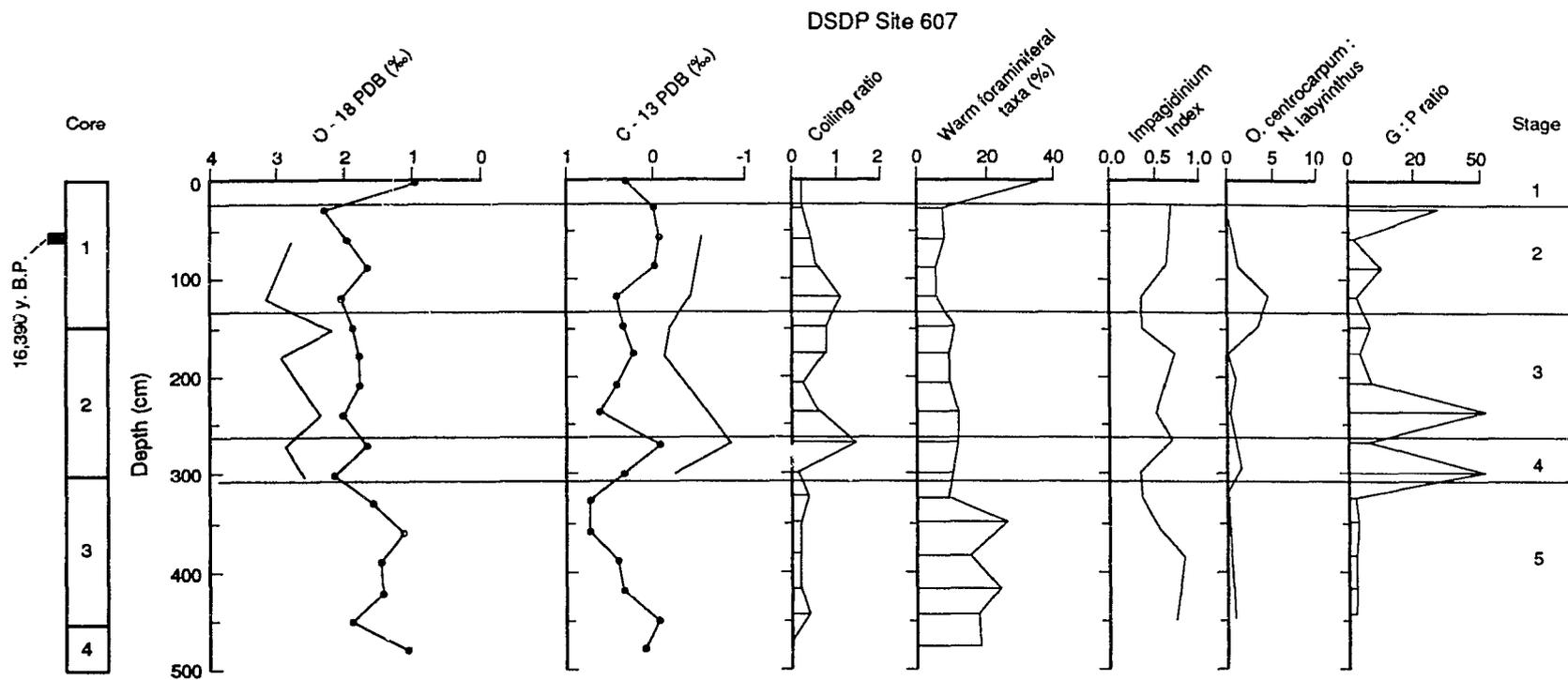


Figure 23. Summary of paleoenvironmental data from Site 607 and climatostratigraphic/paleoceanographic interpretation.

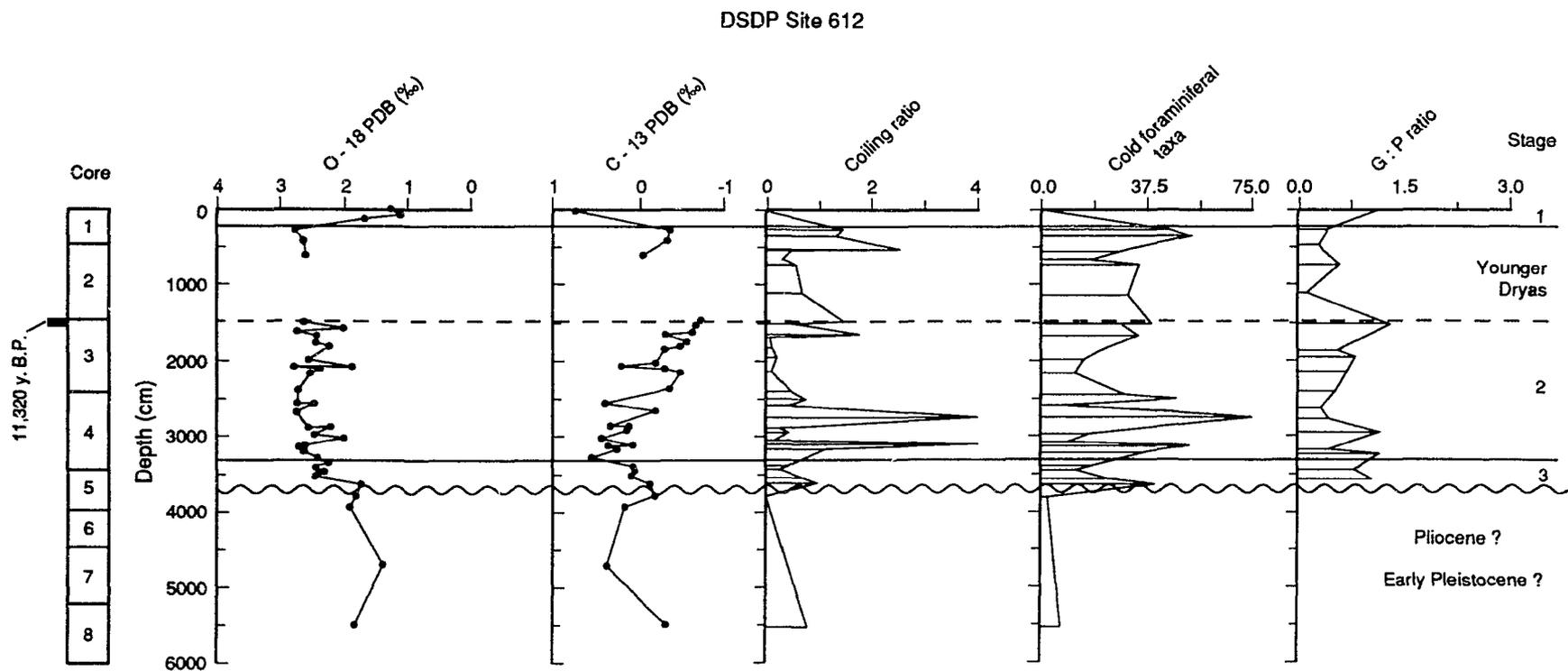


Figure 24. Summary of paleoenvironmental data from DSDP Site 612 and climatostratigraphic/paleoceanographic interpretation.

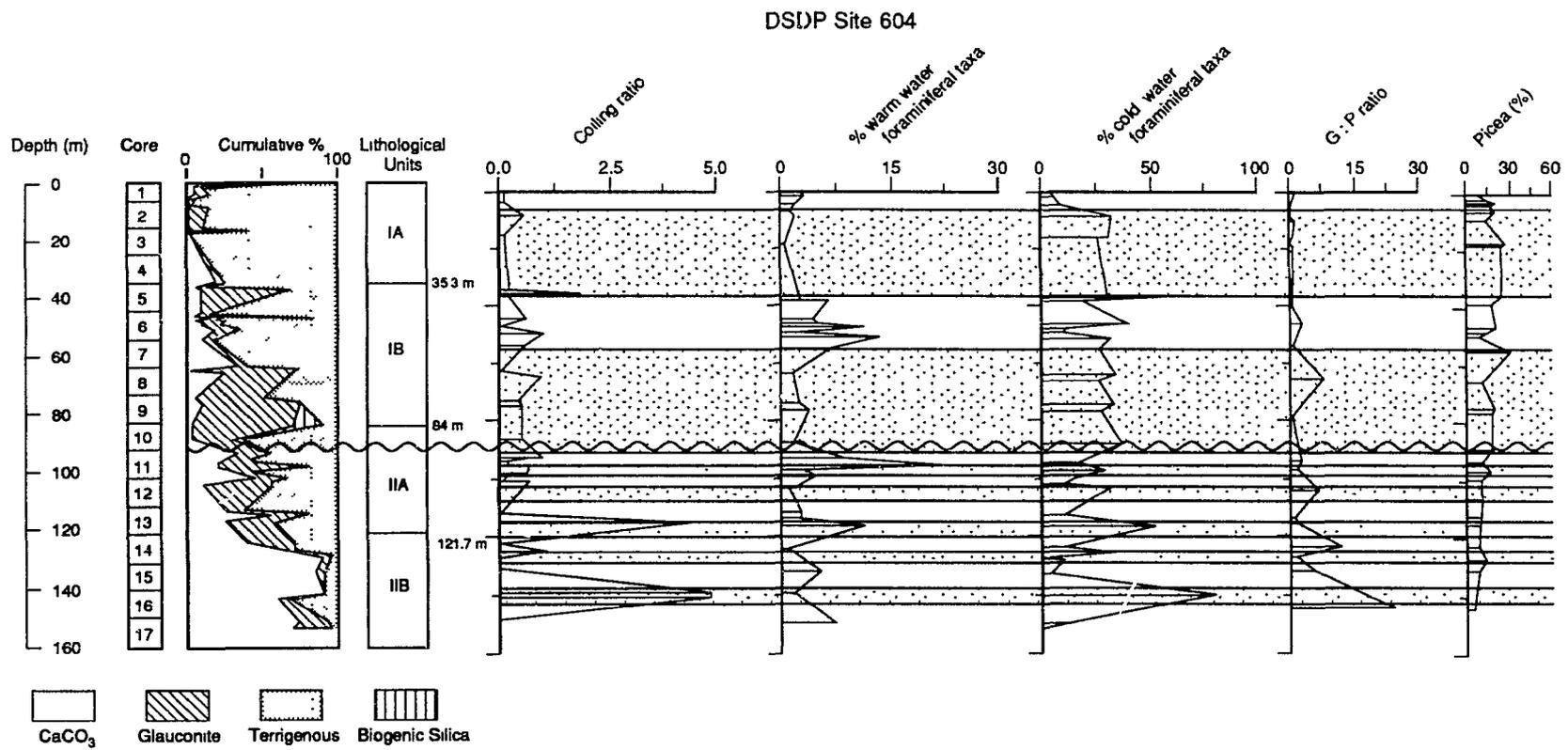


Figure 25. Summary of paleoenvironmental data from DSDP Site 604 and climatostratigraphic/paleoceanographic interpretation. Glacial intervals are stippled.

foraminifera are heterotrophs; the autotrophic dinoflagellates are therefore restricted to the photic zone, generally in the surface 100 m. 2) Planktonic foraminifera are stenohaline and hence tend to be scarce in coastal waters, which usually have less than normal marine salinity (34.92 ‰ is the average salinity of the Atlantic Ocean, according to Fairbridge, 1966); in fact, low ratios of planktonic to benthic foraminifera are frequently interpreted in the rock record to imply proximity to paleocoastlines, while increasing distance from shore is thought to be reflected by increasing planktic/benthic ratios. Planktonic foraminifera are generally abundant in deep sea sediments above the carbonate compensation depth (CCD), where clastic input is low and sedimentation tends to be continuous; their tests frequently dominate the sediments, forming biogenic oozes. The CCD varies from basin to basin depending on factors such as bottom water temperature and chemistry, but generally exceeds 5000 m in the Atlantic Ocean (Fairbridge, 1966). Dinoflagellates, in contrast, are found in all aquatic environments, from normal marine to freshwater. Several dinocyst taxa are abundant in neritic environments; this is important when correlating deep ocean or lower continental margin records with the shelf or upper slope, since the planktonic foraminifera are scarce in shelf and upper slope sediments. 3) Dinoflagellates seem to be more sensitive to small variations in surface water than planktonic foraminifera, since a given foraminiferal zone can contain several dinocyst assemblages, and dinocysts tend to lead planktonic foraminifera in recording environmental/climatic change. 4) Organic-walled dinoflagellate cysts are acid-insoluble, and therefore have a higher preservation potential than the calcareous planktonic foraminifera, *i.e.* they will be present below the CCD and in nearshore areas.

#### 4.1.1 A comparison of dinocyst and planktonic foraminiferal distributions at the three sites studied

The sedimentary record of the upper rise (Site 604) is intermediate between that of the slope (Site 612) and mid-ocean ridge (Site 607). Both the allochthonous and autochthonous components of the sediments on the rise are similar to those on the slope, reflecting the proximity of these two areas, and the existence of the same surface water mass (Slope Water) above both depositional environments; however the completeness of the stratigraphic record on the rise is more similar to that on the mid-ocean ridge, reflecting the lower energy conditions and relative scarceness of erosional agents in these environments as opposed to the slope. The existence of two seismic reflectors within the Quaternary sequence of Site 604 (Figure 8), and the corresponding hiatus between 1.1 and 0.44 Ma at ca 90m corresponding with a strong seismic reflector identified in the calcareous nannofossil data (Lang and Wise, 1987) show that the upper rise has, however, been a site of variable sedimentation, and net erosion at times during the Quaternary.

Because the modern distribution and ecological tolerances of planktonic foraminifera are well understood (*e.g.* Kipp, 1976; Be and Tolderlund, 1971), glacial and interglacials are easily distinguished in the planktonic foraminiferal record at each site. Glacials are recorded by 1) high percentages of "cold water" foraminiferal taxa, *i.e.* those whose modern distributions are north (poleward) of the sites (polar assemblage: *G. quinqueloba* and *N. pachyderma*), and by 2) higher coiling ratios of sinistral to dextral tests of *N. pachyderma*; interglacials can be identified by assemblages similar to the modern distributions, or exhibiting warmer conditions than exist

presently at each site. Another aspect of glacial/interglacial fluctuations that particularly affects the continental margin sites is a change in the "neritic vs. oceanic" character of the surface water, resulting from 1) glacioeustatic changes in sea level which have resulted in shoreline migration through the Pleistocene in the order of 50 km off New Jersey, and 2) the degree of influence of the Gulf Stream over the continental margin (*i.e.* the influx of water from the subtropical gyre into the surface waters overlying the margin). Neritic conditions are indicated by relatively high percentages of taxa common in the Slope Water, such as *G. quinqueloba* and *N. pachyderma*, while oceanic conditions are indicated by high percentages of taxa which characterise the subtropical gyre, such as *G. falconensis*, *G. hirsuta*, and *G. truncatulinoides*.

Several climatic cycles are noted in the foraminiferal records at all three sites (Figures 23-25); in addition, there is an overall trend upcore at Site 604 toward increasing percentages of *G. quinqueloba* relative to subtropical foraminifera (Figure 26). This trend is interpreted as suggesting increasingly neritic conditions, *i.e.* a seaward shifting of the western boundary of the subtropical gyre, since percentages of *G. quinqueloba* are relatively high in the Slope Water, and actually increase shoreward of the Gulf Stream (Be and Tolderlund, 1971). The accompanying decrease in percentages of foraminiferal taxa which characterise the subtropical gyre also reflect decreasingly "oceanic" conditions above the margin (also see chapter 6).

There is a strong similarity between patterns of distribution exhibited by dinocysts downcore with those of planktonic foraminifera (whose living environment they share); both groups of planktonic microfossils tend to exhibit assemblages reflecting the same trends in surface water conditions (temperature and salinity).

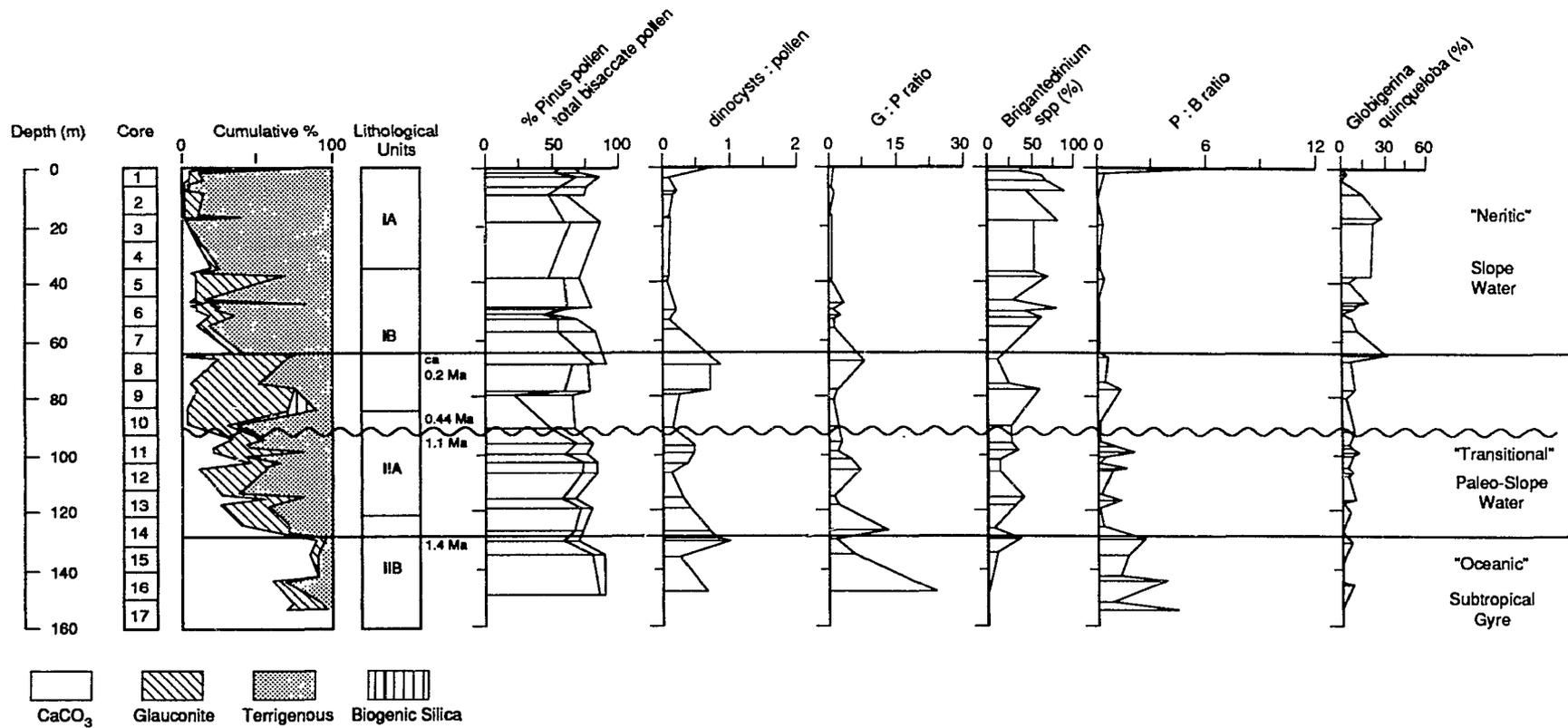


Figure 26. Summary of "oceanic" vs. "neritic" indicators in Pleistocene sediments at DSDP Site 604.

Protoperidinioid dinocysts are generally associated with colder, less saline waters than are the gonyaulacoid dinocysts (Mudie, 1980; Mudie *et al.*, 1989); the gonyaulacoid: protoperidinioid (G:P) ratio therefore serves as a crude paleoclimatic indicator. At all three sites, percentages of protoperidinioid taxa increase dramatically in glacial sediments (*i.e.* those with high coiling ratios and high percentages of "polar" foraminifera; see Figures 23-25). Protoperidinioid dinocysts are also common in neritic environments (Harland, 1983). There are close similarities in the record of Site 604 between the distribution of protoperidinioid dinocysts (*e.g.* *Brigantedinium* spp.), of gonyaulacoid dinocyst taxa characterising the neritic/upper slope environment, and the distribution of the foraminifer *G. quinqueloba*, substantiating the interpretation of increasingly neritic conditions above the New Jersey margin through the Quaternary. The increased amounts of terrigenous clastics in the sand-sized fraction, and the concomitant decrease in biogenic CaCO<sub>3</sub> also reflect more neritic conditions, as fluvial influx of sediments gains increasing prominence over the *in situ* contribution of biogenic calcium carbonate (Figure 26).

Glacial intervals in the lower Pleistocene sequence of Site 604, characterised by high percentages of sinistral *N. pachyderma* and very high coiling ratios (*e.g.* 88.95m, 94.22m, 97.3m, 100.3m, 127.6m, and 146.24m) are rich in *Bitectatodinium tepikiense*, which is a temperate dinocyst according to Wall *et al.* (1977). In the late Pleistocene sequence, high percentages of *B. tepikiense* are restricted to interglacials, while late Pleistocene glacials, characterised by high percentages of *G. quinqueloba* in addition to relatively high percentages of sinistral *N. pachyderma*, are dominated by *Brigantedinium* spp. Modern centres of abundance of *B. tepikiense* include coastal waters off Nova Scotia (Mudie, 1980) and Iceland (Turon, 1984), suggesting that it characterises cool temperate-subpolar surface water of variable salinity. There was a more or less continuous increase through the Quaternary of *Brigantedinium* spp. relative to *B. tepikiense*

at Site 604 during glacial intervals, but the first very high percentages of *Brigantedinium* spp. occur in the late Pleistocene sequence, coinciding with the appearance of significant percentages of the benthic foraminiferal species *E. excavatum* and *C. reniforme* (Scott, in prep.), which characterise "warm ice margin" environments (Scott and Medioli, 1980). *Brigantedinium* spp. are almost the only planktonic microfossils found in terminal Pleistocene sediments on the margin, supporting the default hypothesis (based on the absence of planktonic foraminifera) of lower salinity, cold surface water above the margin over this interval. Planktonic foraminifera are relatively sparse, and where samples contain statistically valid numbers of tests, assemblages are rich in *Globigerina quinqueloba*.

The *Impagidinium* Index, which allows estimates of SST, is plotted for Site 607 in Figure 27, as well as estimates of summer and winter SST calculated from the dinocyst assemblage using the Imbrie-Kipp CABFAC method and the transfer functions of Mudie *et al.* (1990). The rather poor correlation between the two curves may reflect the use of transfer functions which are inadequate to describe conditions at this site, or it may reflect post-depositional alteration of dinocyst assemblages (see below). The G:P ratio correlates well with the SST estimates, which is not surprising since the same dataset is used to generate the two indices.

There is generally good correlation in the upper 3 m with paleoclimatic interpretations based on the foraminiferal record (illustrated by the coiling ratio and isotope ratios measured on tests of *N. pachyderma*; Figure 26); below 3 m, however, dinocysts tend to give a much colder

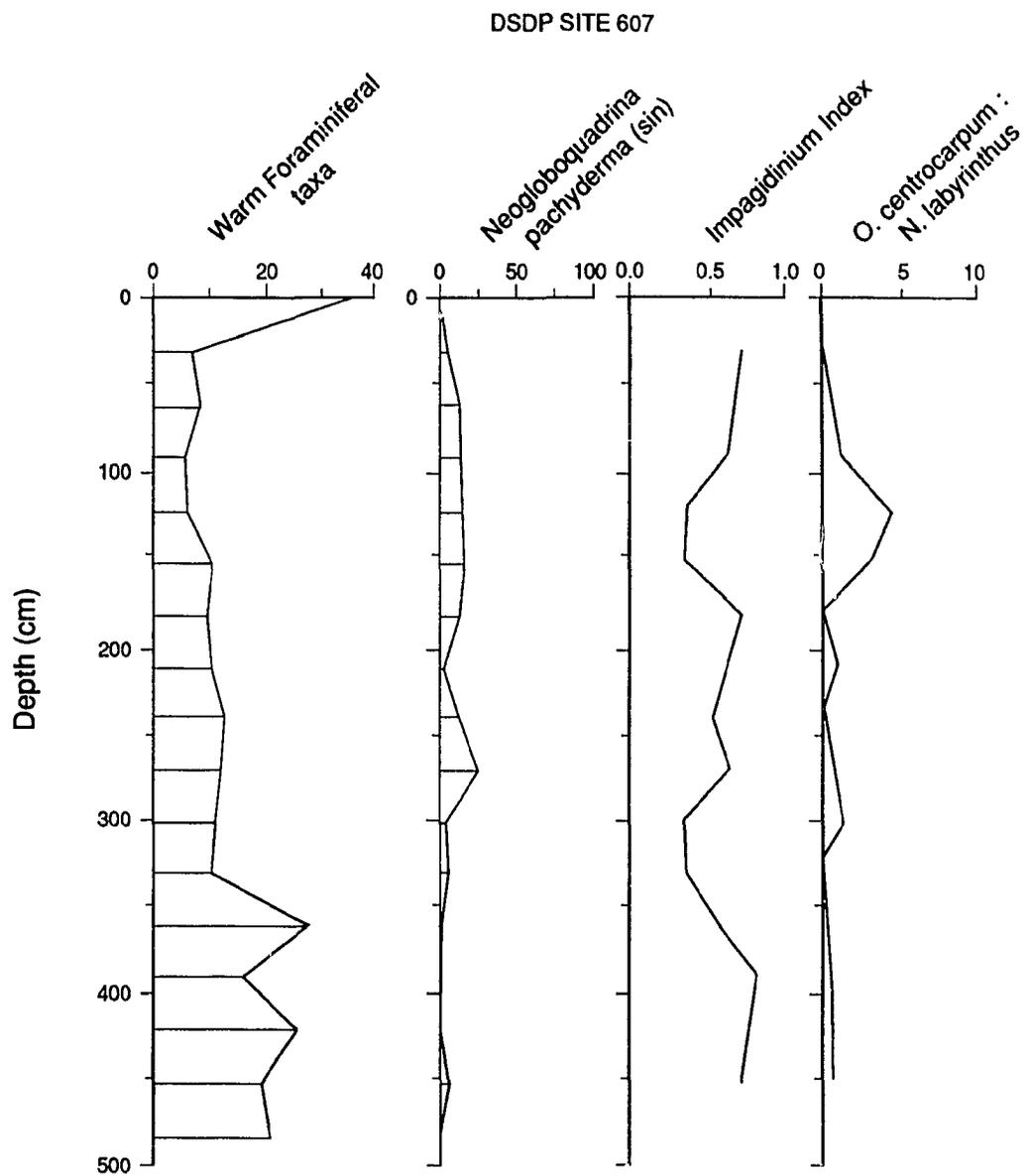


Figure 27. *Impagidinium* Index and ratio of *N. labyrinthus*: *O. centrocarpum* compound with the percentages of sinistral *N. pachyderma* and warm water foraminifera at Site 607.

climate signal than do the foraminifera, which record interglacial conditions. Oxygen isotope ratios below 3 m are also generally light. The reason for this difference must be sought.

The low dinocyst concentrations either reflect low dinoflagellate productivity in the subtropical gyre or higher oxidation of surficial sediments during interglacials. Because some taxa appear to be more resistant to oxidation than others, and therefore may be artificially concentrated in oxidized sediments, giving a false paleoceanographic signal, care must be taken in making paleoceanographic interpretations from dinocyst data, especially where independent evidence of oxidation (*e.g.* sediment colour) exists. The genus *Impagidinium*, for instance, seems to be highly resistant to oxidation, given the very high percentages of *Impagidinium* spp. in samples that are otherwise nearly barren, and the existence of a few specimens of *I. aculeatum* and *I. striolatum* in the surface sample at Site 607, which was essentially barren due to strong oxidation. The existence of high percentages of *Impagidinium* spp. therefore should not necessarily be interpreted as suggesting warm, oceanic conditions, especially where concentrations are low; an index comparing percentages of different species of this genus, therefore, probably gives a better paleoclimatic interpretation than the percentages of *Impagidinium* spp. in the assemblage. The existence of high percentages of *Brigantedinium* spp. with *Impagidinium* spp. in the lower 2 m of this site, in association with foraminiferal and isotopic evidence of warm sea surface temperatures suggest that this taxon is also highly oxidation-resistant.

The dinocyst *Operculodinium centrocarpum*, in contrast seems highly susceptible to oxidation, and is absent from the surface sample and present only in low percentages below 300 cm, where sediments appear oxidised. Although both *O. centrocarpum* and *N. labyrinthus* are opportunistic "weedy" species tolerant of large ranges of temperature and salinity on modern seasonal ice margins (Aksu *et al.*, in press; Mudie *et al.*, 1989), high ratios of *O. centrocarpum*

to *N. labyrinthus* appear to characterise glacial intervals (both stage 2, and stage 4), characterised by relatively high influx of terrigenous sediments and low oxidation at Site 607. The similarity in downcore abundances of *N. labyrinthus* with *I. aculeatum* support the use of the *N. labyrinthus*: *O. centrocarpum* ratio as a climatic indicator, at least in the western North Atlantic; this ratio may, however reflect the migration of the Gulf Stream across this site, so that it may only serve as a paleoclimatic indicator at sites influenced by the Gulf Stream. The ratio of these taxa is compared with the ratio of *I. aculeatum* + *striatum*: total *Impagidinium* spp. (*Impagidinium* Index) and percentages of sinistral *N. pachyderma* and warm water foraminifera in Figure 27.

Dinocysts tend to lead planktonic foraminifera in recording paleoceanographic changes in the sites studied. At Site 607, for instance, there is a striking similarity between *I. aculeatum* and "warm" planktonic foraminiferal taxa, but the *I. aculeatum* curve leads by ca. 30-40 cm (ca. 6000- 8000 y.) (Figure 27). This trend has also been reported in other areas; at ODP Site 646 (Aksu *et al.*, in press), for instance, peaks in primary productivity also lead ice volume by 2-4 ka.

There are several possible explanations for the more rapid response to climate change of the dinoflagellate record relative to the foraminiferal record. 1) Dinocysts (as well as coccoliths) may be reacting to factors that precede glaciation, or the initial increase in primary productivity may in fact contribute to climatic deterioration and glaciation by fixing carbon in organic matter, thus decreasing atmospheric CO<sub>2</sub>. 2) Phytoplankton may inhabit a different microenvironment, since foraminifera tend to live deeper in the water column, and therefore dinoflagellates and calcareous nannoplankton may be more susceptible to small changes in temperature, salinity, nutrients, insolation, etc. 3) Dinoflagellates and coccoliths may migrate more quickly into newly

established niches. Encysted dinoflagellates will survive adverse conditions for many years (Dale, 1983); Huber and Nipkow (1923) found cysts of *Peridinium cinctum* that were still viable after being encysted for 16.5 years. If conditions become favourable for a given taxon in an area, all of the viable cysts of that taxon will excyst and begin going through their life cycle, and producing numerous offspring. Foraminifera, however, have to migrate into an area; the distance and the ecological barriers would impart a lag in changing foraminiferal assemblages. 4) The most important factor controlling the rapid response of dinocyst assemblages to climatic change is probably a function of taxonomic nomenclature; dinoflagellate resting cysts are known to germinate into very different motile stages, and identical motile stages have produced cysts attributable to different genera (Reid, 1974). The dinoflagellate *Gonyaulax spinifera*, for instance, has produced cysts of *Nematosphaeropsis labyrinthus* (Wall and Dale, 1967), *Spiniferites elongatus*, *Spiniferites mirabilis*, and *Tectatodinium pellitum* (Wall and Dale, 1968), *Spiniferites ramosus* (Wall and Dale, 1970), *Bitectatodinium tepikiense* (Wall et al., 1977), and *Spiniferites membranaceous* (Bradford, 1978). Since these cysts have modern distributions that appear to be ecologically determined, environmental conditions presumably control the morphology of the cyst of theca (i.e. the phenotype) generated by the genotype.

#### 4.2 THE DISTRIBUTION OF PLANKTONIC MICROFOSSILS COMPARED WITH TERRESTRIAL SPORE AND POLLEN AND BENTHIC FORAMINIFERAL DATA

Glacials and interglacials can be recognised in the pollen record of the sites on the continental margin. Stadials are characterised by high percentages of the boreal tree pollen *Picea*

and high percentages of *Alnus* and herb pollen, suggesting colder continental climates; common planktonic microfossils in these sediments are the "polar" indicators, sinistral *N. pachyderma*, *G. quinqueloba*, and *Brigantedinium* spp. Relatively high percentages of *Quercus* and pollen of other temperate hardwood taxa as well as the temperate softwood *Pinus* characterise warmer intervals, along with higher percentages of planktonic microfossils recording relatively warmer surface waters, e.g. *Operculodinium centrocarpum*, *Bitectatodinium tepikiense*, *Spiniferites* spp., and temperate-subpolar planktonic foraminifera.

At Site 604 the pollen record is more complex than that of Site 612. There is an overall trend toward decreasing *Pinus* and increasing diversity in the pollen record, as well as decreasing dinocyst:pollen ratios through the Pleistocene at Site 604 (Figure 26). Microfossil concentrations alone do not provide much useful information because rates of sedimentation vary substantially through the core, with much faster sedimentation during the late Pleistocene (Lang and Wise, 1987); ratios of marine vs. terrestrial palynomorphs (*i.e.* dinocyst:pollen ratios) in the same sample therefore provide more reliable paleoenvironmental indicators. High ratios of dinocysts:pollen, and low diversity in pollen assemblages dominated by *Pinus* (and to a lesser extent by *Picea*, another bisaccate pollen grain suited to long-distance transport) below 60 m at Site 604 are interpreted to suggest "oceanic" conditions above the margin through most of the Pleistocene. In late Pleistocene sediments, percentages of *Pinus* deposited beneath the Slope Water mass are only ca. 55% on average, compared to values averaging ca. 70% in early Pleistocene sediments. Bisaccate pollen, in fact, comprises nearly 100 % of the assemblage in cores 15 and 16. The ratio of dinocysts:pollen grains exceeds 0.3 through most of the early Pleistocene, but in the late Pleistocene record, dinocyst:pollen ratios higher than 0.3 occur only in the surface sample. The lower dinocyst:pollen ratios probably reflect higher pollen influx, and therefore more neritic

conditions. The "oceanic" vs. "neritic" character of water overlying the New Jersey margin is controlled by climate, which determines both the sea level (and position of the coastline) and the path of the Gulf Stream (*i.e.* the edge of the subtropical gyre). If an organised, discrete Slope Water mass overlies the margin, limiting the effect of the Gulf Stream on the margin, fluvial runoff is "trapped" in the Slope Sea, and higher pollen fluxes result and more diverse pollen assemblages would be deposited on the margin. Conversely, if climates are warmer, sea level will be higher, inundating the coast, increasing the distance from the shoreline to the upper rise off New Jersey. In addition, if the Gulf Stream has a greater influence on the margin, as recorded below 35 m at Site 604, pollen assemblages will reflect more "oceanic" conditions, as the contribution of water from the Sargasso Sea will increase. The waters of the subtropical gyre are characterised by very low pollen concentrations, and are strongly dominated by *Pinus* (see Site 607). The interpretation of increasingly neritic conditions through the Pleistocene at Site 604 is consistent with the low planktonic foraminiferal concentrations and high percentages of the planktonic microfossil taxa *Brigantidium* spp., *G. quinqueloba* and *N. pachyderma*, which characterise the Slope Water.

Benthic foraminiferal data available from the DSDP sites (Scott, 1987, and in prep.) provide information on deep currents. Patterns of benthic foraminiferal distribution at the sites on the continental margin compare well with those of planktonic microfossils in late Pleistocene to Recent sediments. The growth of large ice sheets at mid-latitudes on the North American continent induced major changes in the water mass overlying the margin; major changes occurred in assemblages of both planktonic and benthic organisms, suggesting that both surface and deep water circulation was altered. The increasingly cold and neritic conditions suggested by the pollen, dinocyst and planktonic foraminiferal record are correlated with an increase in

benthic foraminiferal taxa which characterise a "warm ice margin" (Scott and Medioli, 1980); *Cassidulina reniforme* and *Elphidium excavatum* become present in low but significant percentages (2-8% *C. reniforme* and 2-10% *E. excavatum*) at Site 604 in lithological unit IB (core 9 to the base of core 6), and percentages of these "warm ice-margin indicators" increase further at the transition to lithological unit IA (up to 10 % *C. reniforme* and 26 % *E. excavatum* in cores 5 through 1). At Site 612, these species characterise the sediments above the unconformity at 37 m up to the surface 2 m.

The percentage curves of foraminiferal and palynological taxa recording the establishment of a relatively cold and fresh water mass above the margin are compared in Figure 26. The colder and more "neritic" character of the surface waters in lithological unit I, especially IA, suggests that the Gulf Stream (*i.e.* the western boundary of the subtropical gyre) influenced waters overlying the margin less during the late Pleistocene than during the early Pleistocene. The paleoceanographic interpretation is discussed in more detail in chapter 6.

The various fossil groups record the reworking of pre-Quaternary fossils into Pleistocene sediments, often in association with either glauconite (Neogene) or chalk (Eocene). Neogene planktonic foraminiferal taxa are especially common in lithological unit IB, and in the upper part of unit II (Figure 9); pre-Quaternary dinocysts (such as *Corrudinium harlandii*) are also present in many of the glauconitic or chalky samples. Because pollen can usually only be identified to genus, and most extant plant genera existed in the Tertiary, pre-Quaternary pollen and terrestrial spores were difficult to distinguish; even in samples known to contain completely reworked planktonic microfossils (as in the upper part of core 10), the majority of pollen and spores can be attributed to extant genera.

