LATE PALEOGENE (EOCENE TO OLIGOCENE) PALEOCEANOGRAPHY
OF THE
NORTHERN NORTH ATLANTIC

by

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A.B., Rutgers College, New Brunswick, NJ
(1978)

SUBMITTED IN PARTIAL FULFILLMENT
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DOCTOR OF PHILOSOPHY

at the

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Submitted to the Woods Hole Oceanographic Institution/Massachusetts Institute of Technology Joint Program in Oceanography on 13 September 1982 in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

ABSTRACT

Seismic stratigraphic evidence from the western and northern North Atlantic indicates that a major change in abyssal circulation occurred in the latest Eocene to earliest Oligocene. In the northern North Atlantic, the widely-distributed reflector R4 correlates with an unconformity that can be traced to its correlative conformity near the top of the Eocene. This horizon reflects a change from weakly circulating (Eocene) to vigorously circulating bottom water (early Oligocene). Sediment distribution patterns provide evidence for strong contour-following bottom water flow beginning at reflector R4 time; this suggests a northern source for this bottom water, probably from the Arctic via the Norwegian-Greenland Sea and Faeroe-Shetland Channel. Erosion and current-controlled sedimentation continued through the Oligocene; however, above reflector R3 (middle to upper Oligocene), the intensity of abyssal currents decreased. Above reflector R2 (upper lower Miocene) current-controlled sedimentation became more coherent and a major phase of sedimentary drift development began. This resulted from further reduction in speeds and stabilization of abyssal currents.

Paleontological and stable isotopic data support these interpretations. In the Bay of Biscay/Goban Spur regions, a major $\delta^{18}O$ increase began at ~ 38 Ma (late Eocene), culminating in a rapid (< 0.5 my) increase in $\delta^{18}O$ just above the Eocene/Oligocene boundary (~ 36.5 Ma). A rapid $\delta^{13}C$ increase also occurs at ~ 36.5 Ma in these sites. Major changes in benthic foraminiferal assemblages also occurred between the middle Eocene and the earliest Oligocene: 1) In the Labrador Sea, a predominantly agglutinated assemblage was replaced by a calcareous assemblage between the middle Eocene and early Oligocene; 2) In the abyssal (> 3km) Bay of Biscay, an indigenous Eocene calcareous fauna including Nuttallides truempyi, Clinapertina spp., Abyssammina spp., Aragonia spp., and Alabamina dissonata became extinct between the middle Eocene and earliest Oligocene; 3) In shallower sites (< 3km paleodepth) throughout the Atlantic, a Nuttallides truempyi-dominated assemblage was replaced by a Globocassidulina subglobosa-Gyroidinoides-Cibicidoides ungerianus-Oridorsalis assemblage in the early late Eocene (~ 40-38.5 Ma). These faunal and isotopic changes represent the transition from warm, old, corrosive Eocene bottom waters to colder, younger (lower CO_2 and higher pH, hence less corrosive) early Oligocene bottom waters.

Thesis Co-supervisor:	William A. Berggren	Title: Senior Scientist
Thesis Co-supervisor:	Brian E. Tucholke	Title: Associate Scientist

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CURRICULUM VITAE

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Family Status: Married to Karen Clark Miller.

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Professional Experience:

1976-1978 Research Assistant to R. K. Olsson, Rutgers University, Department of Geological Sciences.
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1978-1982 Graduate Research Assistant, Woods Hole Oceanographic Institution.
1980 Knorr 80 Cruise (Multichannel Seismic Cruise).
1980 Instructor, Sea Education Association.
1979-present Presented 12 papers at 1 regional, 3 international, and 5 national meetings.

Professional Societies:

American Association of Petroleum Geologists.
American Geophysical Union.
North American Micropaleontological Section of SEPM.
Sigma Xi.
Cushman Foundation for Foraminiferal Research.

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Graduated from Rutgers College highest honors (third in class of 1500) and highest distinction (first in department).
Phillips Petroleum Graduate Fellowship.

I have accepted a post-doctoral fellowship at Lamont-Doherty Geological Observatory of Columbia University following my tenure at WHOI.

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PREFACE

"To many geologists...[the description and correlations of local sections] make up the whole of stratigraphy, which is considered profoundly boring by all but those immediately concerned with the specific descriptions and correlations. But we believe that these, important and indispensible as they are, are but the means to a further end that constitutes *the real core and interest of stratigraphy--namely, the interpretation of the stratigraphic record, both the rocks and their contained fossils, in terms of the past history of the earth*."

Dunbar and Rodgers,
Principles of Stratigraphy
1957
(emphasis mine)

CHAPTER 1: INTRODUCTION

> "Below a depth of 600 feet the ocean floor is...touched by currents so gentle as to have little effect on the bottom sediment...a continuous record of all geologic time may be present in parts of the ocean floor."
>
> Dunbar and Rodgers (1957)

The Deep Sea Drilling Project has demonstrated that the deep-sea stratigraphic record is punctuated by numerous hiatuses, many of which resulted from erosion by vigorously circulating bottom waters. One of the most dramatic hiatuses attributed to such erosion occurred in the Late Paleogene of the Atlantic, Southern, and Pacific Oceans. This thesis investigates the age and cause of this event in the North Atlantic and evaluates its effect on the paleoenvironment of the deep sea.

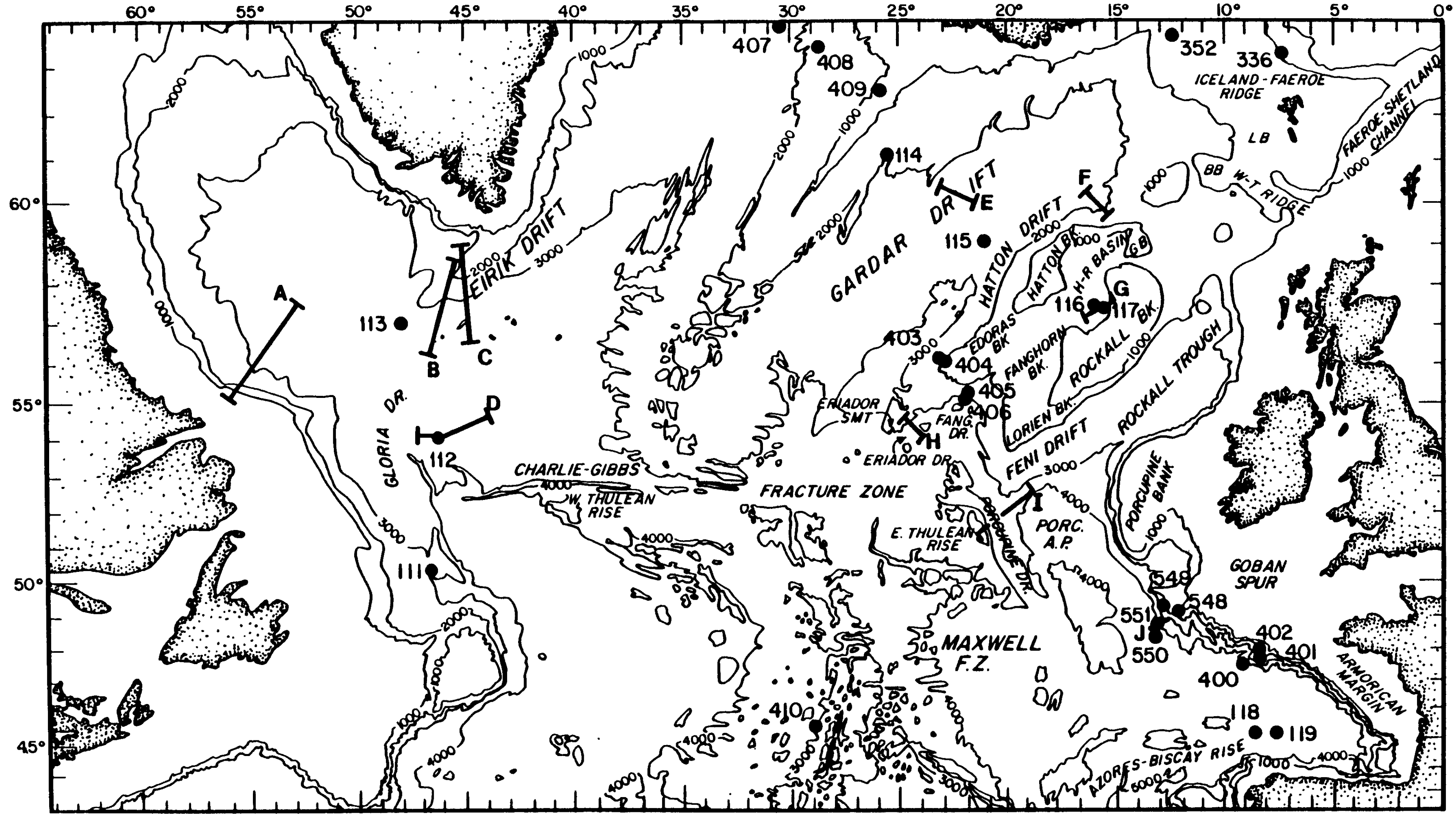
Deep and bottom waters formed in the Norwegian-Greenland Sea today overflow the aseismic ridge between Greenland and Scotland, forming a major constituent of modern North Atlantic Deep Water (NADW). This overflow influences hydrographic properties as far away as the North Pacific Ocean (e.g. Reid and Lynn, 1971). It also profoundly effects sediment distribution, not only through erosional/depositional effects (e.g. Heezen et al., 1966) but also through its importance in fractionation of calcium carbonate and silica between basins (e.g. Berger, 1970). For over a decade, it has been known that overflow of the Greenland-Scotland Ridge had geologically significant effects well back into the Tertiary (Jones et al., 1970). The initial entry of such northern bottom waters into the North Atlantic represents a critical threshold in the development of global bottom-water circulation.

Considerable debate, based upon differing lines of evidence, exists as to the timing of initial vigorous deep circulation in the North Atlantic resulting from bottom water formation in high northern latitudes. Three distinct events in the benthic faunal, isotopic, and lithostratigraphic records have been proposed as representing the initial formation. 1) Based upon increased deposition of biosiliceous sediments in the North Atlantic, Berggren and Hollister (1974; 1977) suggested that the first influx and upwelling of bottom water from the Norwegian-Greenland Sea/Arctic Ocean occurred in the early Eocene. 2) Based upon changes in benthic foraminiferal assemblages and isotopic composition, Schnitker (1979) and Blanc et al. (1980) suggested that the first entry of northern bottom waters into the North Atlantic began in the middle Miocene. 3) Based upon a change from agglutinated to calcareous benthic foraminiferal assemblages in the Labrador Sea near the end of the Eocene, Miller et al. (1982; Appendix 1), suggested that significant influx of northern bottom waters began in the late Eocene to early Oligocene. Such ambiguity results from the nature of the evidence interpreted, for faunal, lithologic, and isotopic evidence are rarely by themselves diagnostic of the nature of abyssal circulation changes (Johnson, 1982; Miller and Tucholke, in press; Appendix 3; Miller, in press a).

The seismic stratigraphic record (and to some extent the rock stratigraphic record) provides less ambiguous evidence for changes in the intensity of bottom water circulation than do faunal or isotopic changes alone, although seismic and rock stratigraphic studies rely upon biostratigraphy to identify the age of events. The development of many regional deep-sea unconformities, especially those that show significant truncation and erosion of lower horizons, are evidence of strong bottom-water flow. Such unconfomities are most clearly manifested in the seismic stratigraphic record (Vail et al., 1977; Tucholke and Mountain, 1979; Miller and Tucholke, in press; Appendix 3). In addition, stratal patterns within seismic sequences such as the development of drift deposits, moats, and sediment waves document the influence of strong currents on deep-sea sedimentation. Regional seismic stratigraphic studies, therefore, are perhaps the most useful tool for delineating the relative timing and importance of current changes.

This thesis addresses the problem of the timing, effects, and causes of the first geologically significant influx of bottom water from northern sources into the North Atlantic. The problem is addressed in three ways. 1) The seismic stratigraphic and lithostratigraphic records of the northern North Atlantic (i.e. the region north of ~ 45^{o}N latitude and south of the Greenland-Scotland Ridge)(Fig. 1) are examined and biostratigraphically dated in order to formulate a testable model for the development of Cenozoic abyssal circulation in the North Atlantic. 2) Eocene to Oligocene deep-sea benthic foraminiferal assemblage and isotopic changes in the northern North Atlantic are examined. This documents the relationship among faunal, isotopic, and abyssal circulation changes in the Late Paleogene. 3) Possible controls (e.g. climatic and tectonic changes) on the development of Cenozoic abyssal circulation in this region are evaluated. This study differs from most earlier studies, in that it does not depend soley upon faunal and isotopic data to interpret the nature of the major abyssal circulation changes. Rather, it uses these data in a supportive role, thus reducing some of the ambiguities that occur in paleoceanographic interpretations based on such records alone.

Most of the detailed data brought to bear on this issue have been published (or are pending publication) elsewhere and are appended here. Appendix 1 proposed that middle Eocene to early Oligocene benthic faunal changes in the North Atlantic were related to the first significant formation of bottom water of northern origin. The circulation model is developed in Appendix 3, while Appendixes 2, 4, and 5 established the relationship of faunal and isotopic changes with respect to the circulation model.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4000
6000

Figure 1. Bathymetric location map of the northern Atlantic with locations of DSDP sites and major sediment drifts. Contour interval 1000m. Sections labeled A through J are illustrated in Figures 4-13 and 15. H-R = Hatton-Rockall Basin; GB = George Bligh Bank; BB = Bill Bailey Bank; FB = Faeroe Bank.

LITHOFACIES VS AGE-WITH CORRELATIONS OF REFLECTORS

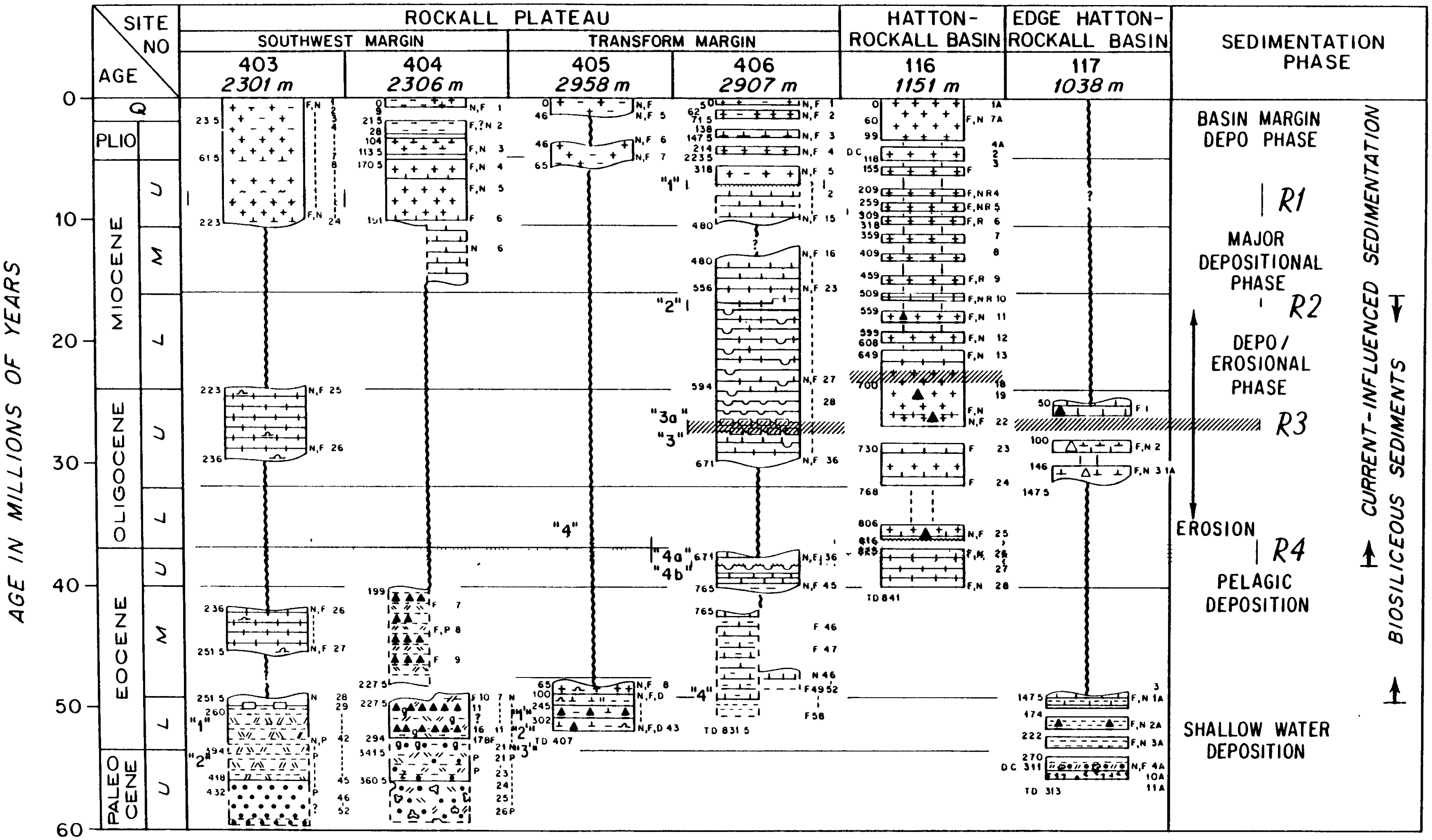

Figure 2. Age versus lithofacies for DSDP sites in Rockall region with correlation of major reflectors. Site locations in Figure 1. Lithologic symbols follow standard "Initial Reports" format. To left of columns are subbottom depths in meters (T.D.= total depth); to right are indicated core number and fossil group on which age is based (N = nannoplankton; F = foraminifers, R = radiolaria, P = palynomorphs, BF = benthic foraminifera, D = diatoms, X/S = extrapolated by sedimentation rate). Heavy wavy lines show hiatuses; dashed lines show inferred continuous sedimentation. Numbers in quotes at left of columns are reflector numbers used in individual site reports by Montadert, Roberts et al. (1979). After Miller and Tucholke (in press; Appendix 3).

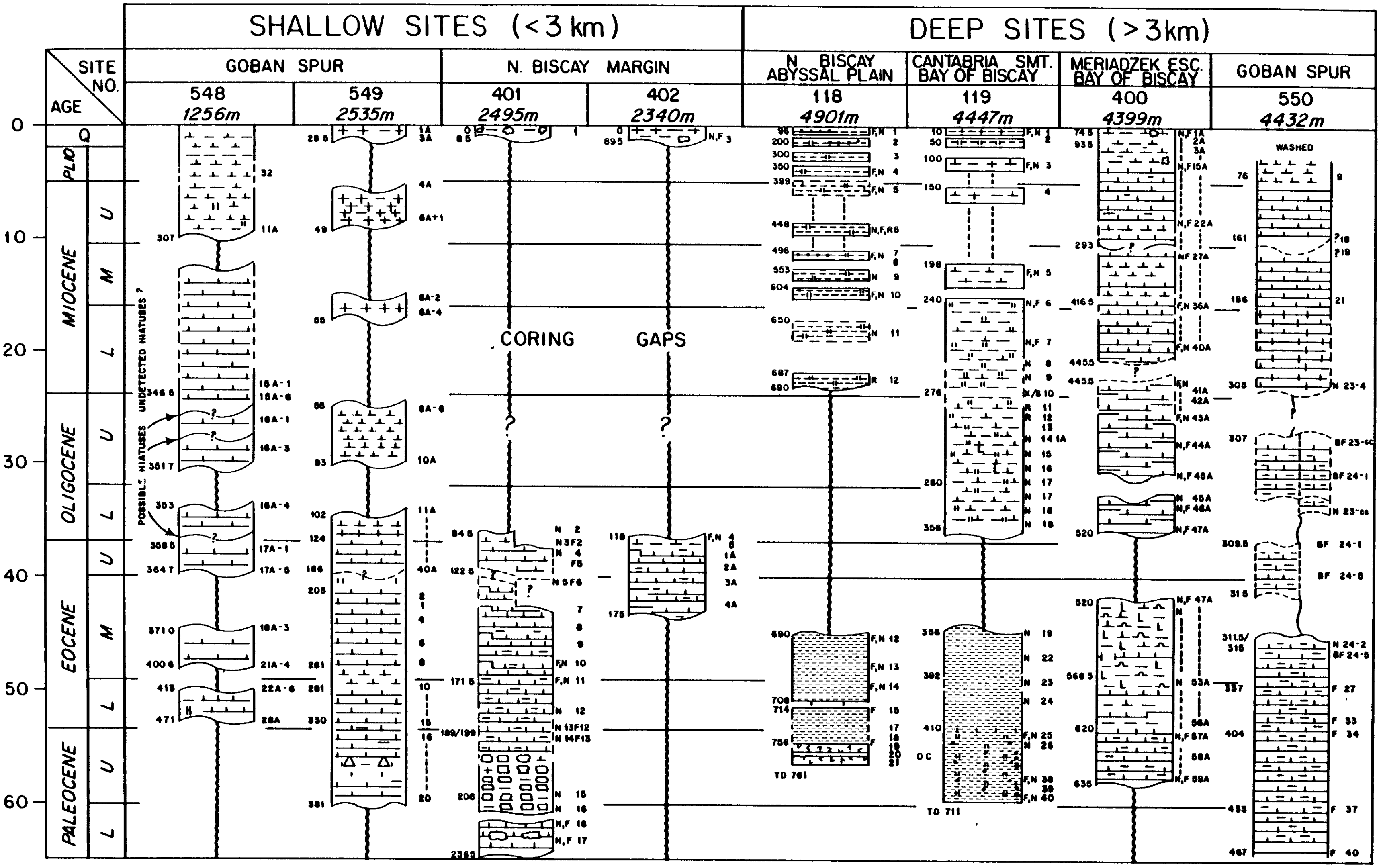
SHALLOW SITES (<3 km)
DEEP SITES (>3km)
SITE NO.
AGE
GOBAN SPUR
N. BISCAY MARGIN
N. BISCAY ABYSSAL PLAIN
CANTABRIA SMT. BAY OF BISCAY
MERIADZEK ESC. BAY OF BISCAY
GOBAN SPUR
548
1256m
549
2535m
401
2495m
402
2340m
118
4901m
119
4447m
400
4399m
550
4432m
0
10
20
30
40
50
60
Q
PLIO
MIOCENE
OLIGOCENE
EOCENE
PALEOCENE
U
M
L
POSSIBLE UNDETECTED HIATUSES ?
CORING
GAPS
?
WASHED
TD 761
TD 711
DC

Figure 3. Age versus lithofacies for Tertiary sediments recovered at DSDP sites in the Bay of Biscay. Explanation in Figure 2. After Miller et al. (in press; Appendix 5).

CHAPTER 2: DEVELOPMENT OF ABYSSAL CIRCULATION IN THE NORTHERN NORTH ATLANTIC: SEISMIC AND LITHOSTRATIGRAPHIC EVIDENCE

Introduction

The presence of strong abyssal currents is best documented by the development of regional deep-sea unconformities that are observed as seismic discontinuities. The first unambiguous evidence of strong bottom-current activity reported in the North Atlantic is the prominent erosional unconformity (Horizon A^u) that occcurs beneath the continental rise of eastern North America. Tucholke (1979) and Tucholke and Mountain (1979) demonstrated that this erosion occurred at some time between the late Eocene and earliest Miocene. They suggested that the unconformity was caused by a precursor to the modern southerly-flowing Western Boundary Undercurrent (Heezen et al., 1966) which contains a significant component of Greenland-Scotland Ridge overflow water. Unfortunately, where cored, the hiatus at Horizon A^u is too long to resolve firmly the timing of this abyssal circulation event.

The sedimentary record recovered in the northern North Atlantic (north of ~ 45^0 N) (Fig. 1; Figs. 2,3) is more complete. A widespread seismic reflector occuring in this region variously termed "R" (Jones et al., 1970; Ruddiman, 1972), or "R4" (Roberts, 1975; Roberts et al., 1979; Miller and Tucholke, in press; Appendix 3), figures prominently in the interpretation of abyssal circulation history because it divides current-influenced sedimentation above from largely pelagic and downslope sedimentation below (Figs. 4-12). Reflector R4 correlates with a prominent unconformity found in boreholes drilled in the Rockall and Biscay regions (Figs. 2,3). In the Rockall Trough and Porcupine Abyssal Plain regions (Fig. 1), it truncates underlying strata (Figs. 5,6; figs. 29,30,32 in Roberts, 1975). Because of these relationships, and because it approximately correlates with Horizon A^u in time, reflector R4 seems to mark initial strong abyssal circulation in the North Atlantic.

I have traced reflector R4, overlying reflectors, and underlying reflectors throughout the northern North Atlantic and dated them at boreholes (Fig. 1) using single channel seismic reflection data of

Lamont-Doherty Geological Observatory (Figs. 4-9,11,12), Woods Hole Oceanographic Institution, U.S. Navy, and DSDP Legs 12 and 48. In addition, multichannel data from the Canadian Geological Survey/Federal Institute for Geosciences and Natural Resources (BGR)(Figs. 10,14), the British National Oil Company (BNOC)/Institute of Geological Sciences (IGS) Edinburgh (Fig. 24), DSDP Leg 48 (Montadert, Roberts et al., 1979), and Leg 80 (Fig. 13)(de Graciansky, Poag et al., in preparation) were examined. This chapter discusses the pre-R4 sequences in the northern North Atlantic, establishing that little evidence of abyssal currents is found prior to reflector R4 time. The age and significance of reflector R4 are then discussed, establishing that this horizon marks the most important change in sedimentary regime in the northern North Atlantic. The post-R4 sequences are then examined and dated, and post-R4 evidence of abyssal circulation changes is investigated. Finally, a model for the development of abyssal circulation in the northern North Atlantic is presented.

Pre-R4 Sequences

The pre-R4 sequences are generally poorly resolved in the single channel and multichannel data examined (Figs. 4-12). In the Labrador Sea, most prominent are strong seismic reflectors thought to be Lower Tertiary basaltic rocks (Figs. 10-12). These basalts are probable age equivalents of some of the extrusives in East Greenland, northern Britain, Davis Straits, and Voring Plateau (Fitch et al., 1974). This "Thulean" volcanic event probably coincided with the initiation of sea-floor spreading in the Norwegian-Greenland Sea about Anomaly 24 time (latest Paleocene to earliest Eocene)(Berggren and Schnitker, in press).

Several strong intra-sedimentary seismic reflectors underlie reflector R4 in the Rockall Trough and Porcupine Abyssal Plain; in general, these pre-R4 reflectors result from pelagic basin fill (Figs. 4,5) or downslope sedimentation. In the southern Rockall Trough and eastern Porcupine Abyssal Plain, Dingle et al. (1982) observed two such reflectors, the deeper termed "Shackleton" and the shallower termed "Charcot." In the Rockall Trough, Roberts (1975) observed three pre-R4 reflectors; he called these Z, X, and Y in ascending stratigraphic order. The relationships between the horizons noted by Roberts (1975) and those of Dingle et al. (1982) are not clear; comparison of their published profiles with profiles herein (V29-09; V23-05) suggests that reflector Y probably corresponds to "Shackleton" and that reflector X probably corresponds to "Charcot."

Reflector Z can be constrained only to the Upper Cretaceous (Roberts et al., 1981); Roberts (1975) noted that sediments below reflector Z consist of pelagic drapes. An exception to this may occur along the margin of the Rockall Trough near Lorien Bank (Fig. 1) where reflector Z is flat lying; Roberts (1975) suggested that this may represent gravity fill. Reflector Y is of late Paleocene age (Roberts et al., 1981); in general, the Z-Y sequence also consists primarily of drapes conformable with basement (Roberts, 1975), although fan deposition may occur in the Z-Y interval (fig. 32 in Roberts, 1975). Reflector X is a post-Paleocene reflector that merges with reflector R4 in the Biscay region (Roberts et al., 1981) and is probably of early or middle Eocene age; it apparently also occurs on the southwest margin of Rockall Plateau, where it drapes over the faulted acoustic basement (fig. 26 in Roberts, 1975).

Evidence of current influences in the R4-X interval are debatable. Dingle et al. (1982) noted apparent lensing between reflector Charcot (= X) and reflector R4 (= their reflector Challenger) in the southern Rockall Trough; they suggested that this indicated current influences on sedimentation in the pre-R4 interval. However, examination of their published data suggests that they traced the top of a high amplitude unit as reflector R4; an overlying reflector (reflector R3, see below) onlaps reflector R4, and generally represents the top of the high amplitude unit. Thus, some of the differential thickening that they noted probably occurs not in the pre-R4 interval, but in the post-R4 interval (between reflector R4 and reflector R3; their figure 4). In addition, reflector X is truncated by reflector R4 (fig. 29 in Roberts, 1975); therefore, apparent lensing in the R4 to X (= Charcot) interval may be due to erosional truncation at the level of reflector R4. This is illustrated by the erosion and truncation of horizons by reflector R4 shown in figure 5; the thickness variations (e.g. Fig. 5) apparent in the pre-R4 interval are not depositional phenomena but are lenslike remnants formed by erosion at the level of reflector R4.

In the Rockall Plateau area, the pre-R4 sequences are poorly-defined (Figs. 4,8,9). Along margins of the Hatton-Rockall basin (Figs. 1,4), Roberts (1975) noted that variations in thickness of the pre-R4 series are attributable to syndepositional faulting. In general, the pre-R4 series on Rockall Plateau is characterized by extremely high sedimentation rates (> 70 m/m.y.; Roberts et al., 1979) associated with the rapid subsidence of Rockall Plateau during the late Paleocene to early Eocene (Laughton, Berggren et al., 1972; Berggren, 1974; Roberts et al., 1979).

West of Rockall Plateau in the Iceland Basin, the pre-R4 interval thickens westward toward the flanks of the Reykjanes Ridge (profile V27-06; profile Kane 70C, fig. 14 in Ruddiman, 1972). The cause of this thickening is not known. Roberts (1975) suggested that it resulted from a westward increase in sedmentation rate in the pre-R4 interval. Still, it is not clear why such a radical east to west change in sedimentation rate occurred over a distance of less than 150km. Erosion of the pre-R4 sequence along the western margin of Rockall Plateau at the level of reflector R4 would seem a more likely explanation for this thickness variation; however, poor resolution of pre-R4 reflector in the Iceland Basin preclude proving this.

In the southern Labrador Sea, several prominent seismic reflectors occur above the presumed upper Paleocene basalts and below reflector R4 in multichannel profiles (Line BGR 1; Fig. 10). Two sedimentary seismic reflectors, designated here P1 and P2, bracket a series of horizons that strongly onlap reflector P2 (Fig. 10). Because of the proximity of the NNW end of BGR 1 to the Greenland Margin (Fig. 1), the progressive onlap of reflector P1 may represent a lowstand deposit (sensu Vail et al., 1977) resulting from progressive outbuilding during a eustatic lowering of sea level. Crustal age underlying reflectors P1 and P2 on BGR line 1 (~ shotpoint 4400; Fig. 10) is Late Cretaceous (pre-Anomaly 31; Srivastava, 1978) and the supposed lowstand fan could be the result of any one of a series of Late Cretaceous to Eocene eustatic lowerings of sea level (Vail et al., 1977). However, if the strong reflectors are assumed to be extrusive (upper Paleocene) basalts, then the lowstand may correspond to an early Eocene major sea-level lowstand (Vail et al., 1977). In any case, the flat-lying reflectors of the pre-R4 interval on BGR 1 (Fig. 7) show no evidence of differential deposition or development of sediment waves; current effects appear to be limited to the post-R4 section.

The best evidence for pre-late Eocene current effects comes from the western basin (North American) of the North Atlantic. On the western Bermuda Rise, Mountain (1981) found unusual wave-like variations in thickness between horizons A^c (lower to middle Eocene) and A^* (upper Maestrichtian). He attributed these to possible Paleocene to early Eocene syndepositional control by bottom currents or to erosion at the level of Horizon A^c. Unfortunately, the upper horizon has not been firmly established as Horizon A^c; it is possible that it is equivalent to Horizon A^u and therefore represents a later erosional event. However, if the "waves" are attributed to early Eocene erosion at the level of Horizon A^c, a major erosional event must have occurred in the early Eocene.

The possibility that such a circulation event occurred in the early Eocene North Atlantic would support Berggren and Hollister's (1974) contention that increased early to middle Eocene deposition of biosiliceous sediments in the North Atlantic resulted from the first formation, sinking, and upwelling of cold, Arctic waters in the northern North Atlantic. This correlation is reasonable, but it is possible that the biosiliceous sedimentation was stimulated by other mechanisms (see Chapter 4). The apparent lack of similar wave-like features like those noted by Mountain (1981) elsewhere in the North Atlantic suggests that the event was not due to the early Eocene entry of bottom waters from the north. Still, pre-R4 data in the northern North Atlantic are generally poorly resolved, and the possiblity exists that some pre-R4 current-influenced deposits remain undetected.

Reflector R4 and Horizon A^u denote the most dramatic change in sedimentary regime in the Cenozoic stratigraphic record of the North Atlantic. Pre-R4 (and similarly, pre-A^u) evidence for any significant abyssal circulation is debatable, but current influences in the post-R4 interval are well developed, as discussed below.

Reflectors R4 and R3

Distribution

Reflector R4 shows the earliest unambiguous and regionally important evidence for strong abyssal circulation in the northern North Atlantic. Differential thickening of sedimentary sequences that is typical of current-controlled deposition occurs above this horizon (Figs. 4-13). In most of the pre-R4 section strata are either largely conformable with basement (typical of pelagic accumulation), attributable to fan development, or attributable to syndepositional faulting. Reflector R4 is an interbasinal reflector which has been reported throughout the Rockall region (Roberts, 1975; Jones et al., 1970), the Iceland Basin (Ruddiman, 1972), the Porcupine Abyssal Plain (Dingle et al., 1982), the Bay of Biscay (Roberts et al., 1979), and the southern Labrador Sea (Egloff and Johnson, 1975). Reflector R4 originally was first defined in the Rockall Plateau region (Jones et al., 1970; Roberts, 1975); it has been identified in boreholes drilled there as a synchronous late Eocene to earliest Oligocene seismic reflector (within uncertainties of correlation into the boreholes and of biostratigraphic age assignments; see below)(Miller and Tucholke, in press; Appendix 3). However, a continuous and unequivocal tracing of reflector R4 from the Rockall Plateau boreholes into these other regions studied is not possible. This situation led Dingle et al. (1982) to use the term "Challenger horizon" for a reflector that they thought was equivalent to reflector R4. However, similar seismic character (amplitude, stratigraphic relations of major reflectors) and stratigraphic position of a major reflector observed throughout the northern North Atlantic suggest that reflector R4 occurs in all these regions. Borehole correlations in the Labrador Sea, Bay of Biscay, and Goban Spur region (Fig. 1) support this contention.

Reflector R4 is overlain by reflector R3 in the Rockall region, Porcupine Abyssal Plain, and Labrador Sea (Figs. 4-13). As with reflector R4, identification of reflector R3 outside of the Rockall region (Fig. 4) is tentative, because a continuous and unequivocal tracing of reflector R3 into other regions is not possible. Reflector R3 is of middle to late Oligocene age (Miller and Tucholke, in press;

Appendix 3). The R3-R4 seismic interval commonly has high amplitude, discontinuous reflectors that tend to obscure the underlying reflector R4 (Fig. 4)(Roberts, 1975). Reflector R3 has no consistent age or lithologic correlation (i.e. it correlates with the middle Oligocene at Sites 406 and 117 and the uppermost Oligocene at Site 116; Fig. 2). In fact, the high amplitude, discontinuous nature of reflector R3 suggests that it may cap a sequence of cherty chalks (Roberts, 1975); the diachronous age correlations of reflector R3 may be explained if it is equivalent to such a diagenetic boundary.

Rockall region

Seismic stratigraphic relationships show that reflector R4 represents the most important erosional event in the Rockall region (i.e. between the Mid-Atlantic Ridge and the Irish Continental Margin; north of the Charlie-Gibbs Fracture Zone and south of the Greenland-Scotland Ridge; Fig. 1). Although reflector R4 generally has not been re-excavated by younger erosional events (with the possible exception of channels such as that noted at the foot of George Bligh Bank; fig. 35 in Roberts, 1975), erosion associated with reflector R4 consistently excavated and truncated older strata (Figs. 5,6; figs. 29,30,32,33 in Roberts, 1975; fig. 4 in Dingle et al., 1982). Reflector R4 in the Rockall area shows a strong correlation to the regionally important hiatus that straddles the Eocene/Oligocene boundary (Fig. 2), except possibly at Site 406 where borehole correlations (Miller and Tucholke, in press; Appendix 3) suggest that it may underlie the unconformity by as much as 45m.

The unconformity generally associated with reflector R4 is best developed along the margins of the basins (e.g. Sites 117, 403, 404, 405; Figs. 2), probably due to local intensification of abyssal currents by topographic boundaries. Away from basin margins the unconformity may not be present, for reflector R4 correlates with the uppermost Eocene to lower Oligocene section that appear to be continous within the resolution of biostratigraphy (e.g. Site 116; Fig. 2). Thus, the unconformity usually equivalent to reflector R4 can be traced to its correlative conformity (= seismic sequence boundary of Vail et al., 1977) near the top of the Eocene. The cause of reflector R4 where the section is apparently conformable is not clear; it may result from an impedance contrast caused by a lithologic change associated with the changing hydrographic regime. It is also possible that the impedance contrast occurs across an as yet unresolved unconformity or diastem.

Reflector R3 onlaps reflector R4 near the margin of Rockall Plateau (Figs. 4,8,9; figs. 19,20,29,30,32,35 in Roberts, 1975); reflector R3, in turn, is often onlapped by overlying horizons (Figs. 4,6,9). The onlap of reflector R4 by reflectors biostratigraphically dated as Oligocene through late Miocene (Miller and Tucholke, in press; Appendix 3; see below) demonstrates that patterns of non-deposition and erosion resulting from vigorous bottom circulation have dominated the margin of Rockall Plateau since R4 time.

Differential thickening of strata between reflectors R3 and R4 indicates that initial development of sediment drifts began in the R3-R4 interval (Figs. 6-9). The development of such current-controlled depositional features is indicated by contorted internal reflectors and possible local development of sediment waves in the R3-R4 interval (Fig. 8). Still, the erosion that created reflector R4 resulted in a hiatus that often encompasses the R3-R4 interval (Figs. 2,3), demonstrating that the R3-R4 interval is primarily erosional, especially along the margin of Rockall Plateau.

In the Rockall region, several sedimentary ridges, both large and small, are built upon reflector R4, including a drift in the Hatton-Rockall Basin (Figs. 4;14), the Feni Drift (Figs. 5,6), Gardar Drift (Fig. 7), Hatton Drift (Fig. 8), and several unnamed drifts east of Eriador Seamount (named Eriador Drift herein; Fig. 9), south of Fanghorn Bank (named Fanghorn Drift herein; Fig. 14), and NW of the "elbow" formed between Bill Bailey and Faeroe Banks (Fig. 1). Accumulation rates increased (Table 1) in the Miocene both on these drifts and on drifts in the western North Atlantic (Blake, Greater Antilles, Bahama, and Hatteras Ridges; Ewing and Hollister, 1972; Tucholke and Mountain, 1979). Therefore, the drifts illustrated here, although present since the Oligocene, are composed primarily of Miocene and younger sediments (i.e. post-R2 sediments on Figs. 4-12). This is especially true along the shallower margins of the drifts, where currents are topographically

TABLE 1

ESTIMATES OF SEDIMENTATION RATES IN BASINS IN THE NORTHERN NORTH ATLANTIC

(given in meters/million years)

BASIN	INTERVAL				
	post-R1 0-8 Ma	R1-R2 8-17 Ma	post-R2[1] 0-27 Ma	R2-R3 17-27 Ma	R3-R4 23-37 Ma
Hatton-Rockall Basin (Site 116; Figs. 2,4)					
estimated[2]	36	33	34	11	12
actual[3]	~ 45	~ 25	--	11	10
Fanghorn Drift (Site 406)					
estimated[2]	44	24	34	~ 7	~ 6
actual[3]	45	34	--	late Oligocene: 22 early Oligocene: 0	
Iceland Basin (Fig. 7)	--	--	36	26	18
Gardar Drift	--	--	46	30	--
Hatton Drift (Fig. 8)	--	--	21-23	12	18
Eriador Drift (Fig. 9)	--	--	31	22-35	~ 13
Feni Drift (Fig. 5)	--	--	--	post R3: 20	31
Porcupine Drift (Fig. 6)	--	--	31	26	18
Eirik Drift[4]	--	--	51	26	13

Table compiled by measuring intervals at depocenters of the sediment drifts for each interval on seismic profiles illustrated in Figures 4-9, assuming average interval velocity of 1750 m/sec. Not corrected for compaction. R3-R4 estimates are minima since no hiatuses are assumed to be present.

[1] Reflector R1 not recognized.
[2] Estimated from seismic stratigraphy.
[3] Actual sedimentation rate computed accounting for hiatuses. (After Laughton, Berggren et al., 1972 and Montadert, Roberts et al., 1979; Site 116 Oligocene rates after Miller and Tucholke, in press; Appendix 3).
[4] Reflector R2 identification speculative on Eirik Drift.

intensified and much of the section has been removed (or was never deposited). Since most DSDP sites have been located on the margins of the drifts (e.g. Sites 117, 403, 404, 405), it is easy to draw the erroneous conclusion that drift deposition began in the Miocene (e.g. Schnitker, 1980a). This is not the case, for drift development can be seen clearly in the Oligocene section between reflectors R3 and R4 and in the Oligocene to lower Miocene between reflectors R2 and R3 (Figs. 4-12).

Bay of Biscay/Porcupine Abyssal Plain/Goban Spur regions

In the Bay of Biscay/Goban Spur region, reflector R4 probably correlates with an unconformity straddling the Eocene/Oligocene boundary at Sites 548, 401, 402, 118, 119, 400A, and 550 (Fig. 3). However, this is not certain due to difficulties in tracing the horizon into these boreholes, incomplete coring, and poor placement of these boreholes for stratigraphic correlations of the Tertiary section. Throughout the abyssal portion of the Bay of Biscay, turbidite deposition dominates the sedimentary record; here, reflector R4 is often not well defined seismically, although it is present (e.g. Site 118; Laughton, Berggren et al., 1972). In areas shielded from turbidite deposition, such as Cantabria Seamount (Site 119; figs. 3,6 in Chapter 10 in Laughton, Berggren et al., 1972), a prominent horizon that correlates in time with reflector R4 is observed; on such topographic highs, the seismic character of reflector R4 is similar to that observed in the Rockall region.

Reflector R4 is well developed in the Porcupine Abyssal Plain region (Dingle et al., 1982; this study). In the western portion of this basin, a large (> 250km long) previously unnamed sediment drift is observed (Fig. 6). The Porcupine Drift (named herein) is a NW-SE trending sediment ridge developed against the eastern flank of the east Thulean Rise south of the Charlie-Gibbs Fracture Zone (Fig. 1). It may be considered a southern extension of the Feni Ridge (although the latter occurs in a different physiographic province, the Rockall Trough, and it has a different orientation, NNE-SSW). The post-R4 interval of the Porcupine Drift shows patterns of differential deposition and sedimentary waves (Fig. 6) that document the presence of strong abyssal circulation.

Labrador Sea

In the southern Labrador Sea, a prominent mid-sediment reflector was first noted by Jones et al. (1970) and Egloff and Johnson (1975), who suggested that it correlated with reflector R (= R4) in the Rockall region. Above this reflector, two large sediment drifts are developed: the Eirik Drift south of Greenland (Figs. 1,10,12) and the Gloria Drift (Figs. 1,11). Seismic comparison with the Rockall region and borehole correlations at Site 112 in the Gloria Drift (Fig. 1; fig. 5 in Appendix 1)(see below) suggest that the top of this mid-sediment reflector may be reflector R3, and that reflector R4 is seismically masked beneath this high-amplitude event over much of the southern Labrador Sea (Figs. 10-12). This conclusion needs to be tested by future drilling. The R3/R4 couplet has been traced from Site 112 throughout the Gloria Drift, but direct correlations into the Eirik Drift are blocked by basement interruptions at the extinct spreading axis. However, the similar seismic character of the R3/R4 couplet in the Gloria Drift and the Eirik Drift (cf. Figs. 11,12) suggests that reflector R4 occurs in the deep basin adjacent to the southern Greenland margin. A major seismic discontinuity thought to be equivalent to reflector R4 has also been noted in the deep basin adjacent to the eastern margin of Greenland by Featherstone et al. (1977).

It is not possible to trace the R3/R4 couplet from the southern Labrador Sea into the central and northern Labrador Sea in available single channel seismic data. North of the Gloria and Eirik Drifts, turbidite deposition dominates the sedimentary record in the upper 1km of section (see also Davies and Laughton, 1972). Unfortunately, seismic reflectors are generally not resolvable below about 1 sec (~ 1km) in the single channel seismic profiles examined; this prevents detecting reflectors R3/R4 which lie deeper than 1 sec. However, three distinct reflectors are observed in multichannel seismic data from the central and northern Labrador Sea (Lines BGR 17, 21, and 12; fig. 1 in Hinz et al., 1979): U, E/O, and E in descending stratigraphic order. It is suggested here that horizon E/O is equivalent in age to reflector R4, and that the event associated with reflector R4 affected the entire deep Labrador Sea including the lower continental rise of Labrador and Greenland.

Age of reflector R4

Jones et al. (1970) cored upper Oligocene sediments from immediately above reflector R4, establishing its age as pre-late Oligocene. Ruddiman (1972) suggested that reflector R4 was early Oligocene based upon its pinchout on crust older than Anomaly 13 (early Oligocene). Based upon corrlations to DSDP boreholes in the Rockall region, Roberts (1975) and Roberts et al. (1979) suggested that reflector R4 is late Eocene to early Oligocene. Miller and Tucholke (in press; Appendix 3) traced reflector R4 to DSDP boreholes throughout the northern North Atlantic; they used velocity-depth data obtained from sonobuoy measurements to determine subbottom depths, and then adjusted the placement of reflectors within geologically reasonable limits. The correlations obtained in this manner agree well with those obtained by Montadert, Roberts et al. (1979) and Roberts et al. (1979) who used synthetic seismograms computed from sonic logs to constrain the depth of reflectors drilled by Leg 48 in the Rockall region. Reflector R4 correlates with a major unconformity at Sites 403 and 404 that separates upper Miocene from Eocene strata, while at Sites 117 and 403 this reflector separates upper Oligocene from lower to middle Eocene strata (Fig. 2). The unconformity associated with reflector R4 can be traced to its apparent correlative conformity (= sequence boundary, sensu Vail et al., 1977) at three locations. 1) At Site 116 the horizon correlates with an apparently continuous section in an uncored interval separating upper Eocene from lower Oligocene strata (Fig. 2). 2) At Site 112 in the southern Labrador Sea, reflector R4 appears to lie at 0.41 sec and to correlate with the lower Oligocene section (Miller et al., 1982; Appendix 1). However, the reflector at 0.41 sec is the top of a high-amplitude unit that may obscure underlying horizons (Fig. 11). As suggested above, top of this high amplitude unit may be reflector R3, and reflector R4 may be masked; in fact, reflector R4 probably underlies the top of the high amplitude unit by 0.1 sec at Site 112 (profile V30-09, Fig. 11). Recomputing the depth of reflector R4 assuming that it lies at 0.51 sec suggests that it correlates with the uncored interval that straddles the Eocene/Oligocene boundary at Site 112 (fig. 5 in Appendix 1). 3) At Site 406 reflector R4 correlates with undifferentiated upper Eocene sediments underlying a major unconformity separating upper Oligocene from upper Eocene strata (Fig. 2). Although the correlation into Site 406 is probably valid (see Appendix 3), correlation of reflector R4 with the lower Oligocene unconformity would still yield geologically a reasonable interval velocity (~ 1850 m/sec).

Given the record recovered, uncertainties in correlation into the boreholes, and biostratigraphic problems, the age of reflector R4 could range from ~ 40-35 Ma (time scale of Hardenbol and Berggren, 1978). However, based upon the correlation at Site 116 and the revised correlation at Site 112, Reflector R4 is probably latest Eocene to earliest Oligocene (~ 38-36 Ma).

Recent drilling in the Rockall region (Leg 81) and the Goban Spur (Leg 80) tends to support the age assignment of reflector R4 to the latest Eocene to earliest Oligocene. However, the middle Eocene to middle Miocene sedimentary section is very thin in the sites drilled by Leg 81 (Roberts, Schnitker et al., in preparation), precluding seismic differentiation of reflector R4 from overlying horizons and preventing a firm age correlation. Reflector R4 has not been unambiguously traced to boreholes drilled by Leg 80 in the Goban Spur region (de Graciansky, Poag et al., in preparation), and correlations there are speculative. Near Site 548 on the Goban Spur (Figs. 1,3), reflector R4 probably occurs at 0.41 sec on Challenger 48 and V28-04 profiles; assuming a seismic velocity of 1750 m/sec, reflector R4 correlates with a probable earliest Oligocene (~ 36.0 Ma) unconformity. At Site 551 reflector R4 occurs just below the seafloor; the uncored section here only constrains the age as pre-Pleistocene and post-middle Eocene. A very similar pattern of onlap onto reflector R4 by the post-R4 series can be seen at Site 551 and in the Hatton-Rockall Basin where reflector R4 and the post-R4 sequences are better defined (cf. Fig. 4 and Fig. 13). Reflector R4 probably occurs at 0.35 sec at Site 550; it correlates with the thin late Eocene to middle Oligocene section. At Site 549, reflector R4 probably lies at 0.14 sec, correlating with the latest Eocene to earliest Oligocene interval (38.0-36.0 Ma; assuming interval velocities between 1.5 and 2.0 km/sec).

Reflector terminations against oceanic basement dated with magnetic anomalies provide an independent, albeit crude, check on the age of the reflectors. Ruddiman (1972) noted the termination of reflector R4 against a crustal high on the Gardar Ridge (Iceland Basin; Fig. 1), which was dated as anomaly 13 (37.0 Ma on the time scale of Heirtzler et al., 1968; 35.5 Ma on the time scale of Hardenbol and Berggren, 1978). Use of the latter time scale would suggest that reflector R4 is somewhat younger than indicated from the borehole correlations discussed here. However, this apparent discrepancy may be resolved. In the Iceland Basin (Fig. 7), the pinchouts of reflector R4 occur between anomaly 13 and anomaly 21 (50.5 Ma, time scale of Hardenbol and Berggren, 1978). The terminations are closer to anomaly 13 than anomaly 21 (all magnetic anomaly identifications after Vogt and Avery, 1974). South of the Charlie-Gibbs Fracture Zone, reflector R4 also terminates between anomaly 13 and anomaly 21, while in the southern Labrador Sea reflector R4 terminates between anomalies 13 and 24. These terminations do not contradict the latest Eocene to earliest Oligocene age of reflector R4.

Correlations into Labrador Margin exploration wells suggest that horizon E/O (Fig. 15) is equivalent to reflector R4. Hinz et al. (1979) suggested that a lower reflector, horizon E, noted in the deep basin (Fig. 15) could be traced to the top of the Eocene in the Labrador Shelf exploration wells; horizon E would thus be the equivalent in age of reflector R4. However, examination of well logs, velocity logs, paleontological data, and seismic data (courtesy of F. Gradstein, S. Srivastava, A. Grant; Canadian Geological Survey) suggests that, due to problems in tracing horizons from the deep basin through the slope, it is equally likely that horizon E/O represents the top of the Eocene (Fig. 15). In the deep basin, horizon E pinches out between anomalies 27 and 28 (early Paleocene) on BGR 17 and BGR 21, and between anomalies 25 and 26 on BGR 12 (middle Paleocene); borehole correlations confirm that horizon E can be no older that early Eocene (Fig. 15). On BGR 17 (~ shotpoint 3500), horizon E/O merges with horizon U; it is believed that this is due to downlap of horizon U onto horizon E/O.[1] Horizon E/O cannot be traced on BGR 17, BGR21, and BGR 12 in the disturbed section between anomaly 24 and the extinct ridge axis; in fact, continuous reflectors generally cannot be traced through this zone (see also Le Pichon et al., 1971). Similarly, reflector R4 terminates against anomaly 24 in much of the southern Labrador Sea (anomalies identified after Srivastava, 1978). The similar pinchouts of reflector R4 and horizon E/O in the Labrador Sea support the contention that the two may be equivalent in age.

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[1] Hinz et al. (1979) showed that on profile BGR 17 horizon E merges with horizon U on their fig. 4, but that an overlying horizon (= E/O) merges with horizon U in their fig. 7; this discrepancy is due to mislabeling of horizon E on fig. 4 (S. Srivastava, personal communication).

Reflectors R2 and R1

Roberts (1975) and Miller and Tucholke (in press; Appendix 3) noted that reflector R2 overlies the R3/R4 couplet in the Hatton-Rockall Basin (Fig. 4) and the southwest margin of Rockall Plateau (Fanghorn Drift; Fig. 14) (Table 1). Reflector R2 is an upper lower Miocene reflector at the upper boundary of silica-rich sediments (Fig. 2). Ruddiman (1972) noted a relatively continuous intermediate reflector (IR) above reflector R4 in the Iceland Basin. He dated this horizon as early Miocene (17 Ma or younger, based on the age of pinchout on oceanic crust), and attributed it to a change in abyssal circulation. This intermediate reflector (IR) therefore correlates in time with reflector R2. In the Iceland Basin, Reflector R2 (= IR) terminates against anomaly 5 on the profiles examined here, consistent with Ruddiman's interpretation and the age assignments for reflector R2 obtained from correlation to the boreholes (Fig. 2). Reflector R2 may correlate with a minor unconformity noted in the boreholes (Miller and Tucholke, in press; Appendix 3). The seismic interval between reflector R2 and the seafloor shows the most coherent pattern of current-controlled deposition in the northern North Atlantic in the form of prominent sediment drifts and associated sediment waves (Fig. 7). Sedimentation rates increased significantly in the Hatton-Rockall Basin and on the Fanghorn, Gardar, Hatton, and Eriador Drifts in the post-R2 interval (Table 1).

Roberts (1975) and Miller and Tucholke (in press; Appendix 3) noted that reflector R1 overlies reflector R2 in the Hatton-Rockall Basin (Fig. 4) and the southwest margin of Rockall Plateau. Reflector R1 falls within upper Miocene calcareous chalks and may represent a lithification boundary (e.g., chalk/ooze) in the calcareous sediments. Along portions of the margin of Rockall Plateau (e.g., Sites 403-405; Fig. 2) deposition did not begin until R1 time (Fig. 2). Sedimentation rates increased in the Hatton-Rockall Basin and on the Fanghorn Drift in the post-R1 interval (Table 1).

The Abyssal Circulation Model

Reflector R4 occupies a chronostratigraphic position similar to Horizon A^u in the western North Atlantic. Stratal patterns associated with these reflectors are similar, i.e. truncation of deeper strata and the development of current-controlled deposition above. Thus, it is likely that both horizons had a similar cause: the development of vigorous bottom water circulation near the end of the Eocene (Miller and Tucholke, in press; Appendix 3).

Sediment distribution patterns provide evidence for a strong contour-following bottom water flow beginning at reflector R4 time. The widespread distribution of reflector R4 and its characteristic correlation with an unconformity (and similarly with Horizon A^u; Tucholke and Mountain, 1979) indicates that strong abyssal circulation affected the North Atlantic basins both east and west of the mid-ocean ridge beginning in the latest Eocene to earliest Oligocene. Along the margins of Rockall Plateau, the unconformity is especially well developed, demonstrating that topographic boundary effects probably intensified the abyssal circulation. The margin-intensified, anticlockwise circulation in the northeastern Atlantic was contained by the topographic barriers of the Rockall margin and the mid-ocean ridge (fig. 12 in Appendix 3). As a result, erosion and current-influenced deposition strongly affected the Porcupine/Biscay regions, despite the fact that these are regions of relatively quiescent bottom-water flow in the modern ocean. These erosional/depositional effects are reflected in a major unconformity noted in the Bay of Biscay/Goban Spur boreholes (Fig. 3) and in the development of the Porcupine Drift along the western margin of the Porcupine Abyssal Plain (Figs. 5,6). Further evidence of post-R4 current-controlled, contour-following deposition is given by the post-R4 sediment isopach map for the Rockall region (Fig. 14). For abyssal circulation to have affected so strongly the sedimentary record in the Rockall region and in the Bay of Biscay, it is necessary that the eastern North Atlantic had a major source of bottom water.

The latest Eocene to earliest Oligocene abyssal circulation event associated with reflector R4 (and Horizon A^u) resulted from the influx of cold bottom water formed in polar-subpolar marginal basins of the North Atlantic. Several topographic features, the Mid-Atlantic Ridge, Azores-Gibraltar Ridge, Azores-Biscay Rise, and the Madeira-Tore Rise, probably prevented significant bottom water exchange between the northeastern North Atlantic and more southerly regions below ~ 3km (Miller and Tucholke, in press; Appendix 3). Although Schnitker (1980a,b) preferred Antarctic sources for Late Paleogene bottom water in the northern North Atlantic, it is unlikely that they could have been geologically significant even in the absence of these barriers, given the minor influence of Antarctic Bottom Water in this region today. Miller and Tucholke (in press; Appendix 3) speculated as to possible sources of the high-latitude, vigorously circulating bottom water: 1) Baffin Bay-Davis Strait; 2) formation of bottom water south of the Greenland-Scotland Ridge; and 3) overflow across the Greenland-Iceland Ridge (= nascent Denmark Straits) and/or across the Faeroe Bank Channel and Wyville-Thompson Ridge. Given the intensity of abyssal circulation in the northeast North Atlantic, especially the Rockall Trough, Porcupine Abyssal Plain, and Bay of Biscay, at least one source was probably overflow across the Wyville-Thompson Ridge. Miller and Tucholke (in press; Appendix 3) noted the time correlation between reflector R4 and the separation of Greenland and Spitsbergen (Talwani and Eldholm, 1977); they suggested that following the opening of this Arctic passage, Arctic waters rapidly entered the Norwegian-Greenland Sea, flowed through the Faeroe-Shetland Channel, across the Wyville-Thompson Ridge and entered the North Atlantic (see Chapter 4 for further discussion). The influx of this vigorously circulating bottom water resulted in strongly erosional conditions that lasted throughout much of the Oligocene (i.e. the R3-R4 interval).

A major early Miocene change in abyssal circulation (at the time of reflector R2) led to increased rates of deposition on the sediment drifts. Previous studies have suggested that increased deposition resulted from intensification of bottom water flow (e.g. Schnitker, 1980a,b; see discussion in Shor and Poore, 1979). In constrast, Miller and Tucholke (in press; Appendix 3) interpreted the change from widespread erosional condition prominent over the R3-R4 interval to coherent deposition of sediment drifts in the R2-R3 and especially in the post-R2 interval to be a result of a general decrease in intensity and stabilization of abyssal circulation. This interpretation is based on the fact that erosion of the major unconformity associated with reflector R4 required much higher current speeds, for speeds were above the limiting velocity separating erosion from deposition (see McCave and Swift, 1976, and references therein). Subsequently, speeds were reduced, resulting in the initiation of deposition above reflector R4.

The change from erosion to deposition above reflector R4 occurred progressively, with deposition beginning first in basins centers and beginning later toward basin margins. This is evidenced by: 1) deposition began in the early Oligocene in sites in the center of basins (e.g. Sites 116 and 112; Fig. 2); 2) deposition began in the middle Oligocene in sites closer to basin margins (about R3 time; e.g. Site 406); 3) deposition lagged until the middle to late Miocene in sites closest to the basin margins (e.g. Sites 403-405).

The progressive narrowing of the erosional zone and change to depositional conditions that occurred from early Oligocene through late Miocene is interpreted as representing a progressive decrease in the speed of abyssal currents. Increased rates of deposition on drifts in the middle Miocene (Table 1) may have resulted, in part, from increased sediment supply resulting in increasing concentration hence greater deposition (McCave and Swift, 1976). Such an increase in sediment supply to the basins may have resulted from lowered sea level in the late Oligocene to early Miocene (Vail et al., 1977). Nevertheless, decreased current velocities are necessary to explain the progressive change from sediment erosion and transport to deposition.

The general trend toward reduced bottom water flow was probably punctuated by brief but important erosional pulses resulting from intensified circulation. Examples of such erosional pulses include development of hiatuses in the latest early Miocene (about the time of reflector R2) and near the middle/late Miocene boundary (Figs. 2,3; fig. 5 in Appendix 3; fig. 3 in Shor and Poore, 1979). Tucholke and Laine (in press) noted two major erosional events in the Miocene of the western North Atlantic: a middle middle Miocene event (= Horizon X) and a middle/late Miocene event. Although the Horizon X event apparently post-dates the R2 erosional pulse, the timing of the middle/late Miocene event correlates between basins.

The overall interpretation of the development of abyssal circulation in the North Atlantic is: 1) a rapid increase in current strength occurred during the latest Eocene to early Oligocene, creating strongly erosional conditions; 2) a general decrease in intensity of flow occurred during the Oligocene to early Miocene, resulting in increased coherence of deposition; and 3) a further decrease and stabilization of abyssal flow occurred in the middle to late Miocene; coherent development of sediment drifts has continued more or less unaltered from this time to the present (see figs. 12a-12c in Appendix 3). This long-term trend in the development of abyssal circulation in the North Atlantic was probably punctuated by the minor erosional pulses such as those discussed above. In addition, other climatic events such as the initiation of northern hemisphere glaciation (~ 2.5-3 Ma; Berggren, 1972) and the intensification of glaciation beginning about one million years ago (Shor and Poore, 1979; Prell, 1980) severely affected the lithostratigraphic record of the northern North Atlantic (Davies and Laughton, 1972). The relationships among abyssal circulation, tectonic, and climatic histories of the North Atlantic will be discussed under paleoceanographic synthesis (Chapter 4).

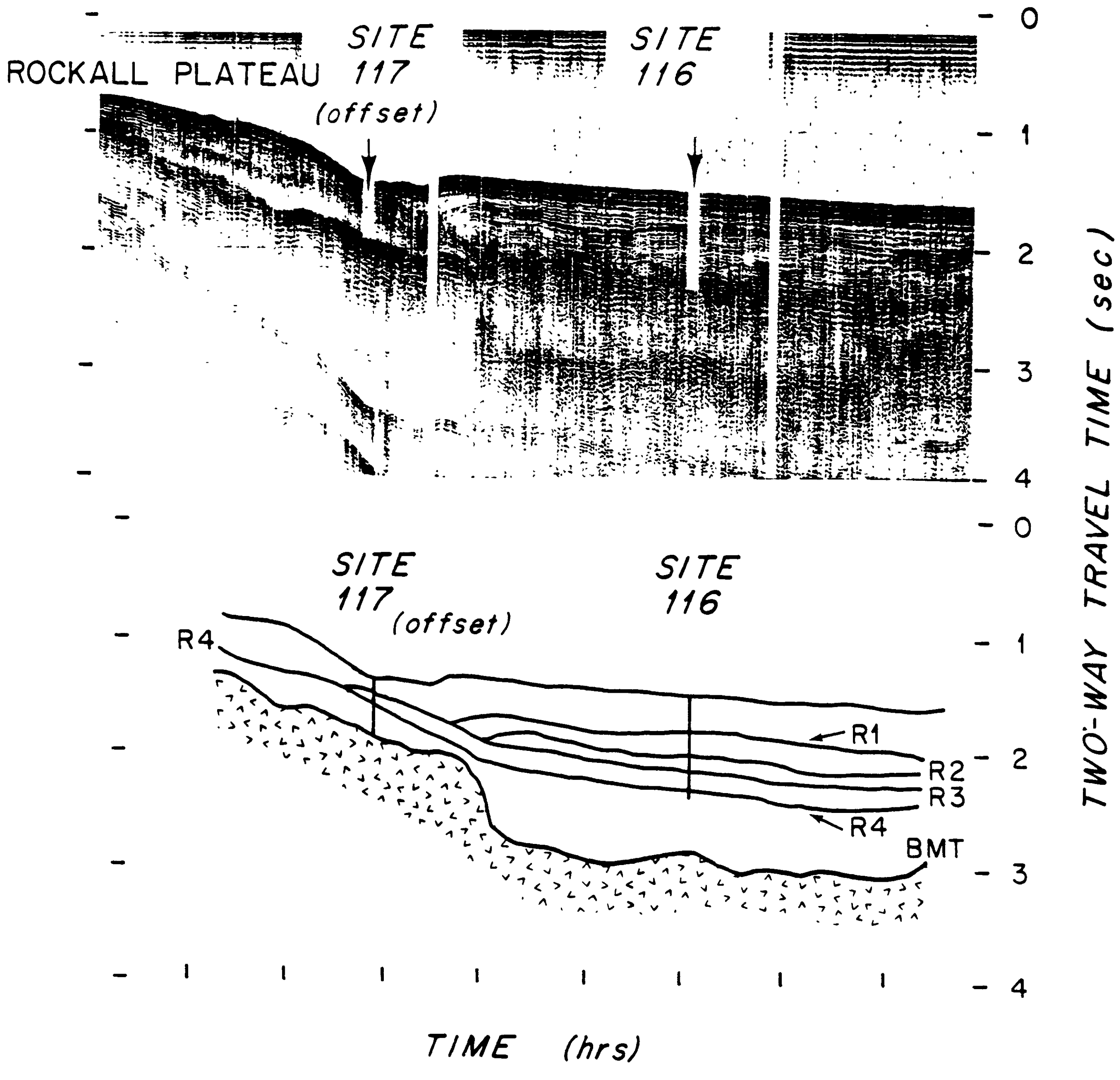


Figure 4. Seismic reflection profile and interpretation of profile V28-04 across DSDP Sites 116 and 117, Hatton-Rockall Basin. Located in Figure 1 as section G. Reflector R4 is upper Eocene to lowermost Oligocene; reflector R3 is middle to upper Oligocene; reflector R2 is uppermost lower Miocene; reflector R1 is upper Miocene. In this and subsequent figures, basement indicated is undifferentiated acoustic basement. After Miller and Tucholke (in press).

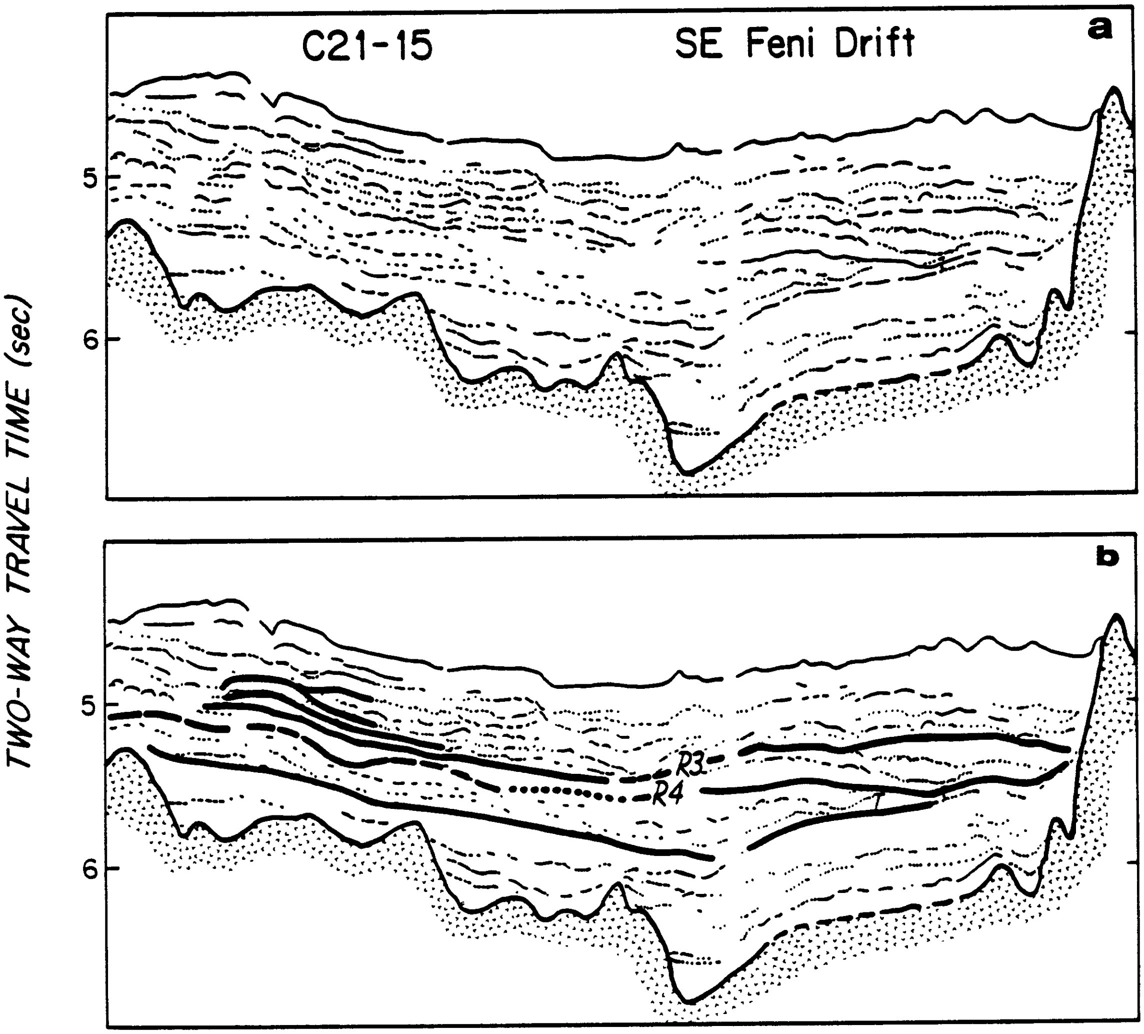

C21-15
SE Feni Drift
a
b
5
6
5
6
TWO-WAY TRAVEL TIME (sec)
R3
R4

Figure 5. Tracing (part A at top) and interpretation (part B on bottom) of profile C21-15 across the SE Feni Drift. Located in Figure 1 as section I.

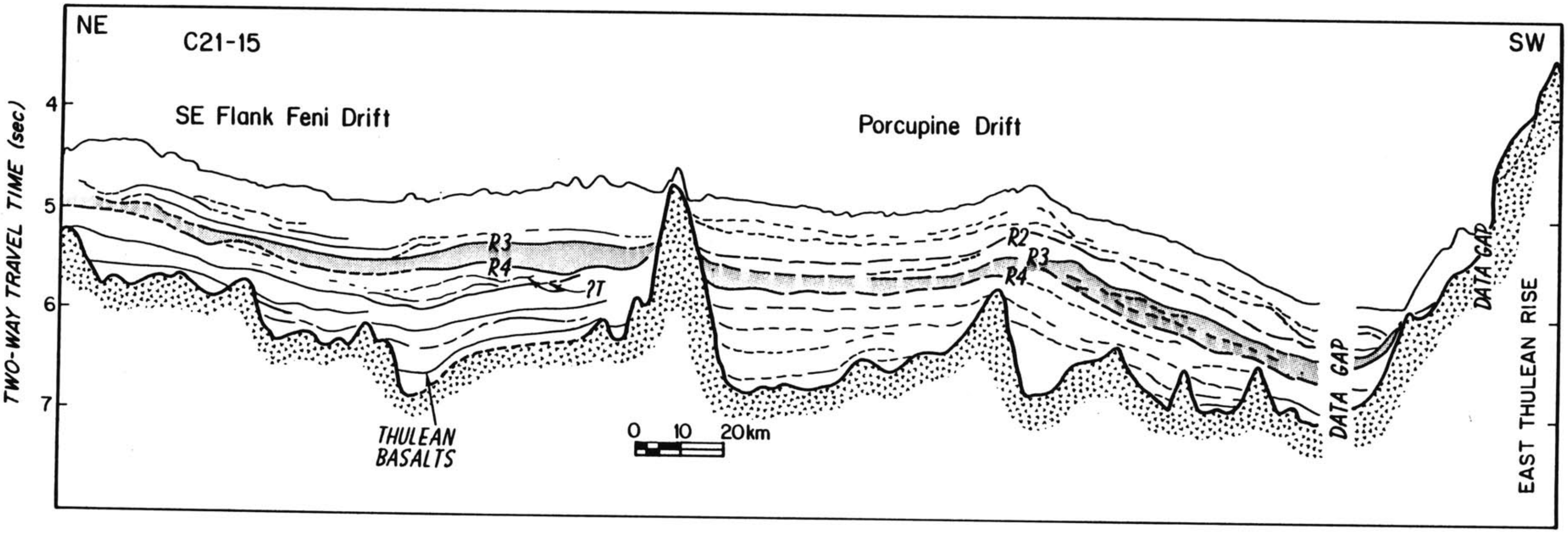

NE
C21-15
SW
SE Flank Feni Drift
Porcupine Drift
TWO-WAY TRAVEL TIME (sec)
4
5
6
7
R3
R4
?T
R2
R3
R4
THULEAN BASALTS
0 10 20km
DATA GAP
DATA GAP
EAST THULEAN RISE

Figure 6. Interpretation of profile C21-15 across the SE Feni Drift and a drift in the eastern Porcupine Abyssal Plain. Located in Figure 1 as section I. NE portion of this line is given in Figure 5. T = truncation of pre-R4 reflectors by reflector R4.