### 4.3 SUMMARY

Comparison of the downcore distributions of planktonic foraminifera, dinocysts, and pollen, as well as the unpublished benthic foraminifera of Scott (1987) shows that:

- 1) The downcore distribution of cysts of the planktonic dinoflagellates compares quite well with the planktonic foraminifera, which inhabit somewhat deeper levels in the water column, except that dinocysts tend to lead foraminifera in reflecting paleoecological/paleoceanographic change.
- 2) An increase in boreal pollen (e.g. *Picea*) on the continental margin correlates with planktonic microfossils indicating cold surface waters of slightly reduced salinity, and with the establishment of an ice-marginal benthic foraminiferal fauna. This probably corresponds to the Illinoian (stage 8) which Morley (1982) found to be associated with very cold, low salinity conditions in the North Atlantic.
- 3) Reworking is evident in the dinocyst, planktonic and benthic foraminiferal records in glauconitic and biogenic silica-rich sediments of Site 604.

## CHAPTER V

### HETEROGENEOUS EXPRESSION OF THE GLOBAL CLIMATE SIGNAL: IMPLICATIONS FOR CLIMATOSTRATIGRAPHIC INTERPRETATION

#### 5.0 BACKGROUND

The dominant factors controlling global climate are the amount of heat energy that reaches the Earth's surface, and the retention and redistribution of that heat. The planet Earth has two sources of heat: the sun (solar radiation) and radioactive isotopes in the crust (radiogenic heat). The rotation of the earth drives atmospheric and oceanic circulation, which redistributes the earth's heat; the patterns of atmospheric and oceanic circulation depend on the heat at different latitudes and on barriers to circulation.

The solar radiation that reaches the top of the atmosphere at different latitudes varies cyclically. Milankovitch (1941) published calculations of the effect of the eccentricity of the earth's orbit around the sun, the obliquity of tilt of the earth's axis of rotation and the precession of the equinoxes on insolation at subpolar latitudes in the northern hemisphere (55°N); these orbital parameters have periodicities of 100 ka, 41 ka and 23 ka (Hays *et al.*, 1976). The use of pestral analysis was able to reproduce these frequencies in proxy climate records in deep sea sediments, showing that the variation in the seasonal distribution of insolation induced by the three orbital periods (eccentricity, obliquity and precession) is the ultimate driving force for Quaternary

ice sheets and climatic fluctuations (e.g. Ruddiman and McIntyre 1981, 1984). The impetus provided by insolation forcing is quite small, however. Positive feedback mechanisms must be invoked if the relatively small variations in insolation are to explain climatic deterioration and glaciation, or conversely amelioration and deglaciation.

The definition of the Pleistocene has been linked with the concept of climatic deterioration since Forbes (1846) associated it with the "Ice Age". At the Eighteenth International Geological Congress in London in 1948 and again, at the 10th INQUA Conference in Birmingham in 1977, emphasis was placed on climatic deterioration in defining the Pleistocene. This climatostratigraphic definition contrasts with the rest of the stratigraphic record which is defined on lithological and paleontologic grounds, based on correlation with a designated stratotype.

The "practical" climatostratigraphic definition for the Plio-Pleistocene boundary, and a conceptual usage of the terms referring to the Pleistocene epoch and its stages evolved for two major reasons:

- 1) Because the evolutionary rate of most groups of plankton is slower than the very high precision demanded by Quaternary geologists (e.g. individual stages, or even events such as the Younger Dryas, spanning a few millenia) paleontological datums defined using planktonic species (favoured as stratigraphic markers due to their world-wide distribution and rapid migration in the open ocean) show some diachroneity at the resolution possible. Baldauf *et al.* (1987) and Hills and Thierstein (1989) for instance, showed that the first appearance of *Globorotalia truncatulinoides*, a datum used informally by marine micropaleontologists since the work of Banner and Blow (1967) to mark the base of the Pleistocene, was diachronous in the North Atlantic.

2) The terrestrial record of the Quaternary has been extensively studied, and these sediments are difficult to date due to the scarcity of fossils and the discontinuous sedimentation. There is a large data base of information in the continental record which is difficult to correlate with stratigraphic zones defined by the presence or absence of planktonic marine microfossils.

Highly accurate paleoenvironmental/ paleoclimatic interpretations are possible because of the abundance of relatively easily accessible sediments containing fossil assemblages for which modern analogs exist, thus there is less reason to question the application of the concept of actualism to Quaternary sediments than to any other time. However, unlike speciation and extinction, climatic change is not a unique event; climatic amelioration and deterioration occur relative to preceding climates. In addition, proxy climate signals based on microfossil assemblages reflect local climates, which are determined not only by the "global climate", but also by geographic and oceanographic factors such as latitude, altitude, continentality, migration of currents, etc. Even stable isotope ratios are affected by local conditions of temperature and salinity. Fairbanks (1983) found that geographic and vertical gradients of the oxygen isotopic composition of seawater in the euphotic zone of the Atlantic Ocean are largely controlled by freshwater (fluvial) discharge. Thus, although it may be possible to accurately reconstruct paleoclimates at any given location by examining biotic, sedimentologic, and isotopic data, correlating between sites according to a stratigraphy based on global climatic change requires an understanding of the local effects of global climatic change. Criteria such as the achievement of a certain temperature, the inception or end of glaciation, or the appearance of taxa adapted to particular climatic conditions, will inevitably be diachronous. "The volume of continental ice is a variable which is not necessarily synchronous with the variations of climate..... The same world-wide thermal fluctuation can have a different manifestation in time from one region to another."

(Occhietti, 1983, page 12). Thus, while global ice volume is a parameter which can be roughly measured (providing temperature and salinity effects on the oxygen isotope signal are known, which is rarely possible), "global climate" is not quantifiable. In addition, it is difficult to justify the association of the Pleistocene with climatic deterioration and simultaneously with the first appearance of *Globorotalia truncatulinoides*, given the growth of large northern hemisphere ice sheets and subsequent onset of ice rafting in the North Atlantic around 2.4 Ma (Raymo *et al.*, 1989; Shackleton *et al.*, 1984), nearly a million years before the FA (first appearance) of *G. truncatulinoides*.

The different local responses to the global climate signal makes correlating between the terrestrial and marine record even more difficult. Glacial sediments (*i.e.* those transported and deposited directly by ice) are easily identified in the terrestrial record, but the majority of these deposits cannot be dated. Glacial-marine sediments, in contrast, can generally be dated biostratigraphically and using oxygen isotope ratios, but it is difficult to identify these records as "glacial" where there is no direct evidence of continental ice sheets (*e.g.* ice rafted debris, ice-marginal faunas, etc.). The terms "glacial" and "interglacial" are qualitative, since continental ice sheets have existed since the Oligocene (Miller *et al.*, 1987); the terms therefore express different degrees of global ice extent, compared to conditions only up to a few millenia before. Some authors have applied the term "interglacial" to intervals with warmer climates, lower continental ice volume, and consequently, higher sea levels than the present; by this definition, oxygen isotope substage 5e constitutes the last interglacial (*e.g.* Kukla *et al.*, 1972). By this definition, the Holocene is not an interglacial, since  $^{18}\text{O}/^{16}\text{O}$  values as light as those found in stage 5e are not found in stage 1. Others take a broader definition of "interglacial", and consider all of stage 5 to be interglacial (*e.g.* Suggate, 1974; Fulton, 1986). Attempts by various workers (*e.g.* Lamb *et al.*,

(Occhietti, 1983, page 12). Thus, while global ice volume is a parameter which can be roughly measured (providing temperature and salinity effects on the oxygen isotope signal are known, which is rarely possible), "global climate" is not quantifiable. In addition, it is difficult to justify the association of the Pleistocene with climatic deterioration and simultaneously with the first appearance of *Globorotalia truncatulinoides*, given the growth of large northern hemisphere ice sheets and subsequent onset of ice rafting in the North Atlantic around 2.4 Ma (Raymo *et al.*, 1989; Shackleton *et al.*, 1984), nearly a million years before the FA (first appearance) of *G. truncatulinoides*.

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1987; Boellerstorff, 1978) to climatostratigraphically correlate between the terrestrial and marine records, therefore, often are very different, depending on the assumptions made.

## 5.1 CLIMATOSTRATIGRAPHY OF THE SITES IN THIS STUDY

The approach taken in this study was to examine more than one group of microfossils to improve the validity of paleoecological interpretations, to evaluate post mortem effects (e.g. dissolution, transportation), and to compare paleoecological changes at sites on the slope, rise and mid-ocean ridge environments. Sedimentological and micropaleontological changes at each site record how the climate evolved in different sectors of the North Atlantic in response to Quaternary changes in global climate. The fossil plankton record reflects conditions in the surface waters (temperature, salinity), and the record of pollen and terrestrial spores reflects the deposition of terrestrial particles which reflect vegetation, and hence climate, on the continent.

The climatic information derived from the micropaleontological, sedimentological and isotopic data permitted the identification of late Quaternary glacial and interglacial stages at each site, and provided the basis for "multiple event stratigraphy", *i.e.* applying "physical happenings and bioevents jointly in delineating Pleistocene stages rather than reliance on a single datum" (Lamb *et al.*, 1987). In this chapter, the climatostratigraphic (= chronostratigraphic) interpretation of the three sites studied, based on the microfossil record and oxygen isotope record, is presented; this will establish the framework for correlation in later discussions.

Most Quaternary marine micropaleontology, especially biostratigraphic work, has been done in mid-ocean sites characterised by continuous sedimentation, "layer cake" stratigraphy, and abundant microfossils. The record of the last glacial cycle at DSDP Site 607 on the Mid-Atlantic Ridge therefore was taken as the reference site against which the late Quaternary record of the more complex sedimentary record of the margin was compared.

### 5.1.1 Mid-Atlantic Ridge: Site 607

The change in lithology at 28 cm at Site 607, from pure yellow foraminiferal ooze to grey foraminifera-rich sediments with higher percentages of terrigenous material and higher dinocyst concentrations appears to mark the glacial/interglacial transition (Termination I). The ratio of  $^{18}\text{O}:^{16}\text{O}$  rises sharply from 0.963 ‰ to 2.305 ‰, and the ratio of  $^{13}\text{C}:^{12}\text{C}$  falls to -0.005 between the surface and 31 cm, suggesting an increase in global ice volume and a decrease in metabolic activity in the oceans (Figure 23). The stage 1/2 boundary is marked a sharp decrease in "warm water" foraminiferal taxa, the establishment of a subpolar-transitional foraminiferal assemblage dominated by *G. bulloides*, dextral *N. pachyderma*, and *G. inflata* at 31 cm. Oxygen isotope ratios measured on dextral tests of *N. pachyderma* show a 5.4°C drop in temperature between the surface and 31 cm. The summer SST (19.7°C), estimated from dinocyst assemblages, is close to the modern value of 19-20°C, but winter SST was calculated to be 11.6°C, considerably cooler than the modern value for SST of approximately 14°C (Figure 18). Although the very low pollen concentrations at this site preclude interpretations of the paleovegetation, the higher concentrations of pollen in relatively dark sediments with polar-subpolar planktonic microfossil assemblages support the interpretation that more terrigenous input was reaching the mid-latitude Mid-Atlantic

### Ridge during glacials.

The very high percentages of *Brigantedinium* spp. at 50 cm (Mudie, 1987) and at 62-65 cm, together with the relatively light oxygen isotope signal measured on dextral tests at 62-65 cm ( $1.998\text{‰}$ ), indicate that meltwater was reaching this site during the summer. Sinistral *N. pachyderma* in this sample record a heavy isotopic signal ( $2.778\text{‰}$ ), suggesting that surface water temperatures were very cold and/or that global ice volumes were still quite large. The convergence of the oxygen isotope curves at this point, relative to the glacial maximum, suggests either a decrease in seasonality or the effect of meltwater overriding temperature signals as suggested by Scott *et al.* (1986) from the Mid-Atlantic Ridge around 45°N. Very low ratios of  $^{13}\text{C}:^{12}\text{C}$  in both sinistral and dextral tests indicate low productivity. It would appear, therefore, that meltwater reached this site several thousand years before the beginning of the Holocene, probably with the inception of deglaciation at mid-latitudes. A radiocarbon age of 16390  $\pm$  110 years B.P. on foraminiferal tests in this sample supports the interpretation that the high concentrations of *Brigantedinium* spp. over this interval records meltwater issuing from the initial pulse of deglaciation.

High numbers of cysts of *Brigantedinium* spp. can indicate influx of cold, relatively fresh water, since this protoperidinioid cyst is abundant in Arctic waters (Mudie and Short, 1985), or alternatively, high numbers of these cysts can be associated with upwelling conditions, as off the northwestern margin of South America (Wall *et al.*, 1977). The cold sea surface temperatures indicated by the planktonic foraminiferal assemblage, as well as the low foraminiferal productivity recorded by carbon isotopes in tests of *Neogloboquadrina pachyderma*, suggest that the influx of cold water of slightly reduced salinity is the most probable explanation. The relatively light

oxygen isotope signal would therefore reflect the influx of isotopically light meltwater at Site 607; the signal is not *very* light, possibly due to considerable mixing with oceanic waters of the central North Atlantic by the time it reached Site 607. Although dinocyst concentrations are high from 50 to 65 cm for this site (ca. 2000 cysts/cc), they are not excessively high relative to dinocyst concentrations in foraminifera-rich sediments in the early Pleistocene sediments of Site 604, for instance. The relatively high primary productivity could result from the release of nutrients from the ice sheets, rather than from upwelling. The inference that high numbers of cysts of *Brigantedinium* spp. in these sediments reflect the influence of meltwater reaching Site 607 is consistent with the inferred location of the glacial Polar Front during the last glacial maximum.

The stage 2 glacial maximum is interpreted to be recorded at 122 cm, based on the higher percentage of *Brigantedinium* spp., the low *Impagidinium* Index (Edwards *et al.*, 1991), the higher ratio of sinistral:dextral *N. pachyderma* and very low percentages of *G. inflata* and "warm water" planktonic foraminifera and very heavy oxygen isotope values (2.091‰ in dextral *N. pachyderma* and 3.199‰ in sinistral tests).

The stage 2/3 boundary is placed between 122 and 150 cm; the low seasonality indicated by the difference in isotopic signals from tests of dextral and sinistral *N. pachyderma* of only 0.296‰ also characterises the stage 2/3 boundary at site HU71-377 of Scott *et al.* (1986).

Interstadial conditions (stage 3) are indicated to below 241 cm by relatively low coiling ratios and high percentages of "warm water" taxa, reflecting conditions similar to (but somewhat cooler than) the modern subtropical gyre. Relatively light oxygen isotope ratios of around 1.8‰ were measured in dextral *N. pachyderma* at 181 and 212 cm. The stage 3/4 boundary is placed

below 241 cm, where the ratio of  $^{18}\text{O}:^{16}\text{O}$  in dextral *N. pachyderma* rises sharply to 2.074‰; the ratio in sinistral tests is only 0.33‰ heavier, again suggesting either low seasonality or the effect of meltwater on the signal. Low seasonality is exhibited at the 2/3 and 3/4 boundaries, but the high seasonality suggested by the large difference between oxygen isotope signals measured in sinistral (2.974‰) vs. dextral (1.833‰) tests at 181 cm is also characteristic of stage 3 at site HU71-377 of Scott *et al.* (1986).

The very high coiling ratio (>1.5) of *N. pachyderma* at 272 cm is interpreted as the glacial maximum in stage 4. *Brigantedinium* spp. and *O. centrocarpum* dominate the assemblage, comprising 24 and 22 % of the sample, respectively. These microfossil assemblages indicate that polar surface waters existed over Site 607; the relatively high percentages of "warm water" plankton (*e.g.* subpolar-tropical foraminifera and *Impagidinium* spp., dominantly *I. aculeatum*), suggest that climatic/oceanographic conditions were variable. Warm waters may have reached the site seasonally, or the position of the polar front may have fluctuated on a longer time frame. The ratio of *N. labyrinthus*: *O. centrocarpum* is low. The ratio of  $^{18}\text{O}:^{16}\text{O}$  in dextral *N. pachyderma* is only 1.736‰, but a very heavy ratio (2.936‰) was measured in sinistral *N. pachyderma* from this sample. The low foraminiferal productivity suggested by very low ratios of  $^{13}\text{C}:^{12}\text{C}$ , especially in the sinistral tests, in addition to the dominance of the dinocyst record by "weedy" opportunistic taxa suggest that the light isotopic signal reflects the influx of meltwater. The much heavier oxygen isotope ratios in sinistral tests relative to dextral tests also suggest that the isotopic composition of the surface waters during the warm summer months was much lighter than it was during the cold winter months, *i.e.* high seasonality. The relatively large range between ratios measured in dextral and sinistral tests is characteristic of glacial intervals at this site, being associated with high coiling ratios at 62 and 122 cm, and again at 272 cm. At 300 cm, no cysts

of *Brigantedinium* spp. were observed; polar foraminifera (sinistral *N. pachyderma* and *G. quinqueloba*) fall to lowest values since the surface. Despite the temperate conditions recorded by the microfossils in the sample at 300 cm (e.g. dinocyst assemblage dominated by gonyaulacoid taxa *Bitectatodinium tepikiense*, *O. centrocarpum* and *N. labyrinthus*, and foraminiferal assemblage dominated by *G. inflata* and dextral *N. pachyderma*), oxygen isotopes in both dextral and sinistral tests are fairly heavy (2.219 and 2.608‰, respectively). This suggests that the growth of continental ice sheets during stage 4 preceded oceanographic/climatic changes at Site 607, i.e. that migration of the Polar Front to these latitudes in the North Atlantic (the stage 4 glacial maximum) occurred some time after widespread continental glaciation. This sample records the brief occupation of the "transitional" water mass over Site 607, as climatic deterioration caused the southward migration of water masses.

The 4/5 boundary is placed between 300 and 331 cm. The brevity of stage 4 relative to stage 2 is consistent with the data of other workers which show that this was a short, but very cold, stage (e.g. Cline and Hays, 1976; Hillaire-Marcel and de Vernal, 1989).

From 331 cm down to the lowest sample at 481 cm, dinocyst concentrations are generally low, possibly reflecting oxidation as in the Holocene sediments in the upper 28 cm of this site; this is consistent with the lighter colour of the sediments over this interval, which is similar to the colour of the surficial sediments. Interglacial conditions are indicated by relatively light oxygen isotope signals, heavy carbon isotope values, very low ratios of sinistral:dextral *N. pachyderma* and very high percentages of "warm water" foraminiferal taxa, but dinocysts show some peaks in *Brigantedinium* spp. and variable percentages of *I. aculeatum* and *Impagidinium* species, sometimes lower than in stage 4. In many areas, glacial-interglacial transitions are marked by

huge increases in primary productivity (Shaffer, 1990); in the Labrador Sea, for instance, glacial-interglacial transitions are marked by increased species diversity, and the protoperidinioid dinocyst genus *Brigantedinium* is replaced by *O. centrocarpum*, *N. labyrinthus*, and *S. ramosus* (de Vernal and Mudie, 1989). At Site 607 Termination II is not marked by an increase in productivity (Mudie, 1987 and this study). Harland (1989), however, found peaks of dinoflagellate cysts associated only with stages 1, 11, 13 and 17 on Rockall Plateau, northeast Atlantic Ocean. The reason for this heterogeneity in dinocyst productivity is not understood, but probably reflects a low salinity (meltwater) influence (*c.f.* Aksu *et al.*, 1992) and oxidizing bottom water (depending on deep-water circulation), or variations in nutrient availability resulting from slightly different circulation of surface currents in different interglacials.

### 5.1.2 Climatostratigraphic interpretation of data from sites on the New Jersey Margin

It is relatively difficult to apply zonations developed for mid-ocean sediments to continental margin sediments, where calcareous plankton are scarce and depositional processes are complex, with frequent erosion and resedimentation. The use of several paleoceanographic indicators is essential in obtaining a reliable paleoenvironmental signal from continental margin sediments; reliance on data from a single source (*e.g.* foraminifera, dinocysts, stable isotopes, etc.) can lead to erroneous interpretations.

The Quaternary record (upper 37 metres) of Site 612, on the mid-slope, is easier to interpret than the more complex record of Site 604 on the upper rise, since core recovery at Site 612 was very good, and the sediments are relatively homogeneous, with no evidence of reworking

or hiatuses above the erosional unconformity at 37 m defined as the Plio-Pleistocene boundary by shipboard biostratigraphers (Shipboard Scientific Party, 1987c). The paleoclimatic/climatostratigraphic interpretation of DSDP Site 612 therefore is outlined first, and then it is compared to the record of Site 604.

#### 5.1.2.1 Site 612 (Mid-Slope)

The surface sample contains an interglacial fauna, characterised by very low percentages of polar planktonic foraminifera and relatively low percentages of protoperidinioid dinoflagellate cysts (*i.e.* relatively high G:P). Percentages of "warm water" foraminifera (e.g. *G. menardii*, *G. ruber*, etc.) and dinocysts (*Impagidinium* spp., especially *I. aculeatum*) are relatively high, although much lower than those found at the mid-ocean site.  $^{18}\text{O}:^{16}\text{O}$  values are light ( $< 2\text{‰}$ ). Percentages of "warm water" taxa are much lower in the next sample at 230 cm, and the oxygen isotope signal becomes much heavier from 133 to 283 cm, increasing from 1.67 to 2.778‰. The upper 2 m are therefore interpreted as Holocene, oxygen isotope stage 1.

Mid Wisconsinan-Holocene sediments are preserved above the erosional unconformity at 37 m; high resolution interpretations are possible, including the glacial maximum as well as the Allerod-Younger Dryas oscillation. The  $^{14}\text{C}$  date of 11,320 +/- 100 years B.P. at about 15m suggests extremely high sedimentation rates in the late Pleistocene/early Holocene. From 2 to 16.5 m, planktonic foraminiferal assemblages dominated by polar taxa, sinistral *N. pachyderma* and *G. quinqueloba*, and the abundance of the protoperidinioid dinocyst *Brigantedinium simplex* suggest that these sediments were deposited during the Younger Dryas. This is consistent with

the very heavy oxygen isotope values obtained where foraminiferal concentrations permitted, suggesting large global ice volumes and low sea level. The *Cassidulina reniforme/ Elphidium excavatum* benthic foraminiferal fauna over this interval (Scott, 1987) also reflect a "warm" ice margin (Scott and Medioli, 1980), consistent with late Wisconsinan ice margins at Long Island.

The very low foraminiferal concentrations in core 2 (6 to 12 m), the dominance of taxa resistant to dissolution and adapted to slightly reduced salinities (*i.e.* *N. pachyderma*, *G. inflata*, and *G. quinqueloba*), as well as the relatively low dinocyst concentrations dominated by *Brigantedinium* spp. suggest that high rates of sedimentation existed during deglaciation, when sediment-laden meltwater reached Site 612; this suggests that approximately between 10.9 and 10.5 ka, large amounts of meltwater from disintegration of the Laurentide ice sheet were passing through the Hudson River into the western North Atlantic above the New Jersey margin.

The microfossil assemblages indicate that glacial conditions returned after the initial deglaciation: percentages of "cold water" (polar) planktonic microfossil taxa (*e.g.* *Brigantedinium* spp. and sinistral *N. pachyderma*) rise sharply in the lower part of core 1 (around 5 m). Planktonic foraminiferal concentrations approximately doubled over values in core 2, suggesting that clastic sediment input was reduced. This probably records the Younger Dryas cooling between about 11.5 and 10.5 ka. Interestingly, the pollen records slightly warmer conditions in the lower part of core 1 than in core 2, with lower percentages of *Alnus*, *Betula* and *Picea*, and higher percentages of *Pinus* and temperate hardwoods such as *Acer* and *Fagus*. This return to full glacial conditions in the Slope Water, together with the identification of terrestrial records of the Younger Dryas oscillation along the Atlantic coast of North America (Mott *et al.*, 1986; Petect *et al.*, 1989) and in western Europe (Mangerud, 1984) suggests that the Younger Dryas is a

climatic change that is either the product of, or at least augmented by surface ocean circulation. The Younger Dryas probably results from the seaward deflection of the Gulf Stream due to the effect of meltwater, as suggested by Broecker and Denton (1989). Cold water of polar origin (Inner Labrador Current) was able to penetrate between the deflected Gulf Stream and the North American continent, allowing full glacial conditions to return to the Slope Water mass, and therefore allowing re-expansion of ice sheets along Atlantic coast, which had quickly disintegrated with the shoreward migration of the Gulf Stream during the Allerod.

Low ratios of sinistral:dextral *N. pachyderma* and high percentages of *O. centrocarpum* and *Spiniferites* spp. are found between 16 and 26 m. These data indicate relatively warm conditions postdating the late Wisconsinan glacial maximum, correlative to the Allerod. Microfossil assemblages are similar to Holocene assemblages, dominated by dextral *N. pachyderma* with significant percentages of "warm water" foraminifera, and a dinocyst assemblage similar to the modern upper slope ecofacies (Wall *et al.*, 1977); microfossil concentrations over this interval, however, are lower than Holocene values. Benthic foraminifera still suggest "warm ice margin" conditions (Scott, 1987). This would be consistent with warming atmospheric and oceanic conditions at mid latitudes associated with large calving continental ice sheets in coastal regions of North America.

Short intervals of much colder conditions are recorded in the microfossil and isotope records around 27 and around 31 m, separated by a longer interval characterised by interstadial-type conditions; because there is no change in sedimentology between these samples, the recurrence of stadial conditions does not appear to be an artifact of re-sedimentation. It is therefore concluded that the late Wisconsinan (stage 2) glacial maximum consisted of two short very cold

stadials separated by a short "interstadial". This feature would not be evident in regions with lower sedimentation rates due to the homogenising effects of bioturbation. This, of course, assumes that there were no errors during coring resulting in the resampling of the same interval.

Planktonic foraminifera at DSDP site 612 suggest that surface waters off New Jersey were colder but more saline during during the late Wisconsinan glacial maximum than during the Younger Dryas. Maximum ratios of sinistral:dextral tests of *N. pachyderma* during the glacial maximum were roughly 8:1, compared to maximum ratios of 1.3:1 during the Younger Dryas. Concentrations of foraminiferal tests are much higher in glacial maximum sediments than in Younger Dryas sediments, however, suggesting lower productivity, higher sedimentation rates and/or more intense dissolution during the Younger Dryas. Dinocyst assemblages dominated by *Brigantedinium simplex* suggest that surface waters off New Jersey during the Younger Dryas were cold and below normal marine salinity, by analogy with the results of Mudie and Short (1985) from Baffin Bay. *Brigantedinium* spp. are less common in samples from the late Wisconsinan glacial maximum. Lower surface salinity during the Younger Dryas would explain the lower concentrations of planktonic foraminifera, high percentages of *G. quinqueloba*, and the dominance of solution-resistant tests in the planktonic foraminiferal assemblage in sediments with moderately high concentrations (over 300/10cc) of benthic foraminifera (Scott, 1987).

The cold, relatively low salinity surface waters at this site through the late Wisconsinan are consistent with a sea level stand 121 +/- 5 m below present during the last glacial maximum (Fairbanks, 1989), which would bring the shoreline to within 50 km of this site. The low salinities in stage 2 are consistent with the very low salinities in the North Atlantic during stage 2 inferred from radiolarian assemblages (Morley, 1982). The even lower salinities during the

Younger Dryas support the hypothesis of influence of meltwater issuing from the Laurentide Ice Sheet on global climate (c.f. Broecker and Denton, 1989). However, the cooling trend precedes the influx of meltwater into the Slope Water, with a marked increase in the cooling rate at 16.22 m. It is possible that the meltwater that entered the Gulf of Mexico when deglaciation began deflected the somewhat cooler and less saline Gulf Stream offshore (c.f. Nurser and Williams, 1990), allowing a greater influx of water of polar origin (the Inner Labrador Current) into the Slope Water early in deglaciation.

None of the samples examined in this study, nor in that of Scott (1987), record conditions as warm as, or warmer than, the present. Poag and Low (1987) report that sample 5-2 (120-124), i.e. at 3630 cm, contains relatively high percentages of "warm water taxa", especially *Globorotalia flexuosa*. They interpret this assemblage as having been deposited during zone X of Ericson and Wollin (1968), which is equivalent to oxygen isotope stage 5e. However, very cold "glacial" conditions are recorded by microfossil assemblages in sample 5-2 (80-86), at 3590 cm (Figure 25) and throughout the rest of the core to ca. 2.5 m. This, in addition to the very high sedimentation rates indicated by the  $^{14}\text{C}$  date of 11,320 y. B.P. at 15m, suggests that only mid-Wisconsinan to Holocene sediments are preserved above the unconformity at 37 m.

The unconformity at 37 m (Figure 24) may separate Pliocene and Pleistocene sediments, as suggested by Poag and Low (1987). Non-glaucconitic samples, or those with rare glauconite, below 37 m are relatively rich in *G. crassaformis* and other "warm water" foraminifera; the sediments were definitely deposited under warmer waters than were the samples in the upper 37 m. However, this could have been during the early Pleistocene, given the lack of distinct Pliocene biostratigraphic markers. The microfossil assemblages below the unconformity are similar to early

Pleistocene assemblages found at Site 604. The erosional surface at 37 m therefore probably represents a significant hiatus, since even assemblages in non-glaucanitic sediments below the unconformity record conditions warmer than those recorded in late Pleistocene sediments at Site 604. Glaucanite-rich samples at Site 612 contain assemblages similar to the glaucanite-rich samples at Site 604 on the upper rise, rich in *Operculodinium israelianum* and *Tectatodinium pellitum*, as well as in pre-Quaternary foraminifera and warm-water foraminifera (e.g. *Globigerinoides sacculifer*, *Pulleniatina obliquiloculata*, and *Globorotalia crassaformis*).

#### 5.1.2.3 Site 604 (Upper Rise)

Unlike Site 612, the sedimentology of the Quaternary sequence at Site 604 is highly variable. There is a general increase in biogenic carbonate downcore, and a decrease in terrigenous sediments (Figure 9). Large amounts of resedimented glaucanite are sporadically present, becoming more common below 65 and 90 m. Moullade's (1987a) biostratigraphic planktonic foraminiferal study of Site 604 did not consider the sedimentology of the samples examined, nor the coiling direction of *Neogloboquadrina pachyderma*, and ignored seismic (Poag and Mountain, 1987; Wise and van Hinte, 1987) and biostratigraphic (Lang and Wise, 1987) evidence suggesting the existence of an unconformity. The chronostratigraphy of the late Quaternary section at this site is therefore reinterpreted in this study.

The surface sediments at both Sites 604 and Site 612 are dominated by subpolar-transitional foraminiferal taxa (mainly dextral *Neogloboquadrina pachyderma* and *Globorotalia inflata*), reflecting the location of both sites beneath the Slope Water. The dinocyst assemblage

in the surface sediments at both sites is dominated by *Brigantedinium simplex* and *Operculodinium centrocarpum*, with significant percentages of *Spiniferites* spp. and *Impagidinium* spp.; this is similar to the upper slope ecofacies of Wall *et al.* (1977). The relatively high percentages of the "polar" protoperidinioid dinocyst, *Brigantedinium* spp. may reflect the influence of the Labrador Current on the Slope Water. The high percentages of this taxon could alternatively be associated with upwelling conditions as off the northwestern margin of South America (Wall *et al.*, 1977). The relatively low dinocyst concentration in these samples, combined with the foraminiferal data indicating cold, relatively fresh surface waters, however, suggest that the high percentages of *Brigantedinium* spp. in the surficial sediments, and especially in Wisconsinan sediments of the sites on the continental margin, are caused by the influence of "polar" surface waters (the Labrador Current?).

Given the similarity in the surface sediments of Sites 604 and 612, it is therefore likely that earlier in the Quaternary, similar conditions existed above the mid-slope and upper rise of New Jersey, so that the record of the two sites should be comparable. The data from lithological unit IA Site 604 therefore are compared with the upper 37 m of Site 612.

Interglacial conditions are indicated in the upper 450 cm of Site 604 by subpolar-transitional foraminifera (dextral *N. pachyderma*, *G. bulloides* and *G. inflata*) with significant percentages of "warm water" foraminifera, and the dinocysts *Operculodinium centrocarpum*, *Spiniferites* spp., *Bitectatodinium tepikiense*, and *Impagidinium* spp. The pollen assemblage is characterised by relatively low percentages of boreal-tundra pollen (*i.e.* *Picea* and herb pollen).

The Holocene record at Site 604 spans the upper 4.5 m, compared to the 2 m record at Site 612; Holocene sedimentation rates on the upper rise are therefore over twice as high as those on the mid-slope, assuming complete preservation and recovery at both sites. It is therefore unlikely that the upper 15 m of Site 604 would record Ericson and Wollin's zones W-7 (late Illinoian-Recent), as suggested by Moullade (1987a), when the upper 15 m of Site 612 record sedimentation since 11370 y. B.P. (near the end of zone Y). In addition, microfossil assemblages recording conditions as warm as, or warmer than, the present were not recovered over this interval in Moullade's (1987a) study nor in this study, placing Moullade's (1987a) interpretation of this interval as including Ericson and Wollin's zone X (Sangamon) in question. On the contrary, all of the samples between 5 and 47 m record conditions colder than the present. Very low concentrations (ca. 1/cc) of "cold water" planktonic foraminifera and very high percentages of the protoperidinioid dinocyst genus *Brigantedinium* from 5.05 to 7.9 m indicate that these sediments were deposited under conditions of low salinity/high sedimentation, *i.e.* during deglaciation, possibly analogous to the meltwater pulse associated with the Younger Dryas at Site 612. Just below this unit, at 8.04 m, relatively high coiling ratios and high percentages of *G. quinqueloba* suggest late glacial conditions, possibly a return to full glacial conditions following the Allerød; the slight degree of reworking of Neogene glauconite (13% glauconite in the sand fraction) in the sample at 8.04 m suggests that the sediments which were not recovered just below this might have been glauconitic, since there are frequently problems associated with recovering glauconite sands.

The poor core recovery below 8.04 m and the seismic data recording probable unconformities around 35 and 90 m (Figure 8) precludes the climatostratigraphic zonation of the record with any degree of confidence. There is no evidence of conditions as warm as, or warmer

than the Holocene in lithological unit IA (upper 35 m); this unit is thought to represent the mid-Wisconsinan to Recent, consistent with the interpretation of Lang and Wise (1987) from the nannofossil record that the upper 90 m at Site 604 were deposited in the last 0.44 Ma. Moullade (1987a) reports a similar planktonic foraminiferal assemblage recording warm conditions in core 4, which may be mid-Wisconsinan. Glacial conditions recorded in core 5 (just below the possible unconformity at 35 m) by high percentages of polar foraminifera and very low percentages of "warm water" foraminifera, high percentages of the dinocyst genus *Brigantedinium*, and high percentages of *Picea* and herb pollen, could therefore be interpreted as early Wisconsinan (stage 4, 74-65 ka) or early Illinoian (stage 8, 302-252 ka), depending on the hiatus, and depending on the age of the single underlying interglacial, *i.e.* whether the interglacial sediments are Sangamon or Yarmouthian. The poor core recovery unfortunately does not permit unequivocal interpretations, but the data suggest that similar environmental conditions are recorded over long depth spans, suggesting rapid sedimentation. Lang and Wise (1987) also interpreted the nannofossil record as indicating rapid sedimentation (min. 201 m/Ma) over this interval.

The assignment of the zonation of Ericson and Wollin (1968) to the sediments of this site by Moullade (1987a) is of questionable value, since "warm water taxa" upon which biostratigraphic and climatostratigraphic zonations are based (such as the marker species for the base of the Pleistocene: *G. truncatulinoides*) are seldom present due to relatively cool, fresh surface waters in the Slope Sea. In addition, the climatostratigraphic zonation of Ericson and Wollin assumes a "layer cake" stratigraphy and a complete sedimentary record with no erosional gaps. A quick examination of the sediments shows that Site 604 is unlikely to contain a complete, undisturbed Quaternary record; the sedimentology is highly variable, and the contorted and convoluted beds of lithological unit IB were interpreted by the Shipboard Scientific Party (1987a)

to represent slump deposits. Lang and Wise (1987) suggest from calcareous nannofossil data at Site 604 that a significant hiatus (max 0.65 Ma duration) occurred in core 10 (around 90m), corresponding to seismic reflection horizon P1. Moullade (1987b, page 1281), however, does not even refer to their calcareous nannofossil data when he concludes "The equivalents of all Ericson's climatic zones are defined in the Quaternary section of Hole 604, which seems to be sedimentarily complete, according to both the foraminiferal and nannofossil records." There may well be additional hiatuses too short to be identified using biostratigraphic markers, but of sufficient duration to upset a climatostratigraphic interpretation, given that the last glacial cycle spanned only ca. 120 ka.