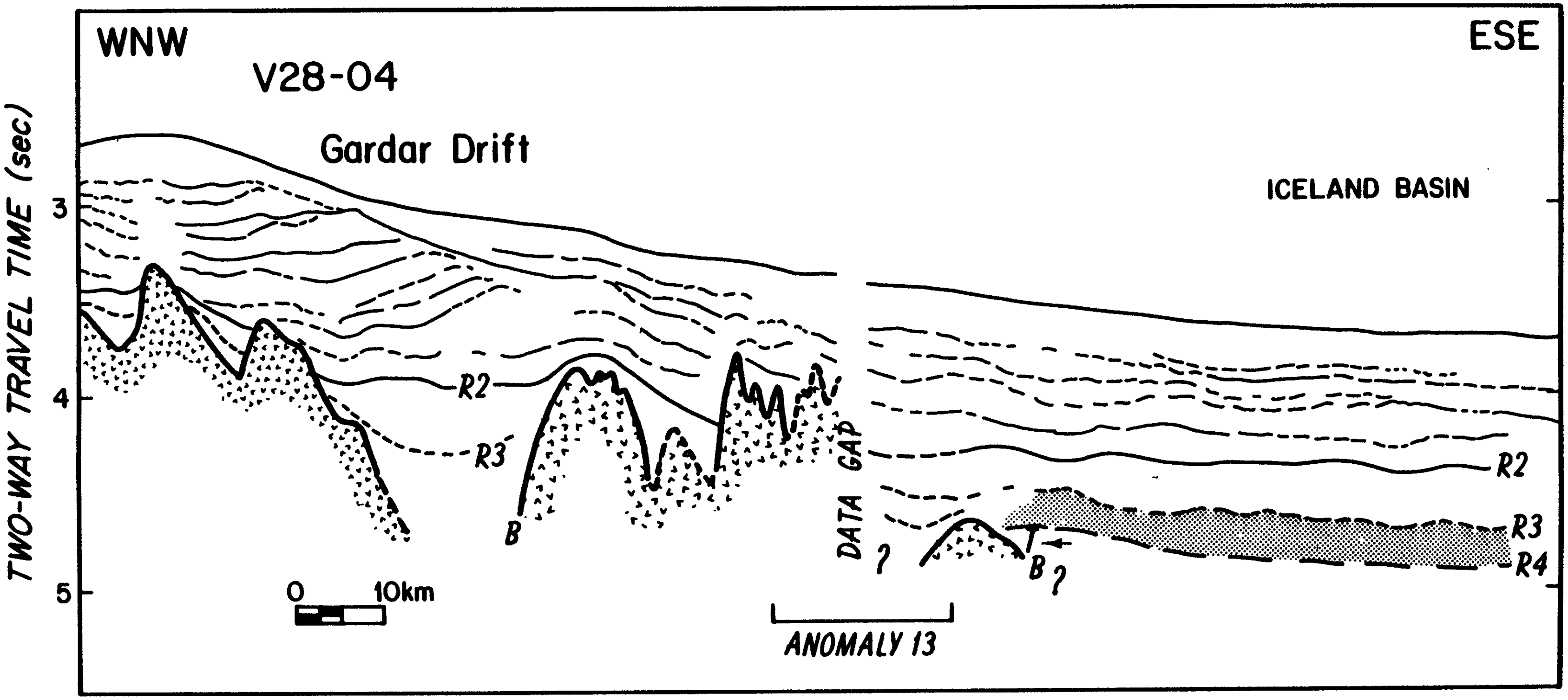

WNW
V28-04
Gardar Drift
ESE
ICELAND BASIN
TWO-WAY TRAVEL TIME (sec)
3
4
5
R2
R3
B
DATA GAP
?
T
B
?
R2
R3
R4
0
10km
ANOMALY 13

Figure 7. Interpretation of profile V28-04 across the Gardar Drift. Located in Figure 1 as section E. Reflectors R3 and R4 are poorly defined in original data. Note development of large waveforms in the post-R2 interval. Position of anomaly 13 is approximate, and follows Vogt and Avery (1974). T = termination of reflector R4. B = basement.

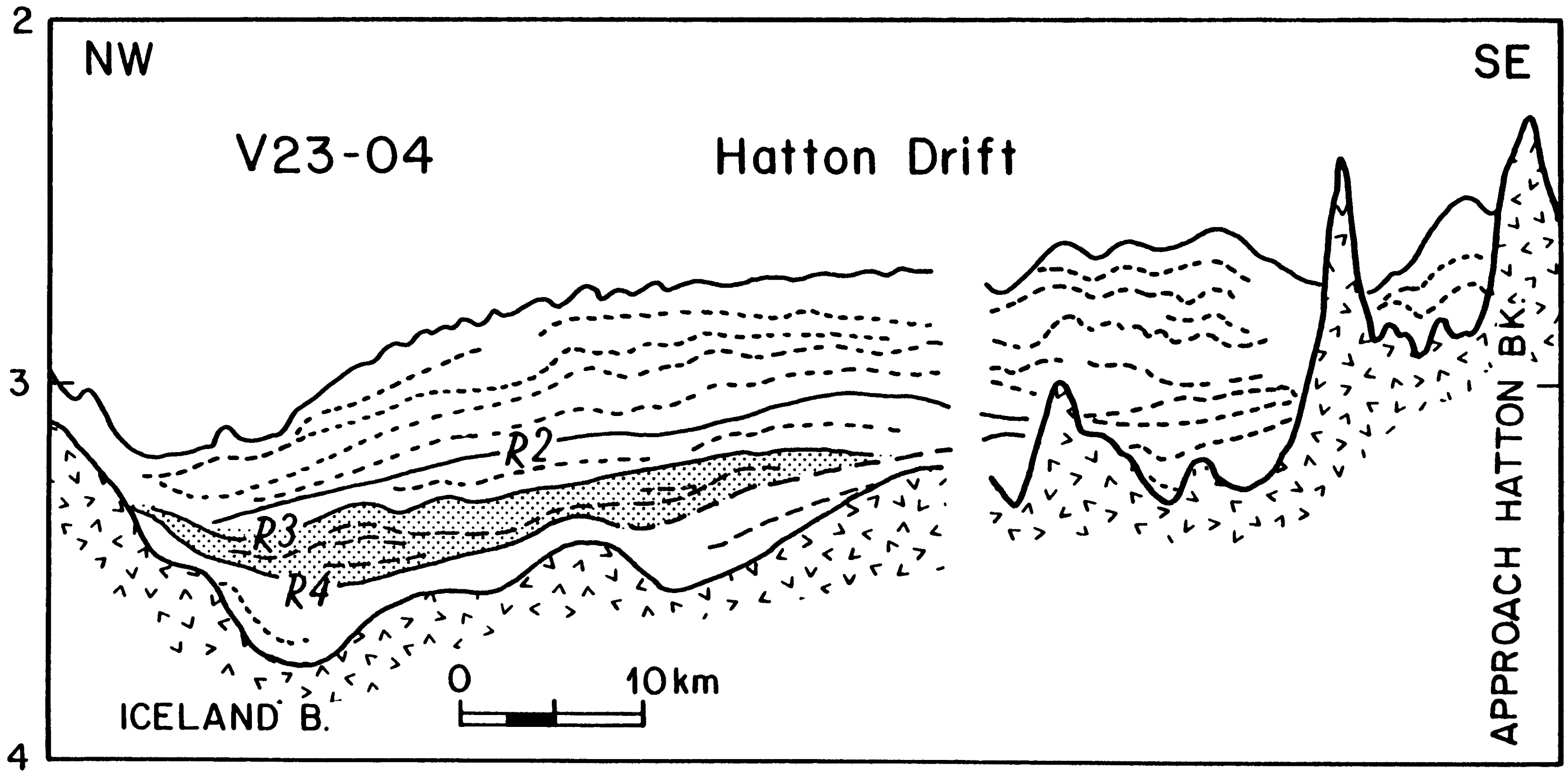

2
3
4
NW
SE
V23-04
Hatton Drift
R2
R3
R4
0
10km
ICELAND B.
APPROACH HATTON BK.

Figure 8. Interpretation of profile V23-04 across the Hatton Drift. Located in Figure 1 as section F.

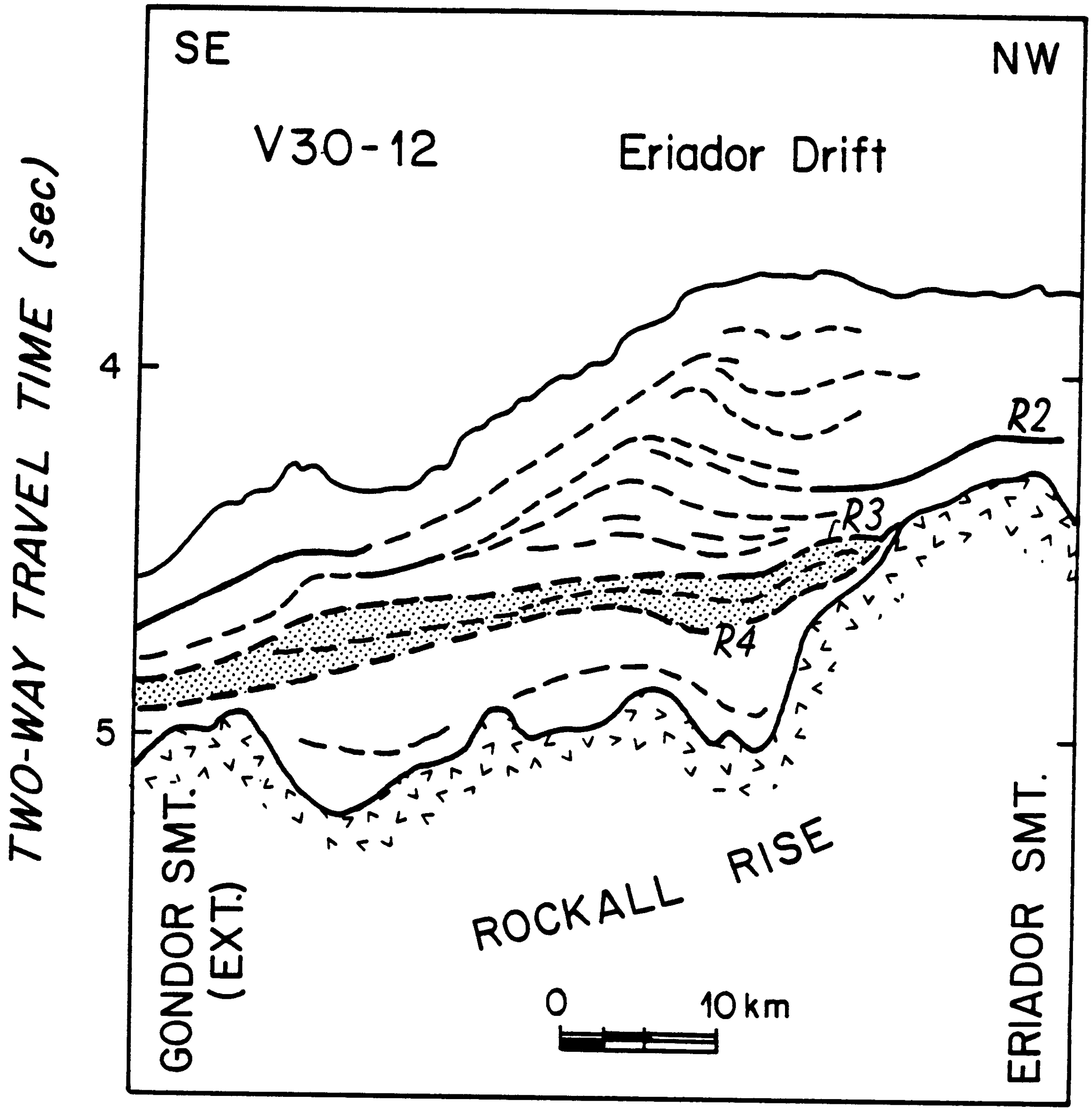

SE
NW
V30-12
Eriador Drift
TWO-WAY TRAVEL TIME (sec)
4
5
R2
R3
R4
GONDOR SMT. (EXT.)
ROCKALL RISE
ERIADOR SMT.
0
10km

Figure 9. Interpretation of profile V30-12 across a drift developed on the flank of Eriador Seamount. Located in Figure 1 as section H.

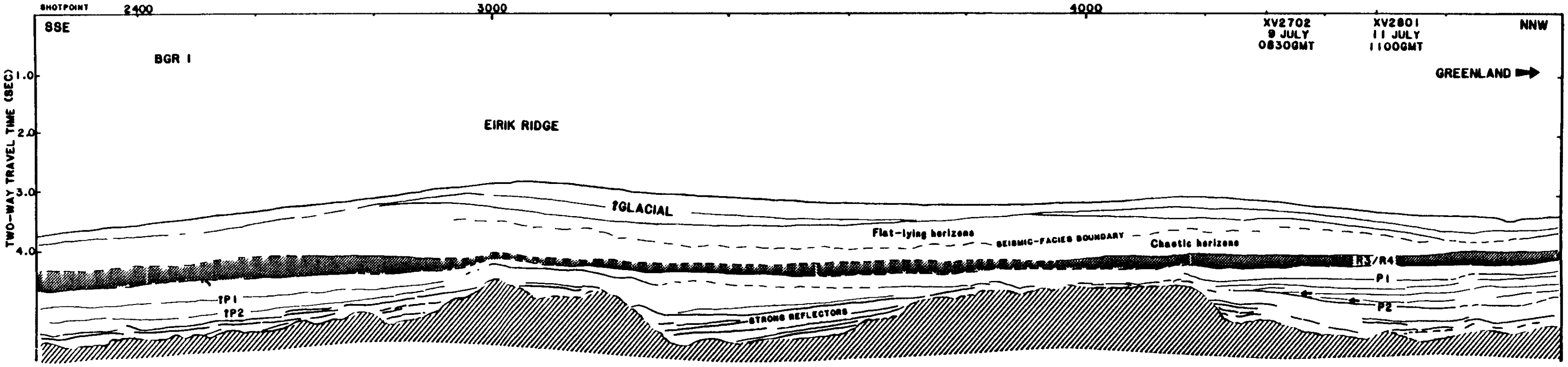

SHOTPOINT
2400
3000
4000
SSE
NNW
BGR 1
XV2702
9 JULY
0830GMT
XV2801
11 JULY
1100GMT
GREENLAND
TWO-WAY TRAVEL TIME (SEC)
1.0
2.0
3.0
4.0
EIRIK RIDGE
?GLACIAL
Flat-lying horizons
SEISMIC-FACIES BOUNDARY
Chaotic horizons
R3/R4
P1
P2
?P1
?P2
STRONG REFLECTORS

Figure 10. Interpretation of multichannel profile BGR 1 across the Eirik Drift. Located in Figure 1 as section C. Data provided by S. Srivastava and K. Hinz.

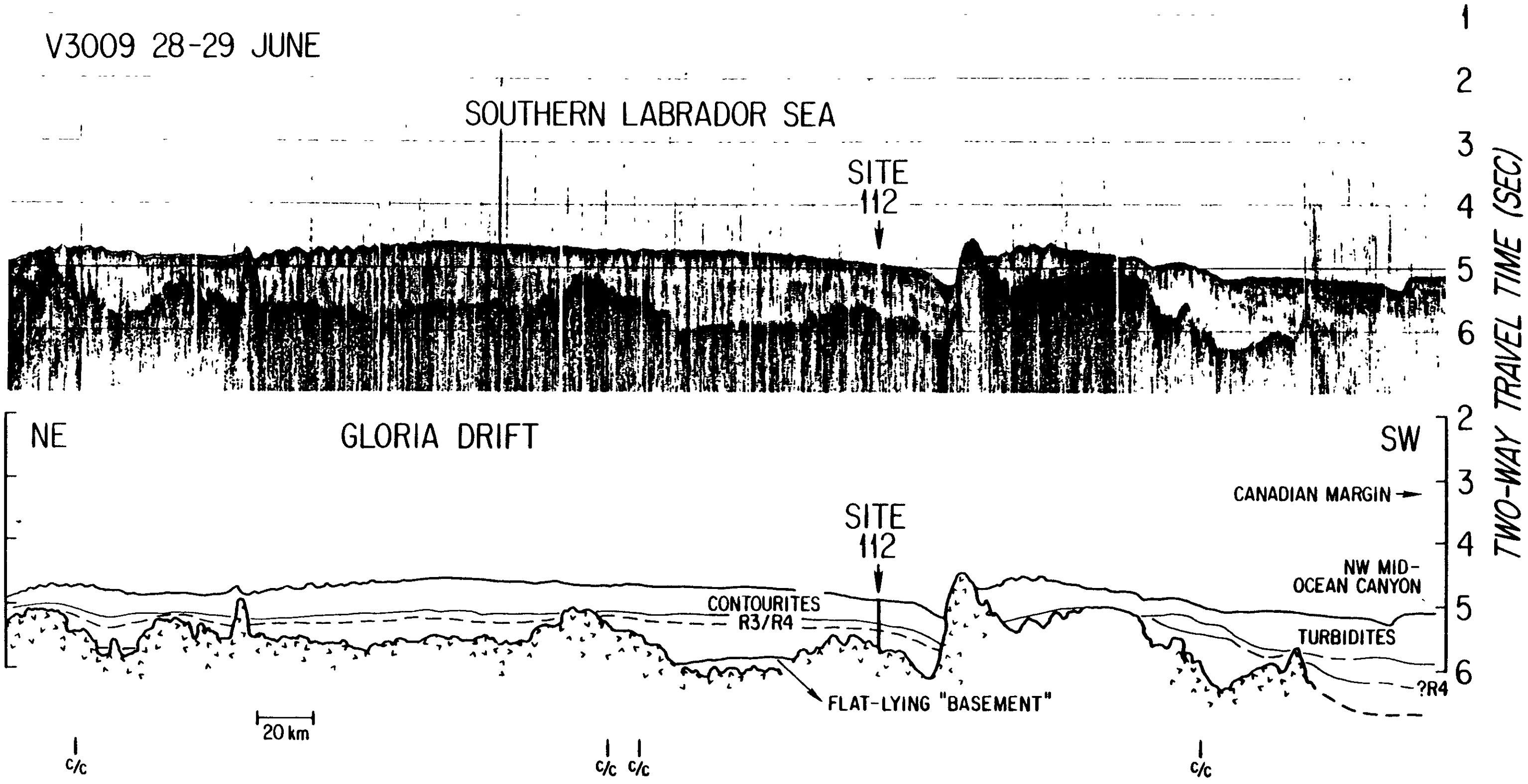


Figure 11. Seismic reflection profile and interpretation of profile V30-09 across DSDP Site 112 on Gloria Drift. Located in Figure 1 as section D.

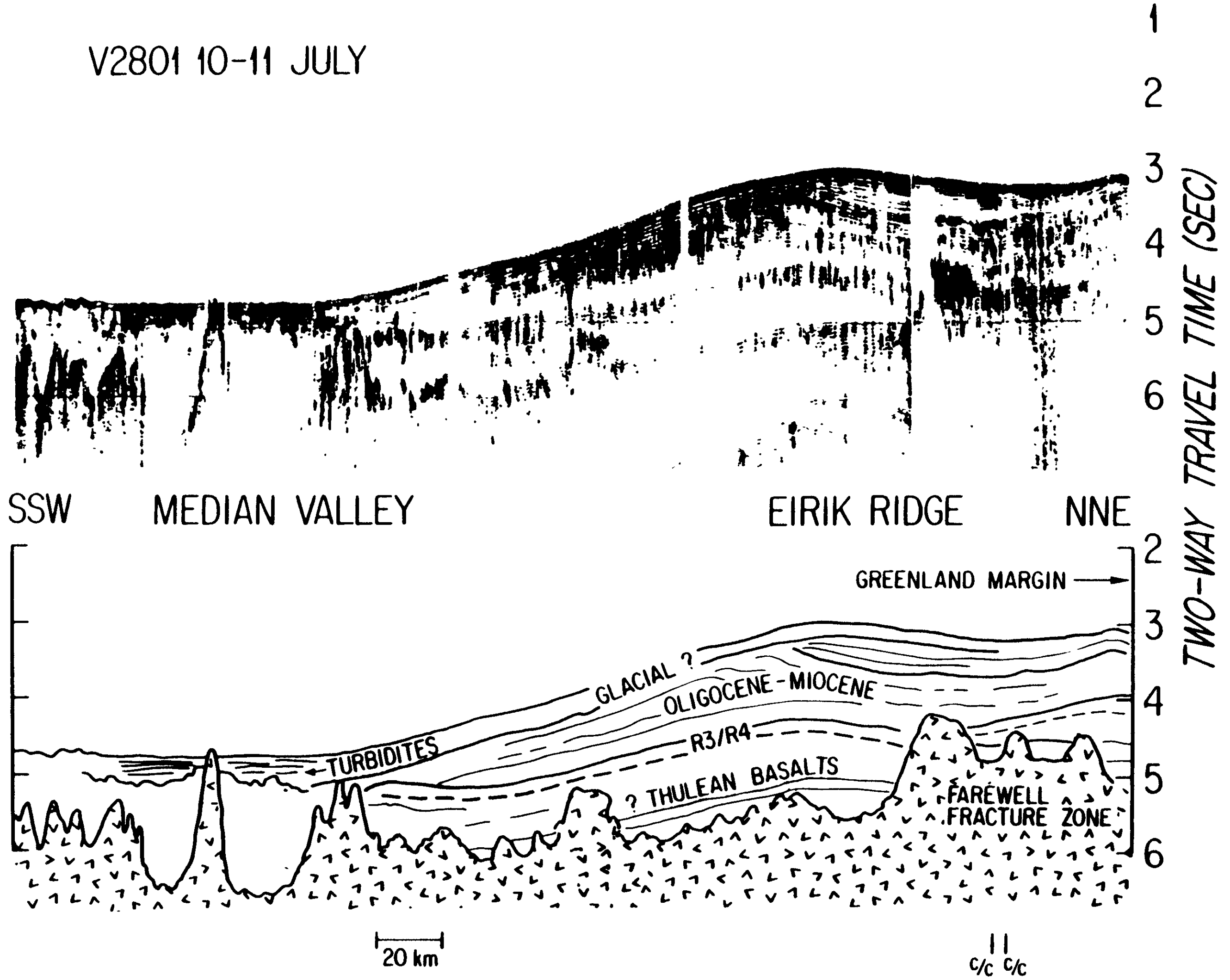

V2801 10-11 JULY
1
2
3
4
5
6
TWO-WAY TRAVEL TIME (SEC)
SSW
MEDIAN VALLEY
EIRIK RIDGE
NNE
2
3
4
5
6
GREENLAND MARGIN
GLACIAL ?
OLIGOCENE-MIOCENE
R3/R4
TURBIDITES
? THULEAN BASALTS
FAREWELL FRACTURE ZONE
20 km
C/C C/C

Figure 12. Single channel seismic reflection profile and interpretation of profile V28-01 across the Eirik Drift. Located in Figure 1 as section B.

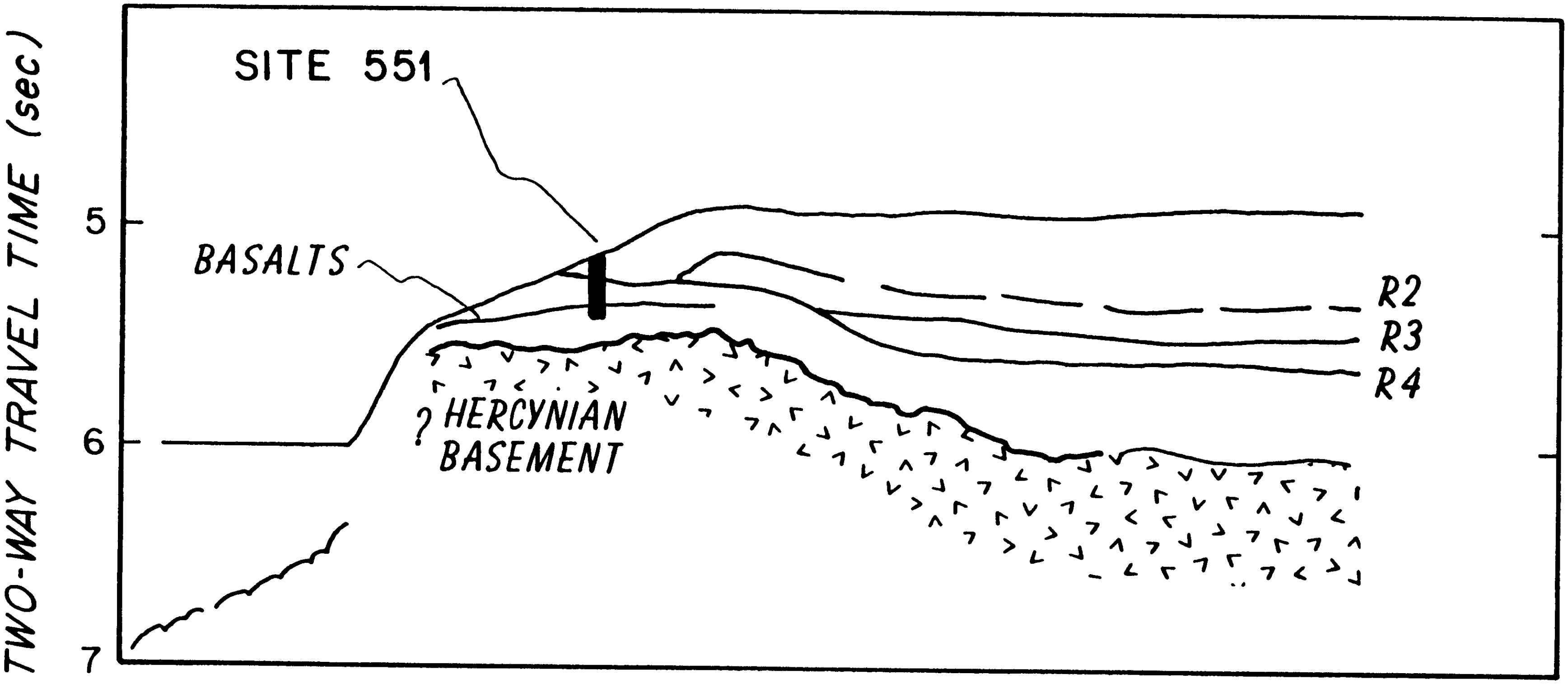

SITE 551
BASALTS
? HERCYNIAN BASEMENT
R2
R3
R4
TWO-WAY TRAVEL TIME (sec)
5
6
7

Figure 13. Interpretation of profile through Site 551. Located in Figure 1 as section J. Data provided by C.W. Poag and P.C. de Graciansky.

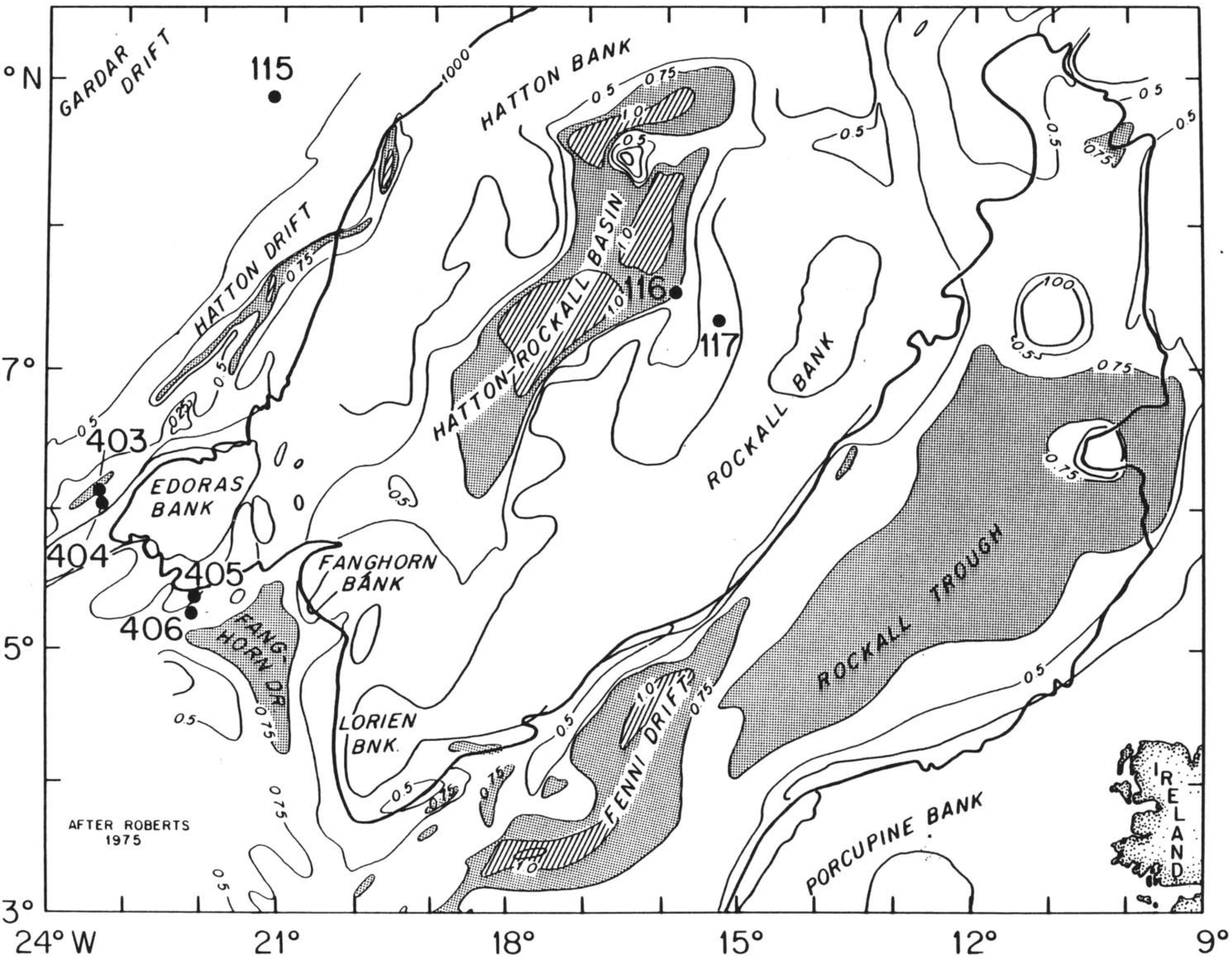

GARDAR DRIFT
115
HATTON BANK
1000
0.5
0.75
1.0
HATTON DRIFT
HATTON-ROCKALL BASIN
116
117
ROCKALL BANK
100
403
404
405
406
EDORAS BANK
FANGHORN BANK
FANG-HORN DR
LORIEN BNK
FENNI DRIFT
ROCKALL TROUGH
PORCUPINE BANK
IRELAND
AFTER ROBERTS 1975
59° N
57°
55°
53°
24° W
21°
18°
15°
12°
9°

Figure 14. Post-R4 sediment thickness (in seconds two-way travel time) of the Rockall Plateau and environs showing location of DSDP sites. Modified after Roberts (1975).

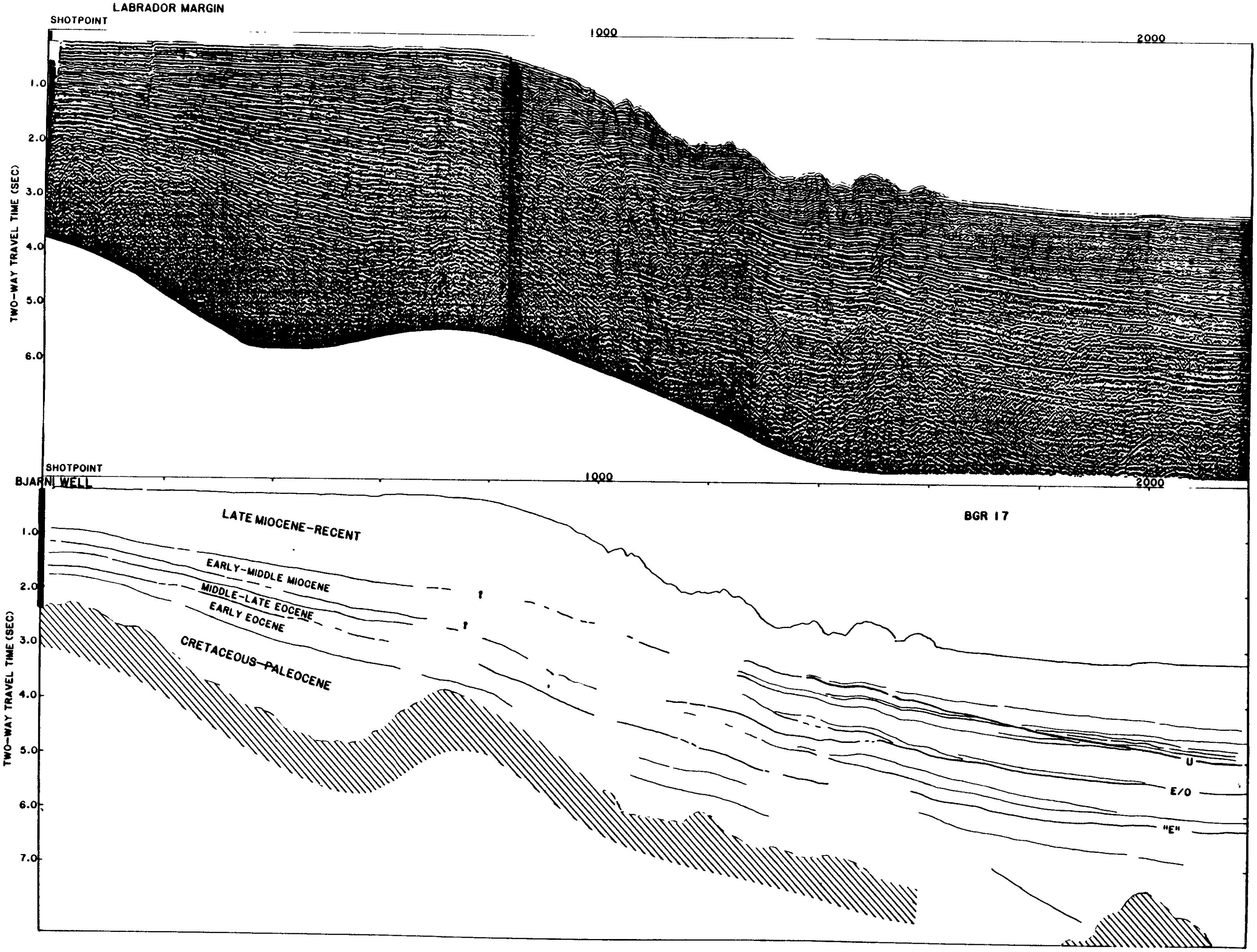

LABRADOR MARGIN
SHOTPOINT
1000
2000
1.0
2.0
3.0
4.0
5.0
6.0
TWO-WAY TRAVEL TIME (SEC)
SHOTPOINT
BJARNI WELL
1000
2000
LATE MIOCENE-RECENT
BGR 17
EARLY-MIDDLE MIOCENE
MIDDLE-LATE EOCENE
EARLY EOCENE
CRETACEOUS-PALEOCENE
U
E/O
"E"
1.0
2.0
3.0
4.0
5.0
6.0
7.0
TWO-WAY TRAVEL TIME (SEC)

Figure 15. Multichannel seismic reflection profile and interpretation of profile BGR 17 across the Labrador Margin into the southern Labrador Sea. Bjarni well age assignments after Gradstein and Williams (1981). Data after Hinz et al. (1979).

CHAPTER 3: RELATIONSHIPS AMONG FAUNAL, ISOTOPIC, AND ABYSSAL CIRCULATION CHANGES

The distribution patterns of modern deep-sea benthic foraminifera often correlate with water masses and properties that co-vary with water masses (Streeter, 1973; Schnitker, 1974; Lohmann, 1978; Corliss, 1979a; Bremer and Lohmann, in press). Interpretations of water-mass/foraminiferal relationships have been extended to the Tertiary fossil record in order to infer changes in paleocirculation (Schnitker, 1979; Tjalsma and Lohmann, in press; Douglas and Woodruff, 1981). However, paleoceanographic interpretations of changes in deep-sea benthic foraminiferal assemblages often conflict. This is illustrated by disputes as to the nature, timing, and causes of Late Paleogene benthic foraminiferal changes. Douglas (1973), Boersma (1977), and Schnitker (1979) suggested that a major benthic foraminiferal turnover occurred near the Eocene/Oligocene boundary, while Corliss (1979b, 1981), Tjalsma (1982), and Tjalsma and Lohmann (1982) suggested more gradual, sequential changes. Benthic foraminiferal assemblage data from both the Miocene (Schnitker, 1979, 1980a,b) and the Late Paleogene (Appendixes 1,4,5) have been cited as evidence for the first significant influx of bottom water into the North Atlantic from northern sources (Arctic and/or Norwegian-Greenland Sea). The seismic stratigraphic record provides less ambiguous evidence for changes in abyssal circulation than can be inferred from benthic foraminiferal assemblage data. Benthic foraminiferal isotopic evidence can provide valuable supportive evidence of changes in abyssal circulation; this evidence can be used to evaluate and expand upon the abyssal circulation model developed from seismic stratigraphic studies (Chapter 2; Appendix 3). Late Paleogene benthic foraminiferal changes in the North Atlantic are used to interpret possible changes in water masses; the interpretations obtained from such water mass-foraminiferal relationships are evaluated in view of changes in abyssal circulation circulation inferred from the seismic stratigraphic and isotopic studies.

A major latest Eocene to earliest Oligocene enrichment of benthic foraminiferal ^{18}O had been noted previously from the Southern and

Pacific Oceans (Shackleton and Kennett, 1975; Savin et al., 1975; Kennett and Shackleton, 1976; Keigwin, 1980). Such ^{18}O enrichments may possibly represent a global change in the isotopic composition of seawater or a drop in bottom-water temperature or both. Initial studies suggested that the major buildup of continental ice in the Tertiary did not occur until the middle Miocene, and therefore, that the enrichment near the Eocene/Oligocene boundary must represent a major cooling (Shackleton and Kennett, 1975; Savin et al., 1975; Kennett and Shackleton, 1976; Savin, 1977). Matthews and Poore (1980), on the other hand, suggested that the enrichment may represent the first major buildup of continental ice on Antarctica.

Prior to the studies conducted here (Appendixes 2,5), it was believed that the major enrichment of ^{18}O occurred near the middle/late Eocene boundary in the North Atlantic at Sites 400A and 398 (Fig. 1) (Vergnaud-Grazzini et al., 1978, 1979; Vergnaud-Grazzini, 1979). Miller and Curry (1982; Appendix 2) analyzed oxygen and carbon isotopic composition of benthic foraminifera from Sites 119 and 401 (Fig. 16). They found that $\delta^{18}O$ values increased ~ 1.9 $^{o}/oo$ and $\delta^{13}C$ values increased ~ 0.8 $^{o}/oo$ between the early middle Eocene (Zones NP13-NP15) and the earliest Oligocene (Zone NP21) of Site 119. They combined isotopic records for Sites 119 and 401 and concluded that ~ 1.4 $^{o}/oo$ of the $\delta^{18}O$ increase occurred in the late Eocene to earliest Oligocene (Fig. 17). They could not resolve biostratigraphically the exact timing of the enrichment, which may have occurred over a period of less than one million years (within Zone NP21) or as long as 4 million years (between Zones P15 and NP21). Miller and Curry suggested that the low Eocene benthic foraminiferal $\delta^{18}O$ values at Sites 400A and 398 (as much as 2 $^{o}/oo$ lower than at Sites 119 and 401) resulted from diagenetic alteration of foraminiferal tests in these more deeply buried sites (see Vergnaud-Grazzini et al. (1978, 1979), Vergnaud-Grazzini (1978), Renard et al. (1979a, 1979b), and Arthur et al. (1979) for evidence of diagenetic alteration of the Eocene sections at these sites).

Miller et al. (in press; Appendix 5) examined the $\delta^{18}O$ record of Site 549 on the Goban Spur (Fig. 17). Site 549 is the first North Atlantic DSDP site with continuous recovery of well-dated upper Eocene through lower Oligocene sediments, apparently unaltered by diagenetic or dissolution effects. This record allows a firm determination of isotopic changes across the Eocene/Oligocene boundary. Here, a major $\delta^{18}O$ increase began at ~ 38 Ma (late Eocene) and culminated in a rapid (< 0.5 my) increase in $\delta^{18}O$ just above the Eocene/Oligocene boundary (~ 36.5 Ma)(Fig. 17). Miller et al. (in press; Appendix 5) established that the ^{18}O enrichment correlates with an enrichment noted by Keigwin (1980) from the Southern and Pacific Oceans. Assuming that the extinctions of the planktonic foraminifera used to recognize the Eocene/Oligocene boundary (viz. Hantkenina, Cribrohantkenina, and Globorotalia cerroazulensis) are synchronous, the ^{18}O enrichment is a chronostratigraphic marker. Whether this enrichment is attributable to ice-volume buildup or to a temperature decrease of bottom waters, remains in debate. Keigwin argued that the $\delta^{18}O$ increase in benthic foraminifera represented mostly a temperature decrease because of the lack of covariance of the enrichment in planktonic and benthic foraminifera at tropical Site 292 (Phillipine Sea). Assuming buildup of glacial ice from an ice-free Eocene world to a fully-glaciated Oligocene world (with ice volume equal to present-day ice volume), Miller and Curry (1982; Appendix 2) argued that an excess $\delta^{18}O$ increase remains, representing at least a 2°C temperature drop. Thus, even if the enrichment reflects ice-volume buildup, a temperature drop of bottom water must have occurred in the latest Eocene to earliest Oligocene.

A rapid major $\delta^{13}C$ increase correlates with the earliest Oligocene $\delta^{18}O$ increase at Site 549 (fig. 3 in Appendix 5). A similar drop occurs in the composite Site 119/401 record (fig. 2 in Appendix 2). Since these records differ from those of Keigwin (1980), they probably cannot be attributed to global changes in $\delta^{13}C$; rather, the $\delta^{13}C$ increases were interpreted as reflecting a decrease in the age of bottom water. Such a decrease in age would have resulted in increased O_2, decreased CO_2, increased pH, and bottom waters less corrosive to calcium carbonate (Miller and Curry, 1982; Miller et al., in press; Appendixes 2,5). Distinct minima of $\delta^{13}C$ occur in the middle Eocene and middle Oligocene of Site 119 (Fig. 16), and were interpreted as reflecting older (high CO_2 and low pH, hence more corrosive) bottom water.

Faunal studies of northern North Atlantic sites indicate that major benthic foraminiferal changes occurred at abyssal depths between the middle Eocene and the earliest Oligocene (Miller et al., 1982; Miller, in press b; Appendixes 1,4). In the deep southern Labrador Sea, Eocene predominantly agglutinated assemblages are replaced by an early Oligocene calcareous assemblage; lithology, percent carbonate, and percent carbon are relatively constant across the faunal change (fig. 5 in Appendix 1). Miller et al. (1982; Appendix 1) suggested that certain hydrographic properties (viz. low O_2, low pH, high CO_2, more corrosive waters) promote the development of predominantly agglutinated benthic foraminiferal assemblages, and attributed the change in assemblages in the Labrador Sea to a decrease in age of bottom waters resulting from the influx of northern bottom water. Unfortunately, coring gaps prevented establishing the exact timing of this faunal change.

A major faunal change in calcareous benthic foraminifera also occurred between the early middle Eocene and earliest Oligocene at two abyssal (> 3 km paleodepth) sites in the Bay of Biscay (Sites 119 and 400A; in the > 150 μm size-fraction)(Miller, in press b; Appendix 4; Schnitker, 1979). Nuttallides truempyi, Clinapertina spp., and Abyssamina spp., which dominated the Eocene abyssal benthic assemblage, were replaced by increasingly abundant, bathymetrically wide-ranging and stratigraphically long-ranging taxa: Oridorsalis spp., Globocassidulina subglobosa, Gyroidinoides, and the Cibicidoides ungerianus plexus (Figs. 18,19). Many abyssal taxa became extinct prior to the early Oligocene. However, a late Eocene hiatus encountered at Sites 119 and 400A prevents dating the timing of these extinctions (Figs. 18,19).

Eocene to early Oligocene benthic foraminiferal assemblage changes are well defined at Site 549 (Miller et al., in press; Appendix 5). At Sites 549 (~ 2.0-2.5km paleodepth), the major faunal change is the replacement of the Nuttallides truempyi assemblage just above the middle/late Eocene boundary (Fig. 20); this event can be shown to have occurred throughout the North Atlantic, South Atlantic, Caribbean, and Gulf of Mexico at this time (figs. 10,11 in Appendix 5; Tjalsma and Lohmann, in press). Other faunal changes at Site 549 include 1) a series of extinctions and local last appearances of taxa in the late Eocene; 2) a series of first appearances (some of which are local) in the late Eocene to earliest Oligocene; and 3) the replacement of a buliminid assemblage just below the Eocene/Oligocene boundary (Fig. 20). The record at Site 549 firmly established that no major benthic foraminiferal changes (in the > 150 μm size fraction) are associated with the Eocene/Oligocene boundary in the North Atlantic; instead, benthic foraminiferal changes occurred throughout the late Eocene to earliest Oligocene interval (~ 40-36 Ma). Many of the first and last appearances and the replacement of the buliminid assemblage are probably local phenomena; however, the decrease in abundance and extinction of N. truempyi and associated abyssal taxa represent a dramatic benthic foraminiferal change that occurred throughout the Atlantic Ocean (Tjalsma and Lohmann, in press; Miller in press b; Miller et al., in press; Appendixes 4,5). In the evolution of deep-sea benthic foraminifera during the Cenozoic, only the massive extinctions of the latest Paleocene (Tjalsma and Lohmann, in press) and assemblage changes of the middle and late Miocene (Douglas and Woodruff, 1981) rival changes that occurred between the middle Eocene and the early Oligocene in importance.

Major benthic faunal changes also occurred in the Late Paleogene in ostracodes. Between the middle Eocene and the early Oligocene, the modern cold-water ostracode fauna developed; this was attributed to the development of cold bottom water circulation and the associated development of the oceanic cold-water sphere (= psychrosphere)(Benson, 1975). Ducasse and Peypouquet (1979) similarly noted that the psychrospheric ostracode fauna appeared in the late Eocene at Site 406 (SW margin of Rockall Plateau). The exact timing of the change in ostracodes is debatable (cf. Benson, 1975 with Corliss, 1981); however, it is clear that between the middle Eocene and the early Oligocene benthic faunas, both ostracodes and foraminifera, underwent a major reorganization.

The relationships of the Eocene to early Oligocene benthic foraminiferal isotopic and assemblage changes are well defined at Site 549. The late Eocene faunal changes began ~ 39.5 Ma, prior to the 1 $^{0}/oo$ oxygen isotopic enrichment which occurred between ~ 38.5 and 36.5 Ma (late Eocene to earliest Oligocene). Although several extinctions and local last appearances occurred in the late Eocene to earliest Oligocene, the majority of taxa range through the Eocene to Oligocene temperature drop.

With the resolution provided at Site 549, the changes in faunal isotopic and assemblage composition can be compared with the abyssal circulation model. The change from sluggish to vigorous, intense bottom water circulation associated with reflector R4 occurred in the late Eocene to earliest Oligocene (Miller and Tucholke, in press; Appendix 3; Chapter 2). Because of the uncertainties in assigning an age to reflector R4, the erosional event associated with this reflector may be coincident with either the early late Eocene assemblage change or the latest Eocene to earliest Oligocene isotopic changes. Still, reflector R4 is probably latest Eocene to earliest Oligocene (~ 36-38 Ma) (see Chapter 2 for discussion), although uncertainties in the biostratigraphy and the correlation of reflector R4 into the various boreholes prevent absolutely establishing this. If this age assignment is correct, reflector R4 and the associated major abyssal circulation change correlate with the major ^{18}O enrichment, but they post-date the initial faunal changes.

The diachrony of the major faunal change with the isotopic and seismic stratigraphic changes poses a problem. An explanation may be provided by considering the overall development of Cenozoic deep-sea benthic foraminiferal assemblages. The major Late Paleogene faunal changes are the progressive restriction of abyssal Eocene taxa to greater depths and the ultimate extinction of these abyssal taxa, Nuttallides truempyi, Clinapertina spp., Abyssammina spp., Alabamina dissonata, and Aragonia spp. (Tjalsma and Lohmann, in press; Miller, in press b; Appendix 4). The depth-restriction apparently began in the middle Eocene, climaxing in the replacement of the abyssal Eocene taxa just above the middle/late Eocene boundary (40-38 Ma)(Tjalsma and Lohmann, in press; Miller et al., in press; Appendix 5). In addition, several species important in the modern abyssal ocean (e.g. Nuttallides umbonifera, Epistominella exigua, Eggerrella bradyi) made their appearance in the late Eocene to early Oligocene (Miller, in press b; Miller et al., in press; Appendixes 4,5). These changes represent the transition from assemblages dominated by Cretaceous and Paleocene relict taxa (Tjalsma and Lohmann, in press) toward the development of modern benthic foraminiferal assemblages. Although they occur at intermediate (e.g. Sites 549 and 401; 2.0-2.5km paleodepth) and shallower (e.g. Site 548; ~ 1km paleodepth) sites, these changes are most dramatic at deepest sites (e.g. Sites 119, 400A, 550; all > 3km paleodepth; Miller, in press b; Miller et al., in press; Appendixes 4, 5). In fact, intermediate-depth Eocene assemblages are very similar in abundance composition (with the exception of abundant *Nuttallides truempyi* in the Eocene) to deep Oligocene assemblages (Schnitker, 1979; Miller, in press b; Appendix 4), with dominant *Globocassidulina subglobosa*, *Gyroidinoides* spp., *Cibicidoides ungerianus*, and *Oridorsalis umbonatus*. These taxa are bathymetrically wide-ranging and stratigraphically long-ranging, and may be interpreted as tolerant of environmental changes.

I speculate that the abyssal Late Paleogene fauna (*Nuttallides truempyi* and associated *Clinapertina* spp., and *Abyssammina* spp., *Alabamina dissonata*, *Aragonia* spp.) was adapted to the warm, corrosive, older bottom waters of the Eocene. The deep-sea environment began to change gradually in the middle to late Eocene, resulting in their demise. This paleoenvironmental change may be due to climatic cooling resulting in cooling of bottom waters, leaking of bottom waters from the Norwegian-Greenland Sea or Arctic Ocean, or other as yet undetermined causes. Circumstantial evidence of paleoenvironmental changes include a δ^{13}C increase of ~ 0.6 ^{0}/oo that correlates with the replacement of the *N. truempyi* assemblage (Miller et al., in press; Appendix 5) and a general cooling of bottom waters that began in the middle Eocene (Shackleton and Kennett, 1975). As a result of these initial paleoenvironmental changes, the Eocene abyssal fauna was replaced by *Globocassidulina subglobosa*, *Gyroidinoides* spp., *Cibicidoides ungerianus*, and *Oridorsalis umbonatus*. These taxa increased in abundance to form the nucleus of the abyssal late Eocene fauna. Since these taxa probably had wide environmental tolerances, they were not seriously affected by the culmination of the paleoceanographic changes near the Eocene/Oligocene boundary, viz. the rapid bottom water temperature drop, increased intensity of bottom water circulation, and decrease in the age of bottom water (higher O_2, lower CO_2, higher pH, hence less corrosive).

In the Bay of Biscay, Globocassidulina subglobosa, Gyroidinoides spp., Cibicidoides ungerianus, and Oridorsalis umbonatus continue to dominate in the early Oligocene at both intermediate (~ 2km) and deep (> 3km) sites. Subsequently in the middle Oligocene, Nuttallides umbonifera expands in abundance in the deepest sites in the northern North Atlantic (> 3 km; Sites 119, 400A, 550), becoming the dominant benthic foraminifera (Fig. 16). In the modern oceans, the abundance of N. umbonifera is positively correlated with increased corrosiveness of bottom water (Bremer and Lohmann, in press); at Site 119 the greatest abundances of Nuttallides spp. are associated with the lowest $\delta^{13}C$ values in benthic foraminifera (Fig. 16) (Miller and Curry, 1982; Appendix 2). Lower $\delta^{13}C$ values are often associated with older water masses that are, in turn, more corrosive to carbonate (Kroopnick et al., 1972; Kroopnick, 1974, 1980; Lohmann and Carlson, 1981) Thus, the middle Oligocene of Site 119 was interpreted as reflecting older, more corrosive bottom water (Miller and Curry, 1982; Miller, in press b; Appendixes 2,4). This suggested increase in age and corrosiveness correlates with a general reduction of intensity of abyssal circulation associated with the post-R3 depositional pulse.

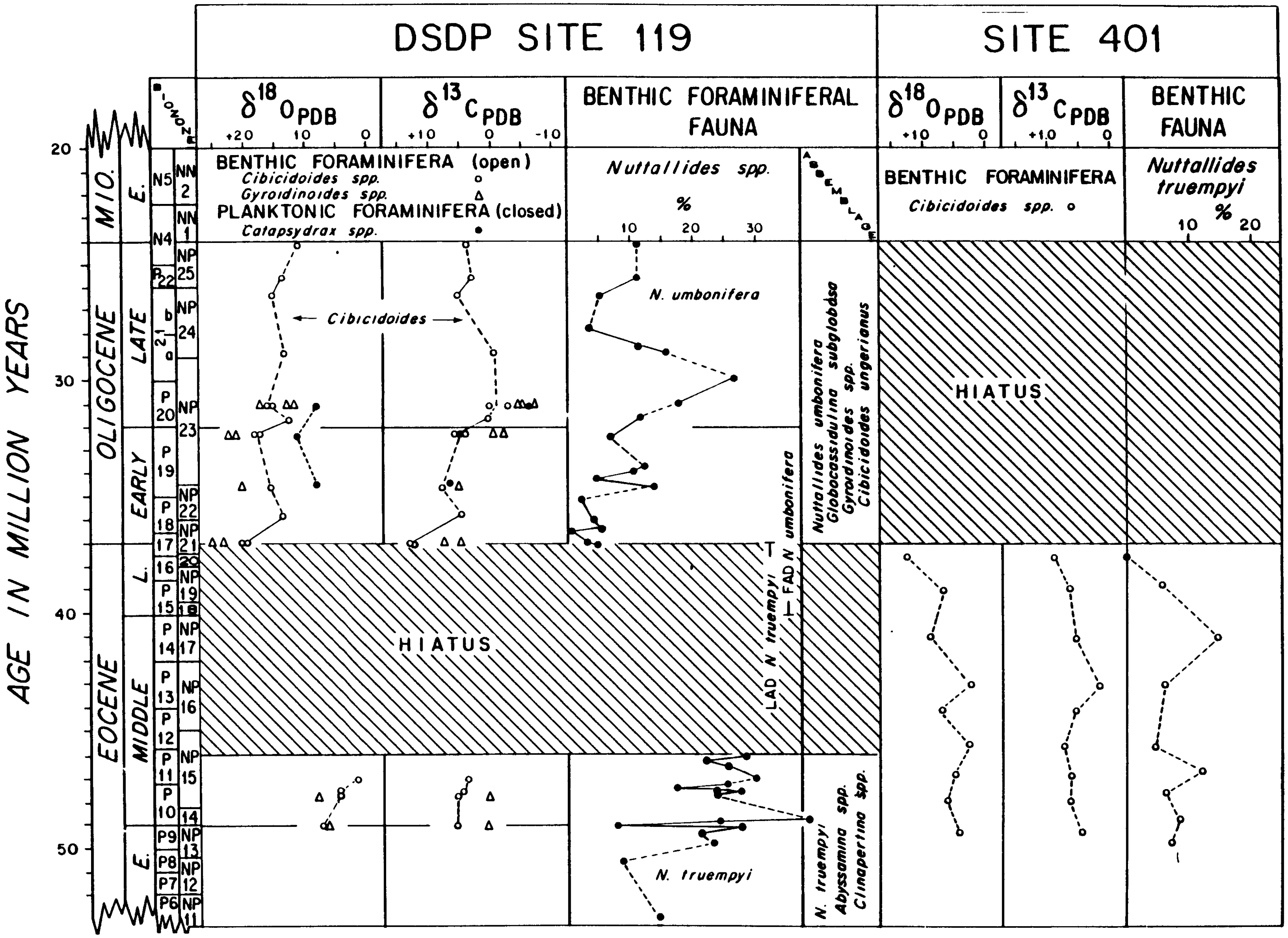


Figure 16. Isotopic composition and distribution of Nuttallides spp. at Sites 119 and 401 in the Bay of Biscay. Site location given in Figure 1. After Miller and Curry (1982) and Miller (in press).

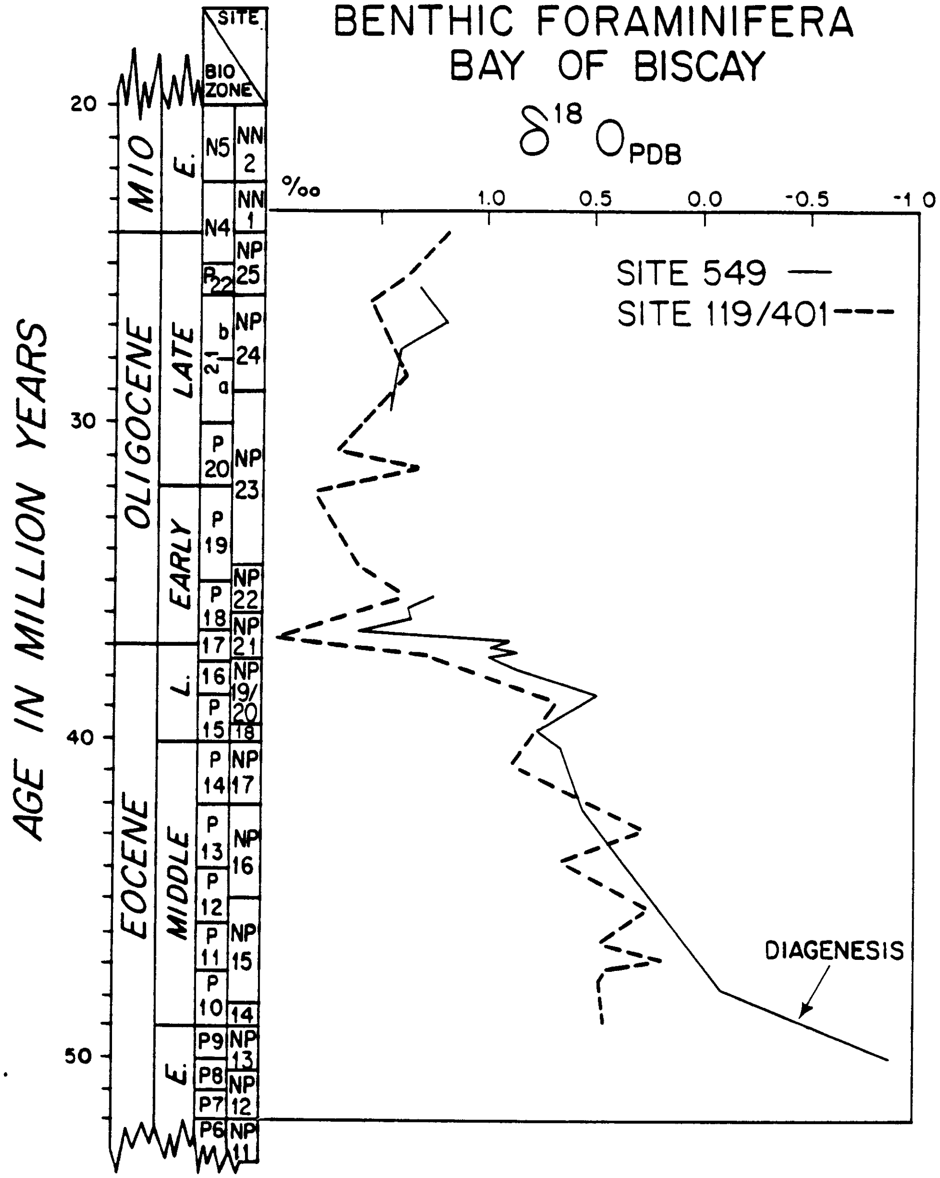


Figure 17. Comparison of Site 549 benthic foraminiferal oxygen isotopic composition with composite record from Bay of Biscay Sites 119 and 401. After Miller et al. (in press).

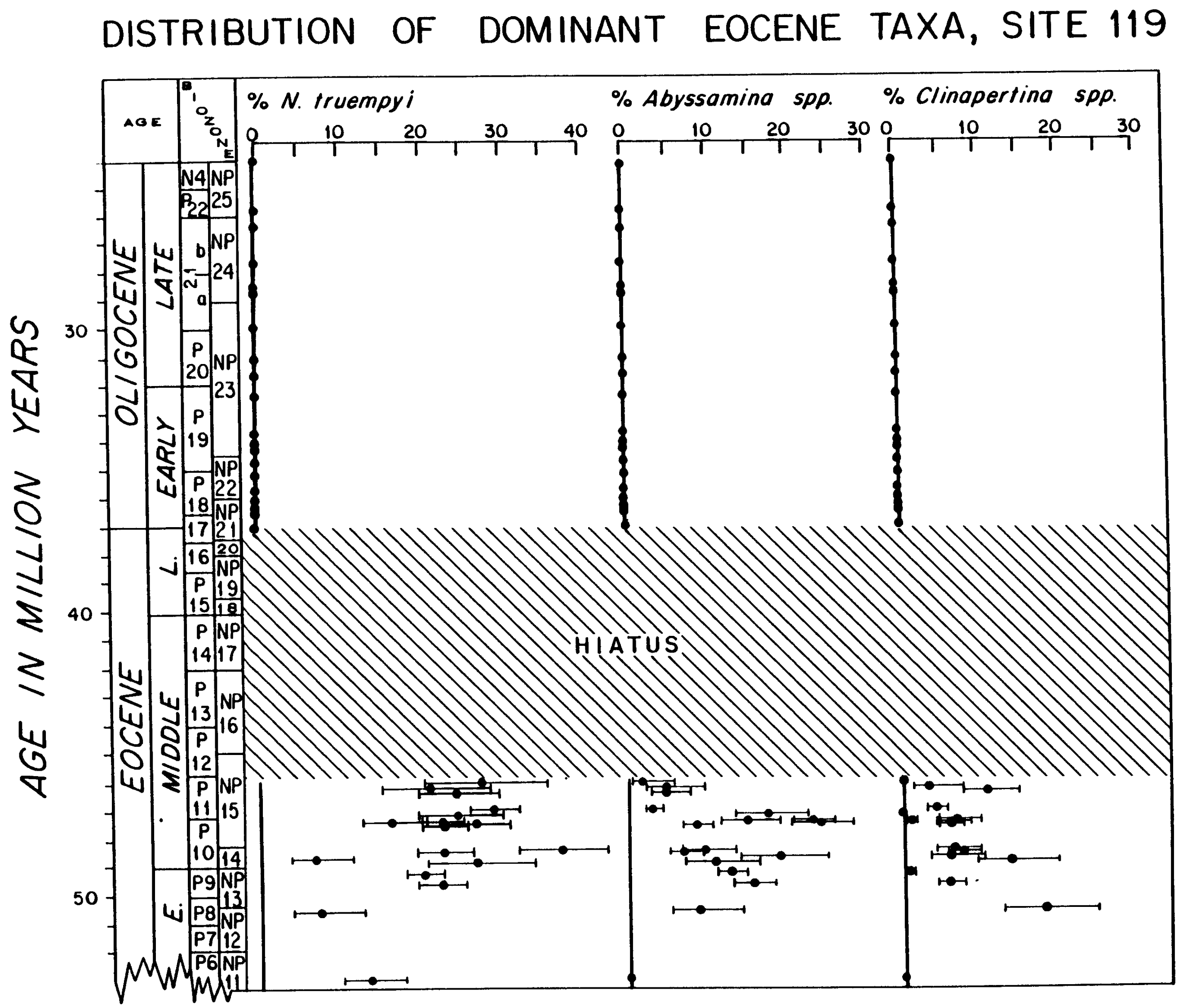
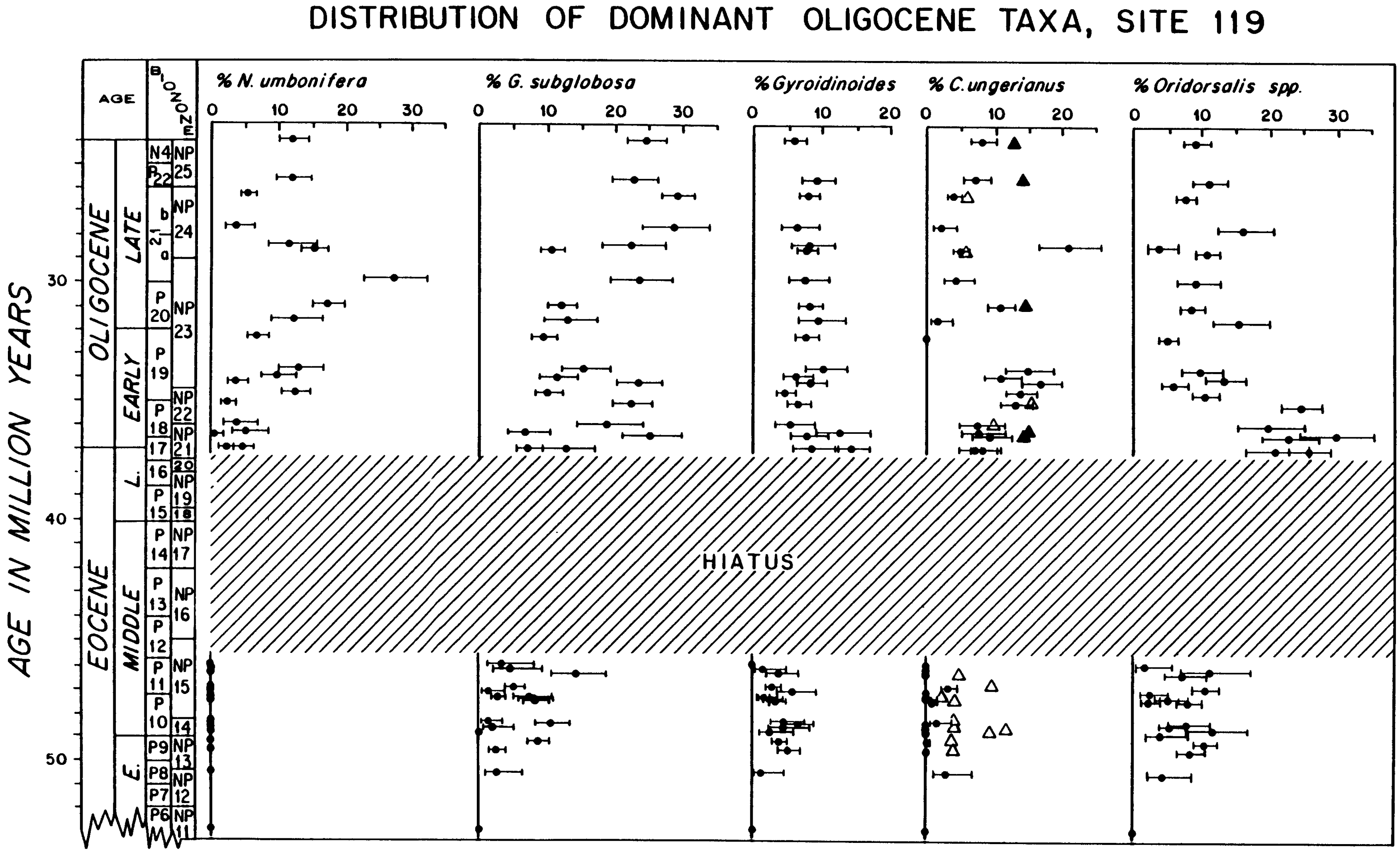


Figure 18. Distribution of dominant Eocene abyssal taxa, Site 119. 80°/o confidence interval is indicated. After Miller (in press).



DISTRIBUTION OF DOMINANT OLIGOCENE TAXA, SITE 119

Figure 19. Distribution of dominant Oligocene taxa, Site 119. 80°/o confidence interval is indicated. After Miller (in press).

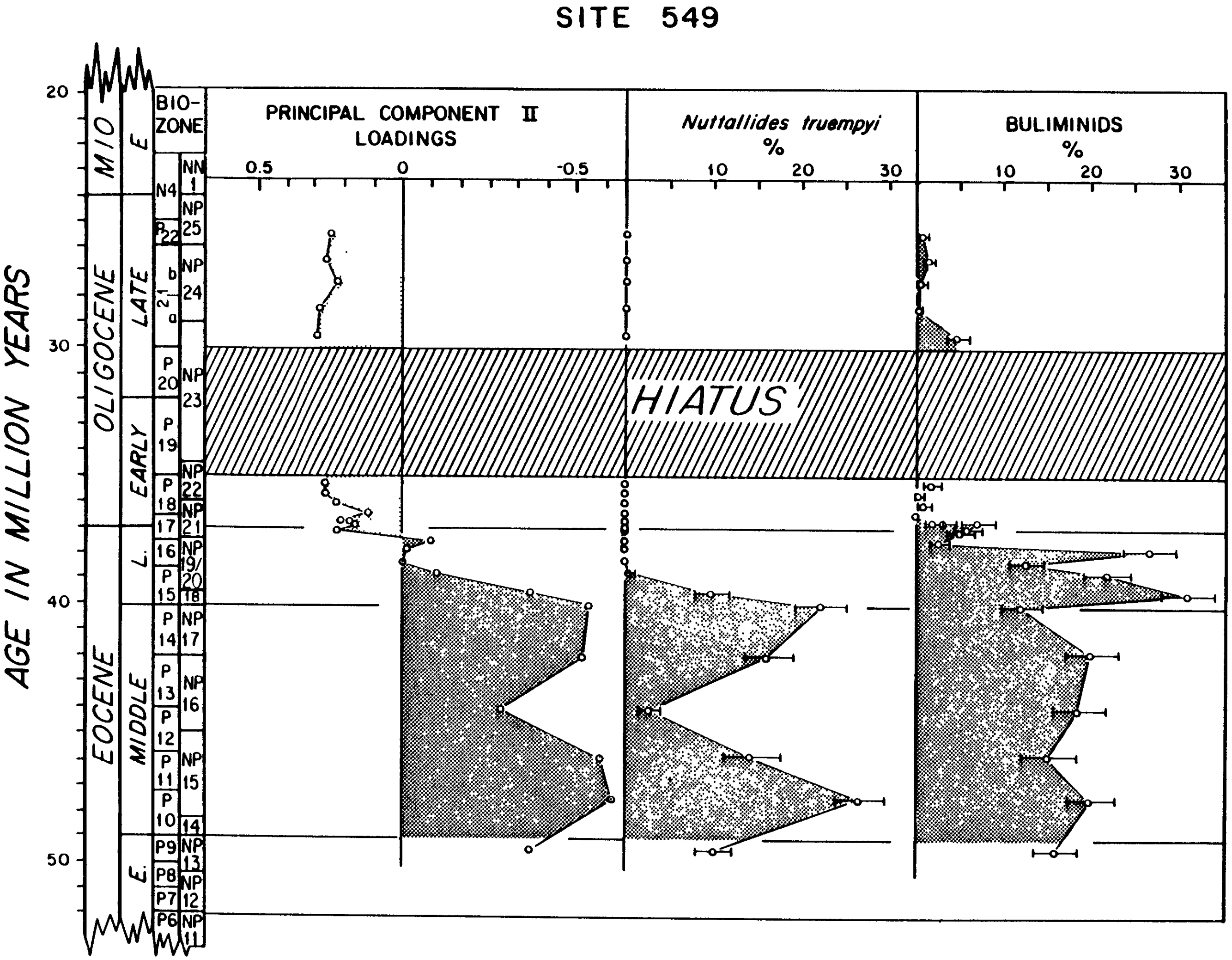


Figure 20. Benthic foraminiferal assemblage composition, Site 549. Principal component analysis was performed on 25 samples from this site. Error bars on percentages indicate 80°/o confidence interval. After Miller et al. (in press).