The poor core recovery may have caused the Sangamon interglacial (stage 5e or zone X) to be missed in the upper 35 m, but it is less likely that stages 5 and 7 would be missed due to poor recovery. Wisconsinan sedimentation rates at Site 604 are unlikely to be significantly lower than at Site 612, given that Holocene rates were over twice as fast at Site 604. The best interpretation of the data would be to equate the upper 35 m (lithologic unit IA) at Site 604 with the upper 37 m at Site 612, *i.e.* mid Wisconsinan-Recent. The very low microfossil concentrations from 5 to 8 m might represent the meltwater spike associated with the Younger Dryas between 10.9 and 10.5 ka at Site 612. Although the poor core recovery in the upper 35 m at Site 604 does not really permit unequivocal climatostratigraphic assignments, the interpretation presented here would be more consistent with the interpretation of Lang and Wise (1987) which suggests that cores 1 through 10 were deposited over the last 440 ka, or since oxygen isotope stage 12. I do not feel that sufficient information is available to allow me to better interpret the chronostratigraphy of lithological units IB and II. The calcareous nannofossil record (Lang and Wise, 1987) appears to provide the best biostratigraphic information.

The abundance of reworking is also a problem at this site; Tertiary radiolarians are found in samples 2-CC and 6-CC, 10-CC, and 11-CC; Pleistocene radiolarians are absent from the upper 13 cores, and Pliocene radiolarians occur in cores 14-24 (Shipboard Scientific Party, 1987a). Moullade (1987a) reports Eocene foraminifera in sample 2-1 (90-94), and Miocene foraminifera in some samples in core 11 and most of core 10. This is especially a problem since many Neogene taxa are still extant, and therefore may be indistinguishable from the *in situ* signal.

Cores 8 through 10 are rich in glauconite, reworked from Miocene strata which are presently exposed on the mid- to lower slope at the edge of the Eocene outcrop belt. Significant percentages (>1%) of identifiably reworked foraminifera are present in these samples, and foraminiferal tests in which the calcium carbonate has been partly or completely replaced by glauconite are common. Foraminiferal assemblages in the glauconite sands that characterise most of core 10 are partly due to totally reworked Miocene foraminifera; this was also identified by Moullade (1987a). Dinocyst assemblages in glauconite-rich sediments are generally rich in *Operculodinium israelianum*, suggesting that this taxon may be reworked. It is occasionally common in samples with little glauconite, however, which could either mean that it can be an *in situ* signal, reflecting warmer surface waters than at present at times in the early Pleistocene, or alternately, that some reworked Miocene sediments are non-glauconitic, and therefore are more difficult to identify. The difficulty of identifying Neogene reworking in Quaternary sediments is that most Neogene taxa are still extant; reworking may therefore be impossible to identify, will almost inevitably be underestimated, and may result in erroneous paleoclimatic interpretations.

The very large sampling interval which Moullade (1987a) used in applying the zonation of Ericson and Wollin is also a problem. This study has indicated, for instance, that cores 11

through 9, assigned by Moullade (1987a) to zone R, which corresponds to a global climatic warming (Ericson and Wollin, 1968; Briskin and Berggren, 1975), actually record dominantly cool conditions, indicated by high percentages of sinistral *N. pachyderma*, and ratios of sinistral:dextral tests of *N. pachyderma* in sample 10-4 (84-87), for example, of nearly 1 (0.958), and also fairly high in samples 9-1 (140-143), 9-2 (131-134), and in the lower half of core 11 (Figures 9 and 11); a few samples within this interval record warm conditions (e.g. 11-1 (132-135) and 11-2 (140-143)), but so does the surface of core 12 (sample 12-1, 32-35 cm), which Moullade (1987a) included in zone Q. The stable isotope work of Ganssen (1987) also shows relatively heavy isotopic values in *G. inflata* in cores 9-11, as heavy as those in cores 8 and 12, and sample 11-3 (90-94) gave the heaviest O18:O16 ratios that he measured in the core, again suggesting that the assignation of these cores to zone R is probably incorrect.

From 84 to 160 m (lithological units IIA and IIB) large-scale climatic fluctuations are recorded with a greater frequency, suggesting lower sedimentation rates; at least 4 glacial-interglacial cycles exist in the lower 76 m compared to only 2 identified in the upper 84 m (Figure 24).

Moullade's (1987a) determination of the Plio-Pleistocene boundary on climatic grounds is also weak, being based on the lowest occurrence of *G. truncatulinoides* in sample 17-1 (91-94) and the subjective determination that "the final 'warm' event of the Pliocene seems to correspond better to an interval ending with sample 17-3 (91-94)" (Moullade, 1987a, page 497). This contrasts with the calcareous nannofossil zonation of Lang and Wise (1987), which placed this boundary at 162.1 m, within core 18, at the base of the *Calcidiscus macintyreii* zone.

The boundary between the upper and lower Pleistocene is also controversial; Poag (1987) considers the late/early Pleistocene boundary at 84 m, while this same boundary is placed between cores 11 and 12 (102m) in Poag and Mountain (1987), in the same volume. Lang and Wise (1987) consider the mid-Pleistocene to be missing within core 10 (ca. 90 m), making this the boundary between early and late Pleistocene sediments. Moullade (1987b) places this boundary at the base of core 9 on the basis of the last occurrence of *G. tosaensis*, a very rare species anywhere at this site due to the relatively cold surface waters overlying the margin.

## 5.2 SUMMARY

The calcareous oozes at Site 607 record continuous pelagic sedimentation through the Quaternary, and the magnetostratigraphy is well established (Ruddiman *et al.*, 1987), with the upper 140 m (Miocene-Recent) readily correlated with the magnetic polarity time scale of Berggren *et al.* (1980). At Site 607 there is little post-depositional sediment disturbance or contamination by reworked microfossils. Planktonic foraminifera are abundant, and there are generally sufficient dinoflagellate cysts to permit paleoceanographic interpretations. The distance from land results in very low concentrations of pollen, however.

Climatostratigraphic interpretation of the record at Site 607 on the mid-ocean ridge is relatively straightforward, since sedimentation is continuous, and the "warm water", oceanic taxa used to define the Quaternary and its stages are abundant; oxygen isotope stages 5 to 1 are easily identifiable in the planktonic foraminiferal record, by examining coiling ratios of sinistral:dextral *N. pachyderma* and percentages of "warm water" vs. "cold water" taxa. In the dinocyst record,

the *Impagidinium* Index of Edwards *et al.* (1991), as well as the ratio of gonyaulacoid to protoperidinioid taxa allowed the distinction of glacial and interglacials; the ratio of *O. centrocarpum*: *N. labyrinthus* also seems to reflect glacial and interglacial conditions at this site (Figure 27).

The interpretation of the upper 480 cm at this site as recording oxygen isotope stages 5-1 is consistent with the calcareous nannofossil zonation at this site which places the NN21/NN20 boundary (0.27 Ma) around 10.5 m, and with the magnetostratigraphy which places the Brunhes/Matuyama boundary (0.73 Ma) at ca. 36 m (Baldauf *et al.*, 1987).

Climatostratigraphic interpretation of continental margin sites is difficult, because the Slope Water is relatively cold and fresh, so that the foraminiferal taxa used to define the Quaternary and its stages tend to be scarce, and at times during the Quaternary, oceanographic conditions prohibited the growth of calcareous plankton. The presence of euryhaline dinocysts, e.g. *Brigantedinium* and *Operculodinium* (Mudie, 1987) allow the interpretation that very cold surface waters of reduced salinity, probably with sea ice, existed above the margin when calcareous plankton were scarce (*i.e.* during stadials). The relatively high rates of homogeneous sedimentation in the upper 37 m at Site 612 permitted the identification of sediments spanning the mid Wisconsinan to Holocene, including resolution of the two-pronged glacial maximum, and the late glacial Allerod/Younger Dryas climatic reversal.

The interpretation of the upper 37 m of Site 612 as recording late isotope stage 3 to the present is consistent with the very high sedimentation rates indicated by the low concentrations of microfossils in these sediments. Unfortunately, the biostratigraphic data at this site appear to

be nondiagnostic; while the site report (Shipboard Scientific Party, 1987c) identifies the upper 36.92 m as belonging to the *Emiliana huxleyi* zone (CN15), and the sample at 36.92 as belonging to the *Gephyrocapsa oceanica* zone, Abbott (1987), identified the diatom assemblages from 17.9 to 72.2 m as being Pliocene to late Miocene.

The complex sedimentology at Site 604, the erosional hiatus suggested by the seismic reflector and the calcareous nannofossil evidence of Lang and Wise (1987) for a hiatus spanning 0.44-1.1 Ma, precludes the assignment of any climatostratigraphic zonation to the record of this site. Interglacial conditions are not recorded below the Holocene (upper 4 m) in lithological unit IA; seismic data suggest an unconformity around 35 m, however, so that the interglacial spanning ca. 47-51 m might represent stage 5 or any other post-440 ka Pleistocene interglacial.

The upper 35 m of Site 604 (lithologic unit IA) are interpreted as recording mid Wisconsinan to Holocene sedimentation, equating this unconformity-bound unit with the upper 37 m at Site 612 above the erosional unconformity previously interpreted by most workers as the Plio-Pleistocene boundary. Rapid sedimentation in the Upper Pleistocene sequence (upper 84 m), consistent with the interpretation of rates of at least 201m/Ma by Lang and Wise (1987) based on calcareous nannofossils, is evidenced by the expanded records of only 2 glacial-interglacial cycles. Much slower rates of sedimentation in the Lower Pleistocene sequence are evidenced by microfossil concentrations on average five times higher than in the upper 84 m, and the existence of at least 4 glacial-interglacial cycles over 76 m in contrast to the 2 glacial cycles recorded in the upper 84 m. This interpretation differs substantially from that of Moullade (1987a, b), but seems to be more consistent with the sedimentological, micropaleontological and palynological data, and with descriptions of other sites on the margin.

## CHAPTER VI

### QUATERNARY PALEOCLIMATES AND THE PALEOCEANOGRAPHY OF THE WESTERN NORTH ATLANTIC OCEAN

#### 6.0 INTRODUCTION

Globally, the Paleocene and Eocene were probably substantially ice free; high productivity in surface waters, shelf erosion during sea-level lowstands, and rapid sediment accumulation on the slope resulted in extensive mass wasting on the U.S. Atlantic margin (Mountain and Tucholke, 1985). High benthic foraminiferal  $^{18}\text{O}$  values suggest the presence of continental ice sheets during much of the last 36 Ma, *i.e.* Oligocene-Recent (Miiler *et al.*, 1987). Cool, strongly circulating bottom water entered the North Atlantic in the early Oligocene due to global cooling and the opening of the Greenland Sea to the Arctic Ocean, and an abyssal boundary current formed along the U.S. margin (Mountain and Tucholke, 1985). Climatic deterioration accelerated in the late Miocene, and the surface waters of the northeast Atlantic changed in a series of steps from warm temperate to subpolar between 3.4 and 2 Ma, and extensive northern hemisphere glaciation began ca. 2.4 Ma B.P. (Shackleton *et al.*, 1984; Loubere and Moss, 1986). Prior to 2.4 Ma, global ice sheet volumes 1/4 to 1/2 as large as late Pleistocene volumes are suggested by  $^{18}\text{O}$  variations, while after 2.4 Ma, ice sheets appear to be 1/2 as large as late Pleistocene ice sheets (Raymo *et al.*, 1989).

Climatic deterioration accelerated again in the late Pleistocene. Sea Surface temperatures

in the mid-latitude North Atlantic Ocean were relatively warm until 1.05 Ma (Ruddiman *et al.*, 1978). Oxygen isotope values have fluctuated throughout the Quaternary, but were relatively light during the Matuyama reversed epoch, until the Jaramillo normal event, dated at 0.92 Ma B.P. (Van Donk, 1976). The first  $^{18}\text{O}$  maximum occurred in isotopic stage 22 at 0.85-0.8 Ma (Shackleton and Opdyke, 1973), which Boellstorff (1978) correlates with the Nebraskan glaciation in the land record. Around this time (ca. 0.9 Ma) changes in  $^{18}\text{O}$ , %  $\text{CaCO}_3$ , and estimated SST increased in amplitude by roughly a factor of 2, suggesting that northern hemisphere ice sheets grew to maximum values during the Brunhes chron around twice those attained during the Matuyama (Ruddiman and Wright, 1987). Sudden deglaciations every 100 ka are the most salient features of late Pleistocene  $^{18}\text{O}$  records, suggesting that ice sheet growth is a long and gradual process, while ice decay is rapid and sudden (Ruddiman and Wright, 1987).

Ruddiman *et al.* (1989) concluded from the coherent, in-phase responses of  $^{18}\text{O}$ ,  $^{13}\text{C}$ ,  $\text{CaCO}_3$  and SST at the rhythms of orbital forcing that northern hemisphere ice volume changes have initiated most of the North Atlantic surface-ocean and deep ocean responses during the Pleistocene. The results of this study show that northern hemisphere glaciation also had a profound effect on water masses and circulation in the western-central North Atlantic; furthermore, these oceanographic changes may have controlled sedimentary processes on the United States margin north of Cape Hatteras. The results presented in the next section will examine these responses in detail.

## 6.1 PALEOCEANOGRAPHY OF THE SUBTROPICAL GYRE, RECORDED IN LATE QUATERNARY SEDIMENTS OF DSDP SITE 607, MID-ATLANTIC RIDGE

Oxidising bottom waters appear to have existed at Site 607 through the Holocene. Harland (1989) reports that the sediments of Rockall Bank are barren of dinoflagellates from stage 11 to lower stage 4; oxidising conditions may therefore have existed in different parts of the North Atlantic at different times. Planktonic foraminiferal concentrations at Site 607 are high, either suggesting high productivity and/or little dilution by other types of sediment, e.g. terrigenous. Percentages of "warm water" planktonic foraminifera (gyre-margin to subtropical taxa of Kipp, 1976) are high, and sinistral tests of *Neoglobobulimina pachyderma* comprise only 1.3% of the assemblage.

The high percentage of *Impagidinium aculeatum*, high *Impagidinium* Index, and high percentages of *Nematosphaeropsis labyrinthus* at 31 cm characterise the latest Pleistocene; these taxa are "oceanic", and characterise the subtropical North Atlantic today. The sample contains no *Operculodinium centrocarpum*, an exception to the ubiquitous presence of this taxon in other samples from Site 607 (Figure 16 and Mudie, 1987). The foraminiferal record is characterised by a "cooler" transitional-subpolar assemblage, but contains few polar foraminifera, and the coiling ratio is only 0.151. Oxygen isotope ratios indicate that there was a 5.4 °C temperature drop between the surface and this sample. SST (summer) was calculated from the dinocyst assemblage at 19.7°C, approximately the modern value for summer SST in this area of 19-20°C. SST (winter) was calculated to be 11.6°C, which is colder than the modern value for winter SST in this area of approximately 14°C, according to the *World Ocean Atlas*.

Heavy  $^{18}\text{O}$  values ( $2.30\text{‰}$ ) in dextral *N. pachyderma* appear to indicate a climatic reversal following the warming trend upcore recorded by the oxygen isotope values and by planktonic microfossil assemblages. Ruddiman and Duplessy (1985) concluded that there was a two-step decrease in the global volume of ice, with the first around 17-15 ka and the second around 10-7 ka. Although this sample would be calculated to have been deposited 8400 years B.P., assuming constant sedimentation rates, it is unlikely that the top was completely recovered, so that the true age of these sediments would be older than the estimate. It is also probable that Holocene rates were considerably lower than Wisconsinan rates, given the marked decrease in terrigenous sedimentation at this site at the end of the Wisconsinan. This sample therefore could well have been deposited between 15 and 10 ka, between the two phases of ice volume decrease.

Samples at 50 cm (Mudie, 1987) and 62 cm are dominated by *Brigantedinium* spp. (nearly 90 % at 62 cm), suggestive of cold water of less than normal salinity; this is interpreted as meltwater issuing from melting of the Laurentide Ice Sheet. A radiocarbon age of 16390  $\pm$  110 years B.P. at 62-65 cm supports the meltwater theory; this date falls within the first "step" of deglaciation, from 17 to 15 ka B.P. Relatively high dinocyst concentrations occurred during meltwater spike, probably the result of the influx of nutrients from the ice sheet. These findings are in agreement with the frequent high productivities at glacial-interglacial transitions reported by other authors (e.g. Schaffer, 1990), but the dinocysts characterising the meltwater at Site 607 are the heterotrophic protoperidinioid *Brigantedinium* species; there is no peak in the autotrophic gonyaulacoid *N. labyrinthus/O. centrocarpum* reported for the Labrador Sea (Aksu *et al.* submitted).

If this interpretation of the microfossil assemblage is correct, the "meltwater" signal is

recorded over at least 12 cm, which corresponds to over 2000 years at the average sedimentation rates at Site 607. This long-lived spike could be due to the influx of meltwater or icebergs into the central gyre of the North Atlantic south of the glacial Polar Front since the inception of deglaciation, while meltwater would only reach sites at higher latitudes in the North Atlantic late in deglaciation when the polar front had retreated to this position. It is possible that lower salinity meltwater and icebergs would reach Site 607 over a span of a few thousand years, since the Gulf Stream would transport meltwater issuing first from the Mississippi River (*c.f.* Broecker and Denton, 1989) and then via rivers entering the Atlantic from the eastern United States (*e.g.* Susquehanna River, Mohawk-Hudson River system, and then the St. Lawrence River), as glacioisostatic rebound altered drainage patterns.

High percentages of *Operculodinium centrocarpum*, *Brigantedinium* spp., *Impagidinium patulum*, and the presence of *Spiniferites ramosus* characterise stage 2 prior to the influx of meltwater. Relatively high percentages of polar foraminifera, high coiling ratios, low *Impagidinium* indices and heavy  $^{18}\text{O}$  values characterise the interval from 91-153 cm, suggesting the influx of polar surface waters over Site 607, consistent with the existence of the glacial Polar Front just north of Site 607 (around 45°N, core 71-377 of Scott *et al.*, 1986) in the late Wisconsinan (stage 2). The sample at 122 cm is interpreted to be the glacial maximum from the higher percentage of *Brigantedinium* spp. and the low *Impagidinium* index, and the higher ratio of sinistral:dextral *N. pachyderma* and very low percentages of "warm water" planktonic foraminifera. Concentrations of planktonic foraminifera are relatively low, reflecting high terrigenous component and/or minor dissolution. This again is similar to the findings of Scott *et al.* (1986) in core 71-377.

The high *Impagidinium* index, high ratio of *N. labyrinthus*: *O. centrocarpum* and high percentages of gonyaulacoid dinocysts from 153 to 241 cm suggest interstadial conditions. In the foraminiferal record interstadial conditions are indicated by relatively low sinistral:dextral *N. pachyderma* ratios and high percentages of "warm" taxa. The ratio of *O. centrocarpum*: *N. labyrinthus* may in fact reflect migration of the Gulf Stream in response to changing global climates. *O. centrocarpum* is considered to be a Gulf Stream indicator (Harland, 1983), while *N. labyrinthus* characterises oceanic environments such as the North Atlantic subtropical gyre (Wall *et al.*, 1977). The path of the Gulf Stream is known to be climatically controlled; Crowley (1981), for instance, found that flow increased in the North Atlantic Current at the approximate time of major polar advances of the last 150 ka, and southward deflection of the Gulf Stream in response to increased flow has been reported by Stommel (1965). Circulation in the North Atlantic is known to have been much more zonal during the last glacial maximum, with the Gulf Stream flowing almost directly across the North Atlantic around 42°N latitude to Spain (Keffer *et al.*, 1988). During stage 2, then, the Gulf Stream may have flowed over Site 607, transporting dinocysts from the Slope Water, while during stage 3, the path of the Gulf Stream may have migrated north of Site 607, depositing "oceanic" dinocysts such as *N. labyrinthus* and *Impagidinium* spp.

The ratio of sinistral:dextral tests of *N. pachyderma* taxa rises to highest values at this site (>1.5) at 272 cm, suggesting that the most southerly migration of the Polar Front occurred in the early Wisconsinan (stage 4). This is similar to the Nova Scotia margin (Boyd *et al.*, 1988; Scott *et al.*, 1989b). *Brigantedinium* spp. and *O. centrocarpum* dominate the assemblage, comprising 24 and 22 % of the sample, respectively. The ratio of *N. labyrinthus*:*O. centrocarpum* is again low, suggesting that the Gulf Stream flowed above this site during the early Wisconsinan.

Percentages of "warm water" planktonic foraminiferal taxa are low relative to samples immediately above and below, but much higher than stage 2 values. Similarly, the abundance of *Impagidinium* species, including *I. aculeatum* is somewhat puzzling, given the classic interpretation of *I. aculeatum* as an indicator of interglacial conditions (Edwards *et al.*, 1991) This "mixed signal" could be due to the brevity of the early Wisconsinan glacial and the slow rates of sedimentation at this site, resulting in "homogenisation" of the record, and/or an unstable position of the polar front at this latitude over the duration of the early Wisconsinan, resulting in fluctuations in the fauna over the 1 1/2 cm (ca. 300 year) span of the sample. The two short intervals of very cold, saline surface water during the late Wisconsinan glacial maximum in the record from Site 612 (Figure 24) support the fluctuation of climatic and accompanying oceanographic conditions over short time spans. A final possibility is greater productivity of subtropical-temperate plankton within the central gyre of the North Atlantic during stage 4 than stage 2. The influx of low percentages of *I. aculeatum* early in late Pleistocene glacial intervals, especially prior to stage 2 in the Labrador Sea is interpreted by Mudie and Aksu (1984), de Vernal (1986) and de Vernal and Mudie (1989) as indicating continued advection of warm surface currents into the Labrador Sea during the early parts of glacials. If the production of *Impagidinium aculeatum* was very high in central gyre during earlier glacials, however, the influx of low percentages of this species into the Labrador Sea may not reflect as great a component of water of southern origin carried north by the North Atlantic Drift as previous workers have suggested.

*Spiniferites* species, indicators of Slope Water together with *O. centrocarpum* (Wall *et al.*, 1977; Harland, 1982) are present in low but significant quantities from 272-362 cm. The dominant dinocyst taxa at 303 cm, *Bitectatodinium tepikiense*, *N. labyrinthus*, and *O. centrocarpum*, are indicators of the Gulf Stream/North Atlantic Drift system (Wall *et al.*, 1977);

a similar assemblage exists at 331 cm, but with much lower percentages of *B. tepikiense*. This suggests a seaward deflection of the Gulf Stream/North Atlantic Drift system prior to stage 4 at mid-latitudes, possibly by high latitude glaciation, or possibly by a change in the prevailing wind patterns.

From 331 cm down to the lowest sample at 481 cm, sediments are calcareous oozes with little terrigenous sediment, and dinocyst concentrations are generally low, possibly reflecting oxidation, but less extreme than at the surface. Interglacial conditions are indicated by very low ratios of sinistral:dextral *N. pachyderma* and very high percentages of "warm" foraminiferal taxa. Oxidising conditions appear to have existed at Site 607 during interglacials. Dinocyst assemblages are more variable and equivocal over this interval, however, with high percentages of *Brigantedinium* spp. in association with *Impagidinium* spp., especially *I. aculeatum*.  $^{18}\text{O}$  values are also variable, with light isotopic values recorded in most samples, but a fairly heavy ratio was measured at 450 cm. SST estimates in stage 5 were affected by the use of the more productive *Brigantedinium*-rich samples; sediments deposited under warmer conditions had very low dinocyst concentrations, and so were not employed in the statistical analysis. These may have been relatively cold, but given the cautions about using dinocysts as paleoecological indicators where oxidation appears to have occurred, these estimates may not be reliable. In addition, the factor analysis used had relatively low communalities, suggestive of a no-analog situation; the data set used in designing the transfer function may not be applicable to this area.

## 6.2 PALEOCEANOGRAPHY OF THE SLOPE WATER, RECORDED IN QUATERNARY SEDIMENTS FROM SITES 604 AND 612 ON THE NEW JERSEY MARGIN

Lower Pleistocene sediments on the upper rise (lithological unit II at Site 604) are foraminifera-rich muds with glauconitic intervals. Relatively high concentrations of planktonic foraminifera on the upper rise in lower Pleistocene sediments, especially in sediments deposited prior to 1.4 Ma, are similar to modern high calcium carbonate concentrations (averaging about 30%) in a band extending beneath warm Gulf Stream waters on the lower continental rise north of Cape Hatteras. In contrast, carbonate comprises only roughly 5-20% of the sediments beneath the modern Slope Water mass (Emery & Uchupi, 1972), where 80-90 % of particles collected in sediment traps on the continental margin are smaller than  $63\mu\text{m}$  (Gardner *et al.*, 1985). The relatively low pollen concentrations and the dominance of bisaccate pollen grains (especially *Pinus*) which are adapted for long-distance transport, in the foraminifera-rich muds in unit IIB at Site 604 are typical of "oceanic" conditions.

The planktonic microfossils in early Pleistocene interglacial sediments of Site 604 reflect relatively warm water, normal marine salinity, and more "oceanic" conditions than are presently found above the upper rise; significant percentages of "warm water" foraminifera and relatively high planktonic foraminiferal species diversity are found, especially those characterising the subtropical gyre, and high G:P ratios and high percentages of "warm-water dinocysts", *e.g.* *O. israelianum*, *T. pellitum*, also occur. The data therefore suggest that there was no water mass comparable to the Slope Water separating the western boundary of the subtropical gyre from the continental margin during early Pleistocene interglacials. The "north wall" of the Gulf Stream was therefore several hundred kilometres shoreward of its present position off New Jersey during early

Pleistocene interglacials (Figure 28A).

The position and intensity of the Gulf Stream is sensitive to climatic conditions. During early Pleistocene glacials, the Gulf Stream was deflected from the continental margin due to two factors: 1) Cooling in itself would strengthen and deflect the Gulf Stream; Nurser and Williams (1990) showed that cooling the separated boundary current of Parsons (1969) results in a 15 Sv strengthening and southward/eastward movement of up to 3000 km. 2) Glacioeustatically-lowered sea levels expose a large percentage of the continental shelf, increasing deflection of the Gulf Stream by bathymetric highs like the Charleston Bump (Pinet *et al.*, 1981; Pinet and Popenoe, 1985).

The position and intensity of the southward-flowing Labrador Current is also sensitive to climatic change. During the mid-Holocene it was deflected seaward, allowing warm water to penetrate northward along the continental shelves of the United States and Canada to southern Newfoundland, and subsequently assumed its present position closer to the coast of Atlantic Canada in response to climatic deterioration (Fillon, 1976; Scott *et al.*, 1984). Climatic deterioration during the early Pleistocene would shift the path of the Inner Labrador Current (or its precursor) landward, thus allowing the cold surface water of Arctic (or Subarctic) origin to penetrate between the Gulf Stream and the margin. The influx of polar surface water over the margin is indicated by high percentages of polar-subpolar plankton at Site 604 (e.g. *N. pachyderma*, *G. quinqueloba*, *G. bulloides*, *Brigantedinium* spp., *Bitectatodinium tepikiense*, etc).

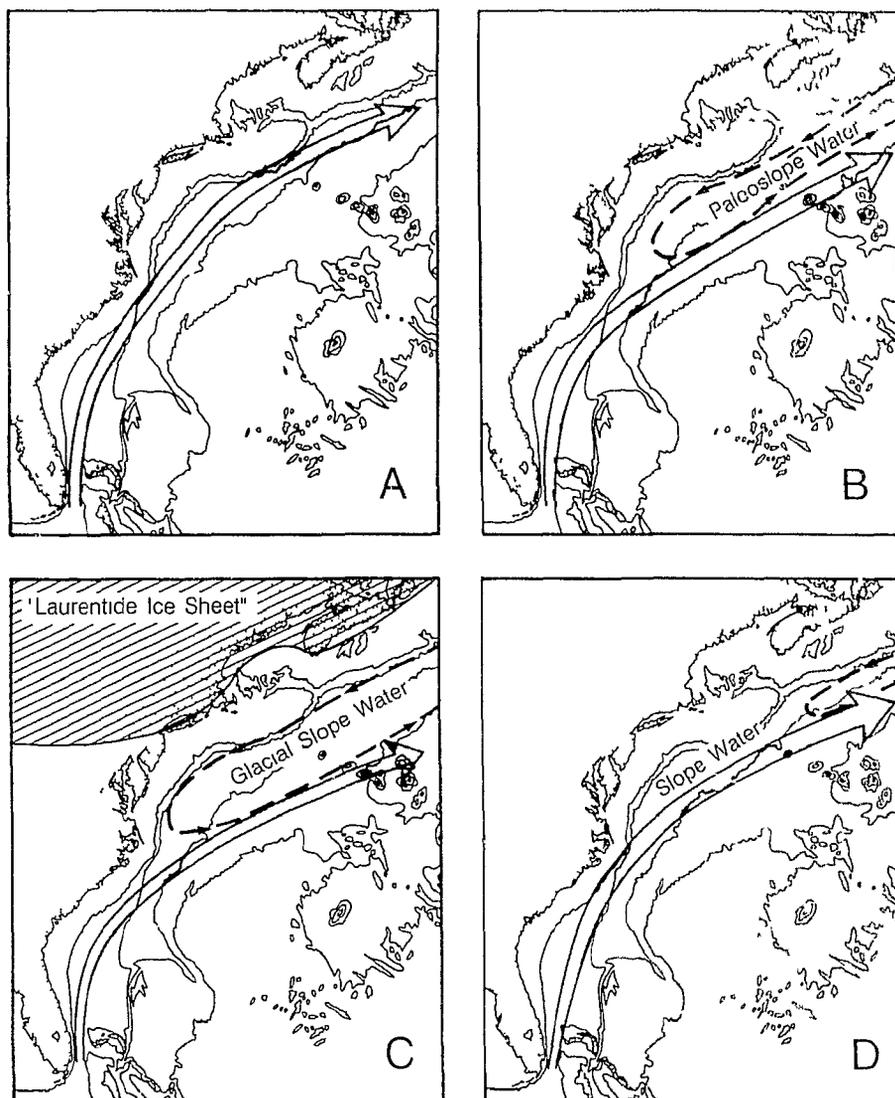


Figure 28. Path of the Gulf Stream through the Quaternary in response to climatic change.

- A) Early Quaternary: Gulf Stream flowed above the lower continental margin north of Cape Hatteras; no water mass separated the Gulf Stream from the margin. B) Late early Quaternary (ca. 1.4 Ma): Gulf Stream was deflected offshore as water of polar origin penetrated between the Stream and the margin. The "Paleo-Slope Water" mass developed above the margin. C) Late middle Quaternary: mean path of the Gulf Stream was deflected further offshore and eddy mixing mechanisms were suppressed by the oceanographic conditions in the "Glacial Slope Water". D) Modern mean path of the Gulf Stream.

During early Pleistocene glacials, significant percentages of warm-water taxa persisted, despite high cooling ratios and high percentages of "cold water" taxa; these "mixed" signals are consistent with fluctuations in the location of the Gulf Stream or ensuing from frequent incursions of Gulf Stream water into the colder, fresher water on the margin. This suggests that there was continued influence of the Gulf Stream above the margin even during early Pleistocene glacials. The influx of warm water from the Gulf Stream was suppressed only when polar conditions (cold water with slightly reduced salinity) existed over the margin, indicated by high percentages of *Brigantedinium* spp. and by intensive dissolution of calcareous plankton and sparse foraminiferal assemblages dominated by solution-resistant taxa, such as *N. pachyderma*, especially sinistral tests (Berger, 1968), or higher foraminiferal concentrations and more diverse foraminiferal assemblages rich in *G. quinqueloba* (as in the upper part of core 16). During the early Pleistocene, these oceanographic conditions were relatively rare above the New Jersey margin, and during subsequent deglaciations, the Gulf Stream migrated shoreward, and "oceanic" conditions were re-established above the margin.

The early Pleistocene "Slope Water" mass, resulting from the mixing of water from the Labrador Current with fluvial influx and with considerable contribution of water from the Sargasso Sea (contributed by instabilities, or eddies, of the Gulf Stream) is referred to as "Paleo-Slope Water" (Figure 28B). The overall trend toward increasing diversity in the pollen assemblages on the upper rise, and decrease in plankton characterising the subtropical gyre upcore suggests that the "Paleo-Slope Water" gradually became established above the margin as a permanent distinct, relatively cold water body, as an expression of the cumulative climatic deterioration through the Pleistocene. A sharp increase in the "neritic" character of the surface waters overlying Site 604 occurred around 1.4 Ma (at around 128 m in the core), assuming constant sedimentation and using

the ages based on calcareous nannofossil data reported by Lang and Wise, 1987). At this point, the terrigenous input increases, indicated by much higher percentages of terrigenous clastics and higher diversity in pollen assemblages; this corresponds roughly to the boundary between lithological units IIB and IIA. The G:P ratio also rises sharply at this time.

By 1.05 Ma, SST values in the mid-latitude North Atlantic had dropped (Ruddiman et al., 1987) and by 0.85 Ma there was an accelerated climatic deterioration documented by Ruddiman and Wright (1987). By 0.44 Ma, the influence of the Gulf Stream above the New Jersey margin had significantly decreased relative to contribution from polar and fluvial sources. This evolution is indicated by a nearly complete disappearance of subtropical ("Sargasso Sea") planktonic microfossils beginning in core 9. At this time, relative to early Pleistocene sediments, there was an overall increase in diversity in the pollen record and a decrease in *Pinus*, suggesting more "neritic" conditions. Increased percentages of the protoperidinoid dinocyst *Brigantedinium* spp., decreased diversity in planktonic foraminiferal assemblages, lower concentrations of planktonic foraminifera, and an increase in percentages of the foraminifer *G. quinqueloba* suggest that a distinct water mass was established above the continental margin with a salinity somewhat less than that of the subtropical gyre.

The contribution of water from the subtropical gyre into the Slope Sea was further suppressed since the mid Wisconsinan; this corresponds to the upper 35 m (roughly, lithological unit IA) at Site 604 (Figure 28C). The suppression of Gulf Stream influence over the margin is indicated by the following observations: 1) lower species diversity (c.f. Balsam and Flessa, 1978) and relatively low percentages of "warm water" foraminifera (including the tropical and gyre margin taxa which are more abundant on the western boundary of the subtropical gyre than the

subtropical taxa) even in Holocene sediments, 2) the low dinocyst:pollen ratio and low G:P ratio and 3) low percentages of *Pinus* and more diversity in the pollen assemblages.

Morley (1982) suggests that isotopic stages 8 and late stage 3-2 were marked by maximum density stratification with coldest shallow subsurface temperature minima and low surface water salinities in the western North Atlantic, based on abundances of the radiolarian *Cycladophora davisiana*. Planktonic microfossils in sediments deposited during late stage 3-2 from Sites 604 and 612 suggest reduced salinity, very cold surface water, probably with floating sea-ice. Concentrations of planktonic foraminifera are very low (generally only around 1/cc), and samples with somewhat higher foraminiferal concentrations are rich in *G. quinqueloba*, which appears to tolerate slightly reduced salinities and cold temperatures (*c.f.* Be and Tolderlund, 1971). Ice-marginal conditions were established above the New Jersey margin at this time, probably with the ice margin at Long Island, indicated by the benthic foraminiferal assemblage, with high percentages of *Cassidulina reniforme* and *Elphidium excavatum* (Scott, unpublished data), which is the "warm" ice margin fauna (Scott and Medioli, 1980); this is consistent with records of ice sheets along the eastern coast of the U.S. by the 300 ka (Richmond and Fullerton, 1986). The dinocyst assemblage in these samples contains relatively high percentages of the protoperidinioid taxa, *Brigantedinium* spp. and *?Multispinula minuta*, which comprise the neritic/upper slope arctic assemblage of Mudie and Short (1985). Pollen concentrations are high, with high diversity, and assemblages are relatively rich in *Picea*. "Warm water" planktonic microfossils are present in relatively low numbers in lithological unit IA at Site 604; even the surface sediments, subtropical, tropical and gyre margin taxa together constitute less than 5 % of the assemblage. Concentrations of planktonic foraminifera are very low throughout unit IA below the surface, again supporting the isolation of the Slope Water mass from the effects of the warmer, more saline Gulf Stream,

and increased input of cold, relatively fresh surface water (from the Inner Labrador Current?). The conclusion reached by Balsam (1981), that  $\text{CaCO}_3$  content appears to be the best indicator of climate change in cores on the U.S. margin, may be due to the fact that the glacial Slope Water mass was very cold and low in salinity, and hence sustained little calcareous plankton; earlier in the Pleistocene, however, glacial sediments had fairly high  $\text{CaCO}_3$  contents (e.g. upper part of core 12, sample 14-4, 140-143 cm, etc.), suggesting more saline surface waters above the continental margin.

The changing oceanographic conditions were associated with a change in the sedimentology as well as the micropaleontology of the Quaternary section at Site 604. Glauconite tends to be associated with unstable water mass configurations, indicated by the "mixed" planktonic microfossil assemblages. The glauconite is thought to be reworked from local outcrops of Tertiary strata (Milliman, 1972); glauconitic Neogene sediments outcrop on the mid-slope, at the edge of the Eocene outcrop belt (Figures 1 and 2). Glauconitic intervals become more frequent upcore in the lower Pleistocene section (common in lithological unit IIA), and glauconite is pervasive through the slumped sediments in lithological unit IB, ca. 0.44- 0.065 Ma; this lithologic transition suggests that erosion of Neogene strata on the mid-slope (at the edge of the Eocene outcrop belt) increased through the early Pleistocene, peaked between 0.44 and ca. 0.2 Ma, and decreased again thereafter. Erosion of glauconite appears to be climatically controlled (Figures 26,29); glauconite is rare to absent in early Pleistocene interglacial sediments as well as in samples recording very cold, low salinity surface water (i.e. samples rich in *Brigantedinium* spp. and in *G. quinqueloba*), throughout the upper 35 m of Site 604, but sporadically abundant below 35 m.