CHAPTER 4: DISCUSSION AND PALEOCEANOGRAPHIC SYNTHESIS

The major abyssal circulation events deduced from the seismic stratigraphic record and various benthic foraminiferal assemblage and isotopic events span approximately 4 million years (Fig. 21). Nevertheless, there is some evidence that the gradual changes occurring over this interval were punctuated by geologically rapid events as suggested by Kennett and Shackleton (1976). Although the overall ^{18}O enrichment occurred gradually over ~ 2 million year interval (assuming constant sedimentation rates, fig. 2 in Appendix 5), the early Oligocene ^{18}O enrichment occurred at Site 549 in an interval of < 0.5 million years. This agrees well with previous estimates (Kennett and Shackleton, 1976; Keigwin, 1980) for the duration of the $\delta^{18}O$ change in the Southern and Pacific Oceans. In addition, the erosional event associated with reflector R4 probably was also a geologically rapid event, with the change from a sluggishly circulating Eocene ocean to a vigorously circulating Oligocene ocean occurring in less than 2 million years.

In Chapter 2, it was suggested that the northeastern North Atlantic had a major northern source of cold bottom water, and that bottom waters may have entered the North Atlantic by flowing over the Greenland-Scotland Ridge. Presently the Greenland-Scotland Ridge separates the Norwegian-Greenland Sea from the North Atlantic (Fig. 22), and bottom water entering the North Atlantic flows over four sills in this aseismic ridge. These are the Denmark Strait (between Greenland and Iceland), Iceland-Faeroe Ridge, Faeroe Bank Channel, and Wyville-Thompson Ridge (Figs. 22,23) (Crease, 1965; Worthington, 1969, 1970, 1976; Ellett and Roberts, 1973). It has been suggested from thermal subsidence models that the Iceland-Faeroe Ridge did not subside below sea level until the early Miocene (Talwani, Udintsev et al., 1976; Eldholm and Thiede, 1980). This estimate was supported by initial paleontological evidence that indicated that the first surface-water connection of the Norwegian-Greenland Sea with the North Atlantic did not occur until the early Miocene (Bjorklund, 1976; Schrader et al., 1976; Van Hinte In: Talwani, Udintsev et al., 1976). However, reconsideration of the nannofossil, planktonic foraminiferal, and terrestrial floral and faunal evidence suggest that the first marine connection of the North Atlantic and Norwegian-Greenland Sea across the Greenland-Scotland Ridge occurred in the early Eocene, soon after the opening of the Norwegian-Greenland Sea (Berggren and Schnitker, in press; McKenna, in press; Miller and Tucholke, in press; Appendix 3). This older age estimate for the connection does not conflict with subsidence models, because the Iceland-Faeroe Ridge is the shallowest of the modern sills, and connections may have existed at the westernmost (the Denmark Straits) or easternmost (Faeroe-Shetland Channel) end prior to connections across the Iceland-Faeroe Ridge (Fig. 22)(Miller and Tucholke, in press; Appendix 3).

The most likely candidate for an Early Tertiary seaway linking the North Atlantic with the Norwegian-Greenland Sea is the Faeroe-Shetland Channel (Fig. 23; Ridd, 1981, in press; Miller and Tucholke, in press; Appendix 3; Berggren and Schnikter, in press). The Faeroe-Shetland Channel is a northwest-southeast trending marine channel forming a gateway between the Norwegian-Greenland Sea and the North Atlantic (Fig. 23). It is separated from the North Atlantic by two sills: the Faeroe-Bank Channel (sill depth ~ 830m) and the Wyville-Thompson Ridge (sill depth ~ 350m). Ridd (1981) discussed the origin and history of the Faeroe-Shetland Channel, noting that the channel contains thick (up to 2.5km) post-Paleocene (mostly Eocene) sediments lying upon uppermost Paleocene (= Thulean) basalts. Some debate exists as to the nature of the underlying crust, for Talwani and Eldholm (1972) and Ridd (1981) suggested that the channel is floored by continental crust, while Roberts (1975), Bott (1975), and Russell and Smythe (1978) favored an oceanic origin. As a result, the pre-Paleocene subsidence history of this region is unconstrained. Still, similarities in the seismic and rock stratigraphic records in the North Atlantic and the Faeroe-Shetland Channel suggest that these regions were connected by the end of the Eocene. Ridd (1982) observed a major change in sedimentary regime in the Faeroe-Shetland Channel associated with a prominent seismic horizon. The sequence overlying this reflector is thinned into lenticular bodies thought to reflect the influence of strong currents (Ridd, 1981). The reflector can be traced to the north flank of the Wyville-Thompson Ridge (Fig. 24), while reflector R4 can be traced to the south flank of the ridge (Ellett and Roberts, 1973). It seems likely that the two horizons are related in time and genesis. However, the age control on the horizon in the Faeroe-Shetland Channel is poor. Although, Ridd (1981) suggested that the change in sedimentary regime correlates in time with reflector R4, Ridd (in press) favored a later timing (near the end of the Paleogene) for this event.

Berggren and Hollister (1974) and Miller and Tucholke (in press; Appendix 3) suggested that the major change in abyssal circulation in the North Atlantic was related to an Arctic connection. Berggren and Hollister (1974) used an early Eocene age for the separation of Greenland and Spitsbergen, the Arctic connection, and the abyssal circulation change. However, Talwani and Eldholm (1977) later concluded that Greenland separated from Spitsbergen prior to Anomaly 13 time (in the late Eocene to early Oligocene). Miller and Tucholke (in press; Appendix 3) noted the time correlation between reflector R4 and the opening of this Arctic passage, and they suggested that following the separation of Greenland and Spitsbergen, Arctic waters rapidly entered the Norwegian-Greenland Sea, flowed through the Faeroe-Shetland Channel, across the Wyville-Thompson Ridge and entered the North Atlantic. The history of connection of the North Atlantic and the Arctic Ocean is a controversial issue (cf. Gartner and Keany, 1978; Gartner and McGuirk, 1979; Thierstein and Berger, 1978; with Ling, et al., 1973; Clark and Kitchell, 1979; Gradstein and Srivastava, 1981). However, shallow-water connections probably existed as early as the Mesozoic through the Labrador Sea (Gradstein and Srivastava, 1981) and possibly through epicontinental seas such as the Barents Sea (F.M. Gradstein, personal communication). Still, tectonic control (viz. opening of a deep passage to the Arctic Ocean) on introduction of bottom water to the North Atlantic would explain both the geologic suddenness and intensity of the abyssal circulation event that created reflector R4 (Miller and Tucholke, in press; Appendix 3).

The late Eocene to early Oligocene ^{18}O enrichment has been observed in several different ocean basins over a wide range of paleodepths, and probably cannot be attributed only to initial entry of Arctic/Norwegian-Greenland Sea sources of cold bottom water. There is evidence that initial formation of cold, vigorously circulating bottom water from both northern sources (as denoted by reflector R4 and Horizon A^u) and southern sources (as denoted by erosion of widespread unconformities and other changes in the Southern and Pacific Oceans; Kennett and Shackleton, 1976; Moore et al., 1978) began near the end of the Eocene. These events were reflected by a major ^{18}O enrichment. The synchronous ^{18}O enrichment associated with cold bottom water formation in both the Antarctic and the polar/subpolar marginal basins of the North Atlantic may argue against a tectonic cause of the abyssal circulation change. Still, high-salinity water provided by modern North Atlantic Deep Water is important in the formation of Antarctic Bottom Water (Foster and Carmack, 1976); such linkages or "teleconnections" (see Johnson, 1982; Schnikter, 1980a,b) might be invoked to explain the formation of southern bottom-water sources following the tectonically-controlled entry of northern sources of bottom water into the North Atlantic.

The development of cold, vigorous bottom water circulation near the end of the Eocene falls within a period of global climatic cooling. Several lines of evidence, including $\delta^{18}O$ of planktonic and benthic foraminifera, distribution of calcareous nannoplankton, distribution of planktonic foraminifera, and paleobotanical evidence obtained from terrestrial floras indicate that general global cooling and increased latitudinal thermal gradients developed between the middle Eocene and the early Oligocene (see Table 1 in Appendix 3 for full references). However, the details of the climatic record are in dispute. Collison et al. (1981) suggested from palynological studies in southern England that climatic cooling was gradual, beginning in the early Eocene and extending into the early Oligocene. Norris (1982) used palynomorphs from the Mackenzie Delta and Wolfe (1978) used terrestrial floras from the northwestern United States to suggest that a major, sharp cooling occurred at the end of the Eocene. Some of this disagreement may be attributable to the nature of the record studied and differences in interpretation. Still, comparison of various studies (Collison et al., 1981; Norris, 1982; Wolfe, 1978; Haq et al., 1977; Haq, 1981) shows general agreement that: 1) following a climatic optimum of the early Eocene (and possibly earliest middle Eocene), general cooling occurred into the Oligocene; 2) offsets to cooler conditions occurred near the middle/late Eocene boundary, near the Eocene/Oligocene boundary, and in the middle Oligocene; 3) the late Oligocene was a period of warming.

The role of climatic change in the development of abyssal circulation is not clear. The faunal changes discussed here occurred in a 4 million year period and may be related, not to a sudden climatic or bottom water temperature change, but to a general climatic and bottom water cooling between the middle Eocene and early Oligocene (see also Corliss, 1979b; 1981). The geologically rapid abyssal circulation event associated with reflector R4 may be explained by tectonic threshold control, and would appear to be causally distinct from the faunal and isotopic changes resulting from gradual climatic cooling. Still, it may be possible to relate the abyssal circulation event to a rapid climatic cooling at the end of the Eocene. Brass et al. (1982) argued that climatic cooling, resulting in increased thermal gradients and the formation of cold bottom water at high latititudes, can result in geologically rapid and catastrophic abyssal circulation changes. The sharp cooling noted near the end of the Eocene may have resulted in the rapid replacement of warm, saline bottom water (WSBW) produced in marginal subtropical seas by cold bottom waters produced in polar/subpolar basins (see Brass et al., 1982 for discussion of WSBW).

This study documents that a major early Oligocene hiautus occurs along the margin of the Rockall Plateau (Fig. 2) and in the Bay of Biscay/Goban Spur region (Fig. 3). Considering the widespread distribution of reflector R4 and its general correlation with an unconformity, it seems likely that the early Oligocene is missing from much of the northern North Atlantic. Similarly, in the western North Atlantic, the Oligocene series is virtually absent along the continental rise and deep basin (Tucholke, 1979). Where are Oligocene sediments The problem is compounded by the scarcity of Oligocene sediments on continental shelves in the North Atlantic (Olsson et al., 1980; Gradstein and Srivastava, 1981). Tucholke and Vogt (1979) suggested that the sediments might have been carried into the South Atlantic. However, recent seismic studies (Gamboa, in press) show that a major unconformity and associated seismic reflector occurs in the Brazil Basin (northwest South Atlantic); this unconformity separates upper Oligocene from lower

Eocene strata. It seems unlikely therefore that the sediments were deposited there. The more southerly Argentine Basin is poorly known, and it is possible that this was the locus of deposition. It is also possible that some of the sediment was deposited on the northern Bermuda Rise (Tucholke and Vogt, 1979). The absence of Oligocene sediments may, in part, be a sampling artifact. Most DSDP holes have been drilled along the margins of the sediment drifts (e.g. Sites 117, 403-405) or where the sediment section is thinned. It has been demostrated that this is where the unconformity is best developed. In sites more toward basin centers, the unconformity becomes an apparent conformity (e.g. Sites 112 and 116). Thus, sediments may have been deposited in drifts in the center of the basins (e.g. the R3-R4 interval in Figures 6,8,9, Table 1) outside the erosional zone. If spread uniformly throughout the western North Atlantic (reconstruction of Sclater et al., 1977), the sediment eroded from the northern North Atlantic (~ 120,000km^3) and the western North Atlantic (~ 200,000km^3; Tucholke and Mountain, 1979) would have formed a layer ~ 60m thick.

The abyssal circulation model presented here requires that the polar-subpolar basins of the North Atlantic and adjacent seas "export" bottom water to other ocean basins beginning in the latest Eocene to earliest Oligocene. Using the model of Berger (1970) it would be expected that such "export" regions should have low siliceous productivity (such as the modern North Atlantic), while high siliceous productivity would be associated with general upwelling regions (such as the modern North Pacific). Yet, widespread deposition of biosiliceous sediments occurred in the early to early middle Eocene of the western and northern North Atlantic, while the patchy deposition of biosiliceous detritus occurs throughout the middle Eocene to early Miocene of most of the northern North Atlantic. The only significant exception occurs in the Labrador Sea (Site 112) where abundant biogenic silica is confined to the Oligocene to Miocene section. Significant biosiliceous productivity terminates at about the level of reflector R2 (upper lower Miocene). This led Schnitker (1980a,b) to suggest that the North Atlantic became an "exporter" of bottom water beginning in the early to early middle Miocene. Thus, the presence of signficant biosiliceous material would argue for upwelling, not export, of bottom waters in the northern North Atlantic during the early Eocene to early Miocene. This apparent contradiction of the abyssal circulation model may be resolved, for the Berger (1970) model is probably an over-simplification when applied to the Late Paleogene North Atlantic. 1) Basin to basin fractionation may have been significantly different in the Late Paleogene, for strong westward-flowing circumeqatorial currents connected the North Atlantic and Pacific Oceans (Berggren and Hollister, 1974; 1977). In fact, the siliceous belt may be related to upwelling resulting from the circumequatorial current (Cita, 1971; Berggren and Hollister, 1974; Tucholke and Vogt, 1979). Restriction of circumequatorial flow by sealing off of the eastern Tethys in the early Miocene (Berggren and Hollister, 1974; Berggren and Van Couvering, 1974) may have resulted in termination of the circumequatorial current, upwelling, and siliceous deposition. 2) Due to upwelling, the Antarctic region is presently a region of very high siliceous productivity (Lisitzin, 1972); yet, formation of Antarctic Bottom Water in the marginal Weddell Sea (Foster and Carmack, 1976) causes the Antarctic region to be an "exporter" of cold, vigorously circulating bottom water. In a similar fashion, upwelling and high productivity may have occurred in the Eocene through Oligocene North Atlantic, but downwelling and bottom-water formation may have occurred in the Oligocene in marginal basins viz. the Norwegian-Greenland Sea or Arctic Ocean. 3) Alternatively, the increase in siliceous deposition may be related to increased volcanism (Gibson and Towe, 1971).

The most dramatic depositional phase of sediment drift and wave development began in the northern and western North Atlantic near the early/middle Miocene boundary (above reflector R2)(Fig. 7; Table 1; fig. 12C in Appendix 3); this depositional phase is interpreted as resulting from a general decrease and stabilization of the abyssal circulation. The major depositional phase correlates with a trend toward east-west basin isolation due to the restriction of fracture zone conduits across the Reykjanes Ridge (Miller and Tucholke, in press; Appendix 3). Still, it is not clear how the closing of fracture zone conduits may have attenuated the abyssal circulation, nor is it clear that this was the sole factor responsible for reduced deep circulation. Shackleton and Kennett (1975), Savin et al. (1975), and Woodruff et al. (1981) noted a dramatic shift in oxygen isotopic composition of benthic foraminifera that began in the middle Miocene. Although they attributed this shift primarily to buildup of the Antarctic ice cap, Woodruff and Douglas (1981) suggested that at least some of the shift represents a major bottom-water cooling and a "thickening of Antarctic Bottom Waters." Such an implied increase of Antarctic sources of bottom water may have resulted in reduced influence of bottom water derived from northern sources.

Many faunal and isotopic problems remain unresolved or remain to be tested by future work. The timing and significance of the major changes in benthic foraminifera (replacement of the Nuttallides truempyi assemblage, acme of Nuttallides umbonifera) need to be corroborated in other locations. The most critical unresolved faunal problem is the timing of the replacement of the Nuttallides truempyi assemblage and the extinction of the Eocene abyssal taxa. Although these events are most dramatic in the deepest sites, no site with continuous recovery of the upper Eocene section from paleodepths greater than 3 kilometers has been studied. Such a site needs to be identified and studied in order to resolve the timing of these events.

Two fundamental problems in using benthic foraminifera to interpret paleoceanographic changes are illustrated by the faunal studies contained here (Appendixes 1,4,5). 1) Size-fraction biases may significantly alter results. Comparison of data from Site 548 from Appendix 5 (> 150 μm size fraction) with data of Poag (in preparation) from the greater than 63 μm size fraction shows significant differences. Unfortunately, the choice of size fraction remains arbitrary and there is little uniformity among investigators. 2) The major changes in Cenozoic deep-sea benthic foraminifera often do not coincide with other evidence of paleoceanographic changes. This is illustrated by four cases. a) Tjalsma and Lohmann (1982) noted a major benthic foraminiferal extinction event near the end of the Paleocene; no change in $\delta^{18}O$ or other known paleoceanographic event is associated with this change. b) The major benthic foraminiferal changes discussed here began prior to the major

$\delta^{18}O$ increase and the change in intensity of abyssal circulation. c) Important changes in benthic foraminiferal assemblages in the middle Miocene apparently post-date a major enrichment of ^{18}O, inferred to be a major temperature drop (Douglas and Woodruff, 1981). d) There is little correlation between glacial and interglacial cycles and benthic foraminiferal response in the Pleistocene Atlantic Ocean (Lohmann, 1978; Peterson and Lohmann, 1981; Bremer, 1982). These cases illustrate that benthic foraminiferal changes alone cannot be used to interpret abyssal circulation changes unequivocally. Future studies need to use other corroborative data including isotopic studies, planktonic stratigraphy, seismic stratigraphy, lithostratigraphy, geophysically-constrained subsidence models ("backtracking") and well histories ("backstripping") to help gain an understanding of the benthic foraminiferal response to oceanographic changes.

The abyssal circulation model presented here is not completely developed or correct in detail, and continued testing of the model is needed. Several important questions associated with the model remain unanswered or need further documentation. A better understanding of the abyssal circulation history obviously can be developed by the acquisition of additional, carefully placed geological (borehole) and geophysical data. In particular, three kinds of analyses would significantly improve the perception of the abyssal circulation history. 1) Basin-wide mapping and sequence analyses using the seismic stratigraphic record in the North Atlantic, Faeroe-Shetland Channel, and Norwegian-Greenland Sea would provide clearer definition of the intra-basin geometries of current effects, and would allow an inter-basinal comparison that should result in a better understanding of tectonic threshold events. Eventually this approach could be extended to the Brazil and Argentine Basins of the South Atlantic in order to establish the relative importance and timing of northern and southern sources for vigorously circulating bottom water. 2) Improved subsidence models of the Greenland-Scotland Ridge would allow better definition of the timing of shallow and deep marine connections across the ridge. 3) Studies, such as those conducted here for the Late Paleogene, of existing Miocene DSDP material from the North Atlantic would allow the testing of the abyssal circulation model using isotopic and lithostratigraphic data, and a determination of the relationship of Miocene benthic foraminiferal changes with the abyssal circulation model and isotopic changes. 4) Continued benthic faunal, planktonic floral and faunal, and isotopic studies of critically placed DSDP holes proposed for the last phase of GLOMAR Challenger drilling and for GLOMAR Explorer drilling in the southern Labrador Sea, the Greenland-Scotland Ridge, and the northeastern North Atlantic would help clarify the Tertiary patterns of paleocirculation, the potential northern-hemisphere climatic effects on bottom-water formation, and the history of surface and deep water connections of the Norwegian-Greenland Sea with the North Atlantic.

Future work will address the age of the circulation event in the Faeroe-Shetland Channel and its relationship to the change in abyssal circulation associated with reflector R4 in the North Atlantic. This should conclusively show if the events are related in time and genesis, and if the Faeroe-Shetland Channel was a conduit for northern bottom waters flowing into the North Atlantic at the time that reflector R4 was formed.

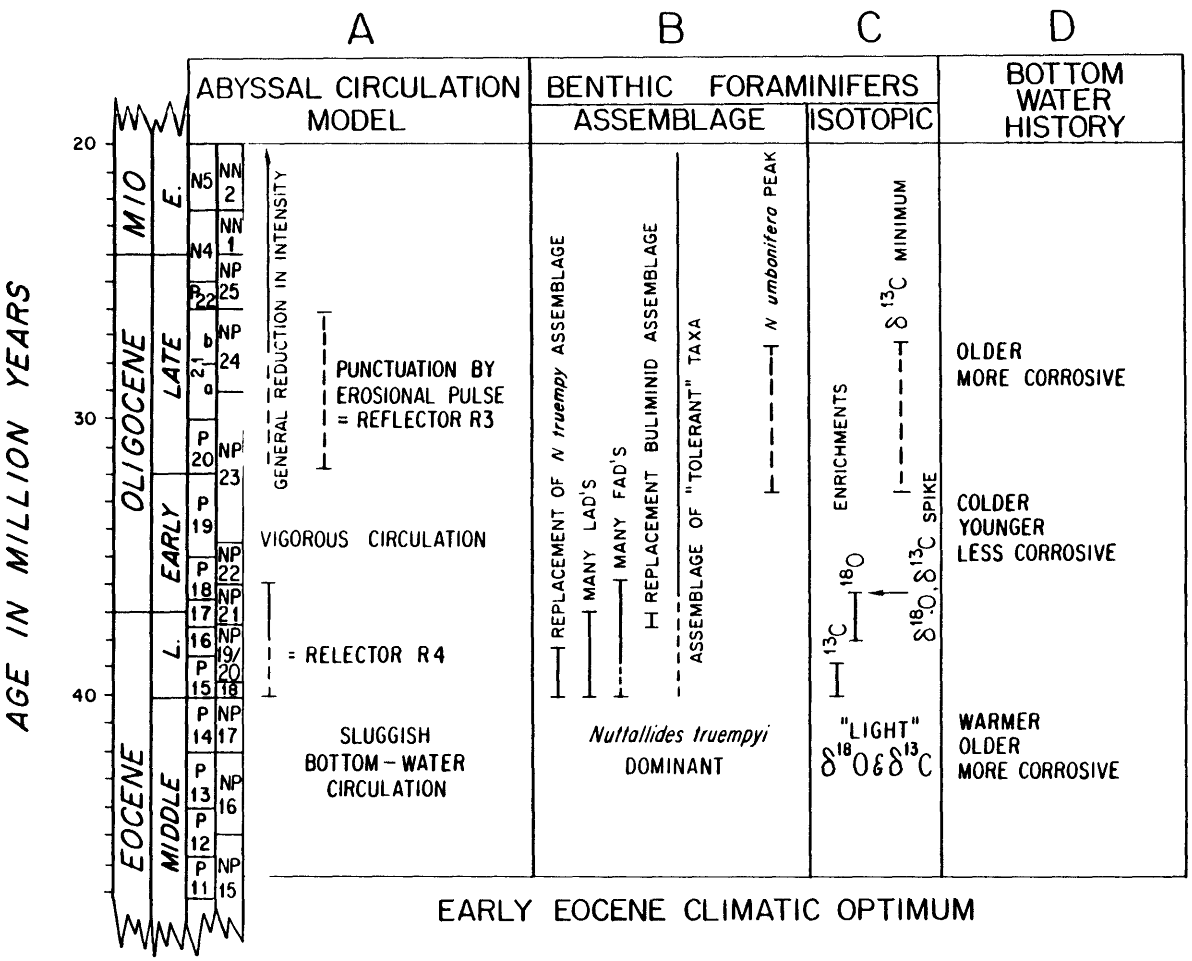

A
B
C
D
ABYSSAL CIRCULATION MODEL
BENTHIC FORAMINIFERS
ASSEMBLAGE
ISOTOPIC
BOTTOM WATER HISTORY
AGE IN MILLION YEARS
20
30
40
MIO
OLIGOCENE
EOCENE
E.
LATE
EARLY
L.
MIDDLE
N5
N4
P22
b
21
a
P 20
P 19
P 18
17
16
P 15
P 14
P 13
P 12
P 11
NN 2
NN 1
NP 25
NP 24
NP 23
NP 22
NP 21
NP 19/20
18
NP 17
NP 16
NP 15
GENERAL REDUCTION IN INTENSITY
PUNCTUATION BY EROSIONAL PULSE = REFLECTOR R3
VIGOROUS CIRCULATION
= RELECTOR R4
SLUGGISH BOTTOM - WATER CIRCULATION
REPLACEMENT OF N truempy ASSEMBLAGE
MANY LAD'S
MANY FAD'S
REPLACEMENT BULIMINID ASSEMBLAGE
ASSEMBLAGE OF "TOLERANT" TAXA
N umbonifera PEAK
Nuttallides truempyi DOMINANT
ENRICHMENTS
^{13}C
^{18}O
δ^{13}C MINIMUM
δ^{18}O, δ^{13}C SPIKE
"LIGHT" δ^{18}O & δ^{13}C
OLDER MORE CORROSIVE
COLDER YOUNGER LESS CORROSIVE
WARMER OLDER MORE CORROSIVE
EARLY EOCENE CLIMATIC OPTIMUM

Figure 21. Summary of paleoceanographic events. Column A indicates abyssal circulation events inferred from Chapter 2, column B indicates benthic foraminiferal changes, column C indicates timing of major benthic foraminiferal isotopic events, and column D indicates bottom water history inferred from the data in columns B and C. After Miller et al. (in press).

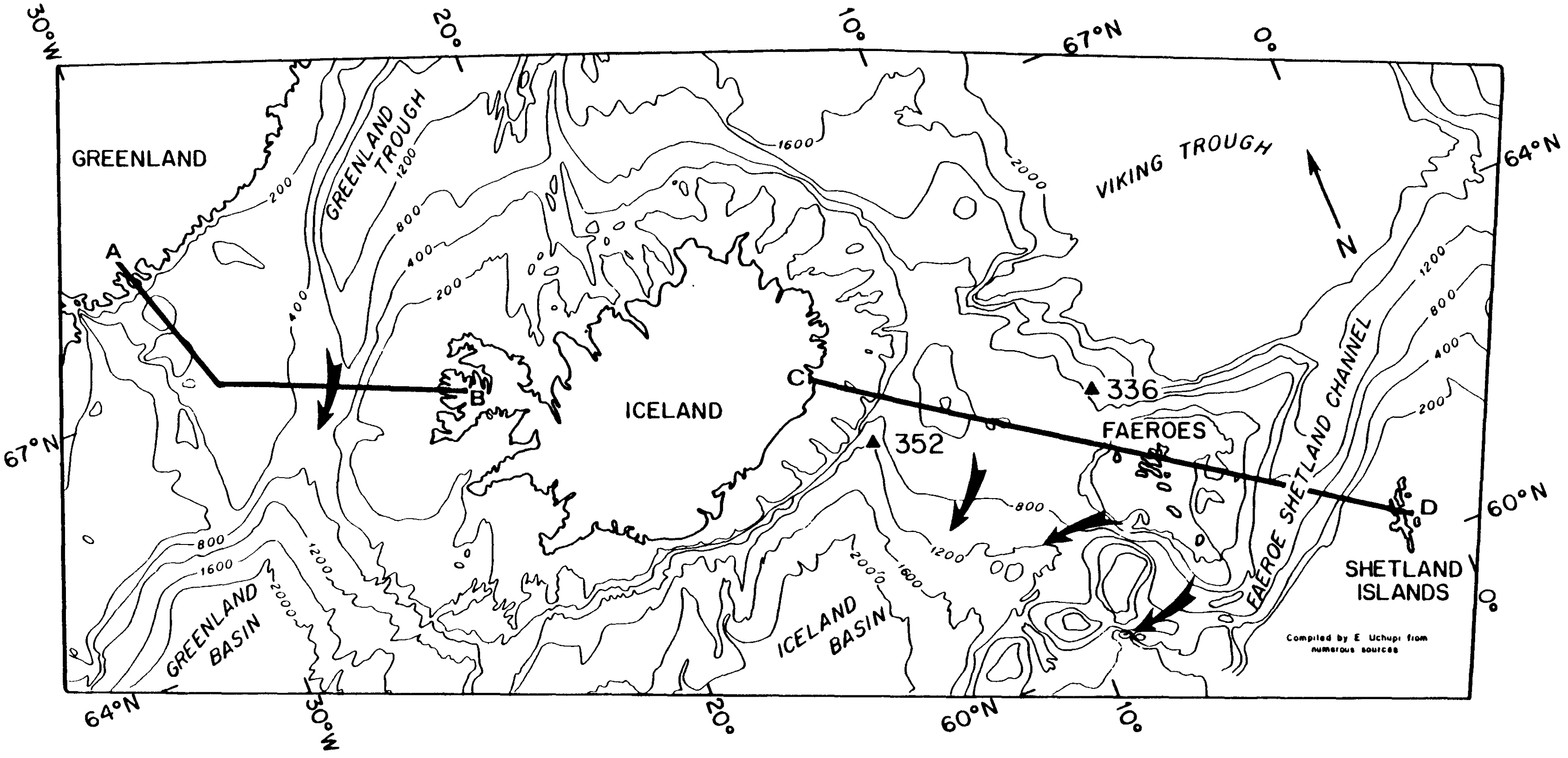
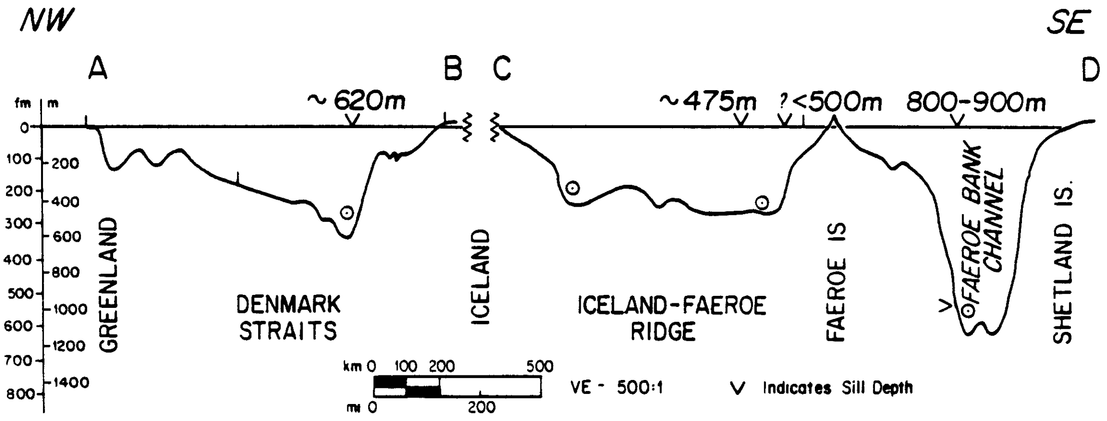

30°W
20°
10°
67°N
0°
64°N
GREENLAND
GREENLAND TROUGH
1200
800
400
200
1600
2000
VIKING TROUGH
N
A
B
C
D
ICELAND
336
352
FAEROES
FAEROE SHETLAND CHANNEL
67°N
60°N
SHETLAND ISLANDS
GREENLAND BASIN
ICELAND BASIN
64°N
30°W
20°
60°N
10°
0°
Compiled by E. Uchupi from numerous sources

Figure 22a. Bathymetry of Greenland-Scotland Ridge, in meters (from Uchupi and Hays, unpublished data). Sections A-B and C-D are illustrated in Figure 22b. DSDP sites 352 and 336 are indicated with triangles. Arrows indicate modern overflow routes (see text).



CROSS SECTION: GREENLAND-SCOTLAND RIDGE

Figure 22b. Bathymetric cross-section of the Greenland-Scotland Ridge (located in Figure 22a) showing maximum present sill depths.

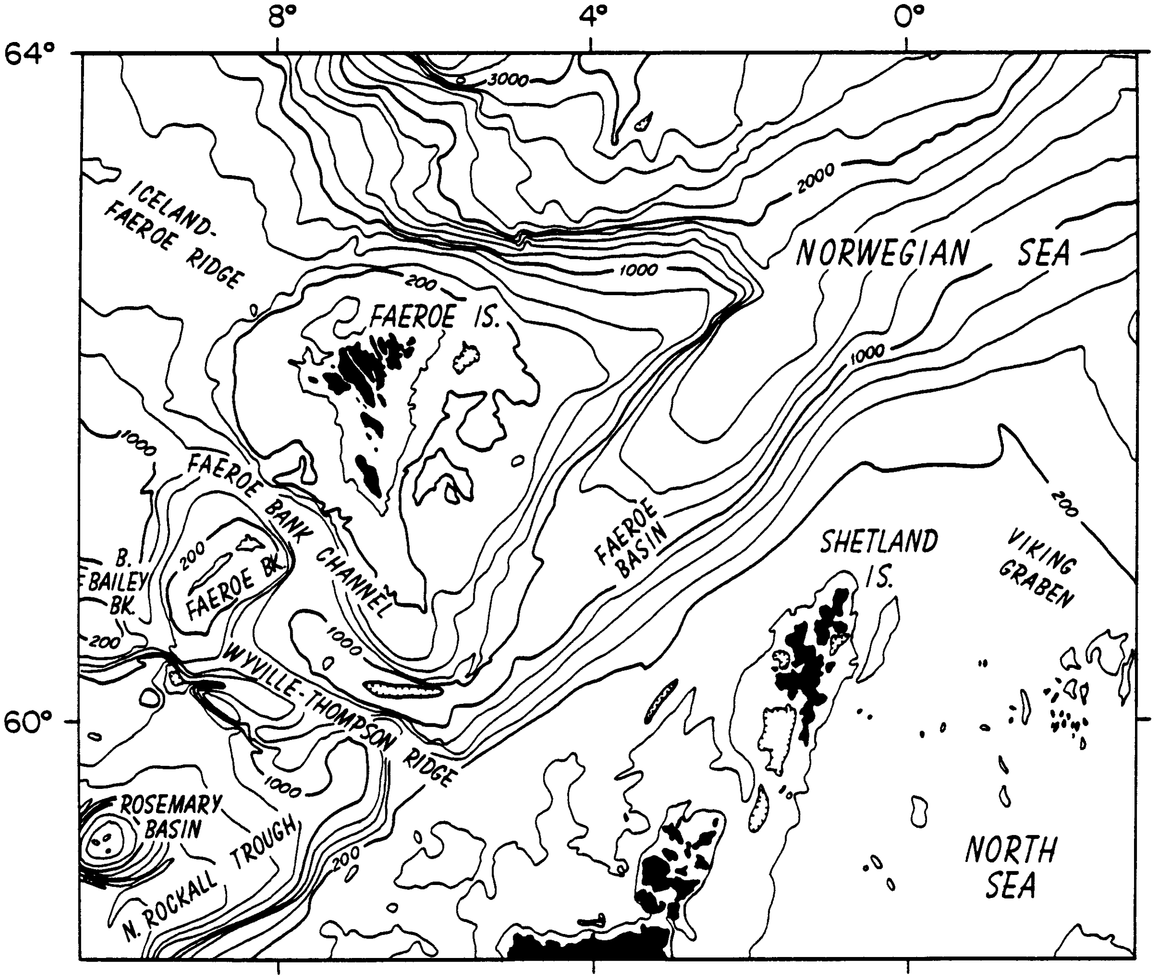

8°
4°
0°
64°
60°
3000
2000
1000
200
ICELAND-FAEROE RIDGE
NORWEGIAN SEA
FAEROE IS.
FAEROE BANK CHANNEL
FAEROE BASIN
SHETLAND IS.
VIKING GRABEN
B. BAILEY BK.
FAEROE BK.
WYVILLE-THOMPSON RIDGE
ROSEMARY BASIN
N. ROCKALL TROUGH
NORTH SEA

Figure 23. Bathymetric map of Faeroe-Shetland Channel and environs. After NAVOCEANO Map NA5 (1977).