Beginning in the Wisconsinan, the presence of ice sheets calving into the North Atlantic at mid-latitudes, and the establishment of a strong zonal Polar Front (as has been suggested for the late Wisconsinan: CLIMAP, 1981; Keffer *et al.*, 1988), further deflected the Gulf Stream. This deflection expanded the area between the margin and the subtropical gyre, increasing the influence of polar surface water into the Slope Water. These oceanographic conditions suppressed eddy formation (*c.f.* McIntyre *et al.*, 1976), so that little water entered the Slope Sea from the subtropical gyre, indicated by the virtual absence of subtropical foraminifera and the low numbers of "warm water" planktonic foraminifera and dinocysts in the upper 35 m of Site 604. Cold water, and relatively low salinity, marks the record of lithological unit IA.

Glauconite is very rare in the upper 35 m, suggesting that erosion of intercanyon areas on the slope was minimal; this interval is thought to represent the early Wisconsinan-Recent, consistent with the interpretation of Lang and Wise (1987) from the nannofossil record that the upper 90 m at Site 604 were deposited in the last 0.44 Ma (see chapter 5). Glauconite is rare to absent in late Pleistocene sediments recording full-glacial conditions (*i.e.* high percentages of polar foraminifera and protoperidinoïd dinocysts). Those rare samples in lithological unit IA with minor glauconite (<15%) contain microfossil assemblages indicative of interstadial conditions (subpolar-transitional foraminifera, low coiling ratio of *N. pachyderma*). Lower percentages of *Brigantedinium* spp., higher *Operculodinium centrocarpum* and *Spiniferites* spp. characterise the dinocyst assemblages in late Pleistocene sediments with minor glauconite. High percentages of *Quercus* in these samples suggest warm continental climates. These data suggest that the more frequent input of Gulf Stream waters into Slope Sea during warmer intervals resulted in minor erosion of intercanyon areas of the mid-slope.

The sediments transported to the rise are of terrigenous origin, and contain occasional Eocene biosiliceous clasts (Shipboard Scientific Party, 1987a). This suggests that the terrigenous sediments carried to the shelf edge by ice sheets were transported to the rise predominantly through submarine canyons as turbidite flows, and the turbidity currents ripped up clasts of Eocene chalk from the canyon floors as they travelled downslope to the rise.

The sediments above the erosional unconformity at 37 m at Site 612 record deposition since the end of the mid Wisconsinan. The upper 15 m are dated as younger than 11.5 ka; these represent nearly half of the "Pleistocene" record at this site. Most of these sediments were deposited prior to 10 ka, during the Younger Dryas. Sedimentation rates between 11.5 and 10 ka were almost 1 meter per century, or 1 cm/yr. Holocene (stage 1) rates were only 0.2 m/ka, in contrast, and even if there was some loss of surficial sediments during coring, it is highly unlikely that this would account for a significant proportion of the difference in glacial/postglacial sedimentation rates. Sedimentation in intercanyon areas of the mid-continental slope off New Jersey appears to have been rapid during stadials, and particularly during late glacials (deglaciations), but slow during interglacials, again suggesting that the suppression of Gulf Stream influence over the margin as a major factor in allowing sediment accumulation in intercanyon areas of the mid-slope off New Jersey. This is consistent with the cessation of reworking of Neogene glauconitic sediments from the slope onto the rise at this time.

### 6.3 GLOBAL CLIMATE, GLACIATION, AND THE POSITION OF THE GULF STREAM: POSITIVE FEEDBACK?

The path of the Gulf Stream is affected by global climate, but the Gulf Stream itself may in turn influence climate, so that positive feedback mechanisms may exist between circulation of the Gulf Stream and global climate. The data in this study suggest that the progressive shifting of ice centres south-eastward on the North American continent since the late Pliocene (Flint, 1971; Richmond and Fullerton, 1985) is linked with the shifting path of the Gulf Stream (Figure 29).

The Gulf Stream existed along the eastern United States since the Eocene (Pinet *et al.* 1981; Pinet and Popenoe 1985). The northward flow of warm water probably intensified in response to the closing of the Isthmus of Panama in the late Pliocene (*c.f.* Berger *et al.*, 1981). This forcing of warm water into polar latitudes would have resulted in an increase in precipitation at these latitudes, allowing the high rates of precipitation needed to build up Northern Hemisphere ice sheets (Berger, 1981).

There is strong evidence that during the early Pleistocene (Matuyama chron) relatively warm water was able to penetrate into the Arctic Ocean, at least at times (Gilbert and Clark, 1983; Mudie and Short, 1985; Scott *et al.*, 1989a). The relatively high rates of precipitation resulting from this advection of relatively warm water to the pole may have permitted the accumulation of ice sheets in the circum-polar region (Mudie, 1990); Andrews *et al.* (1986) for instance, cite evidence for early Pleistocene and possible late Pliocene ice sheets in the Canadian Arctic (*i.e.* Hudson Bay/James Bay Lowlands, along the coasts of Banks Island, and along the eastern coast

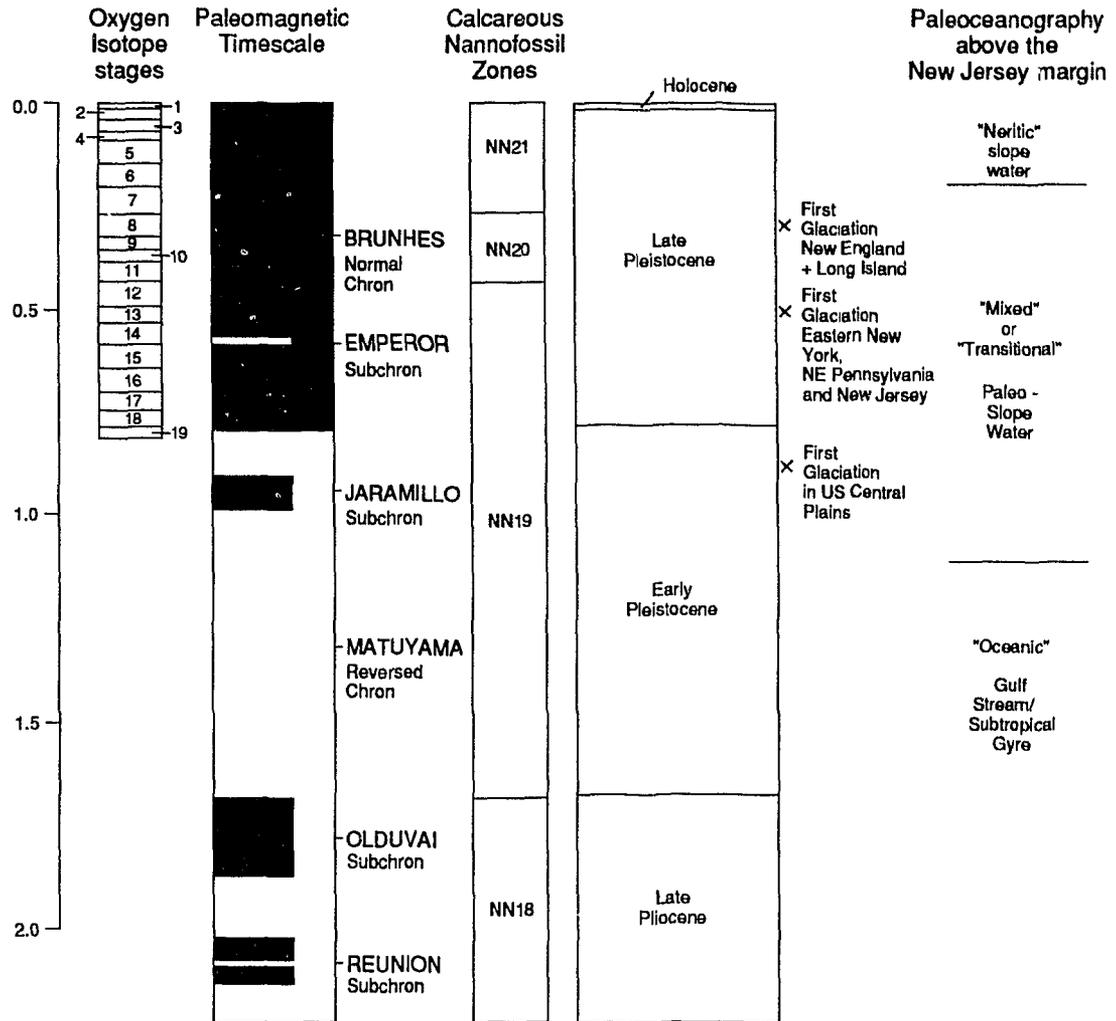


Figure 29. Comparison of paleoceanographic changes in surface water masses above the New Jersey margin, demonstrated by the sedimentary and microfossil record at Sites 604 and 612, with the record of continental glaciation in eastern North America from Richmond and Fullerton (1986).

of Baffin Island). Early Pleistocene ice sheets existed at high altitudes in western North America (including the Sierra Nevada) (Richmond and Fullerton, 1986). However, large ice sheets extending to mid-latitudes did not form in North America until the Brunhes Chron (*e.g.* Ruddiman and Wright, 1987; Ruddiman *et al.*, 1989; Mudie, 1990).

The centres of accumulation of the first large "mid-latitude" ice sheets beginning in the late Pleistocene were in the Keewatin district, with ice sheets extending southward to the Great Plains in the continental interior. Although large ice sheets extended to mid-latitudes in the continental interior during the Nebraskan, they did not reach the east coast of the northern U.S. or the Canadian Maritime Provinces (*c.f.* Flint, 1971; Richmond and Fullerton, 1986); this trend again suggests that the east coast of North America remained quite warm even after much of North America was glaciated early in the late Pleistocene. Willman and Frye (1970) interpreted the glacial history on the basis of ice flow directions in Illinois, and found that during the Nebraskan and Kansan glaciations, ice sheets from the Keewatin center entered Illinois from the west and northwest. Later in the Pleistocene, ice centres had moved east, so that during the Kansan, Illinoian and Woodfordian glaciations, ice from the Labradorian center entered Illinois from the east and northeast, and ice from the Hudson Bay region entered Illinois from a northerly direction during Illinoian, Altonian and Woodfordian times. The Labradorian centre has therefore been occupied for the last ca. 550 ka.

The first good evidence for ice sheets in Long Island and New England is from stage 8, ca. 300 ka (Stone and Borns, 1986) and the first ice sheets in New Jersey, eastern New York and northeastern Pennsylvania occurred in stage 14, ca. 510 ka (Fullerton, 1986). The Maritime Provinces and Maine record no pre-Illinoian glacials on shore, despite the existence in some areas

of topography conducive to stratigraphic preservation, *e.g.* karst, etc. The fact that Tertiary coprolites are preserved in some localities suggests that the lack of Pre-Illinoian glacial sediments is not an artifact of preservation (Stea, pers. comm.). Alam *et al.* (1983) also report Illinoian-Recent sediments offshore in Atlantic Canada. Along the eastern coast of North America at mid-latitudes (*e.g.* Maine and Nova Scotia), an eastward-migration of ice centres is also apparent; ice flow directions due east in Maine and Nova Scotia in the Illinoian suggest an ice centre west of Maine, while early Wisconsinan ice flow directions in Nova Scotia are to the east and southeast, indicating an ice centre in New Brunswick, or perhaps Laurentide ice. Ice centres then migrated to Prince Edward Island, resulting in striations in Nova Scotia from NNE, and then to local ice caps, resulting in northward flow from an ice divide in Nova Scotia (Stea pers. comm., unpublished data; Scott *et al.*, 1987).

The migration of ice centres toward the Atlantic coast in North America through the Pleistocene may be related to the gradual deflection of the Gulf Stream from the coast of the northern United States documented in this study. Paleoceanographic evidence presented in this study suggests that the Gulf Stream continued to flow along the North American margin at least as far north as Long Island through most of the early Pleistocene, until 1.4 Ma. The proximity of the warm Gulf Stream to the coast of the northeastern United States (and, presumably, Atlantic Canada) would have moderated climates, so that even when orbital parameters were conducive to ice sheet growth, ice sheets would have been restricted to high latitudes (*e.g.* Andrews *et al.*, 1986) or at high altitudes in continental areas of western North America (*e.g.* Richmond and Fullerton, 1986) during the early Pleistocene (and latest Pliocene). There is palynological evidence from the Cape May Formation on the Atlantic Coastal Plain that the continental climate in the New Jersey region was significantly warmer than present, *i.e.* warm-temperate (Groot,

1991).

With continued climatic deterioration, large ice sheets began to accumulate "downwind" from the Arctic Ocean and Hudson Bay by 0.85 Ma. Ice centres developed in the Keewatin district and in the Cordillera during glacial intervals when orbital parameters decreased summer insolation. Ice centers migrated through the late Pleistocene from an early center in the Keewatin district southeast to the Labradorean and Hudson Bay centers (Willman and Frye, 1970), reaching the northeastern U.S. by ca. 500 ka (Fullerton, 1986).

There appear to be positive-feedback mechanisms re-enforcing the deflection of the Gulf Stream and the southern migration of continental ice sheets. Shifts in wind patterns accompanying the buildup of polar ice sheets probably initially deflected the Gulf Stream from the margin. This deflection allowed water of polar origin (the Labrador Current?) to flow along the eastern North American margin south to Cape Hatteras by ca. 1.4 Ma. By 1 Ma, pack ice had built up in the Arctic Ocean, restricting the advection of warm water to the poles; ice centres therefore migrated south and east, within reach of warm, moist air masses. Warm North Atlantic water still penetrates into the Labrador Sea/Baffin Bay, even early in glacials, providing a source of moisture for ice sheets in eastern North America (Aksu and Mudie, 1984). The increased deflection of the Gulf Stream also decreased the maritime influence, and increased the continentality of the terrestrial climate of North America. Both the increase in precipitation at mid-latitudes and the decrease in winter temperatures promoted the growth of large ice sheets in the late Pleistocene.

When ice sheets reached Long Island and New England (ca. 300 ka; Stone and Borns, 1986), eddy-mixing mechanisms (which contribute warm Gulf Stream water to the Slope Water)

were suppressed during glacials by the ensuing oceanographic conditions (*i.e.* cold water of reduced salinity with floating icebergs). The modern Slope Water mass formed and the reworking of glauconite from the slope onto the upper rise decreased dramatically.

Ice-marginal conditions are recorded through much of the record of the upper ca. 37 m at Sites 604 and 612 by the following lines of evidence: 1) benthic foraminiferal fauna = *Elphidium excavatum/Cassidulina reniforme* (Scott, 1987, and in prep.), 2) very low concentrations of planktonic foraminifera, dominated by polar taxa: sinistral *Neogloboquadrina pachyderma* and *Globigerina quinqueloba*, 3) *Brigantedinium*-dominated dinocyst assemblage, low G:P ratio), and 4) dominance of *Picea*, together with tundra-indicators, *Alnus* and herb pollen. Neritic conditions are recorded by 1) higher pollen diversity (decreased dominance of bisaccate grains, especially *Pinus*), 2) lower dinocyst:pollen ratios, 3) very low concentrations of planktonic foraminifera, and 4) high percentages of plankton tolerant of cold water of slightly reduced salinity (*e.g.* *Brigantedinium* spp. and *G. quinqueloba*). These factors are consistent with the increased proximity of these sites to the coastline imposed by sea levels over 100 m below present and with the presence of the ice margin in the vicinity of Long Island, a late Wisconsinan-aged terminal moraine.

The existence of a permanent water mass between the U.S. margin and the Gulf Stream since 1.4 Ma may be a function of the growth of northern hemisphere ice sheets increasing in size and extending south in North America. The existence of this water mass in turn helps in maintaining the continent as a site susceptible to the growth of large mid-latitude ice sheets when the orbital parameters are conducive to such growth. The Slope Water mass occupies the space between the margin and the subtropical gyre; the size of the water mass therefore depends on the

size/geometry of the gyre. Crowley (1981) found increased flow in the North Atlantic Current at the approximate time of major polar advances of the last 150 ka, which would permit faster advection of warm water to mid-latitudes where precipitation would fall as snow, thus allowing major ice advances. Nurser and Williams (1990) found that the cooling of the separated boundary current results in 15 Sv strengthening and southward/eastward movement of up to 3000 km. Eventually, the southward/eastward deflection of the Gulf Stream in response to increased flow, reported by Stommel (1965) as well as Nurser and Williams (1990), and the southward migration of the Polar Front (McIntyre *et al.*, 1976; CLIMAP, 1981) would cause the central gyre to become increasingly smaller and faster flowing, more self-contained. Eventually, the North Atlantic Drift would cease to be operational, and all the water in the Gulf Stream System would recirculate south in the central gyre. Moisture starvation resulting from this cessation of advection of warm water to northern latitudes during glacial maxima (Ruddiman and McIntyre, 1981) would result in cessation of ice accumulation, and deglaciation would ensue, particularly rapidly if orbital configurations are such that summer insolation is maximised. In addition, deglaciation may be accelerated because the fast flow of the Gulf Stream causes it to be deflected zonally, forming one central gyre, and ceasing to advect warm water to northern latitudes; at this point, the Polar Front runs almost parallel to 42°N latitude (c.f. McIntyre *et al.*, 1976; CLIMAP, 1981; Keffer *et al.*, 1988).

The record of the Younger Dryas climatic reversal between 16.5 and 2 m at Site 612 (see Chapter 5) suggests that the Gulf Stream was deflected offshore by the vast amounts of meltwater issuing from the Laurentide Ice Sheet. During the late glacial, vast amounts of meltwater entered the Slope Water via the Hudson River System, resulting in very rapid sedimentation on the New Jersey slope adjacent to the Hudson Canyon. This was followed by a return to full glacial

conditions in the Slope Water, indicated by high percentages of polar planktonic microfossils, but the pollen record simply records climatic amelioration from the beginning of deglaciation to the Holocene. This suggests that the vegetation around 38°N was mainly responding to the orbital changes resulting in higher amounts of summer insolation; the existence of "polar" water in the Slope Sea affected vegetational succession in southern New England, as documented for New Jersey by Peteet *et al.* (1990). Further north (*e.g.* in Atlantic Canada; Stea, pers. comm.), ice sheets did readvance and vegetation returned to glacial assemblages between 11 and 10 ka B.P. At these latitudes (around 45°N), the deflection of the Gulf Stream would have been enough to allow expression of the Younger Dryas.

#### 6.4 SUMMARY

Climatic change has affected the position of the Polar Front and ocean currents through the Quaternary, affecting the characteristics (temperature, salinity, nutrient availability, etc.) of the surface waters overlying the sites in this study. Climatic deterioration deflects the Gulf Stream offshore and causes the Labrador Current to flow closer to the North American continent. In the earliest Pleistocene (before *ca.* 1.4 Ma, *i.e.* below 128 m, most of lithological unit IIB at Site 604), the Gulf Stream flowed along the continental margin during interglacials, depositing sediments rich in biogenic CaCO<sub>3</sub>, with relatively high percentages of "warm water taxa" esp. subtropical taxa on the upper rise, even in samples deposited during glacials. Later in the early Pleistocene, sediments rich in CaCO<sub>3</sub> and "warm water taxa" were less frequent, separated by intervals rich in glauconite and terrigenous sediments, and occasional biogenic silica. The increase in clastics, higher pollen diversity, as well as an increase in planktonic microfossils which presently characterise the relatively fresh and cold Slope Water suggests that the Gulf Stream was deflected

from the margin, allowing a cold water mass between the subtropical gyre and the margin into which water and sediments from rivers could settle. Extensive erosion of the slope through most of the Pleistocene is suggested by the long hiatus below mid Wisconsinan sediments at Site 612 on the mid-slope, and extensive reworking of glauconite onto the rise between ca. 1.4-0.2 Ma. The hiatus between 1.1 and 0.44 Ma identified in the calcareous nannofossil data (Lang and Wise, 1987) associated with a strong seismic reflector show that the rise has also been a site of net erosion at times during the Quaternary.

The overall increase in terrigenous sediments, and accompanying decrease in biogenic  $\text{CaCO}_3$  upcore, suggests progressive deflection of the Gulf Stream. The microfossil record indicates a trend toward increasingly cold and fresh surface waters, more "neritic" conditions, and decreased input from the Gulf Stream; there is an increase in *Brigantedinium* spp., *G. quinqueloba*, *Picea*, and a decrease in "warm water" planktonic microfossil taxa and *Pinus*, which characterise the subtropical gyre. The overall trend toward more neritic, "fresh" cold water was accelerated at the beginning of the late Pleistocene and was again accelerated in the mid Wisconsinan (lithological unit IA at Site 604), when benthic foraminifera record high percentages of *Elphidium excavatum* and *Cassidulina reniforme* (Scott, in prep), suggesting ice-marginal conditions (Scott and Medioli, 1980). Glauconite is rare in this unit, especially in samples recording full-glacial conditions. The suppression of slope erosion during glacials is also supported by the thick units of late Wisconsinan sediments at Site 612 on the mid-slope.

The progressive deflection of the Gulf Stream from the coast of eastern North America at mid-latitudes may have controlled the location of ice centres through the Pleistocene. This study shows that the Gulf Stream was progressively deflected from the eastern U.S. margin

through the Quaternary, and that there were two large "steps" in this deflection: 1) late in the early Pleistocene (ca. 1.4 Ma), when the Paleo-Slope Water began to form, separating the western boundary of the subtropical gyre from the U.S. margin, and 2) ca. 0.2 Ma, when the incursion of water from the subtropical gyre *via* meanders, rings and filaments, was suppressed by oceanographic conditions accompanying the presence of ice sheets along the coast at mid-latitudes. While the Gulf Stream flowed along the eastern U.S. margin at least as far north as Long Island, ice sheets were prevented from growing into mid-latitudes; ice sheets were restricted to high latitudes/altitudes. The progressive deflection of the Gulf Stream from the margin (and formation of the Paleo-Slope Water over the margin), which began ca. 1.4 Ma permitted the growth of large ice sheets into mid-latitudes by the Brunhes Chron. By 0.2 Ma, the Gulf Stream was further deflected (and the modern Slope Water formed over the margin), and ice sheets reached the eastern coast of North America at mid-latitudes.

The unidirectional trend suggests that there is positive feedback between the path of the Gulf Stream and global climate. Northward transport of warm water to polar latitudes allowed the initial ice sheets to form in the Arctic and in the Cordillera, but the warmth of the current flowing along the eastern margin prevented the growth of ice sheets to mid-latitudes. The polar ice sheets may have altered wind patterns sufficiently to allow water of polar origin to penetrate between the subtropical gyre and the margin during glacials by ca. 1.4 Ma, forming the Paleo-Slope Water. The deflection of the Gulf Stream increased the continentality of the North American climate, so that large northern hemisphere ice sheets were able to form by the Brunhes Chron. These ice sheets further deflected the Stream, so that by 0.2 Ma, the influence of water from the subtropical gyre into the Paleo-Slope Water was suppressed, and the modern Slope Water formed as ice sheets reached the coast of North America at mid-latitudes.

Disintegration of the North American ice sheet resulted in large amounts of meltwater reaching the Slope Water via the Hudson River System during the terminal Pleistocene. This was followed by a return to full glacial conditions in the Slope Water, suggesting that the Gulf Stream was deflected offshore by the meltwater. The penetration of water of polar origin (Inner Labrador Current) between the Gulf Stream and the coastline did not result in a return to glacial vegetation in New Jersey; further north, however, as in Atlantic Canada, the deflection of the Gulf Stream could have been sufficient to allow ice sheets to readvance, and for vegetational succession to be interrupted during the Younger Dryas.

## CHAPTER VII

### EFFECT OF THE GULF STREAM AND ASSOCIATED INSTABILITIES ON THE U.S. CONTINENTAL MARGIN NORTH OF CAPE HATTERAS THROUGH THE QUATERNARY

#### 7.0 EROSION OF THE CONTINENTAL MARGIN

The deep ocean was long regarded as a quiet, low energy environment. Deep sea drilling programs (DSDP, ODP, etc.), however, have revealed "unexpected abundances of hiatuses in a setting which had been thought to be the ultimate sedimentary sink" (Hay, 1988, p. 1934). Abyssal storms, interpreted to be the result of downward propagation of eddy energy from the upper ocean, were reported from the Nova Scotia rise by the HEBBLE project (*e.g.* Hollister and McCave, 1984; Weatherley and Kelley, 1985), and Culver *et al.* (1988) observed a burst of current activity strong enough to transport surficial sediment and to construct and modify bedforms in a region of slow average current at 2300 m in South Wilmington Canyon.

One of the most striking features of the geology of the United States continental margin north of Cape Hatteras is the outcrop of middle Eocene chalks, marls and shales on the lower continental slope off New Jersey (Figure 5), separating accretionary prisms on the upper slope/shelfbreak and on the rise (Figure 4). The Pleistocene section measures more than 400 m at the shelfbreak, but on the mid-slope, Pleistocene sediments exist only as fingerlike ridges trending downslope, separated by outcrops of Eocene to Miocene sediments in canyon or valley bottoms (Robb *et al.*, 1981; Hampson and Robb, 1984). On the upper rise, Pleistocene sediments

form a pronounced seaward-thickening wedge described as a complex association of slump deposits, turbidites, hemipelagic clays, and channel fill (Hampson and Robb, 1984). Although this is not illustrated in Figure 5, large patches of the outcrop belt are known from borings and surface samples to be covered by a thin, acoustically invisible, layer of silty Pleistocene ooze (Poag, 1985b). In relatively small areas on the slope, areas of both positive (ca. 10-40 cm/1000 years estimated by Emery and Uchupi, 1972) and negative rates of Holocene sediment accumulation are found (Tucholke, 1987).

Some workers maintain that erosional processes continue to be important in shaping the margin today (*e.g.* Stanley *et al.*, 1984). Others (*e.g.* Doyle *et al.*, 1979) suggest that the physiography of the margin is relict. The seaward-thinning geometry of late Tertiary strata has been interpreted as suggesting that mid-Eocene strata have been exposed on the lower slope since the end of the Miocene (Pricr *et al.*, 1984) or even the Oligocene (Farre and Ryan, 1987). Farre and Ryan (1987) suggest that erosion was initiated during the Oligocene by along-slope boundary currents which filled the western Atlantic basin sometime near the Eocene-Oligocene boundary, or downslope-directed currents associated with the mid- Oligocene low sea level stand, or a combination of the two processes. Tucholke (1987) concludes that the present outcrop of mid-Eocene strata on the lower slope is a "direct legacy" of this Oligocene erosion.

There is controversy over the processes responsible for the geology and physiography of the margin; evidence has been presented to support both gravity-driven mass wasting processes and bottom currents as the dominant process. Poag (1985c) and Mountain and Tucholke (1985), for instance reached different conclusions regarding the evolution of the U.S. Atlantic continental margin in the same volume; Poag (1985c) suggested that downslope processes dominated, while

Mountain and Tucholke (1985) concluded that along-slope processes dominated from the end of the Eocene until the inception of northern hemisphere glaciation in the late Pliocene, when downslope processes became more important.

Because of the relatively steep gradients, gravity-driven mass transport processes are more important on continental slopes than anywhere else in the marine realm (Nardin *et al.*, 1979). The relative importance of gravity-driven mass transport processes in this region is a subject of controversy, however. Some workers (*e.g.* Knebel and Carson, 1979; Prior *et al.*, 1984; Stanley *et al.*, 1984) have cited geological evidence which supports mass wastage as an important process on the eastern U.S. margin, and they maintain that mass wasting is ubiquitous on the continental margin between Cape Hatteras and Georges Bank. Recent work has questioned the importance of mass wastage in shaping the margin, however. Pilkey and Cleary (1986) contend that mass wastage is presently minimal in intercanion areas of the continental margin.

The linear geometry of the Tertiary outcrop on the New Jersey margin suggests that along-slope processes were important in shaping the eastern U.S. margin. Mountain and Tucholke (1985, page 332) stated that: "Despite the presumed downslope transport on the upper continental rise off New Jersey, lobate fanlike depocenters are not found there. Instead there are slope-parallel trends that imply continuing current-controlled deposition and redistribution of sediments.... Top-quality seismic profiles (*e.g.* line 25) show evenly structured seismic reflectors above reflector Blue, suggesting that only nonchannelized turbidity currents and bottom currents were significant in affecting depositional patterns there."

At present, however, there is no strong current at the base of the slope north of Cape

Hatteras; the mean path of the Gulf Stream and the relatively high abyssal kinetic energy associated it are deflected from the margin at Cape Hatteras, so that off New Jersey, the Gulf Stream is several hundred kilometers seaward of the Eocene outcrop at the base of the slope. The Gulf Stream appears to be an unlikely candidate to explain the erosion at the base of the slope off New Jersey. The net south to southwest flow in the Slope Water appears to have little discernible geological effect on the continental slope aside from the transport and eventual deposition of suspended particulate matter (Tucholke, 1987).

#### **7.1 EROSIONAL PROCESSES ON THE U.S. MARGIN NORTH OF CAPE HATTERAS: EVIDENCE FROM DSDP SITES 604 AND 612**

Examination of the oceanographic conditions conducive to erosion of the slope could provide insight into the processes responsible for erosion of the margin. The depositional history of the continental margin, and the oceanographic conditions which promote slope erosion can be deduced by examining the microfossil and sedimentological data from Site 612, on the mid-slope, and Site 604, on the upper rise. Sediments on the continental rise originate either "*in situ*" (i.e. the biogenic component), on the continents (transported to the rise primarily by rivers, with ice important at times during the Pleistocene), or on the continental slope (transported to the rise by mass wasting).

The absence of early-mid Quaternary sediments at Site 612 (Chapter 5) suggests that erosion of the mid-slope off New Jersey was not restricted to submarine canyons prior to the late Wisconsinan. Pre-Quaternary microfossils, as well as distinctive sediments reworked from older strata which outcrop on the slope, in Quaternary sediments at Site 604 on the upper rise are also

indicators of slope erosion. Because Eocene chalk and Neogene glauconite outcrop on different parts of the slope, the age of the reworked sediments can also provide insight into mechanisms of erosion; Eocene chalk outcrops on the lower slope and at the bases of deep submarine canyons further upslope, while Neogene glauconite outcrops on the mid-slope. Environmental conditions (e.g. sea surface temperature, salinity) not conducive to erosion can be determined by examining samples in the stratigraphic record of the rise with little or no evidence of reworking; reworked microfossils which are still extant can distort the paleoenvironmental signal, so that samples in which reworking is obvious can not be interpreted paleoclimatically with as much confidence.

Reworking of Neogene sediments onto the rise (indicated by maximum amounts of glauconite and reworked microfossils) was greatest during deposition of sediments between 125 m and 65 m at Site 604 (Figure 8). The seismic data suggesting an unconformity (Poag and Mountain, 1987; Wise and Van Hinte, 1987) and the identification of a hiatus spanning approximately 1.1 to 0.44 Ma in the nannofossil record (Lang and Wise, 1987) within this interval indicates that the continental margin was the site of extensive erosion for a large interval during the mid-Pleistocene, and that the upper rise received large amounts of Neogene sediments eroded from the slope during the late early Pleistocene and early late Pleistocene.

The inception of intensive reworking of Neogene sediments (ca. 1.4 Ma) coincides with a decrease in  $\text{CaCO}_3$  in the coarse fraction, generally lower P:B ratios, an overall decrease in the G:P ratio, a marked increase in percentages of *Brigantedinium* spp., and a marked increase in pollen diversity (i.e. a decrease in the dominance of pine pollen) (Figure 26). These data suggest that a more "neritic" paleoenvironment developed over the rise than had existed previously, suggesting that a water mass was established separating the western boundary of the subtropical

gyre (*i.e.* the Gulf Stream) from the margin (see discussion of Paleo-Slope water, chapter 6).

The reduction in erosion of Neogene sediments from the slope recorded in the upper 35 m of Site 604 (since ca. 0.2 Ma) is characterised by a marked decrease in the dinocyst:pollen ratio, increase in *Brigantedinium* spp. and a decrease in the G:P ratio, a marked increase in *Globigerina quinqueloba*, and an increase in pollen diversity (Figure 26), especially in pollen of boreal-tundra taxa *e.g.* *Picea*, *Alnus*, and herbs (Figure 14). Although glauconite and Neogene microfossils are rare to absent in sediments recording glacial conditions in lithological unit IA, clasts of middle Eocene chalk are associated with paleoclimatological data indicating full glacial conditions, *i.e.* high percentages of sinistral *N. pachyderma* and *G. quinqueloba*, protoperidinioid dinocysts, dominantly *Brigantedinium* spp., high percentages of *Picea*, *Alnus* and herb pollen, as well as a benthic foraminiferal fauna indicative of ice-marginal conditions.

At Site 612 on the mid-slope, sediments of late Wisconsinan to Recent age are preserved, separated from sediments recording considerably warmer surface waters (Pliocene? early Pleistocene?) by an erosional unconformity. The preservation of a thick unit of mid to late Wisconsinan sediments below a thin veneer of Holocene sediments supports the suppression of erosion of intercanyon areas of the mid-slope during glacials. The very low sedimentation rates calculated for interglacial sediments suggest that the mid-slope may be erosional at times during late Quaternary interglacials/interstadials, possibly explaining the absence of pre-mid Wisconsinan sediments at Site 612.

It appears, therefore, that climatic/oceanographic conditions controlled slope erosion, or at least that slope erosion responds to the same forcing as the climate. Glauconitic Neogene

sediments which presently outcrop on the mid-slope, along the edge of the middle Eocene outcrop belt, began to be redeposited onto the upper rise in large quantities when the Gulf Stream was deflected from the continental margin sometime during the early Pleistocene. This deflection of the Gulf Stream allowed the establishment of a "Paleo-Slope Water" mass above the margin, and erosion of intercanyon areas of the mid-slope was suppressed when oceanographic conditions in the "Glacial Slope Water" suppressed eddy formation.

## **7.2 EFFECT OF THE GULF STREAM ON THE LOWER CONTINENTAL MARGIN OFF NEW JERSEY: ABYSSAL KINETIC ENERGY ASSOCIATED WITH INSTABILITIES OF THE STREAM**

Because the Gulf Stream is several hundred kilometers seaward of the Eocene outcrop off New Jersey, high abyssal kinetic energy associated with the Gulf Stream therefore appears to be an unlikely candidate to explain the erosion at the base of the slope. The microfossil data, which suggest a strong correlation between the path of the Gulf Stream and erosion of the margin must be reconciled with modern oceanographic observations of the Stream.

Stanley *et al.* (1981) found that periodically intensified slope currents on the mid-slope to uppermost rise off Cape Hatteras result in winnowing and resuspension, and the sand-sized fraction between 1000 and 2500 m is dominantly terrigenous, unlike the upper slope and rise which is dominated by tests of planktonic foraminifera. Csanady *et al.* (1988) showed from current meter observations that rare bursts (less than 1% of the time) of near-bottom currents faster than 0.2 m/s between depths of 500 and 2000 m off New Jersey are invariably associated with topographic waves and superimposed inertial oscillations. These topographic waves result in

"eddy transport processes of geological scientific significance" (Robinson, 1983), that is they are capable of eroding and transporting sediments from the slope.

Topographic Rossby waves are generated when mesoscale eddies, which develop from the Gulf Stream into the Slope Water, encounter the continental margin (Smith, 1983). The term "eddies" refers to all types of variable flow, including the meandering of intense current systems, ring currents, topographic waves and wakes, etc. This variability contains more energy than any other kind of motion in the ocean. The Gulf Stream meanders with a broad spectrum of variability, particularly downstream of Cape Hatteras (Watts, 1983). This type of high abyssal kinetic energy associated with variability is not restricted to the Gulf Stream System. In the East Australian Current system, Mulhearn *et al.*, (1986) found strong abyssal currents (over 10 cm/s) moving on the slope in the same direction as the corresponding surface currents during the passage of a warm core ring.

As the topographic wave energy propagates shoreward, the decreasing depth and amplification due to other geometric effects such as rough bathymetry (Rubenstein, 1988) leads to strong dissipation in the shallow regions, limiting the penetration of topographic wave energy in the upslope direction (Smith, 1983). Rhines (1971) found that for realistic topography, most of the incident topographic wave energy would be reflected from the steep continental slope. Similarly, Schmitz (1974) found that deep kinetic energy achieves a maximum near the slope-rise junction, ca. 2000 m isobath, and Shaw and Csanady (1988) report that the onshore energy flux associated with topographic waves is deflected by the continental slope, and wave energy propagates along isobaths on the lower slope and upper rise. Thus, the energy associated with these topographic waves is concentrated at depths where Eocene strata presently outcrop. The

intense intermittent currents on the slope would resuspend sediments (as they do at the HEBBLE site), and may be another mechanism to initiate slumps, in addition to turbidity currents (Malahoff *et al.*, 1980), biologic erosion (Rowe and Haedrich, 1979), groundwater flow (Robb, 1984), and seismic activity and inherent sediment instability (MacIlvaine and Ross, 1979), which have previously been suggested to initiate slumps. Despite the rarity of eddy-induced current activity on the margin, it is conceivable that rare bursts of strong abyssal kinetic energy would prevent deposition, so that over even a relatively short span of geologic time, the lower slope along the entire margin would be maintained as a site of net erosion.

Rings are the most energetic eddies in the ocean and their thermocline displacements, swirl speed and volume transports are nearly equivalent to those of the Gulf Stream (Richardson, 1983). Warm core rings are bowl-shaped bodies of water 100-200 km in diameter and about 1 km deep. They form in the Slope Water, a triangular region bounded on the south by the Gulf Stream and on the north by the continental slope (Saunders, 1971). Relatively small rings (ca 100 km diameter) generally form west of Georges Bank; their formation resembles an aneurism during which the side of the Stream bulges out to the north and pinches off the main current. Typically, five rings form per year, and approximately three rings exist at a single time (Richardson 1983). Warm core rings are anticyclonic (rotate clockwise) and drift westward across the Slope Water toward the continental margin with speeds of 3-5km/day (Lai and Richardson, 1977).

The geographic distribution of instabilities (eddies) generating these topographic waves is also consistent with the geographic extent of high current activity on the lower slope, with Eocene strata exposed on portion of the margin mapped by Hampson and Robb (1984), and intermittent bursts of high energy cited off Wilmington Canyon by Culver *et al.* (1988) and off

Cape Hatteras by Stanley *et al.* (1981). Warm-core rings tend to follow constant depth contours (Halliwell and Mooers, 1979), and Brown *et al.* (1986) found that the mean long-lived warm core ring track follows the 4250 m isobath east of 67°W, then crosses to the 2750 m isobath at 69°W, southwest of the New England Seamounts, about which it oscillates to 73°W, and finally crosses to the 2250 m isobath for the remainder of its track; this is illustrated in Figure 30, redrafted from Brown *et al.* (1988), which shows the probability of the occurrence of warm core rings off the eastern coast of North America. DSDP Sites 612 and 604 are located near 73°W, where the mean long-lived ring track crosses over to the 2250 m isobath. Brown *et al.* (1986) also show the space-time location probability density function for their set of long-lived warm core rings; high values extend along the continental margin north of Cape Hatteras to off of Cape Cod, with maximum probability off New Jersey. The horizontal distribution of kinetic energy at 700m water depth calculated from SOFAR float data is higher on the New Jersey margin than anywhere else on the U.S. Atlantic coast (Owens, 1991).

Although Popenoe (1985) showed that the position of the Gulf Stream is climatically controlled, and that this is the major control on sedimentation on the North Carolina coast, the influence of the Gulf Stream on the margin north of Cape Hatteras is not obvious, since the north wall of the Stream is several hundred kilometres from the margin off New Jersey, and flows in water *ca.* 5000 m deep. The influence of the Gulf Stream and its associated rings on the sea bed in the Slope Sea, previously tentatively suggested by several authors (e.g. Hill and Bowen, 1983; Weatherley and Keiley, 1985; Mountain and Tucholke, 1985), is also supported by the results of this study.

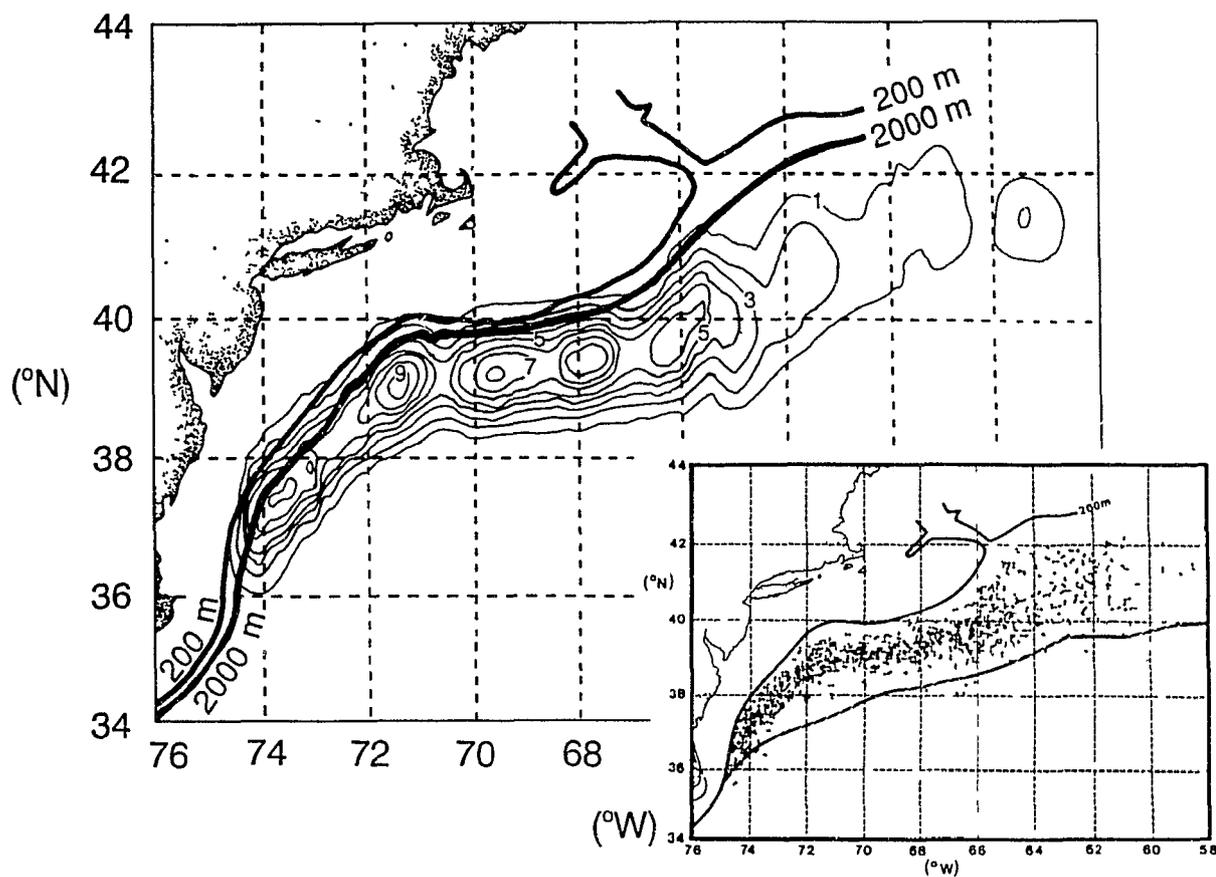


Figure 30. Distribution of warm core rings in the Slope Water; the space-time location probability as calculated by Brown *et al.* (1986) for an 87 ring set spanning 1974-1983. The inset shows the point-location cloud for the time series analysis of Brown *et al.* (1986). The contour interval is  $5.6 \cdot 10^{-6} \text{ km}^2 \text{ day}^{-1}$ . This figure was redrafted from their Figure 5a). Note that the location of New Jersey Slope is shoreward of a "Bull's-eye" indicating high probability of finding warm core rings.

The absence of a water mass separating the subtropical gyre from the continental margin in the early part of the early Pleistocene indicated by microfossil data (Chapter 6, Figure 28A) would have prevented large scale instabilities from developing in the Gulf Stream; the generation of instabilities in the north wall of the Stream requires the existence of a distinct water mass between the Stream and the margin; large-scale instabilities do not form today where the Stream flows along the Florida-Hatteras slope. Later in the early Pleistocene (ca. 1.4 Ma) the Gulf Stream was deflected from the margin, and polar-subpolar surface waters penetrated between the margin and the subtropical gyre and a distinct water mass was established above the continental margin. This "Palco-Slope Water" mass (Figure 28B) received considerable influence from the Gulf Stream via instabilities (meanders and rings), especially during interglacials (Chapter 6). Topographic Rossby waves generated by the instabilities resulted in bursts of strong bottom currents, which were bathymetrically concentrated over the mid- to lower slope; the generation of these topographic waves would have been most frequent off New Jersey, where warm core rings most frequently occur today. Erosion of intercanyon areas of the margin was intensified with accelerated climatic deterioration beginning in the late Pleistocene, marked by slumping resulting in contorted and convoluted glauconite-rich sediments in lithological unit IB at Site 604. Emergence of the shelf at Cape Hatteras during glacioeustatic lowstands could have generated instabilities downstream, especially as the north wall of the Gulf Stream interacted with subpolar waters along the coast, by analogy with the present generation of variability in the Gulf Stream as it interacts with the New England seamounts between 60° and 65°W (*e.g.* Knauss, 1967 by permission in Watts, 1983). Topographic Rossby waves generated by these instabilities resulted in landward erosion and undercutting of the slope, and downslope mass wasting.

The extensive reworking of sediments from the slope onto the rise was responsible for the

extensive Pleistocene overlapping of the Eocene exposure noted by Poag and Mountain (1987), and for the lack of early-mid Quaternary sediments noted by Scott (1987) and in this study (although sediments below the "Plio-Pleistocene boundary" at Site 612 may be early Pleistocene; chapter 5). Extensive deposition occurred on the lowermost slope during the Pleistocene, while there appears to have been increased erosion of the upper slope (Poag and Mountain, 1987).

As ice sheets reached Long Island later in the Pleistocene, the Slope Water mass developed ca. 0.2 Ma (see Chapter 6), the influence of the Gulf Stream on the surface water mass overlying the margin decreased in response to gradual climatic deterioration deflecting the Gulf Stream (Nurser and Williams, 1990). During glacials the Slope Water ("Glacial Slope Sea") was completely isolated from the Gulf Stream; microfossil assemblages indicate that ice-marginal conditions existed in the Slope Sea and that there was increased influx of the Labrador current. Salinities were quite low, indicated by the sparse planktonic foraminiferal assemblages and the dominance of *Brigantedinium* spp. in the dinocyst record. Percentages of the polar planktonic foraminifer *Globigerina quinqueloba* were relatively high in the Glacial Slope Sea; this taxon is common in coastal waters today, with maximum percentages in the Slope Water (Be and Tolderlund, 1971).

These oceanographic conditions accompanying large, mid-latitude ice sheets near the eastern coast of North America, suppressed eddy formation; McIntyre *et al.* (1976) suggested that at 18000 years B.P. (the last glacial maximum), since 1) gradational boundaries were nearly absent between assemblage areas and 2) the structure of the North Atlantic was materially altered by the appearance of a strong polar frontal system running nearly parallel to latitude, this type of wide-scale mixing through eddy formation would not have existed, or would have been severely

suppressed. This is supported by the observations that at present, there appears to be some seasonal control on the mean path of the Gulf Stream (*e.g.* Iselin, 1941; Stommel, 1965; Fugilister, 1955), and on the formation of warm core rings and other instabilities in the Stream. The publications *Gulf Stream* and *Oceanographic Monthly Summary* record the paths of the Gulf Stream and its associated warm and cold core rings for each month since 1975. It is clear from these data that the effect of the Gulf Stream and its associated warm core rings is generally greatest in the Slope water above the U.S. margin in summer and least in winter, and intermediate in spring and fall.

Erosion of intercanyon areas of the slope was suppressed during the late Quaternary, allowing sediments to accumulate rapidly at Site 612 during stadials, but accumulate much more slowly or even be eroded during interglacials/interstadials; only mid Wisconsinan-Holocene sediments were identified in this study and in that of Scott (1987). Reworked Neogene sediments are rare in late Pleistocene unit IA at Site 604 on the upper rise, also suggesting that erosion of intercanyon areas of the slope was suppressed. This pattern of erosion and sedimentation supports topographic waves, generated by eddies of the Gulf Stream, as erosional agents responsible for the present geology of the margin, including the outcrop of Eocene strata on the lower slope north of Cape Hatteras.

At Site 612, excellent recovery and very high sedimentation rates allowed the late Wisconsinan (including a glacial maximum punctuated by a short interval of "interstadial" conditions as well as the Allerod/Younger Dryas climatic oscillation) to Holocene to be resolved. Poor recovery at Site 604 precludes detailed climatostratigraphic interpretations, but the data would be consistent with an assignment of core 5 to isotope stage 4; see chapter 5 for a discussion

of the climatostratigraphy of this core.

Sedimentological evidence suggests that terrigenous sediments which reached the exposed shelfbreak via proglacial streams, were carried to the rise mainly as turbidite flows, depositing layers of silt and clay on the rise. Numerous submarine canyons were incised into the slope; the presence of clasts of Eocene chalk (presently exposed on the lower slope and at the bases of canyons) in latest Pleistocene sediments, support turbidity currents as the erosional mechanism responsible for erosion of Eocene sediments from the lower slope. Thus the suppression of erosion of Neogene sediments from intercanyon areas of the mid-slope when microfossil data indicate the existence of conditions in the Slope Water which would suppress eddy mixing mechanisms, such as warm core ring formation, also suggest that erosion of the margin is linked with instabilities of the Gulf Stream.

The presence of middle Eocene chalk at the bases of submarine canyons in association with rapid deposition in intercanyon areas of the slope (e.g. Site 612) suggests that submarine canyons were cut, or at least extensively occupied, during glacio-eustatic sea-level lowstands, when large amounts of terrigenous sediments were transported to the continental margin by rivers flowing to the coast, whose position was approximately at the modern shelfbreak.

### 7.3 ASSUMPTIONS OF THE VAIL DEPOSITIONAL MODEL: COMPARISON WITH THE RECORD OF THE NEW JERSEY MARGIN

Many authors have cited the similar timing of erosion on continental margins with different tectonic and sedimentologic histories as suggesting a global cause (Poag 1985b,c; Popenoe, 1985, etc). The similarity between the cyclical pattern of regional unconformities on the Atlantic continental margin of the United States and the eustatic sea level curve has been interpreted as supporting the model of Vail *et al.* (1977); Poag (1987) writes that unconformities on the continental margin off New Jersey have equivalents on the shelf and can be traced widely on the seismic profiles. Poag and Low (1987, page 494) conclude that "The growing body of field data... demonstrate a clear link during most of the Cenozoic between widespread shelf and slope (and even abyssal) erosion, increased global ice volumes, cooler global climate and lowered sea levels." Relatively few workers (*e.g.* Watts and Thorne, 1984) question the importance of glacioeustatic sea level changes, proposed by Vail and his colleagues (Vail *et al.*, 1977) as the mechanism controlling patterns of erosion and deposition on continental margins, in shaping the U.S. Atlantic margin. The direct application of the model of Vail *et al.* (1977) to Quaternary margins is especially suspect, given the geometry of Quaternary margins and the high-frequency, high-amplitude changes in sea-level characterising this period.

The assumptions of the model as expressed by Vail *et al.* (1977) led Poag and Low (1987) to suggest that "the middle Eocene outcrop belt, 3 km downdip from Site 612, temporarily shifted updip during glacioeustatic sea level falls." This study has demonstrated, however, that major erosional intervals on the margin in the latter part of the Quaternary occurred during sea-level highstands, not lowstands; intercanyon areas of the New Jersey slope (*e.g.* Site 612) were sites of

rapid deposition during late Pleistocene glacioeustatic lowstands when the shelves were erosional. The middle Eocene outcrop belt did not shift updip during glacioeustatic sea level falls, but rather became largely depositional.

The modifications to the depositional models outlined in Posamentier and Vail (1988) and Posamentier *et al.* (1988) explain the stratigraphy at Site 612 better; the model suggests that slope progradation ceases during glacioeustatic lowstands, and that lowstand fans and wedges develop at the base of the slope. Although this model would explain the rapid rates of sedimentation at Site 612 during the late Wisconsinan (stage 2), it does not provide a good explanation for the unconformity beneath mid Wisconsinan (stage 3) sediments at Site 612. Nor does the model predict the very low rates of Holocene sedimentation at Site 612 and the fact that many parts of the slope are currently areas of net erosion. Non-deposition on the slope would not account for the reworking of Tertiary glauconite during interglacials, although the depositional processes outlined in Posamentier *et al.* (1988) would help to explain accelerated slope erosion and downslope mass wasting during interglacial/glacial and glacial interglacial transitions.

The Vail model also implies that mechanisms of continental margin erosion have remained the same through the Phanerozoic, being related to changes in the "accommodation space" for sediments on the margin (Posamentier *et al.*, 1988), related to glacioeustatic changes in sea level as well as local conditions of uplift, depression, etc. Several studies have demonstrated, however, that the mechanisms and rates of sea level change differed through the Cenozoic, with glacioeustatic changes restricted to the last 36 Ma, and that the different mechanisms and rates of eustatic change may have caused contrasting erosional patterns on passive continental margins (Miller *et al.*, 1987; Katz and Miller, 1987; Mountain and Tucholke, 1985). Similarly, this study

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has shown that erosion of the New Jersey slope is climatically controlled, but appears to depend less on glacioeustatic changes in sea level than on the oceanographic changes in surface waters overlying the margin accompanying glaciation. The sediment record of the rise suggests that the mechanisms and patterns of erosion of the New Jersey margin differ between the early and late Pleistocene. The two lithological units and 4 subunits recognised in the Quaternary record of Site 604 (Figures 8,9) indicate very different sediment sources, and depositional mechanisms. Autochthonous biogenic carbonate dominates in unit IIB, while unit IIA is characterised by alternations between biogenic carbonate, reworked Neogene glauconite, and terrestrially-derived clastics. Contorted beds, interpreted as slump deposits characterise parts of Unit IB, which contains relatively low percentages of biogenic carbonate, and alternates between domination by glauconite (reworked from the slope) and clastics (eroded from the continent). The early and late Pleistocene units are also acoustically different, as seen in the seismic records from the margin; Pleistocene unit A comprises the upper 102 m (cores 1-11) at Site 604 and unit B 102-155 m (cores 12-17). Deposition of unit B is geographically quite homogeneous, while deposition of unit A is far more variable; Poag and Mountain (1987) conclude that thickness variations of the total Pleistocene sequence are due to differential erosion and deposition of Unit A and the filling of downslope channels by Unit B.

#### 7.4 SUMMARY

The oceanographic evidence for strong currents at the base of the slope associated with topographic Rossby waves generated by instabilities of the Gulf Stream, at the depths and geographic range over which Eocene strata outcrop on the U.S. Atlantic margin suggest that warm

core rings and other instabilities of the Stream are the main agents of erosion operating on the lower continental slope. The geological evidence indicates that these instabilities first began to interact with the margin ca. 1.4 Ma B.P. when polar-subpolar water, probably from the Labrador Current, flowed south between the Gulf Stream and the margin; prior to 1.4 Ma, sedimentological and microfossil evidence indicate that the Gulf Stream flowed directly over the margin off New Jersey as it presently does south of Cape Hatteras, so that no large-scale instabilities developed, and hence erosion of the mid-slope was minimal, and hemipelagic muds relatively rich in planktonic foraminifera were deposited on the upper rise. As progressive climatic deterioration deflected the Gulf Stream seaward through the Pleistocene, allowing surface water of polar origin (?the Labrador Current) to flow south along the shoreline to Cape Hatteras, a distinct water mass developed between the subtropical gyre and the continental margin; the Gulf Stream exerted extensive influence on this water mass via instabilities generated along the north wall of the Stream as it interacted with this "Paleo-Slope Water". Topographic Rossby waves associated with these instabilities eroded the mid-slope off New Jersey, the area which is presently geographically and bathymetrically most affected by warm core rings. This erosion of intercanyon areas of the mid-slope is preserved in the geologic record of the upper rise (Site 604) as beds rich in glauconite containing Neogene microfossils.

Erosion of the Neogene strata on the mid slope at the edge of the Eocene outcrop belt was intensified 1.4 Ma resulting in very high percentages of glauconite and extensive contamination by Miocene microfossils in core 10 at Site 604. The upper rise was erosional through much of the Pleistocene, and an erosional unconformity seen in the seismic record (Figure 7) was shown to represent 1.1-0.44 Ma (Lang and Wise, 1987). It was approximately then that climatic deterioration was accelerated and large northern hemisphere ice sheets reached mid-latitudes (*i.e.*

around the Brunhes-Matuyama boundary). This climatic deterioration probably further deflected the Gulf Stream, expanding the area of the Paleo-Slope Water, thus allowing the formation of larger and more frequent instabilities in the Gulf Stream. Subsequent intensification of erosion is marked by large slumps preserved in cores 8 and 9.

When ice sheets reached the coast of the United States during the late Pleistocene, ice-marginal conditions were established in this water mass, and the Slope Sea was almost completely isolated from the Sargasso Sea. This is consistent with the predicted suppression of eddy mixing processes in the Slope Water during the late Wisconsinan glacial maximum due to oceanographic changes (*c.f.* McIntyre *et al.*, 1976) and the indications of a decrease in ring formation and the offshore deflection of the Gulf Stream during modern winters. Erosion of Neogene strata in intercanon areas of the margin was suppressed, and Site 612 on the mid-slope accumulated sediments. Turbidity currents cut submarine canyons into the slope during glacioeustatic sea level lowstands and carried terrigenous sediments brought to the shelfbreak by ice sheets to Site 604 on the upper rise, occasionally eroding clasts of Eocene chalk from the canyon floors, redepositing them on the upper rise.

Rapid sedimentation on the continental slope during the late Quaternary, especially during stadials, indicate that oceanographic conditions above the continental margin, namely the presence or absence of a discrete water mass between the margin and the edge of the subtropical gyre, is the primary control on sedimentary processes on the margin. Glacioeustatic sea level lowstands were associated with rapid sedimentation in intercanon areas of the mid-slope, while higher energy conditions existed on the slope during interglacials/ interstadials, indicated by low sedimentation rates at Site 612, and minor reworking of Neogene glauconite from the slope onto

the rise (Site 604). Although the low rates of sedimentation on the slope could be a function of the geometry of the slope resulting in by-passing, evidence for accelerated slope erosion during interglacial, and especially during glacial/interglacial and interglacial/glacial transitions is indicative of an erosional mechanism operating on the slope during intervals of oceanographic instability on the Slope Water.

## CHAPTER VIII

### CONCLUSIONS

The micropaleontological and sedimentological record of DSDP Sites 604, 607, and 612 provided insight into the paleoceanographic conditions in the North Atlantic accompanying Quaternary changes in global climate. A comparison of the late Quaternary fossil record of marine plankton (planktonic foraminifera and dinoflagellate cysts) and allocthonous terrestrial spores and pollen from three sedimentary environments (the mid-slope, upper rise and mid-ocean ridge flank) examined how environmental conditions were reflected by each fossil group, and how the global climate signal was translated into "local climates".

This multidisciplinary study lead to the following major conclusions:

- 1) While climatostratigraphic interpretation of the microfossil and stable isotope record at Site 607 on the Mid-Atlantic Ridge was easy to relate to classical Quaternary studies based on planktonic foraminifera, interpretation of the relatively complex records from the U.S. continental margin could only be made using a multidisciplinary approach, integrating foraminiferal, dinocyst, pollen, sedimentological, and isotopic data (*i.e.* "multiple event stratigraphy" of Lamb *et al.*, 1987).

Reliance on any single type of data can result in incorrect interpretations. The application of the zonation of Ericson and Wollin (1968) to the Quaternary record of Site 604 on the upper rise by Moullade (1987a) is questioned since "warm water taxa", upon which biostratigraphic and climatostratigraphic zonations are based, are seldom present in the Slope Water. In addition, the

record at Site 604 is demonstrably discontinuous, with sedimentological and microfossil evidence of complex depositional processes, and incorporation of older sediments reworked from the slope. There is biostratigraphic evidence from the record of calcareous nannofossils (Lang and Wise, 1987) that a hiatus spanning approximately 1.1-0.44 Ma at this site, which is supported by the corresponding evidence for an erosional unconformity in the seismic record (Poag and Mountain, 1987). The existence of a long hiatus separating early Pleistocene sediments characterised by low sedimentation rates from late Pleistocene sediments characterised by rapid sedimentation (Lang and Wise, 1987) is substantiated by the foraminiferal, dinocyst and pollen records which record only two glacial cycles in the upper 84m (lithological unit I), and by the much higher microfossil concentrations in early Pleistocene sediments. The climatostratigraphy of this site was reinterpreted, equating lithological unit IA to the mid Wisconsinan-Holocene. The upper 35.3 m at Site 604 are therefore approximately equivalent to the 37 m of "Quaternary" sediments overlying the "Plio-Pleistocene boundary" at Site 612.

Also questioned is the identification of the Plio-Pleistocene boundary with the erosional unconformity at 37 m at Site 612 which separates homogeneous muds from glauconitic muds (Poag and Low, 1987), because the glauconitic sediments in Quaternary strata on the margin contain Neogene microfossils. Comparison of microfossils below the "Plio-Pleistocene boundary" at Site 612 and in the early Pleistocene record of Site 604 indicates that they may have been deposited during the early Pleistocene.

2) The sites on the continental margin reflect changes in the continental climate; pollen and terrestrial spores were present in statistically-significant numbers throughout the Pleistocene record of both sites on the continental margin, allowing general interpretations of the terrestrial

vegetation and climate. Planktonic microfossils record oceanographic changes accompanying the arrival of large continental ice sheets along the coast of the northeastern United States ca. 0.3 Ma, as well as the Younger Dryas climatic reversal.

3) The continental margin was subjected to complex depositional and erosional patterns, and more variable oceanographic conditions than the mid-ocean site; DSDP Sites 604 and 612 show large glacial-interglacial fluctuations in surface water temperature and salinity, rates of clastic sediment input, and rates of bottom erosion and resedimentation. The microfossil and sedimentological data suggest that the Slope Water mass developed gradually through the Quaternary, as climatic cooling deflected the Gulf Stream offshore, allowing cold surface water (the Labrador Current?) to flow along the continental margin south to Cape Hatteras. During the early part of the early Pleistocene (until ca. 1.4 Ma), the Gulf Stream flowed above the lower continental margin off New Jersey, several hundred kilometres inshore of its present position. Foraminifera-rich muds, with relatively high species diversity, rich in planktonics which presently characterise the subtropical gyre, were deposited on the upper rise. Small amounts of glauconite were reworked from the slope during climatic transitions, when water mass configurations were unstable; this is indicated by "mixed paleoclimatic signals" in the microfossil record of glauconitic sediments, with high percentages of "cold water taxa" (e.g. *Neogloboquadrina pachyderma*, *Brigantidium* spp., *Picea*, etc.) as well as "warm water taxa" (e.g. *Globigerinoides* spp., *Globorotalia tumida*, *G. truncatulinoides*, *Tectatodinium pellitum*, *Operculodinium israelianum*, etc.).

Around 1.4 Ma, the Gulf Stream was deflected from the margin and the "Paleo-Slope Water" was established between the margin and the edge of the subtropical gyre. Microfossil

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Around 1.4 Ma, the Gulf Stream was deflected from the margin and the "Paleo-Slope Water" was established between the margin and the edge of the subtropical gyre. Microfossil

assemblages suggest that this "transitional" water mass was more "neritic", *i.e.* cooler and less saline, than that which had overlain the margin in the earliest Pleistocene. This water mass was probably comprised of fluvial runoff, polar surface waters transported south (as the Labrador Current?), and considerable input from the Gulf Stream through eddy-mixing processes. The influx of terrigenous clastics and pollen grains from rivers increased relative to *in situ* CaCO<sub>3</sub>, downslope mass-wasting was accelerated, and reworking of glauconite onto the rise became more frequent. While this water mass existed, the margin was intensely erosional, indicated by 1) a sharp increase in terrigenous sedimentation and reworking of Neogene glauconite from the slope onto the upper rise, 2) the erosional unconformity between at Site 604 (Poag and Mountain, 1987; Wise and van Hinte, 1987) and the hiatus spanning at least 1.1- 0.44 Ma, and 3) the absence of pre-late Wisconsinan sediments at Site 612 on the mid-slope.

A sharp increase in boreal pollen (*e.g. Picea*) on the continental rise around 0.2 Ma correlates with planktonic microfossils indicating colder surface waters of slightly reduced salinity (*e.g. Globigerina quinqueloba* and *Brigantedinium* spp.), more neritic conditions, indicated by lower dinocyst:pollen ratios and by higher pollen diversity, and with the establishment of an ice-marginal benthic foraminiferal fauna. This "neritic" water mass is referred to as the Slope Water; its establishment probably corresponds to the very cold, low salinity conditions which existed in the North Atlantic during the Illinoian (Morley, 1982). A change in lithology also occurred at Site 604 at this time, with the terrigenous input increasing, and glauconite becoming much less frequent in the upper 65 m than in the next 60 m. Resedimentation of glauconite becomes increasingly less frequent upcore, and terrigenous sediments, with little CaCO<sub>3</sub> or glauconite, were deposited in the upper 35 m of Site 604 (lithological unit IA; mid Wisconsinan to Holocene). Clasts of Eocene chalk reworked into lithological unit I probably reflect increased turbidite

activity, as sediment-laden rivers carried their loads to the shelfbreak during glacioeustatic sea level lowstands, and incised canyons down to the chalk.

Considerable variations in oceanographic conditions between glacials and interglacials are indicated by the fossil plankton record (foraminiferal tests and dinocysts) in surface waters at Site 607, in the subtropical gyre just south of the glacial polar front. Sedimentation varied relatively little with changes in global climate, and the Quaternary record is dominated by foraminiferal ooze- unlike the continental margin, there was relatively little variation in planktonic foraminiferal abundance in the subtropical gyre between glacials and interglacials. However, higher amounts of terrigenous muds are found in the foraminiferal ooze deposited during glacials. The increase in terrigenous sediments is accompanied by a slight increase in pollen abundance on the mid-ocean ridge during glacials.

4) The records of the sites on the continental margin provide insight into the interaction between oceanic circulation and the growth of ice sheets. The gradual migration of ice centres toward the Atlantic North American coast may be related to the gradual deflection of the Gulf Stream from the coast of the United States north of Cape Hatteras documented in this study. The proximity of the warm Gulf Stream to the coast of the northeastern United States during the early Pleistocene moderated climates, restricting ice sheets to high latitudes or to high altitudes in continental areas of western North America. The high rates of precipitation at northern polar latitudes allowed the growth of polar and alpine ice sheets when summer insolation values were low, which probably caused shifts in wind patterns, causing the initial deflection of the Gulf Stream from the margin. This deflection allowed water of polar origin (the Labrador Current?) to flow along the eastern North American margin south to Cape Hatteras, hence cooling the North

American continent at mid-latitudes, and allowing large ice sheets to extend into to mid-latitudes in the continental interior. It also reduced the advection of warm water to the pole, allowing pack ice to form in the Arctic Ocean.

The reduction in the advection of warm water to the poles, resulting from deflection of the Gulf Stream, allowed the build up of pack ice in the Arctic Ocean by 1 Ma; this deflection subsequently caused ice centres to migrate south and east, within reach of warm, moist air masses. Progressive deflection of the Gulf Stream through the late Pleistocene may have caused ice centres to migrate from an early center ("Nebraskan", ca. 0.88 Ma) in the Keewatin district southeast to the Labradorean and Hudson Bay centres by 0.5 Ma. The advance of ice sheets to the coast of North America at mid-latitudes and development of an ice marginal fauna on the margin 0.3 Ma is associated with more neritic conditions (*i.e.* less influence from the Gulf Stream into the surface waters overlying the continental margin), recording the formation of a water mass similar to the modern Slope Water.

Extremely high rates of sedimentation existed between 10.9 and 10.5 ka in intercanyon areas of the mid slope (*ca.* 1cm/yr at Site 612, based on an accelerator date of 11,320 +/- 100 y. B.P. on foraminiferal and mollusc shells 13 m below the beginning of the Holocene interglacial). Planktonic microfossil assemblages indicative of reduced salinity, interspersed with assemblages recording cool conditions (equivalent to the Allerod) from 21 to 16 m in the core are followed by a return to planktonic microfossils indicating polar surface waters from 16 to 2 m (the Younger Dryas). The pollen record, however, records a unidirectional vegetational succession reflecting basically uninterrupted climatic amelioration. The Younger Dryas oscillation therefore appears to be an oceanically-induced phenomenon, explaining the location of widely-accepted sites

recording the Younger Dryas in regions with maritime climates (*cf.* Mott *et al.* 1986).

The very high sedimentation rates on the slope during the late glacial indicate that the influx of cold, sediment-laden water issuing from the rapidly-wasting North American ice sheets, through the Hudson River (*cf.* Ogden, 1977) may have triggered the climatic oscillation (*cf.* Broecker and Denton, 1989) by deflecting the Gulf Stream. This deflection allowed the cold inner Labrador Current to flow along the North American coast to Cape Hatteras, re-establishing the "Glacial Slope Water".

5) Because dinoflagellates, as a group, tolerate a wider range of salinity and temperature than do planktonic foraminifera, dinocysts permitted paleoceanographic information throughout the record of Sites 604 and 612, even during stage 2 (the late Wisconsinan stadial), when sediments were nearly barren of calcareous microfossils. Marked discrepancies were noted between paleoenvironmental signals recorded by dinocysts and planktonic foraminifera in pre-Wisconsinan sediments at Site 607, however. These discrepancies appear to be due to post-mortem alteration of dinocyst assemblages due to oxidation resulting from percolation of oxygen-rich waters within the porous nannofossil-foraminiferal oozes deposited on the Mid-Atlantic Ridge; interglacial dinocyst assemblages were especially subject to dissolution because terrigenous influx was very low during interglacials relative to glacials, resulting in higher porosity and permeability within interglacial oozes.

6) Sediments on the rise are more similar to those on the slope than to those on the mid-ocean ridge; the continental margin sediments are dominantly terrigenous while the mid-ocean site is characterised by biogenic calcareous ooze. In addition, the microfossil record from the two sites

on the continental margin show close similarities, but differ substantially from the record of the mid-ocean site. Both the mid-slope and upper rise sites presently are dominated by the planktonic foraminifera dextral *Neogloboquadrina pachyderma* and *Globorotalia inflata*, by the dinoflagellate cysts *Operculodinium centrocarpum*, *Spiniferites* spp. (mainly *S. ramosus*), and *Brigantedinium* spp. (i.e. the neritic-upper slope ecofacies of Wall *et al.*, 1977), and contain a rich pollen assemblage, with high percentages of temperate taxa, e.g. *Quercus*, *Pinus*. The mid-ocean ridge environment, in contrast, contains very little pollen, and contains high percentages of subtropical and tropical foraminiferal taxa (e.g. *Globorotalia truncatulinoides*, *Globigerina falconensis*, and *Globigerinoides ruber*) and the sparse dinocyst assemblage is dominated by "oceanic" taxa such as *Impagidinium aculeatum*. The similarity in the sedimentology and micropaleontology of Sites 604 and 612 reflects the existence of the same surface water mass (Slope Water) above both sites. The Gulf Stream (the fast-flowing western boundary of the subtropical gyre) acts a barrier to the transport of microfossils and sediments from the Slope Water mass, that overlies the continental margin, to the subtropical gyre (Sargasso Sea).

The dinocyst record correlates well with the record of planktonic foraminifera, whose habitat they share, except that the dinocysts tend to lead the foraminifera in recording environmental change. This may simply be due to the fact that resting cyst morphology (and therefore taxonomic assignation) of dinoflagellates appears to be environmentally determined; dinoflagellate thecae assigned to one species can form resting cysts assigned to a number of different species, and even sometimes genera. The dinocyst distributions correlate less closely with those of benthic foraminifera, except in recording major environmental changes, such the arrival of ice sheets along the northeastern U.S. coast.

7) The entire New Jersey margin appears to have been depositional since the late Wisconsinan, when the mid-slope (Site 612) became an area of net deposition, although Holocene rates of sedimentation at Site 612 appear to be very low relative to rates at Site 604 on the upper rise. Sedimentation on the margin was very rapid during stadials but very slow during interglacials, so that glacial conditions are recorded through most of the upper 37 m at both the slope and rise sites. The suppression of slope erosion since the late Wisconsinan, indicated by the thick late Wisconsinan-Recent record at Site 612, and by the reduction in the amount of glauconite reworked from the slope onto the upper rise (Site 604) suggests that the change in the oceanographic character of the surface waters above the margin (*i.e.* less saline, colder Slope Water), and enlargement of the Slope Sea by increased deflection of the Gulf Stream, suppressed erosion of intercanyon areas of the slope.

The apparent climatic control of erosional processes on the margin is best explained by climate-driven surface water processes which affect margin sediments, *i.e.* the Gulf Stream and its associated eddies or instabilities (*e.g.* warm core rings, meanders, etc.). The influence of these instabilities in shaping the U.S. Atlantic continental margin is supported by the following lines of evidence: 1) modern oceanographic measurements and models showing the concentration of abyssal kinetic energy and bursts of strong current activity on the lower continental slope, 2) the highest concentration of warm core rings and the highest eddy kinetic energy on the U.S. Atlantic margin being off New Jersey, where Eocene strata presently are exposed, and 3) the suppression of erosion of intercanyon areas of the New Jersey slope during late Pleistocene glacial maxima, when eddy mixing processes would be suppressed by paleoceanographic changes during glacials (*e.g.* strong zonal polar frontal system, change in Slope Water composition to resemble Labrador Sea Water, etc.).

8) The results disagree somewhat with the predictions of the depositional model of Vail *et al.* (1977). Oceanographic conditions in the surface waters overlying the margin appear to be a more important factor than glacioeustatic sea-level in controlling erosion of the margin off the northeastern United States. Intercanyon areas of the mid- to lower slope off New Jersey were sites of rapid deposition during sea-level lowstands when the shelf was erosional, probably due to the deposition of a lowstand fan at the base of the slope, as predicted by the version of the Vail model put forward by Posamentier *et al.* (1988). Contrary to the predictions of the model, however, the mid-lower slope was dominantly erosional during highstands in sea-level. This does not appear to simply be non-deposition as a result of by-passing due to the geometry of the margin because reworking of Neogene sediments onto the rise also peaks during interglacials. The Vail depositional model, therefore, may not be applicable to continental margins on the western sides of gyres where strong boundary currents exist.