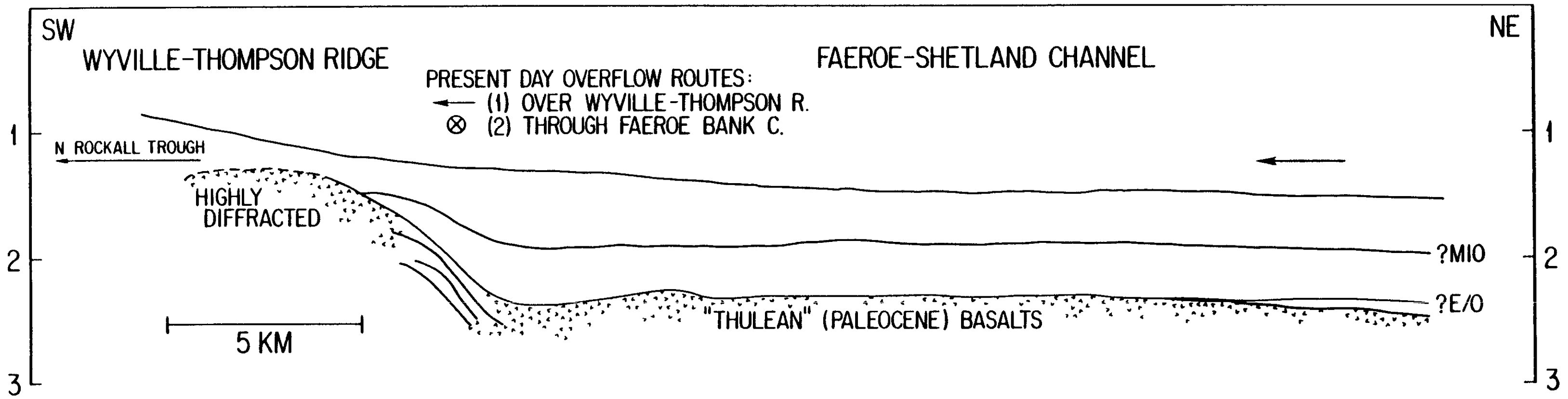

SW
NE
WYVILLE-THOMPSON RIDGE
FAEROE-SHETLAND CHANNEL
PRESENT DAY OVERFLOW ROUTES:
(1) OVER WYVILLE-THOMPSON R.
⊗ (2) THROUGH FAEROE BANK C.
N ROCKALL TROUGH
HIGHLY DIFFRACTED
?MIO
?E/O
"THULEAN" (PALEOCENE) BASALTS
5 KM
1
2
3
1
2
3

Figure 24. Interpretation of multichannel profile in the southern Faeroe-Shetland Channel. Data provided by D. Smythe and M. Ridd.

CONCLUSIONS

1. Based upon seismic stratigraphic studies of the northern North Atlantic, a model for the development of abyssal circulation in the North Atlantic is developed.
   a. The widely distributed reflector R4 correlates with an unconformity that can be traced to its correlative conformity at the top of the Eocene. Reflector R4 reflects a major change in abyssal circulation, from sluggishly circulating bottom water in the Eocene to more vigorously circulating bottom water in the Oligocene. Sediment distribution studies indicate a northern source for this bottom water, probably from the Arctic Ocean via the Norwegian-Greenland Sea and Faeroe-Shetland Channel.
   b. Increased rates of deposition associated with the development of large sedimentary drifts in the later Oligocene (above reflector R3) through Miocene (especially above reflector R2 = late early Miocene) are interpreted as reflecting a general reduction in intensity of abyssal circulation. This general reduction may have been punctuated by brief erosional pulses.

2. A major $\delta^{18}O$ increase begins at ~ 38 Ma (late Eocene), culminating in a rapid (< 0.5 my) increase in $\delta^{18}O$ just above the Eocene/Oligocene boundary (~ 36.5 Ma) in the Bay of Biscay/Goban Spur regions (DSDP Sites 119/401 and 549). A rapid $\delta^{13}C$ increase also occurs just above the boundary in these sites. These changes represent the transition from warm, old (low O_2, high CO_2, low pH, hence more corrosive) Eocene bottom waters to colder and younger (higher O_2, lower CO_2, higher pH, hence less corrosive) bottom waters in the early Oligocene.

3. Benthic foraminiferal changes occurring in the Late Paleogene in the northern North Atlantic include:
   a. In the southern Labrador Sea (DSDP Site 112), an assemblage of predominantly agglutinated benthic foraminifera is replaced by a deep-sea calcareous assemblage between the middle Eocene and the early Oligocene.

b. In abyssal sites greater than 3km, an indigenous Eocene calcareous fauna including *Nuttallides truempyi*, *Clinapertina* spp., *Abyssammina* spp., *Aragonia* spp., and *Alabamina dissonata* becomes extinct between the middle Eocene and early Oligocene.

c. In shallower sites (< 3km paleodepth) in the Goban Spur/Bay of Biscay regions and elsewhere in the Atlantic, a *Nuttallides truempyi*-dominated assemblage is replaced in the early late Eocene (~ 38.5-40 Ma).

d. In the late Eocene to early Oligocene, a *Globocassidulina subglobosa*-*Gyroidinoides*-*Cibicidoides ungerianus*-*Oridorsalis umbonatus* assemblage dominates the northern North Atlantic. These taxa are bathymetrically wide-ranging and stratigraphically long-ranging, and may be interpreted as tolerant of environmental changes.

In general, these changes are thought to reflect the transition from an Eocene ocean more corrosive to calcium carbonate to a less corrosive Oligocene ocean.

e. A middle Oligocene acme in *Nuttallides umbonifera* in the deepest sites (> 3km paleodepth) and a concomittant minimum in δ^{13}C are interpreted as reflecting older, more corrosive bottom water.

4. The faunal and isotopic data are combined with the abyssal circulation model in the following scenario:

a. Old, warm, corrosive, and sluggishly circulating bottom water was replaced by younger, colder, and more vigorously circulating bottom water which had a northern source (Arctic and/or Norwegian-Greenland Sea). This developed throughout the late Eocene (40-37 Ma), but was punctuated by a rapid event (= reflector R4 and the concomittant δ^{18}O increase) just above the Eocene/Oligocene boundary (~ 36.5 Ma).

b. During the middle Oligocene, bottom water circulation was reduced, and the age and corrosiveness of bottom water increased.

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APPENDIX 1

LATE CRETACEOUS TO EARLY TERTIARY AGGLUTINATED BENTHIC FORAMINIFERA IN THE LABRADOR SEA

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Late Cretaceous to Early Tertiary agglutinated benthic foraminifera in the Labrador Sea

ABSTRACT

A major faunal change occurred at the end of the Eocene in the deep Labrador Sea (DSDP Site 112). Predominantly agglutinated Eocene benthic foraminiferal assemblages are replaced by an Oligocene calcareous assemblage This replacement correlates with similar changes in the Canadian margin and the North Sea where the exit of agglutinated foraminifera is associated with a change in depositional environment. In Site 112, however, lithology, percent carbonate, and percent organic carbon are relatively constant across the faunal change. Thus, at least in the deep sea, these properties may not be critical to the development of predominantly agglutinated assemblages. We infer that certain hydrographic properties (low oxygen, high CO_2, low pH, and thus more corrosive waters) favor the development of agglutinated assemblages. We suggest that the replacement of agglutinated foraminifera in the deep Labrador Sea was due to a change in hydrographic properties. Seismic evidence from this region indicates initiation of northern sources of bottom water in the late Eocene to early Oligocene which may have resulted in a change in hydrography and the exit of agglutinated foraminifera. Despite diachrony of certain occurrences in Site 112 and the Canadian margin, compared with previously reported ranges of agglutinated taxa from the North Sea, European flysch basin, and the deep sea, agglutinated foraminifera are useful in the zonation of Labrador Sea sediments.

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INTRODUCTION

Agglutinated (arenaceous) foraminiferal assemblages occur in Maestrichtian to Eocene mudstones of the Labrador Shelf/northeast Newfoundland Shelf and upper Paleocene to Eocene shales of the North Sea. These taxonomically similar assemblages resemble the *Rhabdammina* assemblages described (Brouwer, 1965) from pelitic intervals of flysch sediments throughout the world. Gradstein and Berggren (1981) described the taxonomy, zoogeography, and biostratigraphic utility of these Late Cretaceous to Tertiary "flysch-type" agglutinated assemblages, discussed paleoecologic implications of their distribution, and related their distribution to the tectonic history of the Labrador Sea and the North Sea.

Predominantly agglutinated assemblages resembling the flysch-type *Rhabdammina* assemblages also have been described from deep-sea sediments recovered by the Deep Sea Drilling Project/International Phase of Ocean Drilling (text-figs. 1, 2). Our study compares the distribution of the deep-sea assemblages as reported in the Initial Reports of the Deep Sea Drilling Project, examines in detail Eocene agglutinated assemblages of DSDP Site 112 (Labrador Sea), and compares the Site 112 assemblages with Maestrichtian to Eocene agglutinated assemblages of the Labrador and East Newfoundland shelves. Various paleoecologic models for agglutinated assemblages are evaluated, using lithologic and faunal data from Site 112. We conclude that a late Eocene to early Oligocene change in abyssal circulation resulted in the demise of the agglutinated assemblages in Site 112; local tectonic and sedimentologic events may have been important in the replacement of agglutinated assemblages in the Canadian margin and the North Sea.

PREVIOUS WORK

Berggren (1972) noted relatively diverse assemblages of predominantly agglutinated benthic foraminifera in the ?Paleocene to Eocene sediments of Site 112. Subsequently, Middle Cretaceous to Early Tertiary agglutinated benthic foraminiferal assemblages have been recorded in 30 DSDP sites in various parts of the world (text-figs. 1, 2; Beckman, 1972; Krasheninnikov, 1973, 1974; Webb, 1975; Rögl, 1976; van Hinte,

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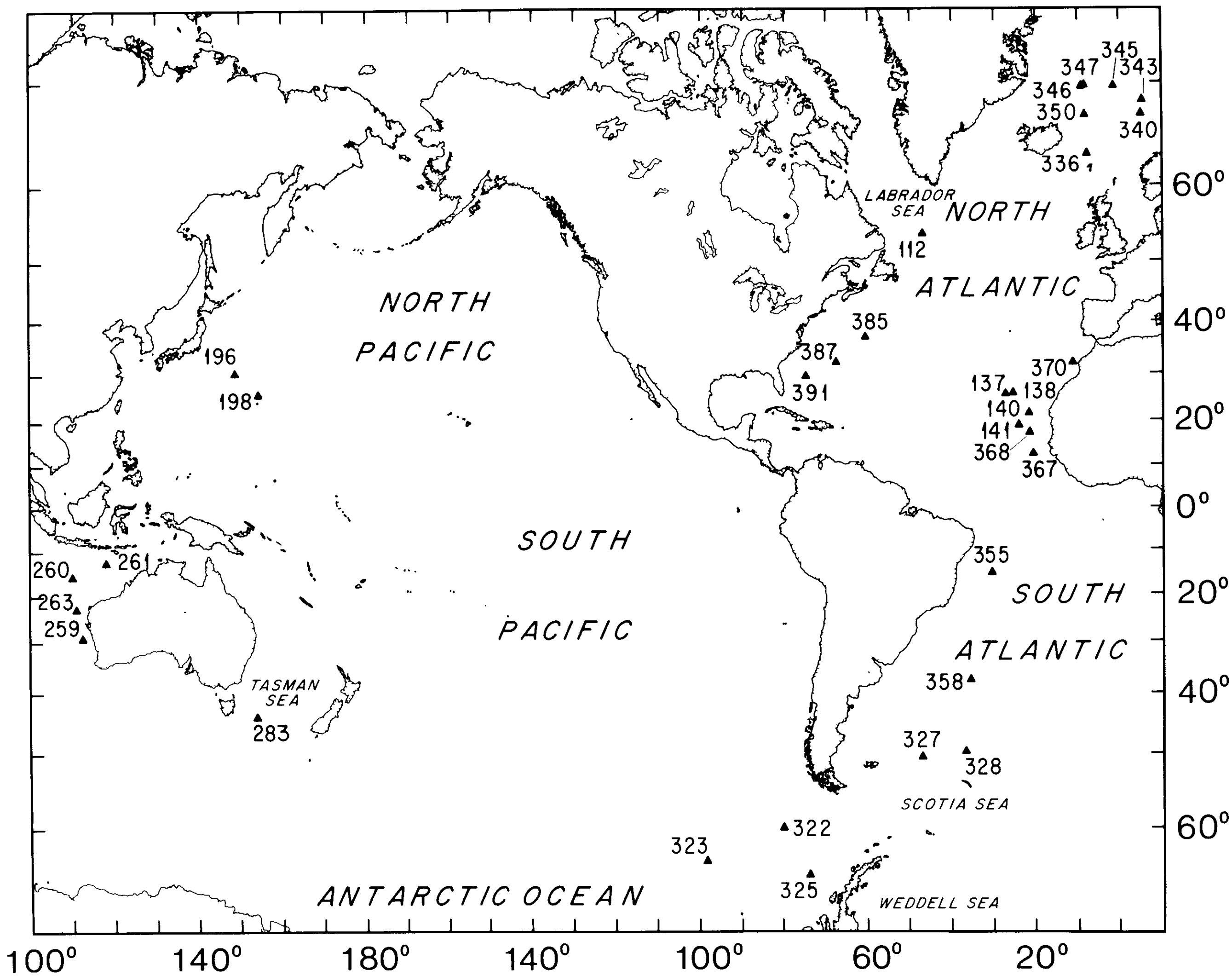

345
347
343
346
350
340
336
60°
LABRADOR SEA
NORTH
112
ATLANTIC
NORTH
PACIFIC
196
198
385
40°
387
391
370
137
138
140
141
368
367
20°
0°
SOUTH
260
261
263
259
SOUTH
PACIFIC
355
SOUTH
ATLANTIC
20°
TASMAN SEA
283
358
40°
327
328
SCOTIA SEA
322
323
60°
ANTARCTIC OCEAN
325
WEDDELL SEA
100°
140°
180°
140°
100°
60°
20°

TEXT-FIGURE 1
Location map of DSDP-IPOD sites containing type A agglutinated benthic foraminiferal assemblages.

in Talwani, Udintsev et al., 1976; Krasheninnikov and Pflaumann, 1977; Tjalsma, 1977; Gradstein, 1978; McNulty, 1979). These assemblages consist of mostly agglutinated taxa, but calcareous benthic forms may be minor constituents. Planktonic foraminifera are, in general, rare or absent. Miller et al. (1979) and Gradstein and Berggren (1981) differentiated two types of deep-sea assemblages of agglutinated benthic foraminifera distinguishable by wall texture, test size, and taxonomic composition. Assemblage type A is found in bathyal and abyssal sediments; it is similar in taxonomic composition to agglutinated assemblages of the Carpathian and Alpine flysch basins (= "flysch-type" agglutinated assemblages of Gradstein and Berggren, 1981), the North Sea, and the Canadian margin. In general, this assemblage is characterized by coarse-grained, larger-sized tests. "Simple" forms including the single chambered and uniserial genera *Rhabdammina, Bathysiphon, Ammodiscus, Lituotuba, Reophax,* and *Hormosina* often dominate this assemblage (pl. 1). Biserial (*Spiroplectammina, Textularia*), multiserial (*Gaudryina, Dorothia*), trochoid (*Trochammina, Recurvoides, Trochamminoides*), and planispiral (*Cribrostomoides, Haplophragmoides, Cyclammina*) forms are less abundant (pl. 2). Assemblage type B has been recorded only from abyssal sediments. This assemblage is composed of small (generally less than 150 μm), thin, smooth walled, predominantly "complex" forms of the families Lituolidae, Trochamminidae, and Textulariidae. Krasheninnikov (1974) noted certain generic similarities between an Upper Cretaceous type B assemblage in Sites 260 and 261 and flysch-type (= type A) agglutinated foraminiferal assemblages. However, such genera as *Plecto-*

recurvoides, Haplophragmium, and *Labrospira,* which are usually rare to absent in type A assemblages, predominate in type B assemblages.

Some of the taxonomic differences between type A and type B assemblages may be due to different stratigraphic and facies distribution. Text-figure 2 shows the distribution of predominantly agglutinated foraminiferal assemblages and lithofacies versus age as reported from 21 DSDP/IPOD sites. Type A assemblages, reported from Sites 112, 137, 138, 140, 141, 367 (cores 15 to 18), 368, 327, 328, 323, and 283 are found in Late Cretaceous to Tertiary sediments. In the Tertiary, the only apparent post-Eocene occurrences of type A assemblages are at Sites 323 and 325 (Bellingshausen Sea; Rögl, 1976) and in the Norwegian-Greenland Sea (van Hinte, in Talwani, Udintsev et al., 1976); however, the Norwegian-Greenland Sea assemblages may be taxonomically distinct from both Type A and Type B assemblages. In contrast, type B assemblages, reported from Sites 196, 198A, 260, 261, 263 (Krasheninnikov, 1973, 1974), 385, 387 (McNulty, 1979), and 391C (Gradstein, 1978) are found only in Cretaceous sediments. A third type of agglutinated assemblage, the *Dorothia praehauteriviana* assemblage, has been reported from the deep sea by Luterbacher (1972), Krasheninnikov and Pflaumann (1977), and Kuznetsova and Seibold (1978) from Sites 101, 105, 367 (cores 26–29), 370 (cores 38–49), and 391C. This assemblage is found only in Lower Cretaceous sediments. The different agglutinated assemblages also occur in different lithofacies. Text-figure 2 shows that both type A and type B assemblages are found primarily in zeolithic and variegated clays, but type B assemblages (with the possible exception of Site 385, core 12) are restricted to such sediments, while type A assemblages are found in a wider variety of lithofacies. The *D. praehauteriviana* assemblage, in contrast, is found only in Neocomian limestones.

Type A and type B assemblages also occur at different paleodepths. We use the "backtracking" method (Sclater et al., 1971; Berger and Winterer, 1974) to estimate paleobathymetry for these deep-sea agglutinated assemblages. With the exception of the Norwegian-Greenland Sea (paleodepths less than 0.7 km to 3.0 km), assemblage A occurs above crust which may be backtracked to paleodepths of 2.5 to 4.5 km (table 1). Assemblage B, however, is restricted to paleodepths greater than 4.0 km.

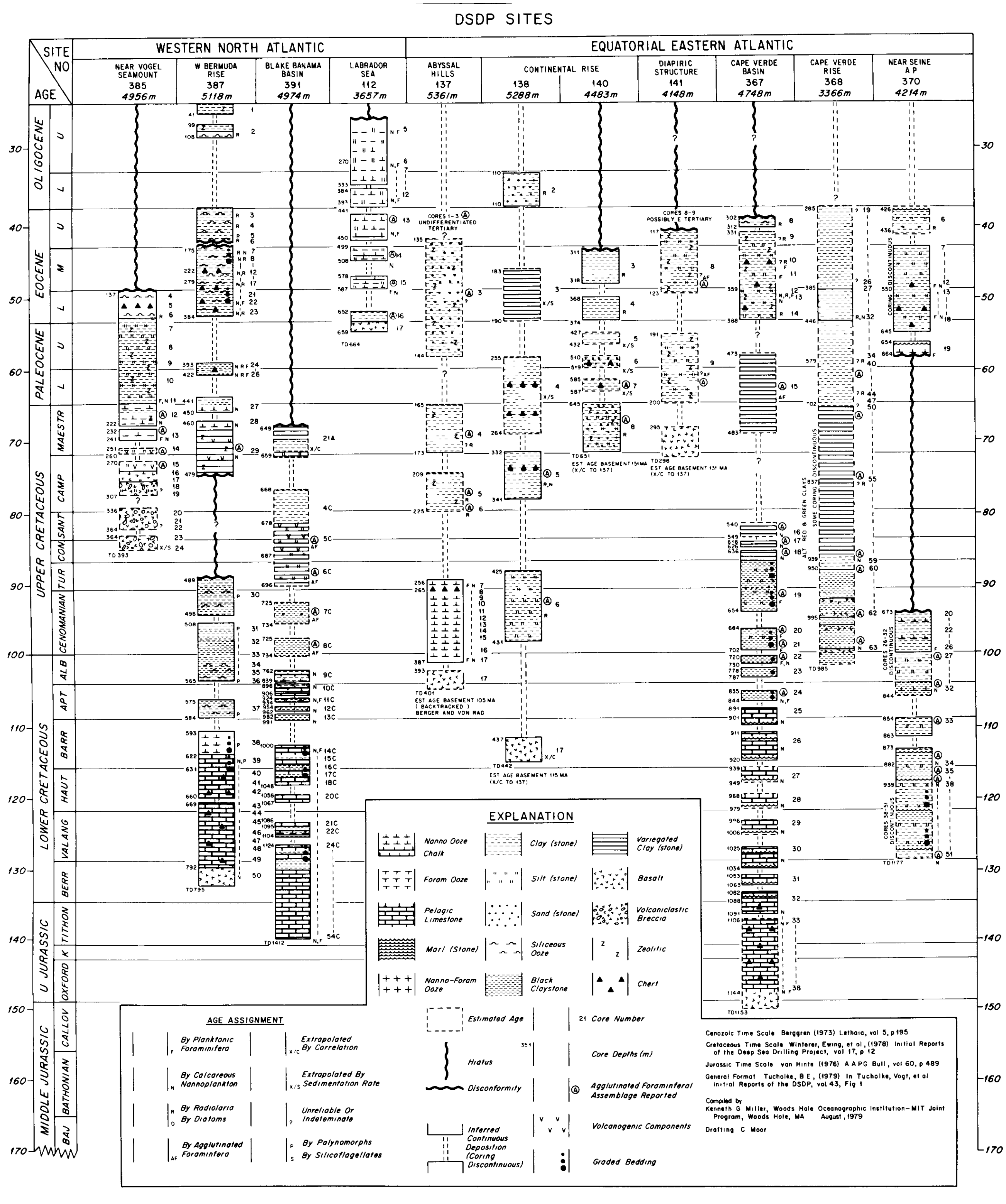
TABLE 1
Paleobathymetry of DSDP sites containing type A agglutinated benthic foraminifera. Pr = present depth (km), Pa = paleodepth (km) obtained by "backtracking."

Area	Site no	Depth in km	
		Pr	Pa
Norwegian-Greenland Sea	343	3.1	3.0
	346	0.7	0.7
	347	0.7	0.7
	349	0.9	0.9
	350	1.3	1.3
Atlantic Ocean	112	3.6	3.0
	140	4.5	4.0
	141	4.1	4.0
	367	4.7	4.5
	368	3.3	4.0 uplifted in Neogene
	328	5.1	3.5–4
Tasman Sea	283	4.7	3.0
Subantarctic	323	5.0	3–3.5

Predominantly agglutinated foraminiferal assemblages have been accounted for by water depth, carbonate availability, hydrographic properties, and substrate characteristics. The extensive paleobathymetric range of type A assemblages suggests that depth alone does not control the distribution of these deep-sea agglutinated foraminifera. Hesse and Butt (1976) suggested that flysch-type agglutinated foraminiferal assemblages result from the exclusion of calcareous benthic foraminifera below a local CCD (Carbonate Compensation Depth). Saidova (1960, 1965) explained the distribution of Recent abyssal agglutinated foraminiferal assemblages from the Pacific Ocean with hydrographic properties (high pressure, low temperature, low dissolved oxygen). Saidova's explanation does not exclude carbonate control, as these properties promote carbonate dissolution. Similarly, Hiltermann (1973) suggested that low temperature and low oxygen favored agglutinated foraminifera in flysch deposits. Ksiazkiewicz (1961, 1975) and Moorkens (1976) invoked substrates containing abundant organic matter as an important control on flysch-type agglutinated assemblages. This explanation does not exclude carbonate or hydrographic limitations, as low O_2, high CO_2, low pH, low Eh, and poor circulation are often associated with high organic matter. These explanations are not mutually exclusive, but they do differ in emphasis. We evaluate them using lithologic and faunal data from DSDP Site 112.

METHODS

Twenty samples from the Eocene to lower Oligocene sediments of Site 112 were examined for benthic foraminiferal content (cores 11–16). All benthic foraminifera were picked from the greater than 150 μm size-fraction. The less than 150 μm size-fraction was examined to ensure that the larger size-fraction con-


DSDP SITES
SITE NO
AGE
WESTERN NORTH ATLANTIC
EQUATORIAL EASTERN ATLANTIC
NEAR VOGEL SEAMOUNT 385 4956m
W BERMUDA RISE 387 5118m
BLAKE BANAMA BASIN 391 4974m
LABRADOR SEA 112 3657m
ABYSSAL HILLS 137 5361m
CONTINENTAL RISE 138 5288m 140 4483m
DIAPIRIC STRUCTURE 141 4148m
CAPE VERDE BASIN 367 4748m
CAPE VERDE RISE 368 3366m
NEAR SEINE A P 370 4214m
AGE IN MILLIONS OF YEARS
OLIGOCENE
EOCENE
PALEOCENE
UPPER CRETACEOUS
LOWER CRETACEOUS
U JURASSIC
MIDDLE JURASSIC
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M
L
MAESTR
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SANT
CON
TUR
CENOMANIAN
ALB
APT
BARR
HAUT
VALANG
BERR
TITHON
K
OXFORD
CALLOV
BATHONIAN
BAJ
30
40
50
60
70
80
90
100
110
120
130
140
150
160
170
CORES 1-3 Ⓐ UNDIFFERENTIATED TERTIARY
CORES 8-9 POSSIBLY E TERTIARY
TD401 EST AGE BASEMENT 105 MA (BACKTRACKED) BERGER AND VON RAD
TD442 EST AGE BASEMENT 115 MA (X/C TO 137)
TD651 EST AGE BASEMENT 151 MA (X/C TO 137)
TD298 EST AGE BASEMENT 131 MA (X/C TO 137)
RED & GREEN CLAYS SOME CORING DISCONTINUOUS
CORING DISCONTINUOUS
CORES 26-32 DISCONTINUOUS
CORES 38-51 DISCONTINUOUS
EXPLANATION
Nanno Ooze Chalk
Clay (stone)
Variegated Clay (stone)
Foram Ooze
Silt (stone)
Basalt
Pelagic Limestone
Sand (stone)
Volcaniclastic Breccia
Marl (Stone)
Siliceous Ooze
Zeolitic
Nanno-Foram Ooze
Black Claystone
Chert
AGE ASSIGNMENT
F By Planktonic Foraminifera
N By Calcareous Nannoplankton
R By Radiolaria
D By Diatoms
AF By Agglutinated Foraminifera
X/C Extrapolated By Correlation
X/S Extrapolated By Sedimentation Rate
? Unreliable Or Indeterminate
P By Palynomorphs
S By Silicoflagellates
Estimated Age
Hiatus
Disconformity
Inferred Continuous Deposition (Coring Discontinuous)
21 Core Number
351 Core Depths (m)
Ⓐ Agglutinated Foraminiferal Assemblage Reported
Volcanogenic Components
Graded Bedding
Cenozoic Time Scale Berggren (1973) Lethaia, vol 5, p 195
Cretaceous Time Scale Winterer, Ewing, et al, (1978) Initial Reports of the Deep Sea Drilling Project, vol 17, p 12
Jurassic Time Scale van Hinte (1976) A A P G Bull, vol 60, p 489
General Format Tucholke, B E, (1979) In Tucholke, Vogt, et al Initial Reports of the DSDP, vol 43, Fig 1
Compiled by
Kenneth G Miller, Woods Hole Oceanographic Institution-MIT Joint Program, Woods Hole, MA August, 1979
Drafting C Moor

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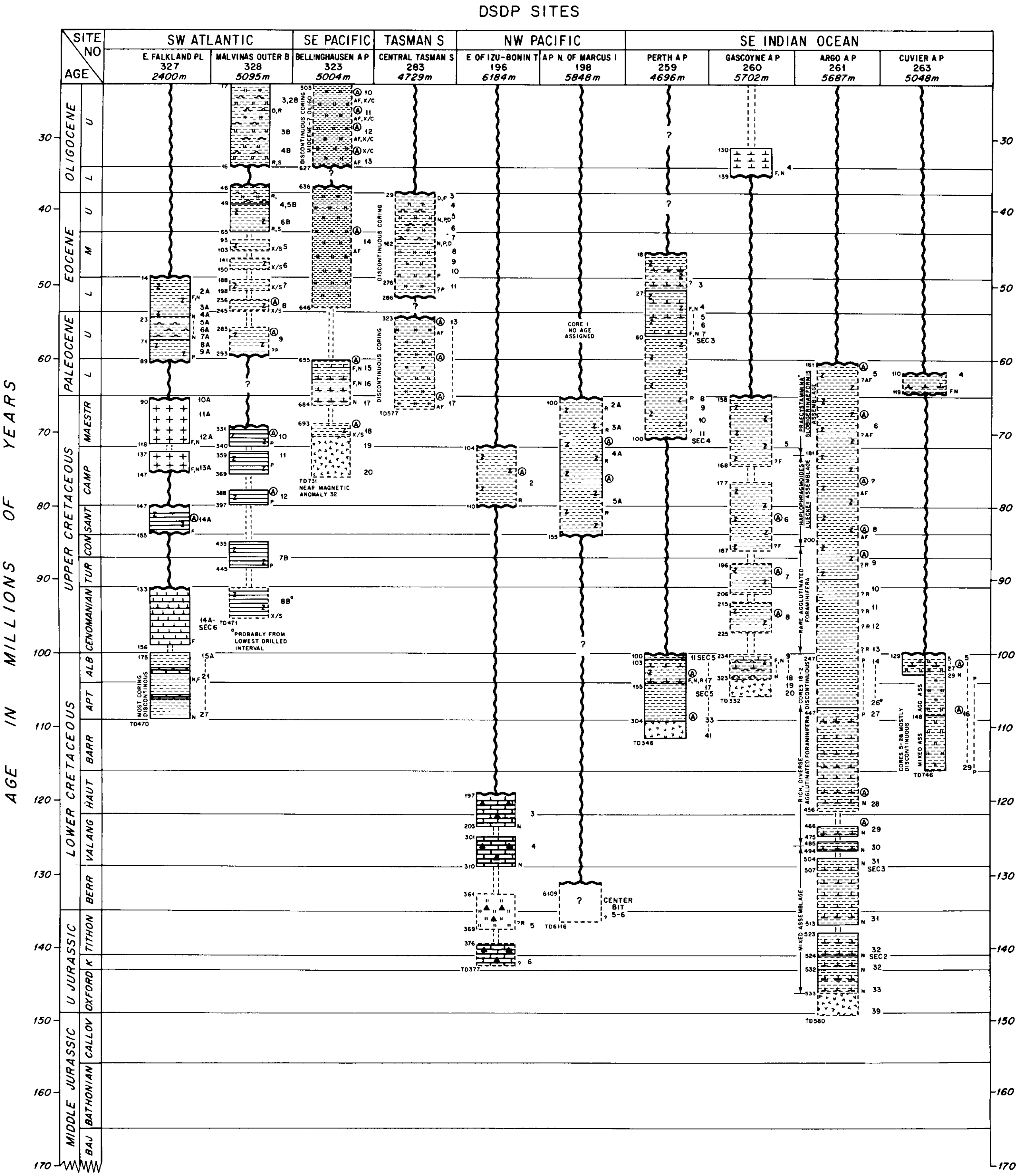
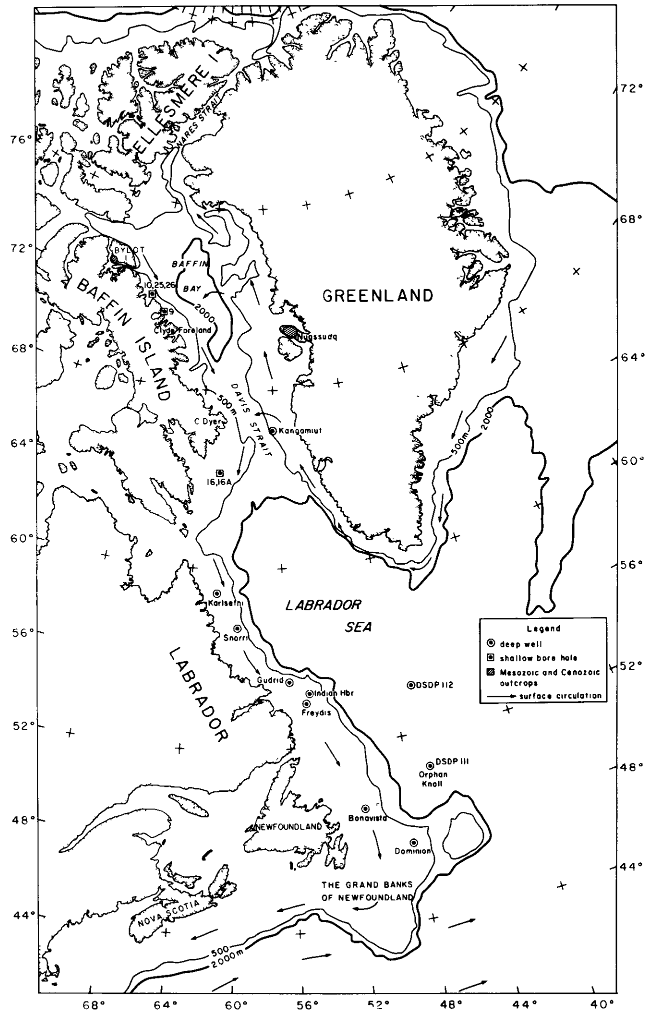


TEXT-FIGURE 2

Stratigraphic distribution of lithofacies and agglutinated foraminiferal assemblages. Sites containing agglutinated foraminifera from Miocene to Eocene of Norwegian-Greenland Sea (Sites 336, 340, 343, 344, 345, 347, 350) and ?Oligocene to early Miocene of the Bellingshausen Sea (Sites 332, 325) not shown.


ELLESMERE I.
NARES STRAIT
BYLOT I.
BAFFIN BAY
BAFFIN ISLAND
10,25,26
9
Clyde Foreland
2000
GREENLAND
Nugssuaq
DAVIS STRAIT
500m
C Dyer
Kangamiut
16,16A
500m
2000
LABRADOR SEA
Karlsefni
Snorri
LABRADOR
Gudrid
Indian Hbr
Freydis
DSDP 112
DSDP 111
Orphan Knoll
Bonavista
NEWFOUNDLAND
Dominion
THE GRAND BANKS OF NEWFOUNDLAND
NOVA SCOTIA
500
2000m
Legend
◎ deep well
▣ shallow bore hole
▨ Mesozoic and Cenozoic outcrops
→ surface circulation
76°
72°
68°
64°
60°
56°
52°
48°
44°
40°

TEXT-FIGURE 3
Location of wells, Labrador Sea and environs

tained representatives of all agglutinated species present in Site 112. This was necessary because species typical of type B agglutinated assemblages are found predominantly in the smaller size-fraction. No species restricted to the smaller size-fraction were noted. Sample size is approximately 25 to 50 g; exact size is not available as most of the samples were taken on shipboard.

Most of the wells available from the Canadian margin (text-fig. 3) contain only cuttings. All foraminifera were picked from the greater than 63 μm size-fraction.

The agglutinated foraminifera were identified using the taxonomy outlined by Gradstein and Berggren (1981), which is further described in the Systematic Paleontology section of this paper. Abundance (table 3), presence/absence data of agglutinated species (text-fig. 4), and percent calcareous benthic foraminifera (table 4; text-figs. 5, 6) have been compiled for Site 112. No estimate of foraminiferal abundance was made for the Canadian margin wells; only exits were used in computing ranges of agglutinated taxa in these wells.

Estimating the relative abundance of agglutinated benthic foraminifera is complicated by the fact that many taxa occur as fragments rather than whole tests. Species of *Rhizammina, Bathysiphon, Hyperammina, Rhabdammina, Hormosina,* and *Reophax* are usually recovered only as fragments. These species, however, are often abundant, and neglecting them results in an inadequate assemblage description. Thus, all identifiable fragments of agglutinated taxa were counted. All identifiable fragments of calcareous taxa were counted for consistency. Although some duplication undoubtedly resulted, we believe that the observed changes in agglutinated assemblages and the relative abundance of agglutinated versus calcareous taxa are valid.