## PLATE 1

## Common Planktonic Foraminiferal Taxa at the Sites Studied

1. *Globigerina quinqueloba* NATLAND, 1938; DSDP Site 607, 1-3 (31-34)
2. *Globigerina falconensis* BLOW, 1959; DSDP Site 607, 1-3 (31-34)
3. *Globigerina cariacensis* ROGL and BOLLI, 1973; DSDP Site 607, 1-3 (31-34)
4. *Globigerina bulloides* D'ORBIGNY, 1826; DSDP Site 607, 1-3 (31-34)
5. *Neogloboquadrina pachyderma* (EHRENBERG), 1861 (sinistral); DSDP Site 607, 1-3 (31-34)
6. *Neogloboquadrina pachyderma* (EHRENBERG), 1861 (dextral); DSDP Site 607, 1-3 (31-34)
7. *Neogloboquadrina humerosa* (TAKAYANAGI AND SAITO), 1962; DSDP Site 604, 12-1 (31-34)
8. *Neogloboquadrina dutertrei* (D'ORBIGNY), 1839; DSDP Site 612, 2-1 (0-6)

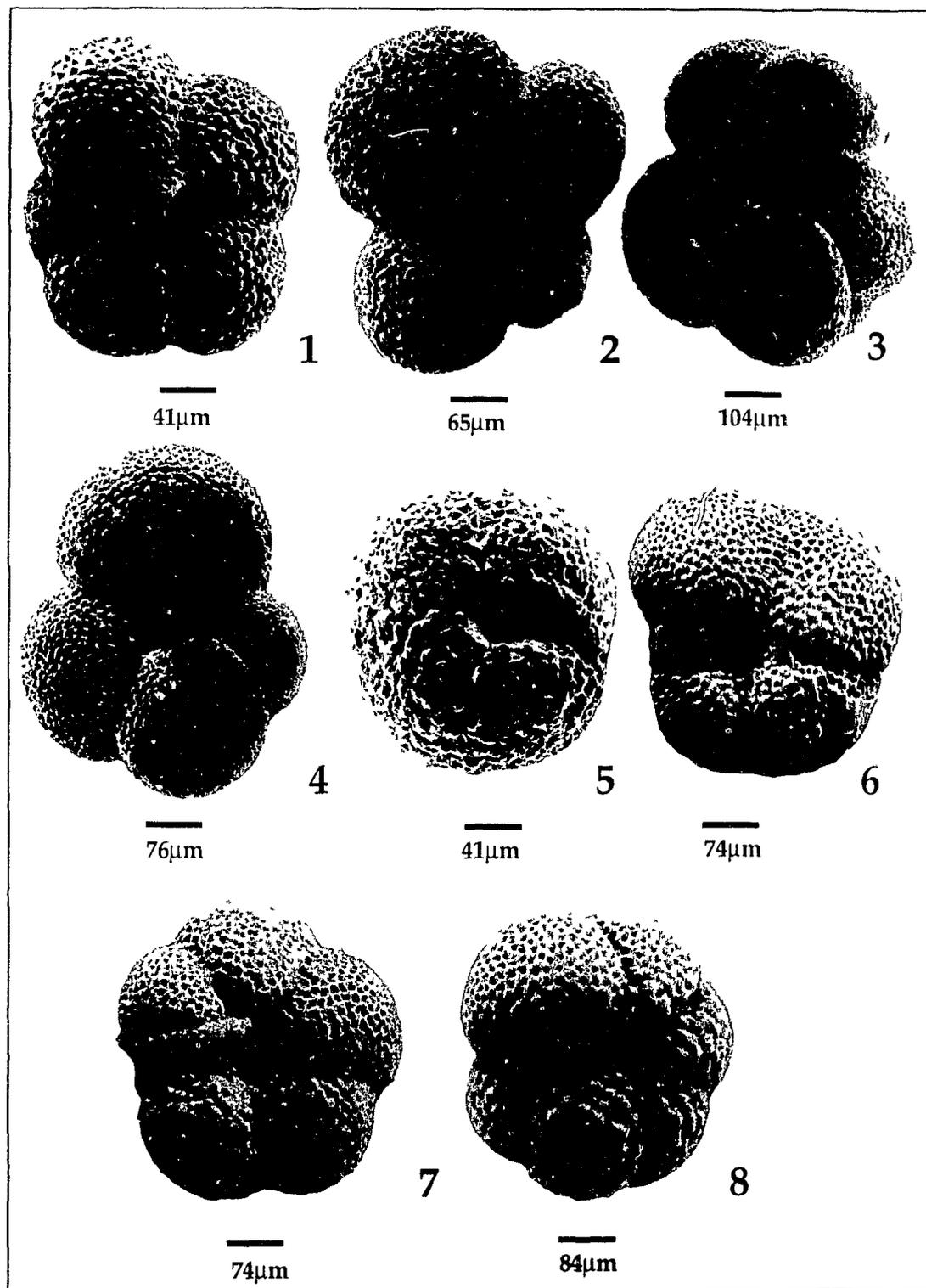


Plate 1

**PLATE 2****Common Planktonic Foraminiferal Taxa at the Sites Studied**

1. *Globigerinoides sacculifer* (BRADY), 1877; DSDP Site 604, 6-3 (16-19)
2. *Globigerinoides conglobatus* (BRADY), 1879; DSDP Site 604, 1-1 (0-3)
3. *Globigerinoides ruber* (D'ORBIGNY), 1839; DSDP Site 604, 6-3 (16-19)
4. *Globigerinoides sacculifer* (BRADY), 1877; DSDP Site 604, 6-3 (16-19)
5. *Globigerinoides pyramidalis* (VAN DEN BROEK), 1876; DSDP Site 604, 6-3 (16-19)
6. *Globigerinoides ruber* (D'ORBIGNY), 1839; DSDP Site 607, 1-3 (31-34)
7. *Globigerinita glutinata* (EGGER), 1893; DSDP Site 607, 1-3 (31-34)
8. *Globigerinita glutinata* (EGGER), 1893; DSDP Site 607, 1-3 (31-34)
9. *Globigerinella aequilateralis* (BRADY), 1839; DSDP Site 607, 1-3 (31-34)

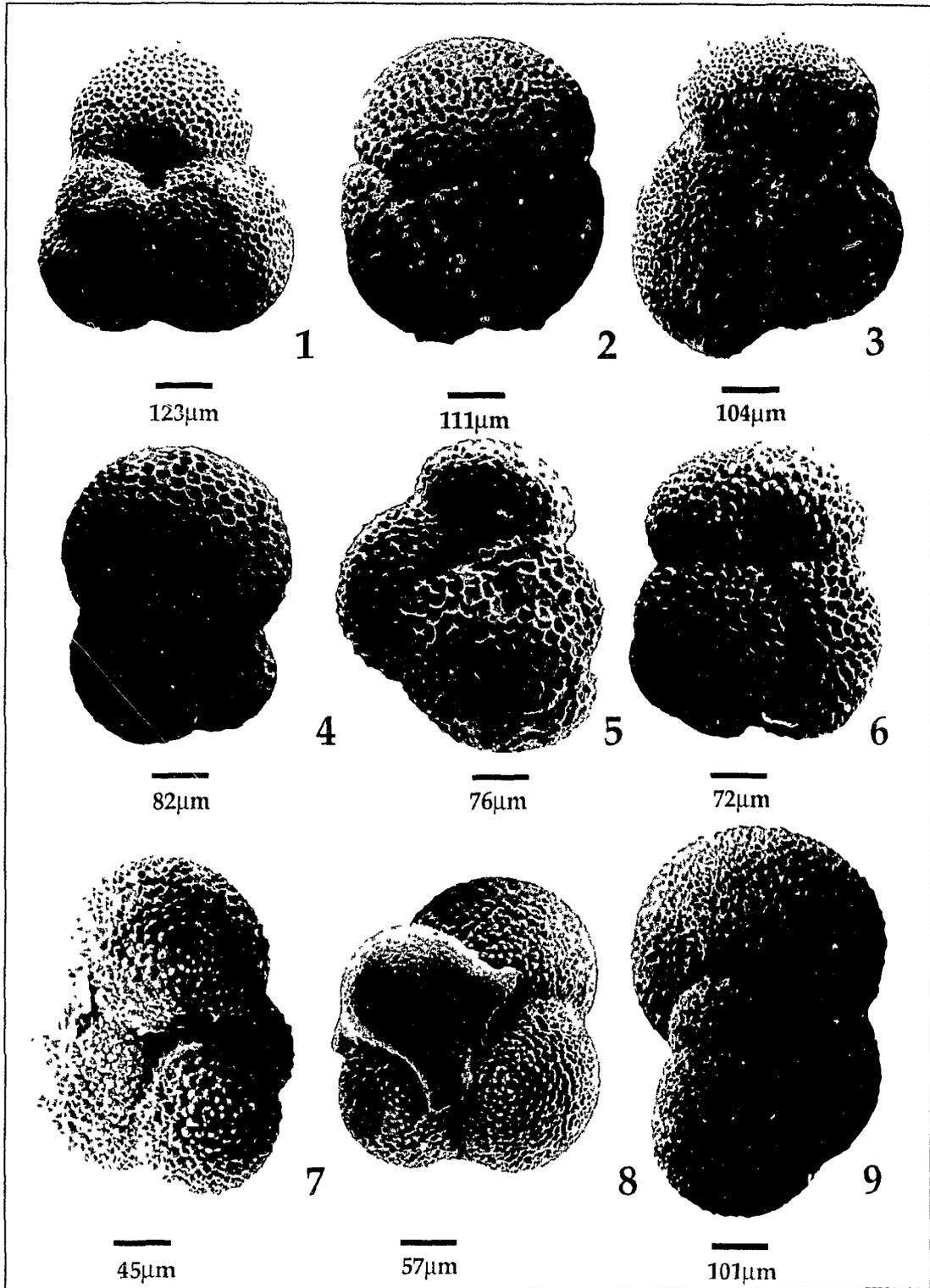


Plate 2

## PLATE 3

## Common Planktonic Foraminiferal Taxa at the Sites Studied

1. *Globorotalia hirsuta* (D'ORBIGNY), 1839; DSDP Site 607, 1-3 (31-34)
2. *Globorotalia scitula* (BRADY), 1882; DSDP Site 607, 1-3 (31-34)
3. *Globorotalia truncatulinoides* (D'ORBIGNY), 1839; DSDP Site 607, 1-3 (31-34)
4. *Globorotalia menardii* (PARKER, JONES AND BRADY), 1865; DSDP Site 612, 2-1 (0-6)
5. *Globorotalia tumida* (BRADY), 1877; DSDP Site 604, 12-1 (31-34)
6. *Globorotalia crassiformis* (GALLOWAY AND WISSLER), 1927; DSDP Site 607, 1-3 (31-34)
7. *Globorotalia inflata* (D'ORBIGNY), 1839; DSDP Site 607, 1-3 (31-34)
8. *Orbulina universa* (D'ORBIGNY), 1839; DSDP Site 607, 1-3 (31-34)

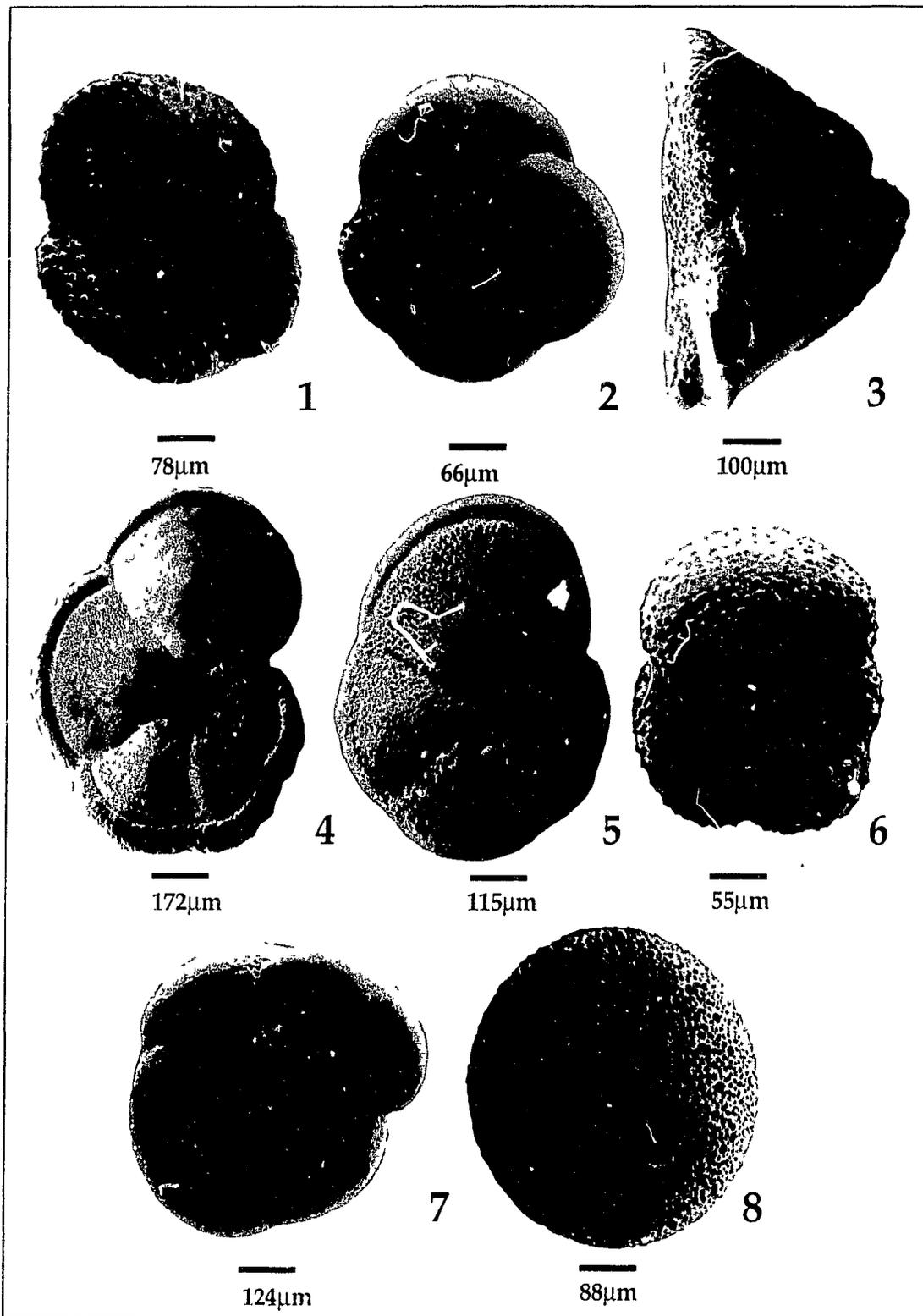


Plate 3

## PLATE 4

## Common Dinoflagellate Cyst Taxa at the Sites Studied

All magnifications are X1000; photographed in normal transmitted light with interference contrast.

(EF= England Finder #)

1. *Brigantedinium simplex*. DSDP Site 607, 1-1 (62-65). Low focus, showing archeopyle and attached operculum. (J53/1)
2. *Brigantedinium simplex*. DSDP Site 607, 1-1 (62-65). Mid focus. (J53/1)
3. ?*Multispinula minuta*. DSDP Site 612, 5-1 (116-118). High focus of typical cyst, partially obscured by organic debris. (X68/2)
4. *Selenopemphix nephroides*. DSDP Site 604, 6-5 (11-13). Low focus. (Q56/4)
5. *Spiniferites pseudofurcatus*. DSDP Site 604, 16-4 (62-65). Low focus showing typical pair of processes in lower right corner of the photo. (S34/1)
6. *Spiniferites pseudofurcatus*. DSDP Site 604, 16-4 (62-65). High focus. (S34/1)

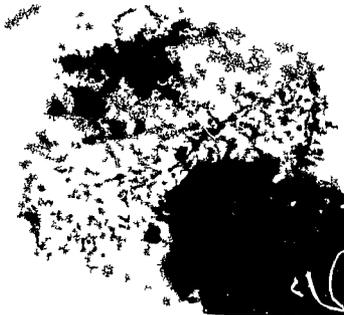


15  $\mu$ m

1



2



3



4



5



6

PLATE 4

## PLATE 5

## Common Dinoflagellate Cyst Taxa at the Sites Studied

All magnifications are  $\times 1000$ ; photographed in normal transmitted light with interference contrast.

(EF= England Finder #)

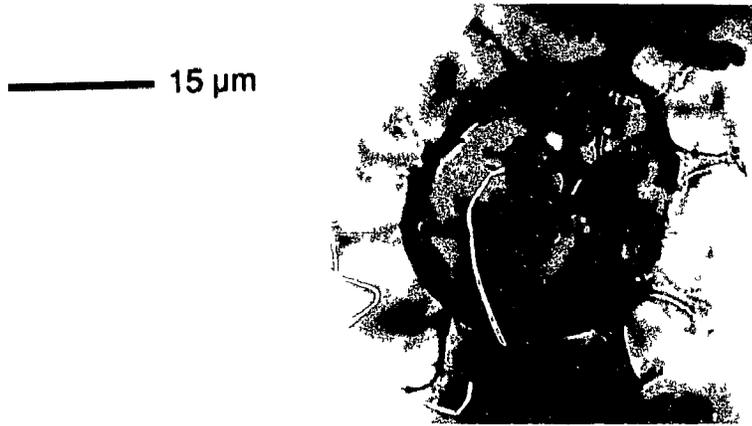
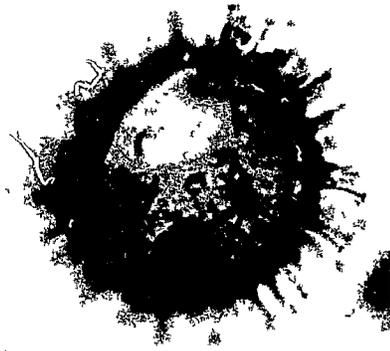
1. *Operculodinium centrocarpum* (Deflandre and Cookson) Wall, 1967. DSDP Site 612, 5-1 (116-118). High focus showing archeopyle. (G42/4)

2. *Operculodinium centrocarpum* (Deflandre and Cookson) Wall, 1967. DSDP Site 612, 5-1 (116-118). Mid focus. (G42/4)

3. *Operculodinium israelianum* (Rossignol) Wall, 1967. DSDP 604, 16-4 (62-65). High focus showing archeopyle. (H23/3)

4. *Operculodinium israelianum* (Rossignol) Wall, 1967. DSDP 604, 16-4 (62-65). Mid focus of broken specimen. (H23/3)

5. *Spiniferites ramosus* (Ehrenberg) Loeblich and Loeblich, 1966. DSDP Site 5-1 (116-118). Mid focus. (H41/4)



15  $\mu$ m

PLATE 5

## PLATE 6

## Common Dinoflagellate Cyst Taxa at the Sites Studied

All magnifications are X1000; photographed in normal transmitted light with interference contrast.

(EF= England Finder #)

1. *Tectatodinium pellitum*. DSDP Site 604, 16-4 (62-65). High focus, showing operculum attached to the cyst wall. (G41/2)

2. *Tectatodinium pellitum*. DSDP Site 604, 16-4 (62-65). Mid focus. (G41/2)

3. *Tectatodinium pellitum*. DSDP Site 604, 16-4 (62-65). High focus view of a typical specimen. (M29/3)

4. *Tectatodinium pellitum*. DSDP Site 604, 16-4 (62-65). Mid focus, showing the spongy appearance of the thick wall. (M29/3)

5. *Bitectatodinium tepikiense*. DSDP Site 604, 16-4 (62-65). High focus. (H35/3)

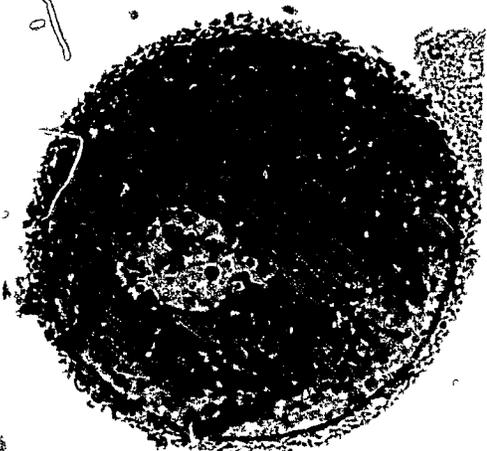
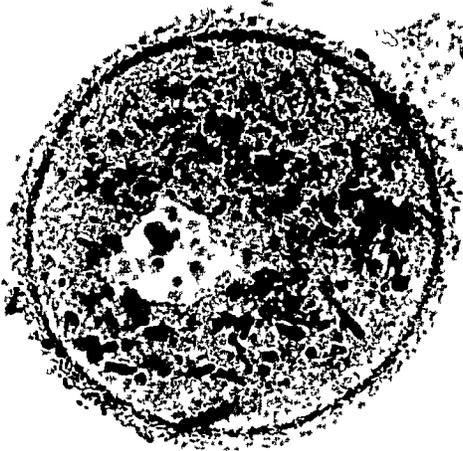
6. *Bitectatodinium tepikiense*. DSDP Site 604, 16-4 (62-65). Mid focus, showing the relatively smooth, thin wall. (H35/3)



15  $\mu$ m

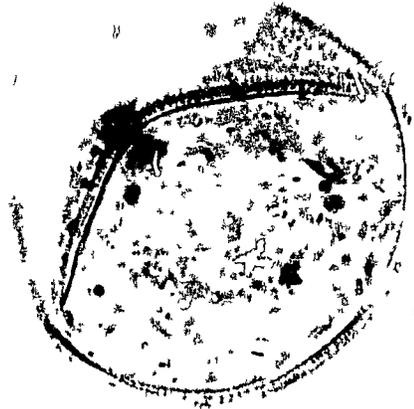
1

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PLATE 6

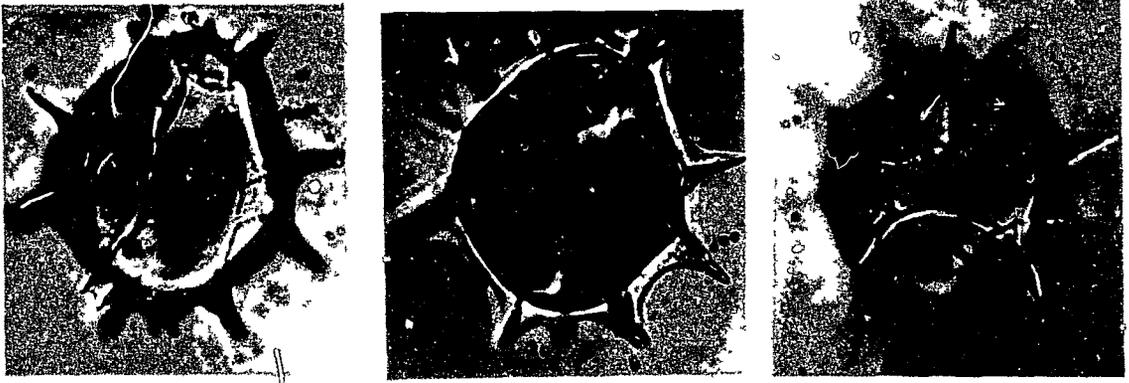
## PLATE 7

## Common Dinoflagellate Cyst Taxa at the Sites Studied

All magnifications are X1000; photographed in normal transmitted light with interference contrast.

(EF= England Finder #)

1. *Impagidinium aculeatum*. DSDP Site 607, 1-1 (62-65). Low focus. (X29/3)
2. *Impagidinium aculeatum*. DSDP Site 607, 1-1 (62-65). Mid focus. (X29/3)
3. *Impagidinium aculeatum*. DSDP Site 607, 1-1 (62-65). High focus. (M22/4)
4. *Impagidinium patulum*. DSDP Site 607, 1-1 (91-94). High focus. (Z51/3)
5. *Impagidinium patulum*. DSDP Site 607, 1-1 (91-94). Mid focus. (Z51/3)
6. *Nematosphaeropsis labyrinthus*. DSDP Site 607, 1-1 (31-34). Mid focus. Most specimens of *N. labyrinthus* in this study were quite corroded like this typical specimen. (G38/2)
7. *Nematosphaeropsis labyrinthus*. DSDP Site 607, 1-1 (31-34). High focus. (G38/2)



15  $\mu$ m

1

2

3



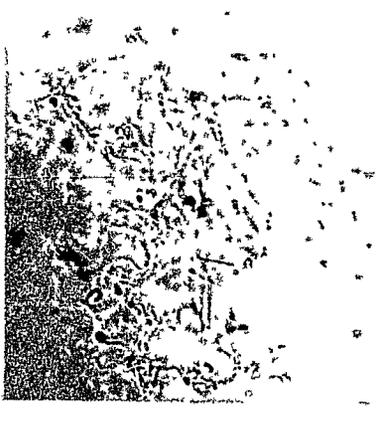
4



5



6



7

PLATE 7

**Appendix A: Effect of spiking palynomorph preparations with different marker particles.**

In order to estimate palynomorph concentrations, samples were "spiked" with a known number of marker grains, following the method of Davis (1965). Because sample preparation was carried out in two different labs, two different types of marker particles were used in preparation, each being in use in each lab: microspheres (Ogden, 1986), and *Lycopodium* spores (Stockmarr, 1971). In theory, the use of different marker particles should not affect resulting estimated palynomorph concentrations. When pollen and dinocyst concentrations were plotted from the first site studied, it was clear that estimated concentrations did differ significantly and systematically between the samples spiked with *Lycopodium* spores and those spiked with microspheres. In order to obtain a general "conversion factor" to apply to the calculated palynomorph concentrations from this site, to make approximate comparisons of palynomorph concentrations, a number of samples were subsequently processed using both types of marker particle. The comparison in estimated palynomorph concentrations obtained using both methods is shown in Appendix Tables 5 and 10.

Although the technique described by Ogden (1986) was carefully followed, the differences in the estimates of concentration probably arose because of the higher specific gravity of the microspheres. The relatively high specific gravity causes microspheres to settle three to four times more rapidly than pollen grains (Ogden, 1986), while *Lycopodium* spores, in contrast, have a specific gravity comparable to common pollen grains. It is therefore more difficult to ensure complete homogenisation of the sample when using marker particles that settles much more quickly. Although Ogden (1986) performed 10 counts on each of 5 samples to which microspheres were added and 5 samples to which *Eucalyptus* pollen was added, and calculated

the statistics (mean, standard deviation, and COV), and found an acceptable subsample range for both techniques, he did not add both types of "spikes" to any samples and compare the values for absolute palynomorph frequency obtained using each technique. Thus, while comparing estimates of concentration between samples spiked with any given type of marker particle appears to yield statistically viable comparisons, it is not possible to compare concentrations estimated using different types of marker particles.

## Appendix B: Systematic Taxonomy of Dinoflagellate Cysts

The dinocyst taxa identified in this study are listed below with remarks where appropriate. Synonyms for most of the species are listed in Lentin and Williams (1989). For a few taxa, the nomenclature of Mudie (1987) is employed rather than that of Lentin and Williams (1989), but the nomenclature of Lentin and Williams is indicated in the descriptions. The illustrated specimens are curated at the Centre for Marine Geology, Dalhousie University, Halifax, Nova Scotia, with locations identified using an England finder.

Genus *ACHOMOSPHERA* Evitt, 1963

*Achomosphaera andalousiense* Jan du Chene, 1977

Genus *BITECTATODINIUM* Wilson, 1973

*Bitectatodinium tepikiense* Wilson, 1973. Plate 6, Figs. 5, 6.

This taxon can be difficult to distinguish from *Tectatodinium pellitum* without a clear view of the archeopyle, but *T. pellitum* characteristically has a thicker, more "spongy" wall than does *B. tepikiense* (see Plate 6).

Genus *BRIGANTEDINIUM* Reid, 1974. Plate 4, Figs. 1, 2.

*Brigantedinium* spp. Cysts with the characteristic brown wall colour and the structure of *Brigantedinium*, but which could not be identified at the species level due to damaged/hidden archeopyle areas were grouped as *Brigantedinium* species. Due to the high numbers of these "unidentifiable" *Brigantedinium* spp. in this study, all these cysts were grouped together in the figures and tables. The vast majority of these cysts in continental margin sediments appear to belong to the species *Brigantedinium simplex* Reid 1974, while high percentages of cysts attributable to *Brigantedinium auranteum* Reid 1974, or *Brigantedinium cariacense* Reid 1974 also were found at Site 607 on the Mid Atlantic Ridge.

Genus *CORRUDINIUM* Matsuoka, 1983

*Corrudinium harlandii* Matsuoka, 1983

Genus *FILISPHAERA* Bujak, 1984

*Filisphaera filifera* Bujak, 1984

Genus *GLENODINIUM* Ehrenberg, 1937

*Glenodinium* sp. A of Mudie 1987

Genus *HYSTRICHOKOLPOMA* Klump, 1953; emend. Williams and Downie in Davey *et al.* 1966.

The cysts attributed to this genus in this study are very small cysts resembling those illustrated in Mudie, 1987, and thus are described as "c.f. *Hystrichokolpoma* sp."

Genus *IMPAGIDINIUM* Stover and Evitt, 1978

*Impagidinium aculeatum* (Wall) Lentin and Williams, 1981. Plate 7, Figs. 1-3

*Impagidinium aliferum* Mudie, 1987

*Impagidinium paradoxum* (Wall) Stover and Evitt, 1978

*Impagidinium patulum* (Wall) Stover and Evitt, 1978. Plate 7, Figs. 4, 5.

*Impagidinium striatum* (Wall) Stover and Evitt, 1978

Genus *LINGULODINIUM* Wall, emend. Wall, Dale and Harada, 1973

*Lingulodinium machaerocarpum* (Deflandre and Cookson) Wall, 1967

Genus *MULTISPINULA* Bradford, 1975

?*Multispinula minuta* Harland and Reid in Harland *et al.*, 1980. Plate 4, Fig. 3.

(= *Algidasphaeridium minuta*, Lentin and Williams, 1989)

*Multispinula quanta* Bradford, 1975

Genus *NEMATOSPHAEROPSIS* Deflandre and Cookson, 1955, emend. Williams and Downie,  
1966

*Nematosphaeropsis labyrinthus* (Ostenfield) Reid, 1974. Plate 7, Figs. 6, 7.

Genus *OPERCULODINIUM* Wall, 1967

*Operculodinium centrocarpum* (Deflandre and Cookson) Wall, 1967. Plate 5, Figs. 1, 2.

*Operculodinium israelianum* (Rossignol) Wall, 1967. Plate 5, Figs. 3, 4.

*Operculodinium longispermum* Matsuoka, 1983

*Operculodinium psilatum* Wall, 1967

Genus *Polysphaeridium* Davey and Williams, 1966. emend. Bujak *et al.*, 1980

*Polysphaeridium zoharyi* (Rossignol) Bujak *et al.* 1980

Genus *SELENOPEMPHIX* Benedek, 1972, emend. Bujak, 1980

*Selenopemphix nephroides* Benedek, 1972, emend. Bujak, 1980. Plate 4, Fig. 4.

Genus *SPINIFERITES* Mantell, 1950, emend. Sarjeant, 1970

*Spiniferites bentorii* (Rossignol) Wall and Dale, 1970

*Spiniferites elongatus* Reid, 1974

*Spiniferites membranaceous* (Rossignol) Sarjeant, 1970

*Spiniferites mirabilis* (Rossignol) Sarjeant, 1970

*Spiniferites ramosus* (Ehrenberg) Loeblich and Loeblich, 1966. Plate 5, Fig. 5. In this taxon are grouped specimens assigned by other workers to *Spiniferites bulloideus* (Deflandre and Cookson) Sarjeant, 1970, since there appears to be complete gradation between both forms in my samples.

*Spiniferites pseudofurcatus* Harland, 1985. Plate 4, Figs. 5, 6.

*Spiniferites* sp. Broken/obscured cysts obviously attributable to this genus, but which could not be identified to species were counted as *Spiniferites* sp.

Genus *TECTATODINIUM* Wall, 1967

*Tectatodinium pellitum* Wall, 1967. Plate 6, Figs. 1-4.

*Tectatodinium simplex* (Harland) Edwards, 1984

Genus *XANDARODINIUM* Reid, 1975

*Xandarodinium variabile* Bujak, 1984

### Appendix C. Abbreviated Systematic Taxonomy of Planktonic Foraminifera

*Globigerina bulloides* D'ORBIGNY, 1826

*Globigerina bulloides* D'Orbigny, 1826, p. 277 (no figures), list 1, d'Orbigny 1839b, p. 132. pl. 2, figs 1-3

*Globigerina bulloides* d'Orbigny- Saito, Thompson and Breger, 1981, pl. 7, figs. 1a-1c.

*Globigerina cariaensis* ROGL and BOLLI, 1973

*Globigerina megastoma cariaensis* Rogl and Bolli, 1973, p. 564, pl. 2, figs 1-10

*Globigerina cariaensis* Rogl and Bolli- Saito, Thompson and Breger, 1981, pl. 5, figs. 1a-1c.

*Globigerina falconensis* BLOW, 1959

*Globigerina falconensis* Blow, 1959, p. 177, pl. 9, figs 40a-c, 41.

*Globigerina falconensis* Blow- Saito, Thompson and Breger, 1981, pl. 7, figs. 2a-2c.

*Globigerina quinqueloba* NATLAND, 1938

*Globigerina quinqueloba* Natland, 1938, p. 149, p. 6, figs. 7a-c.

*Globigerina quinqueloba* Natland- Saito, Thompson and Breger, 1981, pl. 10, figs. 1a-c, 2a-2c.

***Globigerina rubescens* Hofker, 1956**

*Globigerina rubescens* Hofker, 1956, p. 234, pl. 32, figs. 18-21.

*Globigerina rubescens* Hofker- Saito, Thompson and Breger, 1981, pl. 11, figs. 1a-c.

***Globigerinella aequilateralis* (BRADY), 1839**

*Globigerina aequilateralis* Brady, 1879, p. 285, Brady 1884 p. 605, pl. 80, figs 18-21.

*Globigerinella aequilateralis* (Brady) Cushman, 1927, p. 87, Bradshaw, 1959, p. 38, pl. 7, figs. 1-2, Saito, Thompson and Breger, 1981, pl. 2, figs. 2a-2c.

***Globigerinita glutinata* (EGGER), 1893**

*Globigerina glutinata* Egger, 1893, p. 371, pl. 13, figs 19-21.

*Globigerinita glutinata* (Egger)- Parker, 1962, p. 246-249, pl. 19, figs. 1-16.

*Globigerinita glutinata* (Egger)- Saito, Thompson and Breger, 1981, pl. 22, figs. 1a-d, 2-7.

***Globigerinoides conglobatus* (BRADY), 1879**

*Globigerina conglobata* Brady, 1879, p. 286 (no figures)- Brady, 1884, p. 603, pl. 80, figs. 1-5;  
pl. 82, fig. 5.

*Globigerinoides conglobata* (Brady)- Parker, 1948, p. 238, pl. 7, figs. 8,9.

*Globigerinoides conglobatus* (Brady)- Saito, Thompson and Breger, 1981, pl. 14, figs. 1a-1c.

***Globigerinoides obliquus* Bolli, 1957**

*Globigerinoides obliqua* Bolli, 1957, p. 112-114, pl.25, figs 9a-10c.

*Globigerinoides obliquus* Bolli- Blow, 1969, p. 324.

*Globigerinoides obliquus* Bolli- Saito, Thompson and Breger, 1981, pl. 13, figs. 2a-2c.

***Globigerinoides pyramidalis* (VAN DEN BROECK), 1876**

*Globigerina bulloides* d'Orbigny var. *rubra* d'Orbigny subvar. *pyramidalis* van den Broeck, 1876,  
p. 127, pl. 3, figs. 9-10.

*Globigerina pyramidalis* (van den Broeck)- Banner and Blow, 1960a, p. 21.

*Globigerinoides ruber* (d'Orbigny) forma *pyramidalis*- Boltovskoy, 1968, p. 89-90.

*Globigerinoides pyramidalis* (van den Broeck)- Saito, Thompson and Breger, 1981, pl. 16, figs.  
1a-1c.

***Globigerinoides ruber* (D'ORBIGNY), 1839**

*Globigerina rubra* d'Orbigny, 1839a, p. 82-83, pl.4, figs 12-14.

*Globigerinoides ruber* (d'Orbigny)- Parker, 1962, p. 230, pl.3, figs. 11-14; pl. 4, figs. 1-10.

*Globigerinoides ruber* (d'Orbigny)- Saito, Thompson and Breger, 1981, pl. 15, figs. 1a-1c.

***Globigerinoides sacculifer* (BRADY), 1877**

*Globigerina helicina* d'Orbigny- Carpenter, 1862 (not d'Orbigny), pl. 12, fig.11

*Globigerina sacculifera* Brady, 1877, p.535 (no figures)- Brady 1884, p. 604, pl. 80, figs. 11-17;  
pl. 81, fig.2; pl 82, fig. 4.

*Globigerinoides sacculifer* (Brady)- Parker, 1967, p. 156-158, pl. 21, figs. 1,2,4, text-figure 5.

*Globigerinoides sacculifer* (Brady)- Saito, Thompson and Breger, 1981, pl. 17, figs. 1a-1c.

***Globigerinoides tenellus* Parker, 1958**

*Globigerinoides tenella* Parker- 1958, p. 280, pl. 6, figs. 7-11.

*Globigerinoides tenellus* Parker- Parker, 1962, p. 232, pl. 4, figs. 11-12.

*Globigerinoides tenellus* Parker- Saito, Thompson and Breger, 1981, pl. 11, figs. 2a-2c.

*Globorotalia crassaformis* (GALLOWAY AND WISSLER), 1927

*Globigerina crassaformis* Galloway and Wissler, 1927, p. 41, pl. 7, fig. 12.

*Globorotalia crassaformis* (Galloway and Wissler)- Parker, 1962, p. 235, pl. 4, figs. 17, 18, 20, 21.

*Globorotalia crassaformis* (Galloway and Wissler)- Saito, Thompson and Breger, 1981, pl. 43, figs. 2a-d.

*Globorotalia hirsuta* (D'ORBIGNY), 1839

*Rotalina hirsuta* d' Orbigny, 1839b, p. 131, pl.1, figs 37-39.

*Globorotalia hirsuta* (d'Orbigny)- Cushman, 1931, p. 99, pl.17, fig. 6.

*Globorotalia hirsuta* (d'Orbigny)- Saito, Thompson and Breger, 1981, pl. 46, figs. 1a-d.

*Globorotalia inflata* (D'ORBIGNY), 1839

*Globigerina inflata* d' Orbigny, 1839b, p. 134, pl. 12, figs. 7-9.

*Turborotalia inflata* (d'Orbigny)- Bermudez, 1961, p. 1323-1324, pl. 18, figs. 2a-b.

*Globorotalia inflata* (d'Orbigny)- Parker, 1962, . 236, pl. 5, figs 6-9.

*Globorotalia inflata* (d'Orbigny)- Saito, Thompson and Breger, 1981, pl. 41, figs. 1a-d.

***Globorotalia menardii* (PARKER, JONES AND BRADY), 1865**

*Rotalia (Rotalie) menardii* d'Orbigny, 1826, p. 273, no. 26.

*Globorotalia menardii* (d'Orbigny)- Cushman, 1927, p. 175. -Cushman, 1931, p. 91, pl. 17, fig.

1

*Globorotalia menardii* (d'Orbigny)- Saito, Thompson and Breger, 1981, pl. 50, figs. 1a-d.

***Globorotalia scitula* (BRADY), 1882**

*Pulvulina scitula* Brady, 1882, p. 716-717 (no figures)- Banner and Blow, 1960a, p. 27-29, pl. 5, figs. 5a-c (lectotype).

*Globorotalia scitula* (Brady)- Cushman and Henbest, 1940, p. 36, pl. 8, figs. 5a-c.

*Globorotalia scitula* (Brady)- Saito, Thompson and Breger, 1981, pl. 46, figs. 2a-d.

***Globorotalia tosaensis* TAKAYANAGI AND SAITO, 1962**

*Globorotalia tosaensis* Takayanagi and Saito, 1962, p. 81-82, pl. 28, figs. 11-12.

*Globorotalia tosaensis* Takayanagi and Saito- Saito, Thompson and Breger, 1981, pl. 54, figs. 2a-d.

*Globorotalia truncatulinoides* (D'ORBIGNY), 1839

*Rotalina truncatulinoides* d'Orbigny, 1839, p. 132, pl. 1, figs. 25-27.

*Globorotalia truncatulinoides* (d'Orbigny)- Cushman, 1927, p. 176. -Cushman, 1931, p. 97, pl. 17, fig. 4.

*Globorotalia truncatulinoides* (d'Orbigny)- Saito, Thompson and Breger, 1981, pl. 54, figs. 1a-d.

*Globorotalia tumida* (BRADY), 1877

*Pulvulina menardii* (d'Orbigny) var. *tumida* Brady, 1877, p. 535 (no figures). - Banner and Blow, 1960a, p. 26-27, pl. 5, fig. 1 (lectotype).

*Pulvulina tumida* Brady, 1884, p. 692, pl. 103, figs. 4-6.

*Globorotalia tumida* (Brady)- Cushman, 1927, p. 91, pl. 19, fig. 12 (after Brady).

*Globorotalia tumida* (Brady)- Saito, Thompson and Breger, 1981, pl. 50, figs. 2a-d.

*Globorotalia ungulata* Bermudez, 1961

*Globorotalia ungulata* Bermudez, 1961?, p. 1304, pl. 15, figs 6a-b.

*Globorotalia ungulata* Bermudez- Saito, Thompson and Breger, 1981, pl. 53, figs. 2a-d.

*Neogloboquadrina blowi* Rogl and Bolli, 1973

*Globigerina cretacea* Brady, 1884 (not d'Orbigny), p. 596, pl. 82, fig. 10.

*Neogloboquadrina dutertrei* (d'Orbigny) subsp. *subcretacea* (Lomniki) Bandy, Frerichs, and Vincent, 1967, p. 152-157, pl. 4, fig. 2.

*Negloboquadrina dutertrei blowi* Rogl and Bolli, 1973, p. 570, pl. 9, figs. 15-18, 19-20?, 21-22; pl. 17, fig. 12.

*Neogloboquadrina blowi* Rogl and Bolli- Saito, Thompson and Breger, 1981, pl. 37, figs. 1a-d.

*Neogloboquadrina dutertrei* (D'ORBIGNY), 1839

*Globigerina rotunda* d'Orbigny, 1826, p. 27*i*, list no. 6 (nomen nudum)

*Globigerina dutertrei* d'Orbigny, 1839a, p. 84, pl. 4, figs. 19-21 (lectotype).

*Neogloboquadrina dutertrei dutertrei* (d'Orbigny)- Rogl and Bolli, 1973, pl. 9, figs 1-3, 7-10; pl. 17, figs. 1-6.

*Neogloboquadrina dutertrei* (d'Orbigny)- Saito, Thompson and Breger, 1981, pl. 36, figs. 1a-d.

*Neogloboquadrina humerosa* (TAKAYANAGI AND SAITO), 1962

*Globorotalia humerosa* Takayanagi and Saito, 1962, p. 78, pl. 28, fig. 1-2.

*Turborotalia humerosa* (Takayanagi and Saito)- Poag, 1972, p. 512, pl. 2, figs. 9-10.

*Neogloboquadrina humerosa* (Takayanagi and Saito)- Saito, Thompson and Breger, 1981, pl. 37, figs. 2a-d.

*Globigerina cretacea* Brady, 1884 (not d'Orbigny), p. 596, pl. 82, fig. 10.

*Neogloboquadrina dutertrei* (d'Orbigny) subsp. *subcretacea* (Lomniki)- Bandy, Frerichs, and Vincent, 1967, p. 152-157, pl. 4, fig. 2.

*Neogloboquadrina dutertrei blowi* Rogl and Bolli, 1973, p. 570, pl. 9, figs. 15-18, 19-20?, 21-22; pl. 17, fig. 12.

*Neogloboquadrina blowi* Rogl and Bolli- Saito, Thompson and Breger, 1981, pl. 37, figs. 1a-d.

*Neogloboquadrina dutertrei* (D'ORBIGNY), 1839

*Globigerina rotunda* d'Orbigny, 1826, p. 277, list no. 6 (nomen nudum)

*Globigerina dutertrei* d'Orbigny, 1839a, p. 84, pl. 4, figs. 19-21 (lectotype).

*Neogloboquadrina dutertrei dutertrei* (d'Orbigny)- Rogl and Bolli, 1973, pl. 9, figs 1-3, 7-10; pl. 17, figs. 1-6.

*Neogloboquadrina dutertrei* (d'Orbigny)- Saito, Thompson and Breger, 1981, pl. 36, figs. 1a-d.

*Neogloboquadrina humerosa* (TAKAYANAGI AND SAITO), 1962

*Globorotalia humerosa* Takayanagi and Saito, 1962, p. 78, pl. 28, fig. 1-2.

*Turborotalia humerosa* (Takayanagi and Saito)- Poag, 1972, p. 512, pl. 2, figs. 9-10.

*Neogloboquadrina humerosa* (Takayanagi and Saito)- Saito, Thompson and Breger, 1981, pl. 37, figs. 2a-d.

*Neogloboquadrina pachyderma* (EHRENBERG), 1861

*Aristerospira pachyderma* Ehrenberg, 1861, p. 276, 277, 303- Ehrenberg, 1872, pl. 1, fig. 4.

*Globigerina bulloides* d'Orbigny, "Arctic variety", Brady, 1878, p. 435, pl. 21, figs. 10a-c.

*Globigerina pachyderma* (Ehrenberg)- Brady 1884, p. 600, pl. 114, figs. 19-20.

*Neogloboquadrina pachyderma pachyderma* (Ehrenberg)- Rogl and Bolli, 1973, p. 571, pl. 11, figs. 2-6; pl. 16, fig. 12.

*Neogloboquadrina pachyderma* (Ehrenberg)- Saito, Thompson and Breger, 1981, pl. 34, figs. 1a-d.

*Orbulina universa* (D'ORBIGNY), 1839

*Orbulina universa* d'Orbigny, 1839a, p.2, pl. 1, fig. 1

*Orbulina universa* d'Orbigny- Saito, Thompson and Breger, 1981, pl. 19, figs. 1a-6.

*Pulleniatina obliquiloculata* (PARKER AND JONES), 1865

*Pullenia obliquiloculata* Parker and Jones, 1865, p. 183 (nomen nudum)

*Pullenia sphaeroides* (d'Orbigny) var. *obliquiloculata* Parker and Jones, 1965, p. 365, 368; pl. 19, figs. 4a-b.

*Pulleniatina obliquiloculata* (Parker and Jones)- Parker, 1962, p. 234, pl. 4, figs 13-16.

*Pulleniatina obliquiloculata* (Parker and Jones)- Saito, Thompson and Breger, 1981, pl. 31, figs. 3a-3d.

*Sphaeroidinella dehiscens* (PARKER AND JONES), 1865

*Sphaeroidina bulloides* d'Orbigny var. *dehiscens* Parker and Jones, 1865, p. 369, pl. 19, figs 5a-5b

*Sphaeroidina dehiscens* Parker and Jones- Brady, 1884, pl. 84, figs 9-11 (not fig. 8).

*Sphaeroidinella dehiscens* (Parker and Jones) Cushman, 1927, p. 90, pl. 19, fig. 2.

*Sphaeroidinella dehiscens* (Parker and Jones)- Saito, Thompson and Breger, 1981, pl. 20, figs. 2a-2d.

## Appendix D. Data tables.

Table 1. Percentage occurrences of planktonic foraminifera at DSDP Site 604

core section	1-1	1-1	1-3	1-4	2-1	2-1	2-2	3-1	3-1	3-2	5-1	5-1	5-2
Depth in section (cm)	0	140	120	55	32	143	4	48	103	48	124	49	
Depth at site (cm)	0	140	420	505	680	790	804	1655	1710	1764	3578	3654	3730
Individuals/cc	5869.0	50.0	24.0	1.0	0.1	1.0	17.0	7.0	18.0	0.6	19.0	57.0	17.0
<i>Globigerins bulloides</i>	8.2	8.3	2.5			11.0	3.0	1.4	0.5		3.2	0.5	6.5
<i>G. falconensis</i>	0.1												
<i>G. quinquelobe</i>	1.1	3.2	1.7	17.0			16.1	28.4	22.8		21.1	6.8	11.1
<i>G. umbilicata</i>													
<i>Globigerinella aequilateralis</i>													
<i>Globigerinita glutinata</i>	6.8	8.3	2.9				1.2	1.4	2.0		3.7	1.2	7.1
<i>Globigerinoides conglobatus</i>													
<i>G. elongatus</i>													
<i>G. obliquus</i>													
<i>G. pyramidalis</i>								1.4					
<i>G. ruber</i>	1.5	0.8					0.6		0.5			0.9	2.3
<i>G. sacculifer</i>			0.8								2.1		
<i>G. tenellus</i>													
<i>Globorotalia crassaformis</i>	1.0							4.1			3.7		
<i>Globorotalia hirsuta</i>													
<i>G. inflata</i>	28.4	15.9	17.6	17.0		22.0	12.5	23.0	8.8	33.0	15.3	14.4	10.0
<i>G. menardii</i>	0.9	0.8											
<i>G. scitula</i>								1.4					
<i>G. tosaensis</i>													
<i>G. truncatulinoides</i>			0.4										
<i>G. tumida</i>													
<i>Neogloboquadrina</i> Mowbr.													
<i>Neogloboquadrina dutertrici</i>	0.5	2.4	0.4				1.2						0.6
<i>N. humerosa</i>													
<i>N. pachyderma</i> (dextral)	48.4	56.3	66.4	67.0	100.0	33.0	31.3	32.4	59.8	67.0	42.1	26.5	23.5
<i>N. pachyderma</i> (sinistral)	3.1	3.6	6.3			33.0	15.5	2.7	5.4		8.4	50.0	33.5
<i>Orbulina universa</i>		0.4	0.8										
<i>Pulleniatina obliquiloculata</i>											0.5		
<i>Sphaeroidinella dehiscens</i>													
subtropical spp.	0.1	0.0	0.4	0.0	0.0	0.0	0.0	1.4	0.0	0.0	0.0	0.0	0.0
tropical spp.	1.5	0.8	0.8	0.0	0.0	0.0	0.6	0.0	0.5	0.0	2.1	0.9	2.3
gyre margin spp.	1.4	3.2	0.4	0.0	0.0	0.0	1.2	0.0	0.0	0.0	0.5	0.0	0.6
PreQuaternary													
%>63µm		8.0	11.0	8.0	6.0	27.0	18.0		3.0	12.0		18.0	2.0
%CaCO <sub>3</sub>	95.0	5.0	5.0	2.0	2.0	2.0	2.0	2.0	40.0	1.0	20.0	3.0	10.0
%Terrigenous	5.0	90.0	85.0	98.0	95.0	98.0	85.0	90.0	60.0	99.0	75.0	90.0	65.0
%Glauconite		5.0	10.0		3.0		13.0	8.0			5.0	5.0	25.0
%Biogenic silica													
sin:dex	0.1	0.1	0.1	***	***	1.0	0.5	0.1	0.1	***	0.2	1.9	1.4

## Appendix D. Data tables.

Table 1. Percentage occurrences of planktonic foraminifera at DSDP Site 604

core section	5-2	6-1	6-1	6-2	6-2	6-3	6-4	6-5	7cc	8-1	8-1	8-2	9-1
Depth in section (cm)	104	49	136	49	140	16	84	16		32	140	24	140
Depth at site (cm)	3784	4536	4626	4685	4780	4806	4924	5106	5450	6440	6550	6584	7510
Individuals/cc	28.0	19.0	0.0	14.0	24.0	26.0	142.0	296.0	7.0	90.0	1.0	68.0	84.0
<i>Globigerina bulloides</i>	0.7	5.1		18.2	11.9	13.3	6.8	2.8	12.7	8.7	9.1	6.4	2.8
<i>G. falconensis</i>								0.5					
<i>G. quinqueloba</i>	4.3	18.4		7.3	2.5	8.0		8.4	8.5	30.7		5.6	7.5
<i>G. unilicata</i>													
<i>Globigerinella sequilateralis</i>										0.1			
<i>Globigerinita glutinata</i>	7.1	3.1		7.3	11.0	19.0	4.5	3.2	5.6	3.1		8.2	6.0
<i>Globigerinoides conglobatus</i>													0.1
<i>G. elongatus</i>													
<i>G. obliquus</i>													
<i>G. pyramidalis</i>													
<i>G. ruber</i>	0.4	4.1		6.6	9.3	9.5		10.4	7.1	0.1		0.9	1.2
<i>G. sacculifer</i>						1.9	0.3	1.5		0.2			0.8
<i>G. tenellus</i>													
<i>Globorotalia crassaformis</i>				1.4	3.4	4.9	0.6	2.8	4.2			0.3	0.7
<i>Globorotalia hirsuta</i>													
<i>G. inflata</i>	8.9	9.2		15.3	26.3	17.1	8.5	20.2	14.1	5.0		7.3	6.9
<i>G. menardii</i>													
<i>G. scitula</i>													
<i>G. tosacensis</i>													
<i>G. truncatulinoides</i>													
<i>G. tumida</i>							0.3						
<i>Neogloboquadrina blowi</i>						1.5							
<i>Neogloboquadrina dutertrei</i>					1.5			0.9		0.7		0.9	0.2
<i>N. humerosa</i>													
<i>N. pachyderma (dextral)</i>	62.9	37.8		32.8	29.1	20.2	73.0	24.8	35.2	48.4	45.5	47.7	47.9
<i>N. pachyderma (sinistral)</i>	15.7	22.5		10.9	3.8	3.8	6.8	24.0	19.7	2.9	45.5	21.2	25.5
<i>Orbulina universa</i>													0.7
<i>Pulleniatina obliquiloculata</i>					0.8			0.3					0.1
<i>Sphaeroidinella debiscens</i>													
subtropical spp.	0.0	0.0		0.0	0.0	0.0	0.0	0.5	0.0	0.0	0.0	0.0	0.0
tropical spp.	0.4	4.1		6.6	9.3	11.4	0.3	11.9	7.1	0.3	0.0	0.9	2.0
gyre margin spp.	0.0	0.0		0.0	2.3	0.0	0.3	1.2	0.0	0.7	0.0	1.0	0.2
PreQuaternary													
% > 63um	11.0	0.5	27.0		5.0	3.0	35.0	6.0	1.0	0.5	20.0	6.0	12.0
% CaCO <sub>3</sub>	10.0	10.0	5.0	20.0	5.0	10.0	10.0	20.0	10.0	35.0	1.0	25.0	5.0
% Terrigenous	30.0	85.0	10.0	75.0	85.0	75.0	80.0	65.0	85.0	60.0	24.0	55.0	50.0
% Glauconite	60.0	5.0	85.0	5.0	10.0	15.0	10.0	15.0	5.0	5.0	75.0	30.0	45.0
% Biogenic silica													
sin index	0.3	0.6	***	0.3	0.1	0.2	0.1	1.0	0.6	0.1	1.0	0.4	0.5

## Appendix D. Data tables.

Table 1. Percentage occurrences of planktonic foraminifera at DSDP Site 604

core section	9-2	10-4	11-1	11-2	11-3	11-4	11-5	12-1	12-1	12-2	13-1	13-2	13-2
Depth in section (cm)	131	115	132	140	140	110	140	32	140	9	140	32	143
Depth at site (cm)	7660	8895	9422	9580	9730	9580	10030	10280	10390	10490	11350	11390	11500
Individuals/cc	70.0	23.0	67.0	14.0	275.0	960.0	953.0	742.0	1153.0	125.0	125.0	184.0	54.0
<i>Globigerina bulloides</i>	5.1	2.6	2.4	2.9	5.4	5.9	11.1	9.2	4.6		9.6	2.0	0.9
<i>G. falconensis</i>	0.4										0.6	0.4	
<i>G. quinqueloba</i>	2.3	7.0	4.2	5.0	10.9	4.6	3.7	3.0	6.5	1.9	7.0	8.7	1.5
<i>G. umbilicata</i>													
<i>Globigerinella aequilateralis</i>							0.7		0.1		0.2	0.2	
<i>Globigerinita glutinata</i>	3.7	8.3	3.3	0.7	5.7	4.6	9.1	1.1	3.9	5.4	10.1	1.5	3.0
<i>Globigerinoides conglobatus</i>													
<i>G. elongatus</i>				0.7		0.3					0.4		
<i>G. obliquus</i>								0.2					
<i>G. pyramidalis</i>													
<i>G. ruber</i>	1.1	1.3	4.1	10.1	0.3	1.7	3.4	1.1	1.9	0.3	1.1	1.1	1.7
<i>G. sacculifer</i>	0.7		3.3	5.0			0.7	0.2	0.6		0.2		0.4
<i>G. tenellus</i>													
<i>Globorotalia crassaformis</i>			0.5		0.3	0.7		0.4	1.5				0.4
<i>Globorotalia hirsuta</i>													
<i>G. inflata</i>	5.7	11.8	8.1	12.9	5.8	10.0	11.4	17.2	7.8	16.4	8.7	24.3	20.4
<i>G. menardii</i>									0.1				
<i>G. sciuta</i>				0.7		0.7	0.4				0.4	0.2	
<i>G. tosaensis</i>													
<i>G. truncatulinoides</i>			1.1	0.7			0.4						
<i>G. tumida</i>			0.8	1.7	0.3			0.6					
<i>Neogloboquadrina blowi</i>											2.7		
<i>Neogloboquadrina dutertrei</i>	1.4		0.2	4.3	0.7	0.7		0.8	0.1	0.6	0.3		
<i>N. humerosa</i>								0.2					
<i>N. pachyderma (dextral)</i>	55.0	30.7	58.5	49.6	47.9	48.3	46.3	61.6	46.4	43.9	64.1	58.8	66.7
<i>N. pachyderma (sinistral)</i>	25.3	29.4	13.7	5.8	23.2	16.7	11.7	4.3	26.2	30.4	1.1	2.2	0.9
<i>Orbulina universa</i>						0.7			0.2			0.6	0.6
<i>Pulleniatina obliquiloculata</i>									0.1				
<i>Sphaeroidinella debiscens</i>								0.2					
subtropical spp.		0.0	1.1	1.4		0.7	0.8	0.0	0.0	0.0	1.0	0.6	0.0
tropical spp.		1.3	7.4	15.1		1.7	4.1	1.3	2.5	0.3	1.3	1.1	2.1
gyre margin spp.		0.0	1.0	6.0		0.7	0.0	1.4	0.3	0.6	0.3	0.0	0.0
PreQuaternary													
%>63um	10.0	6.0	1.0	2.0	2.0	42.0	25.0	2.0	35.0	38.0	6.0		45.0
%CaCO <sub>3</sub>	10.0		35.0	25.0	20.0	20.0	30.0	30.0	15.0	10.0	25.0	25.0	65.0
%Terrigenous	25.0		45.0	60.0	55.0	20.0	65.0	35.0	45.0	45.0	65.0	65.0	20.0
%Glauconite	65.0		20.0	15.0	25.0	60.0	5.0	15.0	40.0	45.0	10.0	10.0	15.0
%Biogenic silica													
sin:dex	0.5	1.0	0.2	0.1	0.5	0.3	0.3	0.1	0.6	0.7	0.0	0.0	0.0

## Appendix D. Data tables.

Table 1. Percentage occurrences of planktonic foraminifera at DSDP Site 604

core section	13-4	14-2	14-4	14-5	15-3	16-1	16-2	17-1	17-2
Depth in section (cm)	83-86	140	143	140	3	140	140	140	124
Depth at site (cm)	11745	12460	12760	12910	13432	14230	14380	15190	15324
Individuals/cc	16.0	37.0	620.0	1430.0	750.0	24.0	30.0	35.0	43.0
<i>Globigerina bulloides</i>	3.0	1.6	2.1	4.3	9.2		0.3	0.0	1.6
<i>G. falconensis</i>					4.3				1.3
<i>G. quinqueloba</i>	3.5	0.5	3.7	6.2	1.4		8.1	0.6	
<i>G. umbilicata</i>					0.6				
<i>Globigerinella aequilateralis</i>									0.3
<i>Globigerinita glutinata</i>	2.8	2.9	1.4	3.0	12.4	0.4	6.4	18.1	2.8
<i>Globigerinoides conglobatus</i>				0.1	0.2		0.3	0.3	0.2
<i>G. elongatus</i>	0.8			0.2					
<i>G. obliquus</i>							1.0	0.8	
<i>G. pyramidalis</i>					0.2				1.4
<i>G. ruber</i>	10.6	0.6	1.2	0.9	2.2			4.8	
<i>G. sacculifer</i>				0.7	0.4	0.4	2.4		
<i>G. tenellus</i>									0.2
<i>Globorotalia crassaformis</i>					1.4	0.4		2.0	0.2
<i>Globorotalia hirsuta</i>									
<i>G. inflata</i>	8.6	36.7	12.4	7.6	11.2		11.8	26.6	8.3
<i>G. menardii</i>					0.2				
<i>G. scitula</i>									
<i>G. tosensis</i>								0.6	
<i>G. truncatulinoides</i>	0.5							0.6	
<i>G. tumida</i>					0.2				
<i>Neogloboquadrina blowi</i>	7.8			4.0		8.8			
<i>Neogloboquadrina dutertrei</i>			0.7	0.2	3.5	0.8		0.6	1.4
<i>N. humerosa</i>				0.4	3.7				
<i>N. pachyderma (dextral)</i>	12.2	55.0	34.1	69.3	46.4	4.6	3.7	46.3	83.6
<i>N. pachyderma (sinistral)</i>	55.3	2.9	41.4	3.8	1.6	84.4	65.5	7.3	1.6
<i>Orbulina universa</i>	2.5				0.5		0.3		
<i>Pulleniatina obliquiloculata</i>	0.3								
<i>Sphaerolinella dehiscens</i>	0.3			0.1	0.1			1.1	
subtropical spp.	0.5	0.0	0.0	0.0	4.3	0.0	0.0		0.0
tropical spp.	10.6	0.6	1.2	1.6	2.6	0.4	2.4	1.8	0.0
gyre margin spp.	0.3	0.0	0.7	0.2	3.9	0.8	0.0		1.4
PreQuaternary									
%>63µm	8.0	8.0	5.0			1.0	0.5		
%CaCO <sub>3</sub>	25.0	40.0	70.0	90.0	85.0	90.0	60.0	75.0	
%Terrigenous	45.0	30.0	30.0	5.0	10.0	10.0	35.0	5.0	
%Glauconite	30.0	30.0		5.0	5.0		5.0	20.0	
%Biogenic silica									
index	4.5	0.1	1.2	0.1	0.0	18.3	17.7	0.2	0.0



## Appendix D. Data tables.

Table 2. Percentage occurrences of dinoflagellate cysts at DSDP Site 604

Core section	6-3	6-4	6-5	7cc	8-1	8-2	9-1	9-2	10-4	11-1	11-3	11-5
Depth in section	19	87	11		143	27	143	134	118	135	143	143
Depth (cm)	4809	4927	5101	5500	6550	6587	7513	7663	8898	9425	9733	9583
Individuals/cc	260.0	400.0	2349.0	400.0	790.0	567.0	1153.0	309.0	220.0	165.0	305.0	825.0
<i>A. andalousiense</i>			3.0		1.8	2.6			0.5	1.6		
<i>B. tepikiense</i>	9.5	4.5	16.4	2.7	1.9	3.9	1.1	5.6	36.4	23.0	19.6	24.6
<i>Brigantidium</i> sp.	78.6	32.9	55.2	49.5	11.7	9.5	22.0	58.4	28.0	26.2	35.3	14.9
c.f. <i>Cerebrocysta</i> sp.									0.5			
<i>C. cantharellum</i>												
<i>C. barlandi</i>												
<i>F. filifera</i>					1.9					1.6		
c.f. <i>Homotribium</i>												
c.f. <i>Hystriocholopoma</i> sp.												
<i>Impagidium</i> sp. indet			1.5		6.5	1.6		1.1				
<i>I. aculeatum</i>												
<i>I. patulum</i>					1.8			1.1				
<i>I. lachrymosa</i> ?			0.5									
<i>L. machaerocarpum</i>	2.4	3.0								3.2		
<i>Michrystidium</i> sp.		8.3	3.5	0.5	0.5		2.2		0.5	1.5	2.0	0.5
? <i>M. minuta</i>							0.5	1.1	4.7	3.2	3.9	1.8
? <i>M. quanta</i>				0.5		1.6	3.3				2.0	0.5
<i>N. labyrinthus</i>		7.5	0.5			0.5		5.6	0.5	1.6		2.6
<i>O. centrocarpum</i>	3.6	6.0	4.6	10.8	15.5	7.9	3.3	5.6	1.9	8.2		26.3
<i>O. israelianum</i>				6.3	19.4	37.0	16.5		3.7	8.2		
<i>Quinquecuspis</i> sp.												
<i>Spiniferites</i> sp. indet	2.4	19.4	7.8	0.5	14.8	12.6	7.7	8.9	2.8		11.8	9.6
<i>Spiniferites bentorii</i>									1.9			
<i>Spiniferites elongatus</i>		1.5										
<i>S. c.f. membranaceus</i>												
<i>S. mirabilis</i>		3.0	0.9	4.5	3.9	3.2	2.2	1.1	0.5	1.6		2.6
<i>S. ramosus</i>	1.2	9.0		2.7	0.5	0.5	7.7	3.4	0.5	6.5	7.8	6.1
<i>S. splendidus</i>												
<i>S. pseudofurcatus</i>				1.8	1.3		4.4	3.4	2.8	4.9	15.7	4.4
<i>T. pellitum</i>		7.5		16.2	17.4		17.6		3.7	4.9		4.4
<i>S. nephroidea</i>			3.5				5.5	3.4	0.5			
<i>S. quanta</i>												
<i>Selenopemphix</i> Head et al.												
G:P	0.3	2.0	0.7	1.0	7.5	7.5	2.1	0.6	2.0	2.4	1.5	4.8
dino:pollen	0.2	0.1	0.1	0.3	0.9	0.7	0.8	0.3	0.2	0.5	0.5	0.5

## Appendix D. Data tables.

Table 2. Percentage occurrences of dinoflagellate cysts at DSDP Site 604

Core section	12-1	13-1	13-3	14-2	14-4	15-1	16-4
Depth in section	143	143	135	143	145	143	70
Depth (cm)	10393	11350	11645	12463	12762	13273	14610
Individuals/cc	70.0	107.0	512.0	1123.0	1090.0	986.0	1243.0
<i>A. andalusiense</i>		2.9	0.5	0.0	0.7	3.8	2.9
<i>B. tepikiense</i>	16.1	4.4	2.0	0.1	0.0	10.0	24.4
<i>Brigantidium</i> sp.	12.9	39.8	31.4	0.1	0.4	12.5	3.9
c.f. <i>Cerebrocysts</i> sp.							
<i>C. cantharellum</i>	1.1						1.0
<i>C. barlandi</i>						1.3	0.5
<i>F. filifera</i>			2.9	0.0			3.9
c.f. <i>Homotribium</i>							
c.f. <i>Hystriocholopoma</i> sp.							
<i>Impagidium</i> sp. indet		4.4				1.3	1.0
<i>I. aculeatum</i>							
<i>I. patulum</i>							
<i>I. lachrymosa</i> ?							
<i>L. macrocarpum</i>	1.6					2.5	
<i>Michrystidium</i> sp.		5.9	2.9	1.7		2.5	
? <i>M. minuta</i>		5.9	0.5				
? <i>M. quanta</i>			0.5		0.7		
<i>N. labyrinth</i>							
<i>O. centrocarpum</i>	11.3	14.7	13.7	1.7	0.1	3.8	2.9
<i>O. israelianum</i>	17.7	4.4	22.5	44.8	0.0	22.5	4.4
<i>Quinquecuspidata</i> sp.							
<i>Spiniferites</i> sp. indet	8.1	4.4	22.5	1.7	0.0	30.0	24.4
<i>Spiniferites bentorii</i>							1.9
<i>Spiniferites elongatus</i>			0.5				2.4
<i>S. c.f. membranaceus</i>							
<i>S. mirabilis</i>	1.6		2.0		0.0	1.3	8.3
<i>S. ramosus</i>	9.7	4.4	4.9	1.7	0.0		
<i>S. splendidus</i>				1.7	0.7	1.3	1.9
<i>S. pseudofurcatus</i>	8.1	2.9	2.9	3.4	0.0	10.0	8.8
<i>T. pellitum</i>	9.7	1.5	2.0	20.7	0.1	5.0	4.9
<i>S. nephroides</i>			2.0		0.0		
<i>S. quanta</i>							
<i>Selenopemphix</i> Head et al.							0.5
G:P	6.8	1.2	1.9	13.5	1.5	5.7	24.6
dinospollen	0.1	0.4	0.5	0.8	1.1	0.3	0.7

## Appendix D. Data tables.

Table 3. Percentage occurrences of pollen at DSDP Site 604

core section	1-1	1-2	1-2	1-4	2-1	2-2	3-3	3-2	5-1	5-2	6-1
depth in section	3	48	146	58	143	7	103	7	127	107	3
Depth (cm)	3	208	296	508	790	807	1710	1767	3657	3787	4629
Individuals/cc	450.0	9280.0	2130.0	677.0	766.0	1864.0	1360.0	951.0	1136.0	2839.0	650.0
Abies		0.8	1.4							0.5	
Acer											
Alnus	2.4			0.6	2.5	2.4	0.9	0.8	3.2	2.7	1.5
Artemisia											
Betula	3.2	4.3	2.1	1.8	0.6	3.0	1.4	1.7	1.8	0.5	1.5
Carya	0.8	0.8	0.7		1.3	1.2	0.5		0.9		0.7
Caryophyllaceae											
Cephalanthus											
Chenopodiaceae	0.8			1.2		0.6					0.5
Corylus	0.8			3.7	0.6	1.2	0.9	1.7	1.8	1.6	
Cornus											0.5
Cyperaceae	0.8	0.8	3.5	3.7	6.3		0.9	0.8	4.6	1.1	
Dryas				0.6							
Ericaceae				1.8		0.6		0.7			
Fagus			0.7	0.6		1.8	0.5				0.5
Fraxinus					0.6						
Graminiae	0.8	1.7		1.2	3.1	0.6	0.5		2.8	1.1	1.5
Iva											
Juglans											
Liquidambar											
Myrica											
Ostrya		0.8	0.7		0.6	1.2			0.9		1.5
Platanus				0.6							0.5
Picea	4.8	21.6	14.8	20.5	16.4	13.3	26.0	23.7	22.0	15.1	18.2
Pinus	58.1	51.7	68.3	55.9	55.8	45.5	56.7	63.6	48.2	59.1	62.0
Quercus	19.4	6.1	4.9	4.3	9.1	21.8	7.4	6.8	8.3	8.1	8.0
Rosaceae											
Rumex											
Salix	0.8	0.8		1.2		1.2	0.5		0.5	0.5	
Tilia					0.6						0.7
Tsuga	2.4					0.6		0.8	0.5	0.5	0.7
Tubuliflorae				1.2			1.9	1.7		0.5	
Ulmus	0.8			0.6	0.6	1.2					
Umbelliferae											
Spores:											
Adiantum				0.6			0.9				
Dryopteris	2.4	2.6		1.2	1.3	1.8	0.5	2.5	1.4	1.6	
Isetes	0.8										
Lycopodium	0.8	1.7	1.4	5.0	2.5		0.9	3.4	2.8	4.8	1.5
Osmunda	0.8			0.6		1.2				1.1	
Pteridium	4.0	14.7	3.5	3.1	5.7	1.2	0.9	5.9	2.8	0.0	5.1
Sphagnum			0.7		1.2						1.5



## Appendix D. Data tables.

Table 3. Percentage occurrences of pollen at DSDP Site 604

core section	11-3	11-5	12-1	13-1	13-3	14-2	14-4	15-1	16-4
depth in section	143	143	143	143	135	143	145	143	70
Depth (cm)	9733	9583	10393	11350	11645	12463	12762	13273	14610
Individuals/cc	602.0	1763.0	518.0	297.0	1038.0	1434.0	995.0	3055.0	1785.0
Abies		0.6	0.7			0.0		1.9	
Acer	1.4			1.0					
Alnus				1.9					0.6
Artemisia									
Betula	0.7			2.9			0.8	0.6	
Carya				1.0		1.7	0.8		
Caryophyllaceae									
Cephalanthus									
Chenopodiaceae						0.9	0.8		
Corylus	1.4	1.9		2.9	0.7		1.6		1.2
Comus									
Cyperaceae		0.6	1.4		1.4	0.9			
Dryas									
Ericaceae	0.7					0.9	0.8		
Fagus	0.7			1.0					
Fraxinus									
Graminac		1.3		3.8	0.7	1.7	2.4	1.9	
Iva									
Juglans					0.7				
Liquidambar									
Myrica									
Ostrya				1.9	1.4		0.8		
Platanus									
Picea	16.0	8.3	9.3	6.7	6.4	6.0	11.0	6.3	3.8
Pinus	60.4	76.3	77.1	61.0	75.4	70.9	63.0	85.4	90.5
Quercus	13.9	9.6	11.4	11.4	10.6	11.1	15.0	2.5	1.9
Rosaceae									
Rumex	0.7								
Salix	0.7								
Tilia	0.7								
Tsuga		0.6		1.0	0.7	0.9	0.8	1.3	
Tubuliflorae	0.7					0.9	1.6		
Ulmus	0.7					0.9			0.6
Umbelliferae				1.0			0.8		
Spores:									
Adiantum	0.7				0.7				0.6
Dryopteris	1.4		1.4	1.9			2.4	1.3	1.9
Isaetes									0.6
Lycopodium	0.7	1.9		1.0	1.4		0.8	1.3	1.9
Osmunda	0.7	0.6	1.4			0.9		0.6	
Pteridium	5.6	1.3	1.4		7.0	4.3		1.3	2.5
Sphagnum	2.1	1.3				2.6			