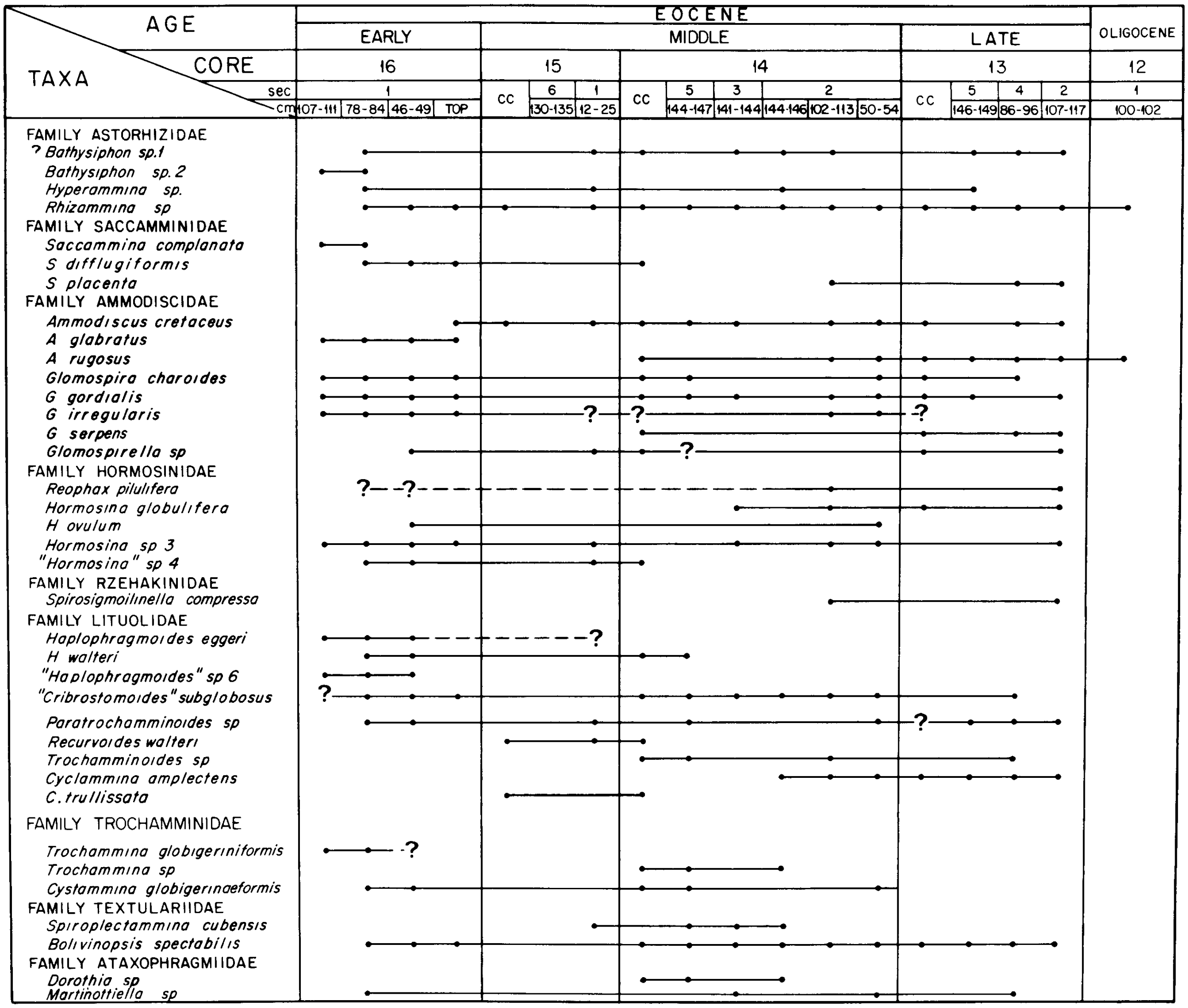
AGGLUTINATED ASSEMBLAGES FROM SITE 112, LABRADOR SEA

Planktonic biostratigraphy and lithology

Site 112 was drilled in 3657-m water depth in the southern Labrador Sea (text-fig. 3). Cores 1A to 5A (79–124 m subbottom) and 5 to 11 (270–293 m subbottom) were continuously cored; however, the Eocene/Oligocene boundary (between cores 12 and 13) was not cored. Basaltic basement was penetrated at 661 m (text-fig. 5).

The cored Paleogene section consists of indurated red clays and claystones with palagonite sills (core 16), indurated, burrow mottled gray pelagic nannofossil clay and marls (cores 12 to 15), and gray pelagic burrow mottled siliceous nannofossil clay and silt and siliceous ooze (cores 3 to 11; text-fig. 5). Cores 12 to 15 are distinguished from the overlying sediments by finer grain size, greater lithification and burrowing, and increased amounts of pyrite, authigenic calcite, and goethite (Laughton, Berggren et al., 1972).

Nannofossils and planktonic foraminifera were originally used to place cores 5 and 6 in the late Oligocene, cores 7 to 12 in the early Oligocene, and cores 13 to 15 in the Eocene. Radiolarians and diatoms also indicated an Oligocene age for cores 5 to 11. Core 16 was placed in the ?Paleocene-early Eocene on the basis of extrapolation of a 1.5 cm/1000 yr (= 15 m/my) sedimentation rate for the Eocene sediments in Site 112. An unconformity was inferred from the low sedimentation rate (less than 5.5 m/my) between the upper Oligocene in the top of core 5 (270 m subbottom) and middle Miocene sediments in the base of core 4 (209 m subbottom). No other evidence for Paleogene unconformities is present (Laughton, Berggren et al., 1972).



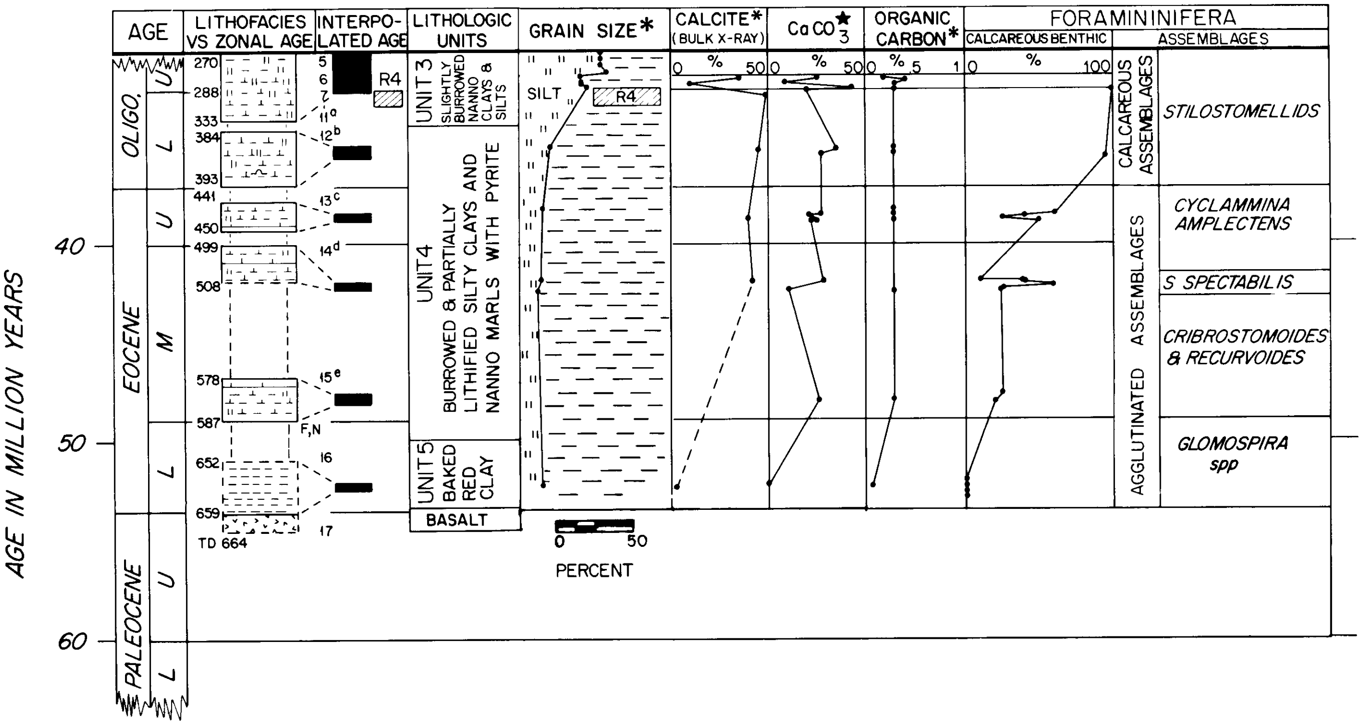
TEXT-FIGURE 4
Range chart of agglutinated taxa, DSDP Site 112. Following agglutinated species not listed: *Labrospira pacifica, Reophax nodulosus, Ammolagena clavata, Uvigeriniformis* sp., *Spiroplectinata* sp., and *Lituotuba* sp.

Type A agglutinated benthic foraminifera assemblages occur in the ?Paleocene to Eocene section of Site 112 (cores 13 to 16; text-fig. 5). We have re-examined planktonic foraminifera and calcareous nannoplankton and have refined the biostratigraphic zonation of these cores as described below.

Core 16 (revised)

No planktonic foraminifera or nannoplankton were originally reported from core 16, although ?late Paleocene-early Eocene nannoplankton were reported from the center bit just above core 16 (Perch-Nielsen, 1972). We note rare specimens of *Subbotina* sp. and *Subbotina* cf. *S. linaperta* Finlay in 16 center bit sample and sample 16-1, 107–111 cm, which agrees with the late Paleocene-early Eocene age assignment. The center bit sample contains a mixture of green marl and red clay, suggesting that the lithologic change from nannofossil clays and marls to baked red clay occurs just above core 16 at approximately 650 m.

Sedimentation rates for the lower Oligocene to Eocene clays and marls were recomputed after assigning core 15 to middle Eocene using planktonic foraminifera (Zone P10; 47–49 Ma; this paper), core 13 to the late Eocene using nannoplankton (Zone NP19, 37.5–39 Ma; Perch-Nielsen, 1972) and core 12 to the



TEXT-FIGURE 5

DSDP Site 112 Lithofacies symbols as in text-figure 2. Interpolated ages determined using a 15 m/my sedimentation rate assuming core 13 and core 15 lie in the middle of zones NP19 and P10, respectively Correlation of reflector R4 shown by cross-hatched areas.

[a,b]*E. obruta, E. subdisticha* Zone—K. Perch-Nielsen (NP21, e. Oligo), *H. reticulata* Zone—D Bukry (NP22, e Oligo)

[c]*I. recurvus* s.l. Zone—K Perch-Nielsen (NP19, late Eocene); *D barbadiensis* Zone—D. Bukry (late Eocene).

[d]*D tani nodifer* s l Zone—K Perch-Nielsen (NP16, middle Eocene), *D. barbadiensis* Zone—D. Bukry (late Eocene), NP17 M -P Aubry (middle Eocene).

[e]P10, this paper, *D. sublodoensis* Zone—K. Perch-Nielsen, D Bukry (NP14, middle Eocene)

*G. W. Bode and R. E Boyce, in· Laughton, Berggren, et al , 1972

★Pow-Foong Fan and I Zemmels, in: Laughton, Berggren, et al , 1972.

early Oligocene using nannoplankton (Zones NP21/NP22; 34–37 Ma; Perch-Nielsen, 1972; Bukry, 1972). The fossil zones are dated using the time scale of Hardenbol and Berggren (1978).

Sedimentation rates range from 12–19 m/my depending on whether an upper or lower boundary age is taken for each zone; this agrees well with the original estimate of 15 m/my.

Extrapolating this rate downward yields an age of 50–53 Ma (earliest Eocene) for the top of core 16. An age of 51–54 Ma is estimated for the base of core 16 assuming a constant sedimentation rate across the lithologic change. Core 16 is thus probably early Eocene, but possibly latest Paleocene in age.

Core 15 (revised)

Subbotina frontosa (Subbotina) and *S. linaperta* are abundant in core 15. The first appearance of *S. frontosa* occurs at the base of Zone P10 (*S. frontosa frontosa/Globorotalia pseudomayeri* Bolli CRZ; Blow, 1979). *Subbotina frontosa* exits at the top of P11 (Blow, 1979).

Acarinina bullbrooki (Bolli) [= *A. densa* (Cushman)] occurs in core 15. Berggren (1971) suggested the first appearance of *A. densa* as an alternative criterion to *Planorotalites palmerae* in the recognition of Zone P9. Stainforth et al. (1975) cited the first appearance of *A. bullbrooki* as an alternative criterion for the recognition of the base of the early middle Eocene *Hantkenina aragonensis* Zone of Bolli (1957a, 1957b, 1966; = P10 of Blow 1969, 1979). The occurrence of both *S. frontosa frontosa* and *A. bullbrooki* suggests an assignment of P10 to P11 (early middle Eocene) for core 15.

The presence of *Pseudohastigerina wilcoxensis* (Cushman and Ponton) (FAD P7; LAD P10[1]; Blow,

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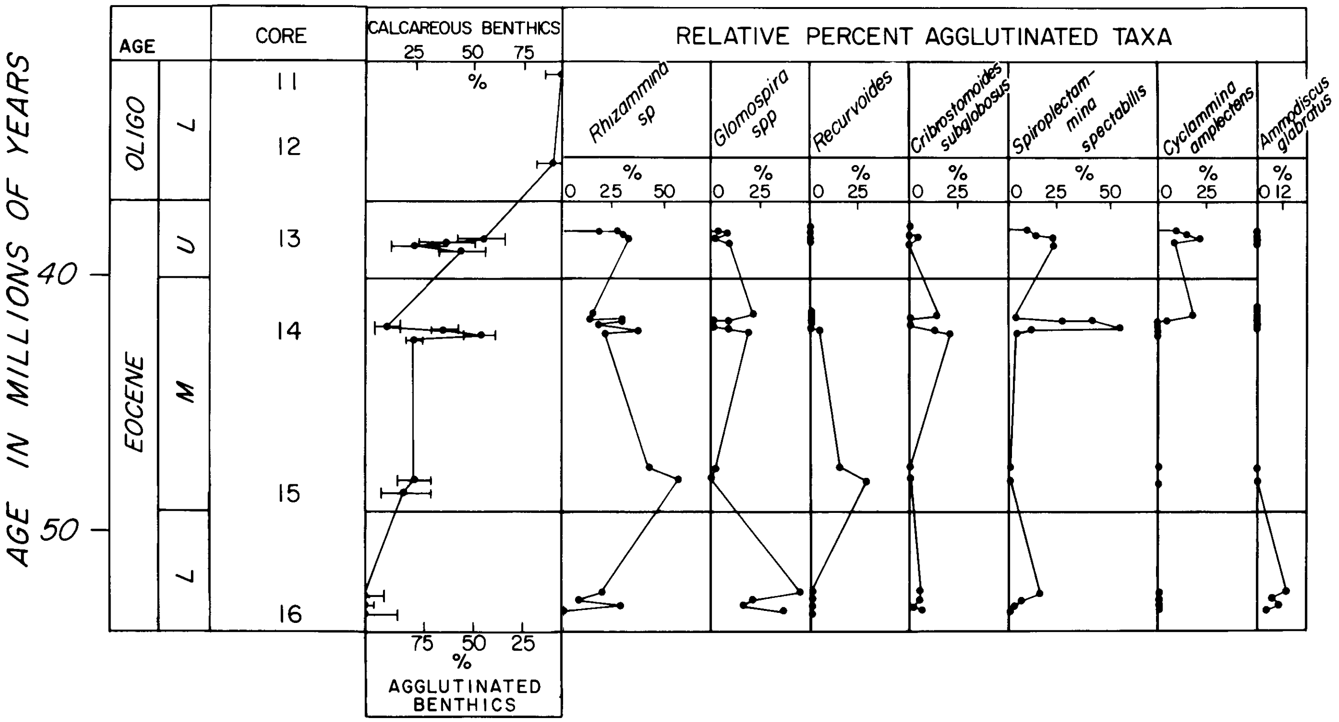
[1] This last appearance datum for *P wilcoxensis* is considered unreliable in light of the discussion below (Core 13)



TEXT-FIGURE 6
Relative percent of agglutinated taxa in DSDP Site 112. *Glomospira* spp. includes relative percent of agglutinated foraminifera of *G. charoides, G. gordialis,* and *G. irregularis.* Error bars on percent calcareous/agglutinated taxa indicate 95% confidence level.

1979), *Catapsydrax turgidus* (Finlay) (FAD P10; LAD P11; Blow, 1979). *Dentoglobigerina galavasi* (Bermudez) (FAD P9; Blow, 1979), and *Acarinina broedermanni* Cushman and Bermudez (FAD P8; LAD P11; Blow, 1979) support the assignment of core 15 to the interval of Zones P10 to P11 of Blow (1979).

Both Perch-Nielsen (1972) and Bukry (1972) placed core 15 in the *Discoaster sublodoensis* (NP14) Zone. Perch-Nielsen followed Martini's (1970) assignment of this zone to the early Eocene (= P9), while Bukry (1972) placed it in the middle Eocene (= P10). Zone NP14 has been correlated with Zone P10 (Bukry and Kennedy, 1969), and nannoplankton typical of this zone have been correlated with the type Lutetian stage (= middle Eocene; Bouche, 1962; Hay et al., 1967). Our study agrees with the correlation of NP14 with P10. Thus, nannofossils and planktonic foraminifera support an early middle Eocene age (P10, NP14) for this core.

Core 14 (revised)

Planktonic foraminifera are rare in core 14; predominantly long-ranging *Subbotina linaperta, Dentoglobigerina galavisi, Turborotalia* sp., and *Catapsydrax* sp. were found. The occurrence of *C. unicavus* Bolli (FAD P14; Stainforth et al., 1975) suggests a late middle to late Eocene age. On the basis of calcareous nannoplankton, core 14 was originally assigned to the late Eocene *Discoaster barbadiensis* (= NP18 to NP20) Zone by Bukry (1972) and to the middle Eocene *D. tani nodifer* s.l. (= NP16) Zone by Perch-Nielsen (1972), although Perch-Nielsen acknowledged possible assignment to the *Reticulofenestra umbilicata* Zone (= NP17 to 18, late Eocene in her study; middle to late Eocene of Martini, 1970). Calcareous nannoplankton from core 14-2 were examined and assigned to the late middle Eocene Zone NP17 of Martini's (1970) zonation (M.-P. Aubry, personal communication, 1980) based upon the simultaneous occurrence of *D. saipanensis* Bramlette and Riedel and *Neococcolithus minutus* Perch-Nielsen and the absence of both *Chiasmolithus oamaruensis* Deflandre and *C. solitus* (Bramlette and Sullivan) (table 2). This latest middle Eocene age assignment agrees well with interpolations of the sedimentation rate. If core 15 is assigned to Zone P10 (48 Ma), a 15 m/my sedimentation rate would place core 14 in the late middle Eocene (40 Ma; text-fig. 5).

TABLE 2
Nannofossil age documentation sample. 12-112-14-2/102–103 cm. Zonal assignment NP17 of Martini's zonation. Age· late middle Eocene

Assemblage
Coccolithus pelagicus (Wallich)
Reticulofenestra umbilica (Levin)
R dictyoda Deflandre
R cf *R. insignata* Roth and Hay
R samodurovi Hay, Mohler and Wade
R cf *R. hillae* Bukry and Perceival
R bisecta Hay, Mohler and Wade
Ericsonia formosa (Kamptner)
Discoaster saipanensis Bramlette and Riedel
D barbadiensis Bramlette and Riedel
Chiasmolithus expansus Bramlette and Sullivan
Zygrhablithus bijugatus (Deflandre)
Markalius inversus Bramlette and Martini
Pemma sp
Neococcolithites minutus Perch-Nielsen
Cyclococcolithus pseudogammation Bouche

Zonal determination based on absence of *Chiasmolithus oamaruensis* Deflandre and *C solitus* (Bramlette and Sullivan) and the simultaneous occurrences of *Discoaster saipanensis* and *Neococcolithites minutus.* (After M -P Aubry, personal communication)

Core 13 (revised)

Subbotina linaperta (LAD P17; Stainforth et al., 1975; Blow, 1979) and *Catapsydrax dissimilis* (Cushman and Bermudez) are the predominant planktonic foraminifera in core 13. The simultaneous occurrence of *C. unicavus* Bolli (FAD P14; Stainforth et al., 1975) and *S. linaperta* suggests latest middle to late Eocene age (P14 to P16). This agrees with the assignment of core 13 to the late Eocene nannoplankton *Isthmolithus recurvus* (= NP19; Perch-Nielsen, 1972) and *Discoaster barbadiensis* (= NP18 to 20; Bukry, 1972) Zones. *Acarinina* cf. *A. rotundimarginata* Subbotina [= *Truncorotaloides collacteus* (Finlay)] also occurs in this core. Berggren (1977) noted that *A. rotundimarginata* ranges into the late Eocene in high latitudes at least as far up as the *I. recurvus* Zone; this supports the assignment of core 13 to this zone.

The presence of *Pseudohastigerina wilcoxensis* in this core is apparently anomalous. Cordey et al. (1970), Stainforth et al. (1975), and Blow (1979) suggested a middle Eocene extinction of this species (middle Eocene; P12, P11, respectively). The specimens we have identified as *P. wilcoxensis* (pl. 3, figs. 11–12) are larger and less compressed than *P. micra* (Cole) and have a much more rounded, inflated axial-apertural aspect to the last chamber which is cited by Blow (1979) as characteristic of *P. wilcoxensis.* Our species can be differentiated from *P. danvillensis* (Howe and Wallace) by lack of an ovoid to pinched axial-apertural aspect (Blow, 1979). Only if we assume that *P. wilcoxensis* extends into the late middle Eocene as inferred by Cordey et al. (1970) and core 13 is reassigned as latest middle Eocene (P14) can this distribution be explained in terms of previous authors' ranges. As this results in a significant disagreement with the nannoplankton age assignment (late Eocene), we accept core 13 as late Eocene (NP19; P15–16) and note the occurrence of *P. wilcoxensis* as anomalous.

Core 12 (revised)

Core 12 is assigned to the early Oligocene *Helicosphaera reticulata* (= NP22) by Bukry (1972) and the *Ericsonia obruta* (= *E. subdisticha* = NP21) Zone by Perch-Nielsen (1972). Planktonic foraminifera include rare *Catapsydrax dissimilis* and *Subbotina* spp. which are not age diagnostic.

The Eocene/Oligocene boundary may lie in a 48 m uncored interval between cores 12 and 13. Assuming a 15 m/my sedimentation rate, the unsampled interval represents 3 my. The assignment of core 13 to NP19 and core 12 to NP21/NP22 requires a separation of 1 to 4 my. This suggests that either continuous sedimentation or only a short hiatus (less than 1 my) occurs across the Eocene/Oligocene in this corehole.

Benthic assemblages of Site 112

We recognize 43 agglutinated species belonging to 23 genera and 8 families in the Eocene section of Site 112 (cores 13 to 16). The generic composition and number of species identified are similar to those reported from other DSDP sites where special studies were made of type A agglutinated assemblages (Sites 367, 368—Krasheninnikov and Pflaumann, 1977; Site 323—Rögl, 1976; Site 283—Webb, 1975; see table 3 of Gradstein and Berggren, 1981). All principal genera (i.e. those noted by more than two authors) noted in previous studies of type A assemblages in the deep sea are present in Site 112. In addition, 20 of the agglutinated species in Site 112 have been noted from the central North Sea, Labrador Shelf, Newfoundland Shelf, and West Greenland (Gradstein and Berggren, 1981). These taxa are indicated with an asterisk in table 3. The presence of the same taxa in Site 112 (paleodepth approximately 3 km; table 1) and the Labrador Shelf and central North Sea (both paleodepths less than 1 km) illustrates the wide bathymetric ranges of these agglutinated taxa.

Several of the stratigraphic ranges (text-fig. 4) of the agglutinated taxa in Site 112 are significantly restricted. The stratigraphic utility of these species is discussed in conjunction with the Canadian margin wells (below).

Abundances of benthic foraminifera range from 0 to 168 individuals/sample in Site 112. In the Eocene section, agglutinated foraminifera comprise from 40 to 100% of the benthic foraminifera, and agglutinated diversity ranges from 5 to 22 species/sample (table 4).

TABLE 3
Abundance of agglutinated species, DSDP Site 112. Number of each specimens/sample is given. Asterisks indicate species noted from Canadian margin or North Sea by Gradstein and Berggren (1981).

SAMPLES SPECIES	13-2 107-117 cm	13-4 86-96 cm	13-5 146-149 cm	13-cc	14-2 50-54 cm	14-2 102-113 cm	14-2 144-146 cm	14-3 141-144 cm	14-5 144-147 cm	14-cc	15-1 12-25 cm	15-cc	16-1 TOP	16-1 46-49 cm	16-1 78-84 cm	16-1 107-111 cm
? Bathysiphon sp.	2	1	1	0	0	6	6	6	0	3	3	0	0	0	1	0
Saccammina complanata	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1
S. difflugiformis	0	0	0	0	0	0	0	0	0	1	0	0	1	4	2	0
S. placenta *	1	1	0	0	0	2	0	0	0	0	0	0	0	0	0	0
Ammodiscus cretaceus *	8	1	0	1	1	7	0	1	1	4	1	2	1	0	0	0
A. glabratus	0	0	0	0	0	0	0	0	0	0	0	0	4	11	11	1
A. rugosus	1	1	2	3	1	4	0	0	0	2	0	0	0	0	0	0
Glomospira charoides *	0	2	0	1	6	0	0	0	1	12	0	0	6	9	1	2
G. gordialis *	1	0	1	2	6	1	0	1	4	8	0	0	5	10	2	2
G. irregularis *	0	0	0	1	5	7	0	0	0	4	1	0	4	12	14	3
G. serpens	1	1	0	1	0	0	0	0	0	3	0	0	0	0	0	0
Glomospirella sp.	2	0	0	1	0	0	0	0	3	3	1	0	0	13	0	0
Reophax pilulifera *	1	0	0	0	0	6	0	0	0	0	0	0	0	1	1	0
Hormosina globulifera	2	0	0	0	0	7	0	1	0	0	0	0	0	0	0	0
H. ovulum *	0	0	0	0	4	0	0	0	0	0	0	0	0	13	0	0
H. sp. 3	2	0	0	2	1	3	0	4	0	0	3	0	1	2	4	2
"Hormosina" sp. 4	0	0	0	0	0	0	0	0	0	3	3	0	0	3	1	0
Spirosigmoilinella compressa *	3	0	0	0	0	2	0	0	0	0	0	0	0	0	0	0
Haplophragmoides eggeri *	0	0	0	0	0	0	0	0	0	0	1	0	0	23	2	1
H. walteri *	0	0	0	0	0	0	0	0	2	1	0	0	0	3	3	0
"Haplophragmoides" sp. 6	0	0	0	0	0	0	0	0	0	0	0	0	0	5	3	2
Cribrostomoides subglobosus *	0	1	0	0	10	1	0	1	7	24	0	0	1	4	1	1
Paratrochamminoides *	1	2	1	3	4	0	0	0	1	7	2	0	0	3	3	0
Reophax walteri *	0	0	0	0	0	0	0	0	0	2	8	5	0	0	0	0
Trochamminoides sp.	0	1	0	0	0	1	0	0	1	2	0	0	0	0	0	0
Cyclammina amplectens *	3	3	5	3	12	3	0	0	0	0	0	0	0	0	0	0
C. trullissata	0	0	0	0	0	0	0	0	0	6	0	1	0	0	0	0
Trochammina globigeriniformis *	0	0	0	0	0	0	0	0	0	0	0	0	0	4	13	1
Trochammina sp.	0	0	0	0	0	1	0	0	1	4	0	0	0	0	0	0
Cystammina globigerinaeformis *	0	0	0	0	2	0	0	0	5	4	0	0	0	11	6	0
Spiroplectammina cubensis	0	0	0	0	0	3	0	1	1	0	1	0	0	0	0	0
S. spectabilis *	3	3	5	9	2	22	9	40	6	2	0	0	5	7	2	0
Dorothia sp.	0	0	0	0	0	2	0	0	2	2	0	0	0	0	0	0
Bathysiphon sp. 2	0	0	0	0	0	0	0	0	0	0	0	0	0	1	1	3
Rhizammina sp.	7	7	8	15	13	15	6	12	24	27	24	11	6	12	32	0
Labrospira pacifica *	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	0
Hyperammina sp.	0	0	1	0	0	5	0	0	0	0	5	0	0	1	0	0
Martinottiella sp.	0	1	0	0	9	0	0	2	0	0	0	0	0	2	0	0
Reophax nodulosus	0	1	0	0	0	1	0	0	0	0	2	0	0	0	0	0
Lituotuba sp. *	0	0	0	0	1	0	0	0	2	0	1	0	0	0	0	0
? Spiroplectinata sp.	0	0	0	0	0	0	0	1	0	0	0	0	0	0	0	0
? Karreriella sp.	0	0	0	0	0	0	0	0	0	1	0	0	0	0	0	0
Ammolagena clavata	0	0	0	0	0	0	0	1	0	1	0	0	0	0	0	0

TABLE 4
Benthic foraminifera from DSDP Site 112

Core	Depth (cm)	% Calcareous benthic foraminifera	# Specimens benthic foraminifera	# Species agglutinated benthic foraminifera
11-1	145–147 5	100	35	0
12-1	100–102	94	71	2
12-CC	—	barren		
13-2	107–117	60	94	15
13-4	86–96	40	42	13
13-5	144–149	25	36	11
13-CC	—	49	91	15
14-2	50–54	10	86	17
14-2	102–113	39	167	22
14-2	144–146	40	37	4
14-3	141–144	60	183	11
14-5	144–149	27	94	14
14-CC	—	24	168	22
15-1	12–25	25	88	16
15-6	130–135	barren		
15-CC	—	16	32	5
16-1	top	0	34	10
16-1	46–49	0	154	22
16-1	78–84	0	108	22
16-1	107–111	0	20	12

Although 43 agglutinated species are found in Site 112, only 9 species constitute more than 10% of agglutinated foraminifera in two or more samples. The relative abundances of these nine species are used to identify four agglutinated assemblages in the Eocene section: the *Glomospira, Cribrostomoides/Recurvoides, Spiroplectammina spectabilis* (Grzybowski), and *Cyclammina amplectens* assemblages (text-fig. 6). *Rhizammina* sp. is abundant (approximately 25% of agglutinated foraminifera) in all assemblages.

The *Glomospira* assemblage is found in core 16 and is composed of highest abundances of *G. charoides* (Jones and Parker), *G. gordialis* (Jones and Parker), *G. irregularis* (Grzybowski), and *Ammodiscus glabratus* Cushman and Jarvis. This assemblage is characterized by exclusively agglutinated foraminifera (table 4; text-fig. 5) and is associated with the baked red clay lithologic unit (unit 5; text-fig. 5). *Haplophragmoides eggeri* Cushman, *H.* sp., and *Trochammina globigeriniformis* (Parker and Jones) are restricted to this early Eocene assemblage.

The following assemblages occur within the silty clays and nannofossil marls of lithologic unit 4.

The middle Eocene *Cribrostomoides/Recurvoides* assemblage occurs in core 15 and core 14 (14-CC; 14-5). It is distinguished from the overlying assemblage by the low percentages of *Spiroplectammina spectabilis* and the high percentages of *"Cribrostomoides" subglobosus* (Sars) and *Recurvoides* ex. gr. *walteri* (Grzybowski). It is associated with approximately 25% calcareous benthic foraminifera.

The *S. spectabilis* assemblage (14-3; 14-2, 144–146 cm; 14-2, 102–113 cm) is composed predominantly of the nominate species (25–50% of the agglutinated foraminifera) and high relative abundances of calcareous benthic foraminifera (39–60% of total benthic foraminifera).

The sample with highest abundance of *S. spectabilis* and calcareous foraminifera (14-2, 140–142 cm) is described in the initial core description as a light gray indurated layer distinct from overlying and underlying lithologies (Laughton, Berggren et al., 1972). We noted abundant sand-sized planktonic foraminifera in this sample which are not found in any other Eocene sample. Since *S. spectabilis* is thought to have calcareous cement, this assemblage may reflect a paleoenvironment favorable for the deposition or preservation of calcium carbonate. Unfortunately, no carbonate percentage data are available for this interval.

The middle to late Eocene assemblage found in core 14 (14-2, 50–54 cm) and core 13 contains abundant *Cyclammina amplectens* and variable percentages of calcareous benthic species (10–60%; table 4). These variable percentages of calcareous species are not correlative with the fluctuations of calcium carbonate (24% ± 3%) in the bulk sediment.

As was first noted by Berggren (1972), the Eocene agglutinated assemblages are replaced in cores 12 and above by an Oligocene calcareous benthic assemblage. This exit is shown in text-fig. 4, where the diverse (15 species/sample) late Eocene *C. amplectens* assemblage is replaced by an early Oligocene benthic foraminiferal assemblage with sparse agglutinated species (0–2 species/sample). Table 4 shows moderate percentages of calcareous taxa in core 13 (25–60%) which increase to nearly 100% in cores 11 and 12. This faunal turnover occurred between the upper Eocene at 444 m subbottom and the lower Oligocene cored at 385 m subbottom (discontinuous coring). Assuming a 15 m/my sedimentation rate, this interval represents approximately 4 my.

AGGLUTINATED ASSEMBLAGES OF THE EASTERN CANADIAN MARGIN AND NORTH SEA

Diverse agglutinated foraminiferal assemblages resembling flysch-type and type A assemblages occur in Maestrichtian through Eocene beds on the Labrador and Newfoundland shelves (text-figs. 3, 8). Similar assemblages have been noted from the late Paleocene through Eocene strata in the central North Sea. Lithologies associated with these assemblages are predominantly mudstones and shales. Localized late Paleocene to early Eocene intercalations termed Gudrid sands occur on the Labrador shelf (Umpleby, 1979); stratigraphic equivalents occurring in the central North

Sea form oil and gas producing horizons in the Forties, Montrose, Frigg, Lomond, Cod, and other oil and gas fields (e.g. Ziegler, 1980). These sandy layers are largely devoid of indigenous microfauna. Both in the North Sea and on the Canadian margin the agglutinated fauna largely disappear at the end of the Eocene.

The stratigraphic ranges of the agglutinated taxa are calibrated to foraminiferal zonations developed for the Canadian margin (16 wells) and central North Sea (15 wells). The central North Sea correlation is based on last appearance levels of planktonic and benthic taxa. However, in the Paleocene section, such exits are rare (see Berggren and Gradstein, 1981); therefore, the dinoflagellate biostratigraphy of Ioakim (1979) was also used. On the Canadian margin, we used a probabilistic approach to Cenozoic foraminiferal stratigraphy. Exits of over 150 benthic and planktonic taxa were used in conjunction with a locally derived planktonic zonation (see Gradstein and Agterberg, in press). Rather than emphasizing individual wells, these zonations aim at a broad regional application. As a result, stratigraphic resolution is relatively conservative with an eight- to tenfold subdivision of the Cenozoic beds.

Approximately 60 agglutinated species representing 32 genera have been recognized in wells from the Canadian margin and North Sea. Most of these taxa have been described by Gradstein and Berggren (1981). Few taxa are endemic within these two regions.

In the Maestrichtian through Paleocene shales of Labrador wells, *Bathysiphon, Saccammina, Glomospira, Karreriella, Trochammina, Haplophragmoides,* and *Recurvoides* are dominant genera. In the Maestrichtian of the Indian Harbour well (text-figs. 3, 8), *Reophax* and *Uvigerinammina* are also important. Eocene mudstones both on the Canadian margin and in the North Sea contain frequent *Bathysiphon, Ammodiscus, Haplophragmoides, Trochammina, Cribrostomoides,* and *Karreriella. Cyclammina* spp. are locally abundant (e.g. Karlsefni H-13); species include *Cyclammina placenta* (Reuss), *C. amplectens* Grzybowski, *C. rotundidorsata* (Hantken), and forms classified as *C. placenta* (Reuss) or *C. cancellata* Brady.