## Appendix D. Data tables.

Table 4. Percentage occurrences of planktonic foraminifera at DSDP Site 607

Core section	1-3	1-3	1-3	1-3	1-4	1-4
Depth in section (cm)	31	62	91	122	3	31
Depth at site (cm)	331	362	391	422	453	481
Individuals/g	4224.0	5312.0	3304.0	4134.0	4624.0	2936.0
<i>Globigerina bermudezi</i>						
<i>Globigerina bulloides</i>	15.0	11.2	10.2	15.2	18.2	22.6
<i>G. cariacensis</i>						
<i>G. falconensis</i>	1.5	14.2	5.6	6.4	6.9	1.4
<i>G. quinqueloba</i>	9.1	11.7	10.2	3.1	9.7	2.7
<i>G. rubescens</i>		1.2	1.9	1.6	0.4	0.8
<i>G. umbilicata</i>			0.7			
<i>Globigerinella aequilateralis</i>	0.2	0.6	0.2	1.9	0.9	0.5
<i>Globigerinita glutinata</i>	15.2	7.2	10.4	9.4	7.4	7.1
<i>Globigerinoides conglobatus</i>			0.5	0.2	0.5	
<i>G. elongatus</i>	1.7	3.0	0.5	1.4	1.4	0.5
<i>G. pyramidalis</i>	0.4					
<i>G. ruber</i>	0.4	3.6	3.1	5.7	1.9	5.5
<i>G. saeculifer</i>				0.6		0.3
<i>G. tenellus</i>		0.6	0.2	0.2		
<i>Globoquadrina conglomerata</i>						
<i>Globorotalia crassaformis</i>	4.4	5.4	3.1	5.3	3.5	5.5
<i>G. crassula</i>			0.2			3.3
<i>G. hirsuta</i>		1.5		1.6	1.2	0.5
<i>G. inflata</i>	10.8	4.5	10.9	13.2	9.0	15.5
<i>G. scitula</i>	5.3	5.7	5.6	4.7	2.4	4.4
<i>G. tumida</i>				2.1	2.4	1.1
<i>G. truncatulinoides</i>	1.9	2.1		2.5	2.1	5.5
<i>G. unguolata</i>				1.9	0.4	
<i>Neocarinina blowi</i>						
<i>Neogloboquadrina dutertrei</i>				0.8	0.9	0.3
<i>N. pachyderma (dextral)</i>	18.0	12.7	20.8	10.5	16.6	17.7
<i>N. pachyderma (sinistral)</i>	4.9	1.2	1.7	0.8	4.7	
<i>Orbulina universa</i>						
tropical spp.	0.4	3.6	3.1	6.3	1.9	5.8
gyre	0.0	0.0	0.0	2.9	3.3	1.4
subtropical spp.	8.7	23.5	11.2	15.2	12.6	11.8
index	0.3	0.1	0.1	0.1	0.3	0.0

## Appendix D. Data tables.

Table 4. Percentage occurrences of planktonic foraminifera at DSDP Site 607

Core section	1-1	1-1	1-1	1-1	1-1	1-2	1-2	1-2	1-2	1-2	1-3
Depth in section (cm)	3	31	62	91	122	3	31	62	91	122	3
Depth at site (cm)	3	31	62	91	122	153	181	212	241	272	303
Individuals/g	6336.0	2811.0	3221.0	4267.0	3704.0	4776.0	3400.0	2696.0	5632.0	5128.0	3184.0
<i>Globigerina bermudezi</i>								0.5	0.6		
<i>Globigerina bulloides</i>	16.2	33.4	19.9	22.8	35.6	24.3	12.9	17.2	23.9	20.0	14.8
<i>G. cariocensis</i>								0.8	1.2	0.2	
<i>G. falconensis</i>	10.4		0.3		0.6	2.0	2.3	3.6	3.1	2.8	3.6
<i>G. quinqueloba</i>	2.5	3.6	14.6	14.3	16.2	10.7	15.5	10.1	11.0	10.0	2.5
<i>G. rubescens</i>	0.3										
<i>G. umbilicata</i>								0.2		0.6	
<i>Globigerinella aequilateralis</i>	1.0							0.7	0.9	0.2	0.8
<i>Globigerinita glutinata</i>	8.6	8.5	5.0	7.5	11.2	9.4	13.2	15.4	10.8	6.9	5.5
<i>Globigerinoides conglobatus</i>	0.3										
<i>G. elongatus</i>	1.0	0.9						2.1		0.7	1.6
<i>G. pyramidalis</i>	0.5								0.9		
<i>G. ruber</i>	11.6	3.6	2.6	1.3		3.0	2.1	3.6	2.1	2.7	2.5
<i>G. sacculifer</i>	0.8			0.3		0.2	0.2			0.2	
<i>G. tenellus</i>	0.3			0.3		0.3	0.5	0.3			
<i>Globoquadrina conglobata</i>								0.0	0.9		
<i>Globorotalia crassaformis</i>	3.5	0.6		1.5	0.9	2.2	3.1	4.2	2.7	0.8	1.6
<i>G. crassaula</i>	0.8										
<i>G. hirsuta</i>	1.8						0.5		0.2		
<i>G. inflata</i>	10.9	16.9	10.6	9.0	3.9	4.4	7.3	6.8	3.1	3.9	22.5
<i>G. scitula</i>	3.5	2.7	5.3	3.8	3.0	4.4	3.3		6.0	4.8	1.9
<i>G. tumida</i>	2.0										
<i>G. truncatulinoides</i>	5.1	0.2				0.1	0.5	1.5	0.6	0.5	1.1
<i>G. unguolata</i>	1.0										
<i>Neocarinina blowi</i>										0.5	
<i>Neogloboquadrina dutertrei</i>	0.8				0.2	0.3		1.2		0.3	1.4
<i>N. pachyderma (dextral)</i>	8.1	23.9	28.8	22.8	12.3	17.1	12.9	18.4	18.8	16.5	38.9
<i>N. pachyderma (sinistral)</i>	1.3	3.6	11.3	12.3	13.6	13.1	9.4	3.0	11.5	23.4	3.0
<i>Orbulina universa</i>	0.5		0.3		0.2	0.3		5.9			1.1
tropical spp.	12.4	3.6	2.6	1.6	0.0	3.2	2.3	3.6	2.1	2.9	2.5
gyre	2.8	0.0	0.0	0.0	0.2	0.3	0.0	1.2	0.0	0.3	1.4
subtropical spp.	20.8	2.9	5.6	3.8	3.6	6.5	6.6	5.1	9.9	8.1	6.6
index	0.2	0.2	0.4	0.5	1.1	0.8	0.7	0.2	0.6	1.4	0.1

## Appendix D. Data tables.

Table 5. Percentage occurrences of dinoflagellate cysts at DSDP Site 607

Core section	1-1	1-1	1-1	1-1	1-1	1-2	1-2	1-2	1-2	1-2	1-3	1-3
Depth in section	3	33	63	93	123	3	33	63	93	123	3	33
Depth	3	31	62	91	122	153	181	212	241	272	303	331
Total/cc	2	248	700	200	103	63	92	52		19	104	40
total/cc (lycop)			4475					180				
correction factor			6.4					3.5				
<i>A. andalousiense</i>												5
<i>B. tepikiense</i>										2.1	29.9	
<i>Brigantedinium</i> spp.			89.7	9.6	28.8	7	22.4	10		23.9		30
? <i>Cannosphaeropsis</i> sp.		1.8										
<i>Glenodinium</i> sp.												
cf <i>Hystrieholpoma</i>		6.3	2.1	16.1	7.2	25.6	7.5	0.9		4.3		
<i>I. aculeatum</i>	50	22.5	1	14.5	9.3	4.7	14.9	23.6	22.7	19.6	6	15
<i>I. aliferum</i>										2.1	3	5
<i>I. multiplexum</i>												
<i>I. paradoxum</i>											1.5	5
<i>I. patulum</i>		6.3		8.8	16.5	9.3	6	15.5	27.3	10.9	6	15
<i>I. striatum</i>	50	0.9					1.5	1.8	9.1	4.3		
<i>L. truncatum</i>												5
<i>L. machaerocarpum</i>			1		1						1.5	5
<i>Michrytridium</i> sp.		2.7	3.1	8.8	3.1	14	3	1.8		4.3		
? <i>M. minuta</i>		0.9	1			2.3	1.5	0.9				
? <i>M. quanta</i>												
<i>M. sp. 1 of Mudie 1986</i>					2.1			0.9				
<i>N. labyrinthus</i>		49.5		19.4	3.1	9.3	23.9	18.2	13.6	2.1	13.4	
<i>N. oblonga</i>					1							
<i>N. sp. 1</i>					2.1							
<i>O. centrocarpum</i>			2.1	22.6	14.4	27.9	6	20	4.5	21.7	20.9	
<i>O. israelianum</i>		2.7			3.1							
<i>O. longispermum</i>												5
<i>O. sp. of Piasecki</i>					1							
<i>O. sp. of Jan du Chene</i>					1							
<i>P. zoharyi</i>											1.5	
<i>S. bentorii</i>												
<i>S. elongatus</i>										4.3		
<i>S. mirabilis</i>												3
<i>S. ramosus</i>		1.8		1.6	1							
<i>T. pellitum</i>					3.1		7.5	5.5				
<i>T. simplex</i>					2.1		3					
<i>Z. lenticulatum</i>							3					
Impagidinium index	1	0.79	1	0.62	0.36	0.34	0.73	0.62	?	0.67	0.36	0.38
Ocentron:NIaby		0	*	1.2	4.6	3	0.25	1.1	0.33	10.3	1.6	0

## Appendix D. Data tables.

Table 5. Percentage occurrences of dinoflagellate cysts at DSDP Site 607

Core section	1-3	1-3	1-3	1-4	1-4
Depth in section	63	93	123	3	33
Depth	362	391	422	453	481
Total/cc	92	12	10	50	20
total/cc (lycop)	148			300	
correction factor	1.6			6	
<i>A. andalusiense</i>	5			1.4	
<i>B. tepikiense</i>	1.7				
<i>Brigantedinium</i> spp.	26.8	26.7		38.6	
? <i>Cannosphaeropsis</i> sp.					
<i>Glenodinium</i> sp.		12.5			
cf <i>Hystriehokolpoma</i>	5			5	6.7
<i>I. aculeatum</i>	15	40	25	15.7	60
<i>I. aliferum</i>				1.4	
<i>I. multiplexum</i>	1.7				
<i>I. paradoxum</i>	1.7				
<i>I. patulum</i>	5	6.7		4.3	
<i>I. striolatum</i>	3.4			2.8	13.3
<i>L. truncatum</i>					
<i>L. machaerocarpum</i>		6.7			
<i>Melchrytidium</i> sp.	3.4			2.9	
<i>M. minuta</i>					
<i>M. quanta</i>				2.1	
<i>M. sp. 1 of Mudie 1986</i>					
<i>N. labyrinthus</i>	20			10	13.3
<i>N. oblonga</i>					
<i>N. sp. 1</i>					
<i>O. centrocarpum</i>	5	6.7		7.2	
<i>O. israelianum</i>					
<i>O. longispermum</i>					
<i>O. sp. of Plaseki</i>					
<i>O. sp. of Jan du Chene</i>					
<i>P. zobaryie</i>					
<i>S. bentonii</i>					
<i>S. elongatus</i>	1.7			0.7	
<i>S. mirabilis</i>		6.7			
<i>S. ramosus</i>				0.7	
<i>T. pellitum</i>	1.7				
<i>T. simplex</i>					
<i>Z. lenticulatum</i>					
Impagidinium index	0.58	0.86	1	0.76	1
Ocentro:NIaby	0.25	*	0	0.72	*



## Appendix D. Data tables.

Table 6. Percentage occurrences of pollen at DSDP Site 607

Core section	1-3	1-3	1-3	1-4	1-4
Depth in section	63	93	123	3	30
Depth	363	393	423	453	483
Pollen grains/cc	70.7	11.4	0	12.2	25.6
Abies					
Alnus				28.6	
Betula					
Carya					
Fagus					
Gramineae				14.3	
Ostrya					
Picea		66.7			
Pinus	100	33.3		57.2	100
Quercus					
Tsuga					
Ulmus					
Dryopteris	100				
Lycopodium				28.6	
Pteridium					
Osmunda					

## Appendix D. Data tables.

Table 7. Stable isotope ratios from DSDP Site 607

Depth Depth (cm)	O-18 O-18	C-13 C-13	O-18 O-18	C-13 C-13
	dextral	dextral	sinistra	sinistra
3	0.963	0.298		
31	2.305	-0.005		
62	1.998	-0.064	2.778	-0.528
91	1.691	-0.007		
122	2.091	0.412	3.199	-0.4
155	1.915	0.343	2.211	-0.161
181	1.833	0.24	2.974	-0.107
212	1.829	0.419		
241	2.074	0.629	2.404	-0.579
272	1.736	-0.045	2.936	-0.835
303	2.219	0.344	2.608	-0.189
331	1.655	0.721		
362	1.193	0.736		
391	1.531	0.415		
422	1.503	0.376		
453	1.972	-0.024		
481	1.149	0.127		

**Appendix D. Data tables.**

Table 8. Sea surface temperatures using the Imbrie-Kipp method with the transfer functions of Mudie (1990)

Core section	1-1	1-1	1-1	1-1	1-2	1-2	1-2	1-2	1-3	1-3	1-4
Depth in section	33	63	93	123	3	33	63	123	3	63	3
Depth (cm)	33	63	93	123	153	183	213	273	303	363	453
SST (sum)	19.7	19.3	21.1	16.9	21.6	16.5	20	18.4	21.9	19.6	17
SST (win)	11.6	6.1	10.5	4.4	10.1	5.7	10	6.8	12.3	7.9	4.2

## Appendix D. Data tables.

Table 9. Percentage occurrences of planktonic foraminifera at DSDP Site 612

core section	4-4	4-5	4-5	4-6	5-1	5-1	5-2	5-2	5-3	8-2
Depth in section (cm)	80	32	80	30	0	47	34	80	78	
Depth at site (cm)	2930	3032	3080	3130	3360	3407	3534	3584	3606	5493
Individuals/cc	140.0	77	20.0	70.0	67.0	35.0	30.0	26.0	4.0	
<i>Globigerina bulloides</i>	35.6	8.1	16.9	14.5	29.5	21.2	31.8	15.2	14.3	14
<i>G. falconensis</i>										
<i>G. quinqueloba</i>	4.9	1.8	0.5	6.7	3.3	2.9	1.8	12.5		
<i>G. umbilicata</i>										
<i>Globigerinella aequilateralis</i>		0.1			0.3		0.5			2
<i>Globigerinita glutinata</i>	0.3	5	1.0	8.5	1.2	2.9	3.2	7.2	7.1	5
<i>Globigerinoides conglobatus</i>										
<i>G. obliquus</i>										1
<i>G. pyramidalis</i>										
<i>G. ruber</i>		0.5	1.5			1.5	1.4	0.4		3
<i>G. sacculifer</i>									4.8	
<i>G. tenellus</i>										
<i>Globorotalia crassaformis</i>		0.4			0.3				4.8	2
<i>Globorotalia hirsuta</i>						1.5				
<i>G. inflata</i>	2.0	28.7	21.9	3.2	8.9	13.3	6.1	6.1		41
<i>G. menardii</i>										
<i>G. scitula</i>				1.7						
<i>G. tosacensis</i>		0.1								
<i>G. tumida</i>										
<i>G. truncatulinoides</i>		0.1								
<i>Neogloboquadrina dutertrei</i>					0.9					1
<i>N. humerosa</i>		0.4								
<i>N. pachyderma (dextral)</i>	42.5	52.5	6.0	29.8	39.9	47.8	33.6	28.4	54.8	8
<i>N. pachyderma (sinistral)</i>	17.2	1.8	52.2	34.9	13.4	8.7	23.2	26.9	2.4	6
<i>Orbulina universa</i>	0.3			0.1		0.6			2.4	
<i>Pulleniatina obliquiloculata</i>		0.1			0.3			0.4	2.4	7
<i>Sphaeroidinella debiscens</i>										1
subtropical spp.	0.0	0.4	0.0	0.0	0.3	1.5	0.0	0.0	4.8	2
tropical spp.	0.0	0.5	0.0	0.0	0.0	1.5	1.4	0.4	4.8	3
gyre margin spp.	0.0	0.2	0.0	0.0	1.2	0.0	0.0	0.4	2.4	8
Pre Quaternary									7.1	
coiling ratio	0.4	0.1	8.7	1.2	0.3	0.2	0.7	0.9	0.1	0.8

## Appendix D. Data tables.

Table 9. Percentage occurrences of planktonic foraminifera at DSDP Site 612

core section	3-2	3-2	3-3	3-4	3-3	4-1	4-1	4-2	4-3	4-4
Depth in section (cm)	32	80	84	44	80	0	80	34	34	30
Depth at site (cm)	1622	1670	1824	1934	2120	2400	2480	2584	2734	2880
Individuals/cc	3	34	28.0	45.0	38.0	34.0	49	71.0	26.0	87.0
<i>Globigerina bulloides</i>	9.4	25.3	32.0	22.0	15.1	14.1	11.8	30.3	15.6	33.6
<i>G. falconensis</i>										0.6
<i>G. quinqueloba</i>	12.5	11.6	16.9	5.1	4.7	14.7	19	5.9	35.0	0.3
<i>G. rubescens</i>										0.3
<i>G. umbilicata</i>										
<i>Globigerinella acquilateralis</i>		1.2								
<i>Globigerinita glutinata</i>	12.5	3		8.2	2.4	0.9	3.9	0.6		
<i>Globigerinoides conglobatus</i>										
<i>G. obliquus</i>										0.1
<i>G. pyramidalis</i>										
<i>G. ruber</i>		1.5		0.7	1.0	1.5	0.8			0.7
<i>G. sacculifer</i>							0.1			
<i>G. tenellus</i>										
<i>Globorotalia crassaformis</i>		1.2								0.3
<i>Globorotalia hirsuta</i>										
<i>G. inflata</i>	15.7	6.5	6.1	4.9	7.0	19.1	2.4	12.2	5.5	7.4
<i>G. menardii</i>										
<i>G. scitula</i>				0.2		0.6				
<i>G. tosensis</i>										
<i>G. tumida</i>										
<i>G. truncatulinoides</i>										
<i>Neogloboquadrina dutertrei</i>	3.1					1.2		1.3		
<i>N. humerosa</i>										
<i>N. pachyderma (dextral)</i>	12.5	46.1	34.2	48.7	63.8	32.8	37.5	39.3	1.6	50.7
<i>N. pachyderma (sinistral)</i>	21.9	3	3.6	9.6	6.0	15.3	28.6	10.4	38.1	5.9
<i>Orbulina universa</i>	3.1									0.2
<i>Pulicatina obliquiloculata</i>										
<i>Sphaeroidinella debiscens</i>										
subtropical spp.	0	1.2	0.0	0.2	0.0	0.6	0	0.0	0.0	0.9
tropical spp.	0	1.5	0.0	0.7	1.0	1.5	0.9	0.0	0.0	0.7
gyre margin spp.	3.1	0	0.0	0.0	0.0	1.2	0	1.3	0.0	0.0
Pre Quaternary	3.1	0.6								
colling ratio	1.8	0.1	0.1	0.2	0.1	0.5	0.8	0.3	23.8	0.1

## Appendix D. Data tables.

Table 9. Percentage occurrences of planktonic foraminifera at DSDP Site 612

core section	1-1	1-2	1-3	2-1	2-1	2-2	2-2	2-5	3-1	3-1
Depth in section (cm)	3	80	30	26	50	30	80	24	30	80
Depth at site (cm)	3	230	330	506	530	660	710	1104	1472	1500
Individuals/cc	448.0	12.0	14.0	2	5	3	7.0	3.0	12.0	14.0
<i>Globigerina bulloides</i>	9.1	6.8	7.2	5	11.1	8	16.9	9.1	23.0	20.0
<i>G. falconensis</i>										
<i>G. quinqueloba</i>	0.5	5.9	5.8	10	11.1	4	11.2	12.1	19.7	7.1
<i>G. rubescens</i>						8				
<i>G. umbilicata</i>										
<i>Globigerinella aequilateralis</i>	0.3	1.6								
<i>Globigerinita glutinata</i>	5.2	5.1	4.3		9.3	16	4.2	15.1	13.1	6.4
<i>Globigerinoides conglobatus</i>										
<i>G. obliquus</i>										
<i>G. pyramidalis</i>										
<i>G. ruber</i>	0.7	0.8	0.7				2.8	12.1	0.8	2.8
<i>G. sacculifer</i>	0.1									
<i>G. tenellus</i>										
<i>Globorotalia crassaformis</i>										
<i>Globorotalia hirsuta</i>										
<i>G. inflata</i>	17.6	5.1	7.9	10	29.6	4	2.8	6.1	1.6	1.4
<i>G. menardii</i>	0.7									
<i>G. scitula</i>		4.2							1.6	
<i>G. tosacensis</i>										
<i>G. tumida</i>	0.3									
<i>G. truncatulinoides</i>										
<i>Neogloboquadrina dutertrei</i>	0.8					4				0.7
<i>N. humerosa</i>										
<i>N. pachyderma (dextral)</i>	60.4	28.8	35.5	20	27.7	40	39.4	27.4	13.1	41.1
<i>N. pachyderma (sinistral)</i>	3.7	41.5	44.9	50	13	12	22.5	18.1	18.9	19.1
<i>Orbulina universa</i>	0.1			5						0.7
<i>Pulleniatina obliquiloculata</i>	0.1									
<i>Sphaeroidinella edwardsensis</i>										
subtropical spp.	0.0	4.2	0.0	0	0	0	0.0	0.0	1.6	0.0
tropical spp.	0.8	0.2	0.7	0	0	0	2.8	12.1	0.8	2.8
gyre margin spp.	1.9	0.0	0.0	0	0	4	0.0	0.0	0.0	0.7
PreQuaternary										
coiling ratio	0.1	1.4	1.3	2.5	0.6	0.3	0.6	0.7	1.4	0.5

## Appendix D. Data tables.

Table 10. Percentage occurrences of dinoflagellate cysts at DSDP Site 612

Core section	1-1	1-2	1-3	2-2	2-4	3-1	3-3	3-4
Depth in section	3	83	148	83	27	35	87	47
Depth in core (cm)	3	233	448	713	957	1475	1827	1937
cysts/cc	3000.0	1293.0	360.0	1525.0	338.0	3627.0	16744.0	8923.0
cysts/cc corrected*							3349.0	1784.0
<i>Bilteatodinium tepikiense</i>	2.0	6.1	2.6	28.8		9.2	8.2	0.5
<i>Brigantodinium</i> spp.	24.4	52.4	64.9	47.5	84.0	23.1	44.0	37.9
<i>Corruodinium harlandii</i>	0.6							
<i>Impagidinium</i> spp	10.9	3.7	1.3	1.3		1.5	0.7	1.0
<i>Impagidinium aculeatum</i>								
<i>Impagidinium psulum</i>								
<i>Impagidinium striolatum</i>								
<i>Lingulodinium machaerocarpum</i>	5.8	3.4	6.5	3.8		1.5	5.2	0.5
<i>Mmichrystidium</i> sp.		8.5	5.2	2.5		6.2		
? <i>Multispinula minuta</i>	9.6	7.3		2.5		6.2	6.7	1.5
? <i>Multispinula quanta</i>	0.6					1.5	2.2	2.5
<i>Multispinula</i> sp. n of Mudie 1986								
<i>Nematosphaeropsis labyrinthus</i>	0.6							
<i>Nematosphaeropsis acquaducta</i>	0.6							0.5
<i>Operculodinium centrocarpum</i>	37.2	8.5	18.2	10.0	8.0	43.1	32.1	31.8
<i>Polykrikos</i> sp.								
<i>Selenopemphix nephroides</i>		1.2	2.6	1.3				0.5
<i>Spiniferites</i> indet.	2.6	4.9	2.6					6.1
<i>Spiniferites bentonii</i>								
<i>Spiniferites elongatus</i>	1.3				4.0	1.5		
<i>Spiniferites membranaceus</i>	0.6							
<i>Spiniferites mirabilis</i>	0.6					4.6	3.0	3.0
<i>Spiniferites ramosus</i>	4.5	2.4		1.3		9.2	7.5	4.5
<i>Spiniferites</i> c.f. <i>tertiaria</i>								0.5
<i>Tectatodinium pellitum</i>	1.2	3.7				2.3	0.7	
G:P ratio	1.9	0.7	0.5	1.0	0.2	2.2	0.9	1.4

## Appendix D. Data tables.

Table 10. Percentage occurrences of dinoflagellate cysts at DSDP Site 612

Core section	3-5	4-2	4-3	4-4	4-6	4-6	5-1	5-2	
Depth in section	83	37	40	83	33	83	117	27	
Depth in core (cm)	2078	2633	2740	2933	3183	3233	3477	3537	
cysts/cc	7335.0	4307.0	346.0	3847.0	6722.0	4948.0	7797.0	5547.0	
cysts/cc corrected*	1467.0	861.0		769.0	1344.0	990.0	1559.0		
<i>Bitectatodinium tepikiense</i>		1.5			0.8		2.0	8.2	10.3
<i>Brigantedinium</i> spp.	45.2	60.8	51.9	29.9	54.6	31.6	34.0	33.8	
<i>Corrudinium barlandii</i>									
<i>Impagdinium</i> spp	0.5	1.5		0.8	0.5	1.3	1.3	4.4	
<i>Impagdinium aculeatum</i>									
<i>Impagdinium patulum</i>									
<i>Impagdinium striatum</i>									
<i>Lingulodinium machaerocarpum</i>	2.5	2.6	3.9	3.1	1.1	1.3	1.9	3.0	
<i>Mmichrystridium</i> sp.	1.0	0.5			0.5		0.6	5.9	
? <i>Multispinula minuta</i>			1.0	2.6	3.1	2.7	1.3	4.4	
? <i>Multispinula quanta</i>	0.5	2.1	2.6	0.8	1.1	0.7	4.4	1.5	
<i>Multispinula</i> sp. n of Mudie 1986		0.5							
<i>Nematosphaeropsis labyrinthus</i>				0.8		1.3			
<i>Nematosphaeropsis aequeducta</i>									
<i>Operculodinium centrocarpum</i>	35.7	14.9	32.5	48.0	29.5	35.9	30.2	33.8	
<i>Polykrikos</i> sp.				0.8					
<i>Scenopemphix nephroides</i>	0.5	0.5		0.8	0.5	2.0			
<i>Spiniferites</i> indet.	2.5	4.6	3.9	2.4	3.8	5.9	2.5	1.5	
<i>Spiniferites bentorii</i>		0.5		0.8	0.5	2.0			
<i>Spiniferites elongatus</i>	0.5			2.4	0.5	1.3	3.1		
<i>Spiniferites membranaceus</i>						0.7	0.6		
<i>Spiniferites mirabilis</i>	1.5	2.6	2.6	3.1	1.6	3.9	5.7	4.4	
<i>Spiniferites ramosus</i>	1.5	4.1		5.5	1.6	10.5	6.9	4.4	
<i>Spiniferites</i> c.f. <i>tertiaria</i>								8.8	
<i>Tectatodinium pellitum</i>							4.4		
G:P ratio	1.2	0.6	0.8	2.0	0.7	2.0	1.3	1.8	

## Appendix D. Data tables.

Table 11. Percentage occurrences of pollen and terrestrial spores at DSDP Site 612

Core section	3-5	4-2	4-3	4-4	4-6	4-6	5-1	5-2
Depth in section (cm)	83	37	40	83	33	83	117	27
Depth at site (cm)	2078	2633	2740	2933	3183	3233	3477	3537
pollen grains/cc	10044.0	13237.0	260.0	3847.0	4346.0	17452.0	10287.0	616.0
corrected* pollen/cc	2009.0			769.0	869.0	3490.0	2057.0	
<i>Abies</i>	0.0	0.0	0.0	0.6	0.0	0.0		0.0
<i>Acer</i>	0.6	0.5			1.0	0.4		
<i>Alnus</i>	3.1	2.7		0.6	2.4	1.4	0.0	
<i>Ambrosia</i>								
<i>Artemisia</i>	1.3	0.5		1.9	1.4	0.4	1.9	
<i>Betula</i>	2.5	4.1	1.4		1.0	1.1	1.9	
<i>Carya</i>	1.9				1.4	0.4	0.4	
Caryophyllaceae		0.5			0.7			
<i>Castanea</i>								
Chenopodiaceae	0.6	0.5	1.4	0.6				
<i>Comus</i>	0.6				0.3			
<i>Corylus</i>	0.6	1.4	1.4	0.6				
Cyperaceae	0.6	2.3		1.9	1.0	0.7	2.7	
<i>Dryas</i>	0.6	0.9		0.6	1.0			
Ericaceae	0.6				0.7		0.4	
<i>Fagus</i>	2.5				0.3			
<i>Fraxinus</i>	0.6			0.6			0.4	
Gramineae	1.9	0.0		0.0	0.0	0.0	0.0	
<i>Ostrya</i>	4.4	0.9			0.0	0.0	0.0	1.2
<i>Picea</i>	14.7	18.9	20.3	19.1	17.9	22.1	13.4	22.2
<i>Pinus</i>	54.4		60.9	67.5	63.6	66.5	65.9	62.9
<i>Platanus</i>	0.6							
<i>Quercus</i>	3.1		11.6	3.8	5.5	3.6	6.1	7.4
<i>Rumex</i>								
<i>Salix</i>								
<i>Tilia</i>								
Tubuliflorae	0.6							
<i>Tsuga</i>	0.6							
<i>Ulmus</i>								
Umbelliferae								
<i>Dryopteris</i>	2.5	1.6	1.5			1.5	0.7	
<i>Lycopodium</i>	1.9		3.0		2.4	0.7		2.6
<i>Pteridium</i>	3.1			1.9	0.6	1.5	4.8	
<i>Sphagnum</i>	1.3	0.6				0.4		
<i>Isoetes</i>	0.6		1.5					

## Appendix D. Data tables.

Table 11. Percentage occurrences of pollen and terrestrial spores at L DP Site 612

Core section	1-1	1-2	1-3	2-2	2-4	3-1	3-3	3-4
Depth in section (cm)	3	83	148	83	27	35	87	47
Depth at site (cm)	3	233	448	713	957	1475	1827	1937
pollen grains/cc	5714.0	4140.0	880.0	2663.0	528.0	9724.0	16539.0	8767.0
corrected* pollen/cc							3308.0	1753.0
<i>Abies</i>	0.0	0.0	0.0	0.0		0.0	0.0	0.0
<i>Acer</i>	2.9	1.3	1.9			0.5	1.2	1.3
<i>Alnus</i>	3.4	0.8	0.6	5.6	3.8	3.1	0.6	2.2
<i>Ambrosia</i>	0.5	0.8						
<i>Artemisia</i>		0.8		1.6	3.8	1.0	1.9	0.4
<i>Betula</i>	2.0	4.6	3.1	7.3	3.8	1.6	1.2	0.9
<i>Carya</i>	1.0	1.7	1.9	0.8	1.9	1.6	0.6	0.9
Caryophyllaceae								
<i>Castanea</i>				1.6		1.0		
Chenopodiaceae		0.4				0.5	0.6	
<i>Cornus</i>	0.5	0.4		0.8	1.9	1.0		
<i>Corylus</i>	0.5	1.7	0.6	1.6		2.1		
Cyperaceae	0.5	0.8	1.3	0.8		2.1		0.9
<i>Dryas</i>	1.0	0.4	1.3		1.9	0.5	0.6	0.4
Ericaceae								
<i>Fagus</i>	1.5	1.3	2.5		1.9	0.5	0.6	0.9
<i>Fraxinus</i>				0.8				
Gramineae	3.4	2.1	0.6			1.6	0.6	0.0
<i>Ostrya</i>	2.9	2.1	0.6	7.3	1.9	3.1	0.6	1.3
<i>Picea</i>	5.1	16.9	15.4	14.5	19.8	12.7	12.7	17.3
<i>Pinus</i>	52.2	54.6	62.9	48.8	50.0	43.8	70.5	70.7
<i>Platanus</i>					1.9			
<i>Quercus</i>	10.7	4.2	5.7	8.1	3.8	10.9	6.3	2.7
<i>Rumex</i>	1.5	0.4						
<i>Salix</i>	0.5	1.7	1.9	4.0	1.9	1.6		
<i>Tilia</i>		0.4		0.8		0.5	0.6	0.4
Tubuliflorae			0.4	0.8		4.1	1.9	0.9
<i>Tsuga</i>	6.3	0.4		1.6				
<i>Ulmus</i>	2.4	0.8		0.8	3.8			
Umbelliferae		0.4						
<i>Dryopteris</i>	1.5	0.8	2.6	0.7		2.2	1.2	1.4
<i>Lycopodium</i>	0.8		1.9	2.1	5.7	0.6	0.6	0.5
<i>Pteridium</i>	3.1	1.5	0.6	1.4		2.8	2.4	3.2
<i>Sphagnum</i>	2.3	1.5	0.6		1.9	0.6	1.2	0.9

## Appendix D. Data tables.

Table 11. Percentage occurrences of pollen and terrestrial spores at DSDP Site 612

Core section	1-1	1-2	1-3	2-2	2-4	3-1	3-3	3-4
Depth in section (cm)	3	83	148	83	27	35	87	47
Depth at site (cm)	3	233	448	713	957	1475	1827	1937
pollen grains/cc	5714.0	4140.0	880.0	2663.0	528.0	9724.0	16539.0	8767.0
corrected* pollen/cc							3308.0	1753.0
<i>Abies</i>	0.0	0.0	0.0	0.0		0.0	0.0	0.0
<i>Acer</i>	2.0	1.3	1.9			0.5	1.2	1.3
<i>Alnus</i>	3.4	0.8	0.6	5.6	3.8	3.1	0.6	2.2
<i>Ambrosia</i>	0.5	0.8						
<i>Artemisia</i>		0.8		1.6	3.8	1.0	1.9	0.4
<i>Betula</i>	2.0	4.6	3.1	7.3	3.8	1.6	1.2	0.9
<i>Carya</i>	1.0	1.7	1.9	0.8	1.9	1.6	0.6	0.9
<i>Caryophyllaceae</i>								
<i>Castanea</i>				1.6		1.0		
<i>Chenopodiaceae</i>		0.4				0.5	0.6	
<i>Cornus</i>	0.5	0.4		0.8	1.9	1.0		
<i>Corylus</i>	0.5	1.7	0.6	1.6		2.1		
<i>Cyperaceae</i>	0.5	0.8	1.3	0.8		2.1		0.9
<i>Dryas</i>	1.0	0.4	1.3		1.9	0.5	0.6	0.4
<i>Ericaceae</i>								
<i>Fagus</i>	1.5	1.3	2.5		1.9	0.5	0.6	0.9
<i>Fraxinus</i>				0.8				
<i>Gramineae</i>	3.4	2.1	0.6			1.6	0.6	0.0
<i>Ostrya</i>	2.9	2.1	0.6	7.3	1.9	3.1	0.6	1.3
<i>Picea</i>	5.1	16.9	15.4	14.5	19.8	12.7	12.7	17.3
<i>Pinus</i>	52.2	54.6	62.9	48.8	50.0	43.8	70.5	70.7
<i>Platanus</i>					1.9			
<i>Quercus</i>	10.7	4.2	5.7	8.1	3.8	10.9	6.3	2.7
<i>Rumex</i>	1.5	0.4						
<i>Salix</i>	0.5	1.7	1.9	4.0	1.9	1.6		
<i>Tilia</i>		0.4		0.8		0.5	0.6	0.4
<i>Tubuliflorae</i>			0.4	0.8		4.1	1.9	0.9
<i>Tsuga</i>	6.3	0.4		1.6				
<i>Ulmus</i>	2.4	0.8		0.8	3.8			
<i>Umbelliferae</i>		0.4						
<i>Dryopteris</i>	1.5	0.8	2.6	0.7		2.2	1.2	1.4
<i>Lycopodium</i>	0.8		1.9	2.1	5.7	0.6	0.6	0.5
<i>Pteridium</i>	3.1	1.5	0.6	1.4		2.8	2.4	3.2
<i>Sphagnum</i>	2.3	1.5	0.6		1.9	0.6	1.2	0.9

## Appendix D. Data tables.

:Table 12. Stable isotope ratios from DSDP Site 612

Depth (cm)	O-18 dextral	C-13 dextral	Depth (cm)	O-18 dextral	C-13 dextral
0	1.287	0.779	2584	2.479	0.271
97	1.127	0.378	2683	2.762	-0.203
133	1.67	0.275	2880	2.217	0.128
283	2.778	-0.375	2883	2.573	0.346
433	2.631	-0.324	2930	2.41	0.163
480	0.834	0.59	2983	2.469	0.254
613	2.613	-0.036	3032	2.016	0.44
1500	2.635	-0.721	3130	2.639	0.085
1573	2.024	-0.662	3133	2.724	0.37
1622	2.728	-0.617	3180	2.657	0.253
1670	2.416	-0.323	3283	2.432	0.553
1772	2.456	-0.559	3360	2.229	0.3
1824	2.229	-0.495	3407	2.44	0.072
1873	2.363	-0.297	3457	2.34	0.062
1990	2.571	-0.244	3493	2.435	0.059
2077	1.896	-0.169	3534	2.465	0.105
2080	2.803	0.205	3643	1.75	-0.133
2120	2.404	-0.302	3793	1.829	-0.178
2173	2.546	-0.49	3943	1.924	0.177
2400	2.74	-0.343	4700	1.397	0.384
2573	2.744	0.412	5493	1.839	-0.303

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