In the Labrador Shelf shales (Indian Harbour M-52 well) agglutinated foraminifera are found associated with *Globotruncanella mayaroensis* (Bolli) (also reported by Dufaur et al., 1976, from the adjacent Gudrid U-55 well) and a rich assemblage of *G. havanensis* (Voorwijk), *Globigerinelloides messinae* Bronnimann, *Rugoglobigerina* aff. *rotundata* Bronnimann, and *Bolivinoides draco* (Marsson). These species indicate a late Maestrichtian age. The agglutinated taxa *Uvigerinammina jankoi* Majzon, *Dorothia oxycona* (Reuss), and *Arenobulimina dorbignyi* (Reuss) are restricted to these Maestrichtian shales. Of these species *U. jankoi* has been reported from many Upper Cretaceous flysch deposits in Europe and from the Upper Cretaceous clays in the Atlantic and Indian oceans (Geroch, 1959; Krasheninnikov, 1974; Krasheninnikov and Pflaumann, 1977). *Dorothia oxycona* (Reuss) is a cosmopolitan Cretaceous form; it locally extends in Paleocene beds (Aubert and Berggren, 1976). *Arenobulimina dorbignyi* (Reuss) was reported by Hanzlikova (1972) from Turonian-Campanian strata in Europe.

In the North Sea, *Rzehakina epigona* (Rzehak) is restricted to Upper Cretaceous deposits. On the Canadian margin *R. epigona* is found more frequently in Upper Cretaceous shales, but also consistently occurs in Paleocene beds. Isolated individuals occur as high as Zone P9 (= *Acarinina densa* Zone; late early Eocene).

A distinctive species of *Ammobaculites* with 10 or more chambers in the last whorl and undulating, dark sutures is common in the Labrador and Newfoundland Shelf wells. It is rare in the North Sea. It ranges from Maestrichtian through Eocene beds and is provisionally assigned to *A.* aff. *A. polythalamus* Loeblich.

Three distinct taxa, important in the North Sea, are absent from the Canadian margin. *Trochammina subvesicularis* Homola and Hanzlikova and a form provisionally assigned to *T.* aff. *T. albertensis* Wickenden are both homeomorphic with "conical" planktonic foraminifera. Occurring in North Sea wells in beds overlying Danian carbonates and underlying *Cyclammina* bearing (Eocene) strata, they are probably Paleocene in age. *Thurammina* sp. has been reported by Bettenstaedt et al. (1962) from lower middle Eocene beds; it has a similar range in the central North Sea.

The absence of pre-Maestrichtian and post-Eocene to Miocene agglutinated foraminiferal assemblages in the Canadian margin and North Sea probably results in the limited stratigraphic ranges of some of the taxa mentioned. These truncated ranges may result from paleoenvironmental changes, and therefore may be useful only in local zonations.

COMPARISON OF SITE 112 WITH EASTERN CANADIAN MARGIN

Calibration of the distribution of benthic agglutinated taxa with the zonations discussed above allowed us to compare their ranges from the deep Labrador Sea (Site 112) with the Canadian margin and previously reported ranges from the North Sea, flysch deposits of Europe, and the deep sea. The long stratigraphic ranges of agglutinated species poses one of the major problems in the use of these taxa in biostratigraphy. For example, 12 agglutinated species range throughout the Eocene in Site 112. Other biostratigraphic

TABLE 5
Stratigraphic ranges of agglutinated taxa

Species	Age
Uvigerinammina jankoi Majzon	Maestrichtian shales (C M) Late Cretaceous (E F , D S)
Dorothia oxycona (Reuss)	Maestrichtian shales (C M) Paleocene (Tunisia)
Arenobulimina dorbignyi (Reuss)	Maestrichtian shales (C M.) Turonian-Campanian (E.F)
Rzehakina epigona (Rzehak)	Late Cretaceous (N S) acme Late Cretaceous (C M) Cretaceous-early Eocene (C M) (rare in Eocene)
Cystammina globigerinaeformis (Krasheninnikov)	Late Cretaceous (D S) Maestrichtian to Eocene (C M , N S.) early to middle Eocene (112)
Spiroplectammina spectabilis (Grzybowski)	Maestrichtian to middle Eocene rare above Paleocene (C M) late Paleocene to early Eocene (N S , E F) Eocene (112)
Cyclammina spp	post-Paleocene (C M , 112, D.S) latest Paleocene and younger (N S)
C amplectens (Grzybowski)	Eocene (C M) early to middle Eocene and possibly late Eocene (N S) middle to late Eocene (112)
C. placenta (Reuss) and *C rotundidorsata* (Hantken)	Eocene (acme in middle Eocene, E.F.) to Oligocene, possibly Miocene (C M , N S)
Spirosigmoilinella compressa Matsunaga	middle Eocene to early Miocene (N S) middle to late Eocene (112) Miocene (Japan)
Ammobaculites aff *polythalamus* Loeblich	Maestrichtian-Eocene (C M , N S)
Trochammina subvesicularis Homola and Hanzlikova and *T.* aff. *albertensis* Wickenden	late Paleocene (N S)
Thurammina sp	early middle Eocene (N S , E F)

C M = Canadian Margin, N.S. = North Sea, E F = European flysch basins, D S = deep sea, 112 = Site 112

problems arise from agglutinated taxa that have diachronous ranges. This is illustrated by the genus *Glomospira,* which is represented in our samples by *G. charoides* (Jones and Parker), *G. gordialis* (Jones and Parker), and *G. irregularis* (Grzybowski). On the Canadian margin these species are fairly common to abundant in Maestrichtian through Paleocene beds; they disappear in the early middle Eocene. These same species are found throughout the Eocene in Site 112, exit in the middle Micene in the North Sea (text-fig. 7), but are found in the present-day Atlantic (Brady, 1884), Pacific (Brady, 1884; Saidova, 1961), and Gulf of Mexico (Phleger and Parker, 1951; Pflum and Frerichs, 1976). Despite these problems, we have been able to use many agglutinated taxa for regional zonations (see text-figs. 4, 7 and table 5 for summary of ranges).

The taxonomic position and stratigraphic range of the genus *Praecystammina* is unclear. Krasheninnikov (1973) used the new taxon *Cystammina globigerinaeformis* (Krasheninnikov) as a Late Cretaceous marker in abyssal Pacific and Indian Ocean deposits. Specimens resembling this form occur in the Maestrichtian of the Labrador Shelf (Indian Harbour M-52 well) and Paleocene to Eocene of the central North Sea. In Site 112, *C. globigerinaeformis* exits at the end of the middle Eocene. Further studies are needed to determine to what extent *Praecystammina* really differs from *Cystammina* and the chronostratigraphic range of *C. globigerinaeformis.*

On the Canadian margin, *Spiroplectammina spectabilis* (Grzybowski), *S. navarroana* Cushman, and *S. dentata* (Alth) have been found from Maestrichtian to middle Eocene beds. *Spiroplectammina spectabilis* is rare above the Paleocene. In the central North Sea *S. spectabilis* and *S. navarroana* were observed in beds overlying Danian carbonates and locally range into or just above sediments with *Subbotina patagonica* (Todd and Knicker) and rare *Acarinina* aff. *pentacamerata* (early Eocene). *Spiroplectammina spectabilis* has been reported from West Germany from upper Paleocene to lower Eocene beds (Bettenstaedt et al., 1962). In Site 112 *S. spectabilis* is found throughout the Eocene, suggesting that it has a longer range in the deep sea.

In the regions under discussion, *Cyclammina* is thought to be restricted to post-Paleocene beds, although in the North Sea some relatively small specimens occur slightly below the tuff marker at the Paleocene-Eocene boundary. Otherwise, in the North Sea *C. amplectens* occurs in strata with *Acarinina pentacamerata* (Zones P8 to P12; lower to middle Eocene; Stainforth et al., 1975) and *Spiroplectammina spectabilis.* Although it may range higher, it is mostly restricted to the Eocene. On the Canadian margin *C. amplectens* occurs throughout the Eocene (associated with *Subbotina patagonica, Acarinina densa,* and *Turborotalia pomeroli*). In Site 112 it is restricted to the middle to upper Eocene section. In Central Europe (Carpathians) *C. amplectens* has been reported from the Eocene; it occurs in greatest abundance in middle Eocene strata (Geroch, 1960).

In the central North Sea and on the Canadian margin both *Cyclammina placenta* (Reuss) and *C. rotundidorsata* (Hantken) are longer ranging than *C. amplectens*; these species extend into beds with *Turrilina alsatica* (Andreae) (Oligocene) and *Globigerinoides/Globorotalia* (Miocene). These species are absent from Site 112.

In the North Sea, *Spirosigmoilinella compressa* Matsunaga extends from the upper part of the range of *Cyclammina amplectens* into beds with *Globigerinoides* and *Globorotalia acrostoma*; the chronostratigraphic range is middle Eocene to early Miocene. This species occurs in the *Areosphaeridium diktyoplokus* Zone of Ioakim (1979) of Bartonian age (middle Eocene). It also occurs in the middle upper Eocene in Site 112; it was not found in the Canadian margin wells. Specimens of *S. compressa* Matsunaga differ from *Rzehakina epigona* by having an evolute rather than involute test, a variable but often slenderer width of the test, and an early stage which coils in a different plane from the last few chambers. *Rzehakina epigona,* in contrast, is planispiral. *S. compressa* was originally described from the Miocene of Japan. *Hormosina ovulum* (Grzybowski) disappears in the lower Eocene in the Carpathians, Caucasus, Eastern Alps (Geroch, 1959), and Site 112.

Of the important indicator taxa noted in the Canadian margin and North Sea, *Trochammina subvesicularis, T.* aff. *T. albertensis, Thurammina* sp., *Ammobaculites* aff. *A. polythalamus, Uvigerinammina jankoi, Dorothia oxycona, Arenobulimina dorbignyi,* and *Rzehakina epigona* were not found in Site 112.

In addition to the replacement of the late Eocene agglutinated assemblage by an early Oligocene calcareous assemblage in Site 112, agglutinated assemblages also exit across the Eocene/Oligocene boundary in many of the Canadian margin wells (text-fig. 8). In the North Sea, Gradstein and Berggren (1981) noted the disappearance of agglutinated foraminifera in most of the wells studied at the end of the Eocene. Only in a few wells in the center of the basin do agglutinated taxa extend into stratigraphically younger beds. Thus, a major faunal turnover occurs in the late Eocene to early Oligocene of the North Sea, Labrador Shelf, Newfoundland Shelf, and the Labrador Sea.

PALEOENVIRONMENT OF AGGLUTINATED FORAMINIFERA

Agglutinated foraminifera in the Labrador Shelf and North Sea are associated with high sedimentation rates, high organic content, and deepest paleowater depths. The best preserved and highest diversity agglutinated assemblages are found in shales high in organic matter in wells in the depositional center of the basins. In general, these centrally situated wells have relatively high sedimentation rates and calcareous benthic foraminifera which suggest deeper water conditions than in updip wells that have a lower diversity agglutinated assemblage dominated by *Bathysiphon* and *Trochammina.*

Paleoenvironmental factors controlling the distribution of flysch-type agglutinated assemblages in the Labrador and East Newfoundland shelves and the North Sea may be related to the model suggested by Ksiazkiewicz (1961, 1975) and Moorkens (1976), which invokes rapid deposition of organic-rich, fine-grained clastics with restricted bottom water circulaton (see discussion in Gradstein and Berggren, 1981). Such conditions developed in the Maestrichtian through Eocene of the Labrador Shelf and in the late Paleocene to Eocene of the central North Sea. Differential subsidence, leading to paleogeographically separated and paleoceanographically restricted basins, coincided with rapid deltaic outbuilding into the deeper, more centrally situated parts of the basins (Gradstein and Berggren, 1981). Nevertheless, evidence from Site 112 suggests that, at least in the deep sea, high organic matter is not a prerequisite for the development of predominantly agglutinated foraminiferal assemblages.

Few cores are available from the North Sea and Canadian margin; therefore, the cored Eocene-lower Oligocene section of Site 112 was used to make a detailed comparison of lithologic characteristics with the distribution of agglutinated benthic foraminifera.

The change from agglutinated to calcareous benthic foraminiferal assemblages occurs in Site 112 within lithologic unit 4. Grain size (silty clays), percent calcium carbonate (approximately 30%), and percent organic carbon (0.3%) do not change in the upper Eocene to lower Oligocene section (text-fig. 5). The constancy of these lithologic parameters across this faunal change shows that the exit of the agglutinated assemblages was not caused by a change in substrate.

The presence and relatively constant values of abundant calcium carbonate throughout the middle Eocene to lower Oligocene section does not support the suggestion by Hesse and Butt (1976) that the dominant factor controlling flysch-type (type A) agglutinated assemblages is deposition below the CCD. With the exception of core 16, which is barren of calcium carbonate and calcareous benthic foraminifera, and possibly the *Spiroplectammina spectabilis* assemblage, there is no clear correlation between percent carbonate and agglutinated and/or calcareous benthic foraminiferal abundance (text-fig. 5). Despite this, carbonate availability may play a role in the development of predominantly agglutinated foraminiferal assemblages, for the exit in Site 112 correlates with a global drop in the CCD (Heath, 1969; Berger, 1973; van Andel et al., 1975; Kennett and Shackleton, 1976). Nevertheless, factors which caused this drop (see below) may also be related to the faunal change.

Due to diagenetic reduction of the carbon content of marine sediments (Heath et al., 1977), the absolute carbon values observed in Site 112 may reflect only a portion of the carbon present at the time of deposition. Nevertheless, the constant values of organic carbon associated with the faunal turnover (text-fig. 5) do not support the idea that organic carbon is the dominant control on flysch-type agglutinated foraminiferal assemblages. Although agglutinated foraminifera may be favored in areas of high organic matter, high organic carbon is not a prerequisite for predominantly agglutinated foraminiferal assemblages. This is shown by the presence of exclusively agglutinated foraminiferal assemblages in the present-day deep North Pacific by Saidova (1960) and Bernstein et al. (1978), where carbon values are invariably low (Heath et al., 1977).

The exit of the agglutinated foraminifera and their replacement by a deep-water calcareous assemblage may have resulted from changes in hydrographic properties. The Tertiary climatic record shows that a major period of cooling occurred in the middle Eocene to early Oligocene, and that the greatest temperature drop occurred close to the Eocene/Oligocene boundary (Dorman, 1966; Devereaux, 1967; Savin et al., 1975; Shackleton and Kennett, 1975; Wolfe, 1978; Vergnaud-Grazzini et al., 1979). The development of the psychrosphere (the deep, cold water layer of the oceans; Benson, 1975), and the formation of cold bottom water production in the southern ocean (Kennett and Shackleton, 1976) also began near the end of Eocene time. This initiation of cold bottom water resulted in an inferred 4 to 5°C temperature decrease in bottom water temperatures in the southern ocean (Site 277) in the earliest Oligocene time (Kennett and Shackleton, 1976).

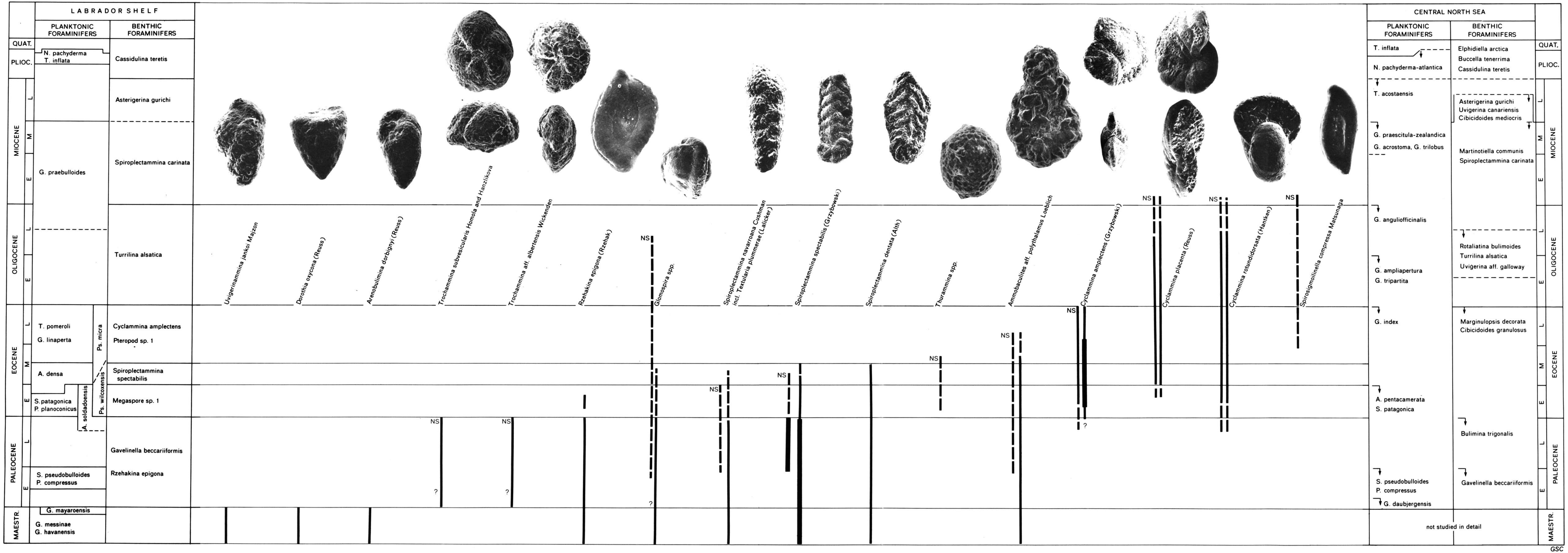
In addition to lowering the temperature and supplying more oxygen to bottom waters, a presumed increase in oceanic turnover in the late Eocene to early Oligocene (Kennett and Shackleton, 1976) may have decreased CO_2 and thereby increased pH. This turnover contributed to the lowering of the CCD which is reflected in the deep-sea stratigraphic record as a well-documented increase in preservation of calcium carbonate (Heath, 1969; Berger, 1973; van Andel et al., 1975; Kennett and Shackleton, 1976).

Such changes in hydrography could have eliminated the indigenous agglutinated benthic foraminiferal assemblages and favored the migration of calcareous benthic assemblages to Site 112. Moorkens (1976) and Saidova (1960, 1965) both explain the preponderance of agglutinated foraminifera in abyssal depths by the tolerance of these foraminifera to low oxygen and pH. The modern deep North Pacific, with its low oxygen, high CO_2, and low pH (Sverdrup et al., 1942), and exclusively agglutinated benthic foraminiferal assemblages (Saidova, 1961, 1965, 1970, 1976; Bernstein et al., 1978) may be a modern analog to the Eocene deep Labrador Sea, although carbonate sediments are absent from the former but are present in the Eocene Labrador Sea.

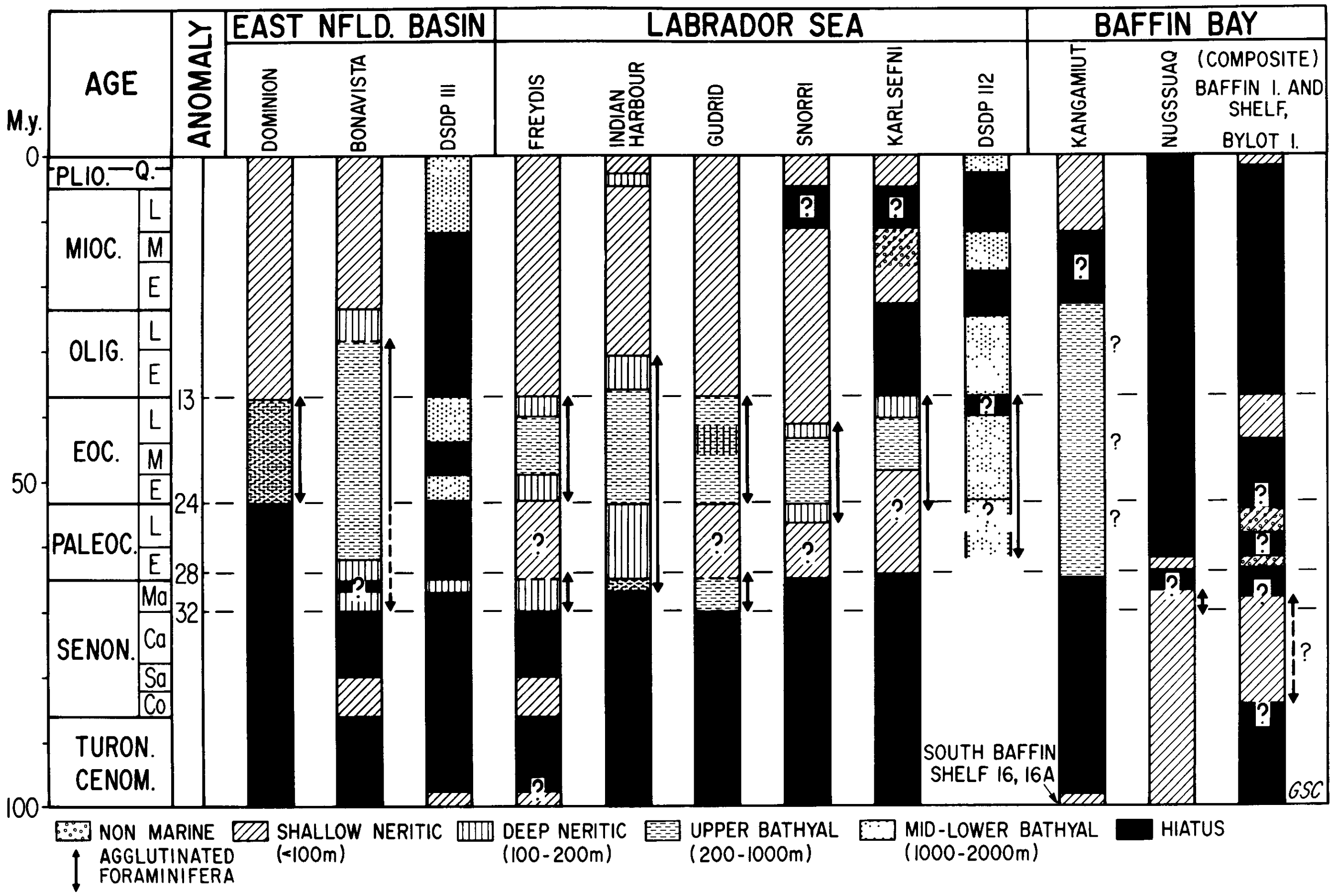
Although the faunal turnover in Site 112 approximately coincides with the formation of the psychrosphere, increased oceanic turnover, and the initiation of southern sources of bottom waters, it is not clear that the only late Eocene to early Oligocene source of low-temperature, high-oxygen bottom water in the Labrador Sea is from the Antarctic region. Despite the contention of some authors (Shor and Poore, 1979; Blanc et al., 1980; Schnitker, 1980a, 1980b) that bottom water from northern sources (i.e. the Arctic, Norwegian-Greenland Sea, or other northern North Atlantic sources including the Labrador Sea) did not form until the early to middle Miocene, seismic evidence from the northern North Atlantic suggests the presence of a northern cold-water source as early as late Eocene.

Johnson and Schneider (1969), Jones et al. (1970), and Ruddiman (1972) have clearly demonstrated that bottom water circulation (predominantly North Atlantic Deep Water, NADW) profoundly influences sedimentation patterns in the northern North Atlantic (north of 45°N). The earliest effect of this deep circulation is in the form of sediment drifts overlying a subhorizontal sedimentary reflector, R4 (Roberts, 1975; = reflector R of Jones et al., 1970). In this region, reflector R4 has been firmly established as pre-late Oligocene by Jones et al. (1970). It has been suggested that reflector R4 is latest Eocene age in the Rockall Plateau and Rockall Trough (Roberts, 1975) and Iceland Basin (Ruddiman, 1972).

In the Labrador Sea near Site 112, reflector R4 lies at 0.41 seconds two-way travel time (Laughton, Berggren et al., 1972). A hard layer encountered in Site 112 in core 10 (upper lower Oligocene) at 315 m subbottom was thought to correspond to this reflector (Laughton, Berggren et al., 1972); however, this correlation is not in agreement with evidence suggested by sediment velocity and changes in grain size. Placing reflector R4 at a depth of 315 m, results in a mean seismic velocity of 1536 m/sec in the drifted sediments. This velocity is considerably lower than that expected for sediments of this thickness. In addition, a grain size change might be expected in the change from pelagic clay to current-controlled deposition. A grain size increase from clays to silty clays and clayey silts occurs between cores 11 and 12 in Site 112 (text-fig. 5), suggesting that a major change in depositional regime occurred in the lower Oligocene section between 333 and 383 m subbottom. Assuming reflector



FORAMINIFERAL ZONATIONS FOR THE LABRADOR SHELF AND CENTRAL NORTH SEA AND STRATIGRAPHICAL RANGES OF SELECTED AGGLUTINATED TAXA



TEXT-FIGURE 8
Late Mesozoic-Cenozoic depositional history of Canadian margin and DSDP Sites 111 and 112.

R4 correlates with this change in grain size (text-fig. 5) in Site 112 we obtain a 1625–1875 m/sec mean seismic velocity which is more realistic for sediments of this thickness from this region (cf. typical mean velocities for the upper 300–350 m of sediment from Leg 48 determined from sonic logs of approximately 1750 m/sec; Montadert, Roberts et al., 1979). This correlation of R4 in the Labrador Sea as a lower lower Oligocene reflector separating pelagic from current-controlled deposition suggests that the formation of northern sources of bottom water influenced the Labrador Sea by early Oligocene time.

Within the biostratigraphic limits of error, reflector R4 has been identified as a synchronous reflector in other parts of the northern North Atlantic. The correlation of reflector R4 in this region as latest Eocene to earliest Oligocene (Ruddiman, 1972; Roberts, 1975) suggests that the change in depositional regime associated with reflector R4 occurred throughout the northern North Atlantic at this time (for further discussion, see Miller and Tucholke, in press).

Hiatuses are widespread in the late Eocene to Oligocene sections of the World Ocean, particularly the North Atlantic (Moore et al., 1978). Of the DSDP sites drilled in the northern North Atlantic (Legs 12, 38, 48, 49) only Sites 112 and 116 may be inferred to have continuous deposition across the Eocene/Oligocene boundary (cf. fig. 3 of Shor and Poore, 1979; figs. 3–6 of Miller and Tucholke, in press). Although factors other than bottom water flow influence the development of deep-sea hiatuses (Moore et al., 1978 and references therein), the development of a regional late Eocene to Oligocene hiatus in the northern North Atlantic supports the idea that bottom water formation began in this region at this time.

The change in depositional regime and the inferred initiation of bottom water in the northern North Atlantic approximately coincides with the disappearance of agglutinated assemblages from Site 112. Still, we cannot firmly extrapolate the faunal change to the entire deep southern Labrador Sea, and therefore the possibility exists that the faunal turnover resulted from a local phenomenon restricted to the immediate environs of Site 112.

The early Oligocene change in depositional regime (between cores 11 and 12) in Site 112 slightly postdates the exit of the agglutinated foraminiferal assemblages (between cores 12 and 13) by 1–7 my (text-fig. 5). This time lag may be due to the gradual intensification of currents during the middle Eocene to Oligocene due to increased climatic cooling. As bottom water began to form in the middle to late Eocene, hydrographic properties began to change, resulting in the replacement of agglutinated foraminifera in Site 112. By the early Oligocene, bottom currents were sufficiently competent to transport silty sediments into the Labrador Sea. There is, in fact, a slight suggestion that percent calcareous taxa gradually increased and replaced the agglutinated foraminifera in Site 112 in the middle Eocene to Oligocene (text-fig. 5). This scenario would agree with the gradual development of the psychrosphere proposed by Corliss (1979). Unfortunately, due to coring gaps and the uncertainties attached to the relative percentages of calcareous species (text-fig. 6) our data are insuffcent to convincingly show a gradual change in benthic foraminifera in Site 112.

CONCLUSIONS

(1) Late Cretaceous to Paleogene predominantly agglutinated benthic foraminifera reported from the deep sea (= flysch-type, type A assemblages) are taxonomically similar to assemblages noted on the Canadian margin, in the North Sea, and in European flysch basins. These assemblages range from bathyal (hundreds of meters) to abyssal (approximately 4.5 km) paleodepths. Smooth-walled, smaller-sized (< 150 μm) Cretaceous assemblages (= type B assemblages) noted only from abyssal paleodepths are easily differentiated from these type A assemblages.

(2) Comparison of ranges of type A assemblages in Site 112, the Canadian margin, and the North Sea shows that some taxa are biostratigraphically useful (text-fig. 4; table 5).

(3) Although high organic content, deposition below the CCD, and inferred poor circulation are often associated with many type A assemblages, these properties are not necessary for faunal development. Rather, we infer that, at least in the deep sea, hydrographic properties (low oxygen, low pH, high CO_2, and thus more corrosive waters) are critical to the development of the predominantly agglutinated assemblages. Since these properties affect carbonate availability, the development of agglutinated assemblages may co-vary with percent calcium carbonate, although this is not the case in Site 112. In addition, similar conditions may develop in reducing substrates associated with high organic matter and poor circulation. Predominantly agglutinated assemblages are probably favored in such substrates and/or in areas with the given hydrographic properties.

(4) The late Eocene exit of agglutinated foraminifera from Site 112, the Canadian margin, and the North Sea coincides with changes in hydrographic properties associated with the development of a more vigorous abyssal circulation and the psychrosphere. We speculate that the late Eocene to early Oligocene initiation of bottom water formation in the northern North Atlantic and/or Norwegian-Greenland Sea, inferred from sediment distribution patterns, caused the replacement of the agglutinated assemblage in Site 112 by a calcareous assemblage, but that local tectonic and sedimentologic changes may have been important in the elimination of agglutinated assemblages in the Canadian margin and North Sea.

SYSTEMATIC PALEONTOLOGY

Agglutinated foraminifera Site 112

Rare = less than 5 individuals in all samples
Common = 5 to 25 individuals/sample
Abundant = greater than 25 individuals/sample

Order FORAMINIFERIDA Eichwald, 1830
Suborder TEXTULARIINA Delage and Hérouard, 1896
Superfamily AMMODISCACEA Reuss, 1862
Family ASTRORHIZIDAE Brady, 1881
Subfamily RHIZAMMININAE Rhumbler, 1895

Rhizammina sp
Plate 1, figure 2

Remarks: Short, curved, coarse-walled tubes. Indistinguishable from *R. indivisa* Brady of Gradstein and Berggren (in press) and *R. algaeformis* Brady of Webb, 1975. Abundant.

?***Bathysiphon*** sp
Plate 1, figure 1

Remarks: Very fine-grained straight tube with lateral projection similar to *Bathysiphon* of the North Sea and Canadian margin; may be a central chamber characteristic of the genus *Rhabdammina.* Common.

Bathysiphon sp.

Remarks: Straight, fine-grained tubes similar to smooth-walled *Bathysiphon discreta* Brady var. A of Gradstein and Berggren (in press). Present only in core 16. Common.

Subfamily HIPPOCREPINNAE Rhumbler, 1895

Hyperammina sp.

Remarks: Initially globular, later as in *Rhizammina* and *Bathysiphon.* Rare.

Family SACCAMMINIDAE Brady, 1884
Subfamily SACCAMMININAE Brady, 1884

Saccammina complanata (Franke)
Plate 1, figure 3

Pelosina complanata FRANKE, 1912, pl 3, fig. 1—CUSHMAN and JARVIS, 1932, p. 5, pl. 1, figs. 4–6.
Saccammina placenta Grzybowski.—GEROCH, 1960, pp. 37–38, pl. 2, figs. 1–6—JURKIEWICZ, 1967, p. 41, pl. 1, fig. 5.—GRADSTEIN and BERGGREN, 1981, pl. 2, fig. 5.
Saccammina complanata (Franke).—RÖGL, 1976, pl. 3, figs. 7–8.—KRASHENINNIKOV and PFLAUMANN, 1977, pl. 1, fig. 14.

Remarks: Franke illustrated a specimen with a pronounced neck and a disk-shaped depression similar to our specimens. Found only in core 16. Rare.

Saccammina placenta (Grzybowski)
Plate 1, figure 4

Reophax placenta GRZYBOWSKI, 1898, pl. 10, figs. 9–10.
Saccammina placenta (Grzybowski).—KRASHENINNIKOV and PFLAUMANN, 1977, pl. 1, fig. 16.

Remarks: Grzybowski did not describe the type of aperture but, as he illustrated a specimen with no neck, we assign specimens without a neck to *S. placenta* and those with a distinct neck to *S. complanata.* As noted by Krasheninnikov, *S. placenta* has a thin lip. Found only in cores 13 and 14. Rare.

Saccammina difflugiformis (Brady)
Plate 1, figure 5

Reophax difflugiformis BRADY, 1879, pl. 4, fig. 3.

Remarks: Vaselike test; may be initial chamber of uniserial form such as "*Hormosina*" sp. 4. Rare.

Family AMMODISCIDAE Reuss, 1862
Subfamily AMMODISCINAE Reuss, 1862

Ammodiscus cretaceus (Reuss)
Plate 1, figure 6

Operculina cretacea REUSS, 1845, p. 35, figs. 64–65.
Ammodiscus cretaceus (Reuss).—FRIZZEL, 1954, p. 58, pl 1, fig. 15.—KRASHENINNIKOV, 1974, pl. 7, fig. 8.—WEBB, 1975, p 834, pl. 1, fig. 9.—RÖGL, 1976, pl. 2, fig. 22.—GRADSTEIN and BERGGREN, 1981, pl. 2, figs. 12–13
Ammodiscus cretaceus cretaceus (Reuss).—KRASHENINNIKOV and PFLAUMANN, 1977, pl. 2, fig. 6

Remarks: Large, very fine-grained, numerous whorls (more than 8), tightly coiled test, biconcave in axial view. Common.

Ammodiscus rugosus Schijfsma (not Terquem)
Plate 1, figure 7

Ammodiscus cretaceus (Reuss) var. *rugosa* SCHIJFSMA, 1946, pl. 6, fig. 2.
Ammodiscus cretaceus rugosus Schijfsma.—KRASHENINNIKOV, 1974, pl. 7, fig. 9 —KRASHENINNIKOV and PFLAUMANN, 1977, pl. 2, fig. 6.

Remarks: Smaller, coarser grained, more loosely coiled, with a tendency toward irregular coiling, and fewer whorls (2½ to 4) than *A. cretaceus.*

Ammodiscus glabratus Cushman and Jarvis
Plate 1, figure 8

Ammodiscus glabratus CUSHMAN and JARVIS, 1928, pl. 12, fig 6.—KRASHENINNIKOV and PFLAUMANN, 1977, pl. 2, figs. 8–9.

Remarks: Small, finer grained than *A. rugosus* and often coarser grained than *A. cretaceus,* very slightly concave in axial view, numerous whorls (more than 9). Common.

Glomospira charoides (Jones and Parker)
Plate 1, figures 10–11

Trochammina squamata Jones and Parker var. *charoides* JONES and PARKER, 1860, p. 304
Ammodiscus charoides (Jones and Parker) —GRZYBOWSKI, 1896, figs. 39–43.
Glomospira charoides (Jones and Parker).—BARKER, 1960, pl. 2, figs. 8–10.—GEROCH, 1960, pp. 46–47, pl. 4, figs. 1, 2, 5.—JURKIEWICZ, 1967, pp. 58–59, text.-fig. 7, pl. 2, figs. 16–17.—WEBB, 1975, pl. 1, figs. 13–14.—RÖGL, 1976, pl. 2, figs. 26–27.—GRADSTEIN and BERGGREN, 1981, pl. 3, figs. 5–7.
Glomospira charoides (Jones and Parker) var. *corona* CUSHMAN and JARVIS, 1928, p. 89, pl. 12, figs. 9–11; 1932, p. 10, pl. 2, figs. 8–10.
Glomospira corona (Cushman and Jarvis).—KRASHENINNIKOV, 1974, pl. 7, fig. 5.—KRASHENINNIKOV and PFLAUMANN, 1977, pl. 2, fig. 2.

Remarks: More regularly coiled than *G. gordialis* (cf. pl. 1, figs. 10–11, figs. 14–15). Krasheninnikov (1974, 1977) used *G. corona, G. charoides,* and *G. gordialis* as progressively irregularly coiling species. In our view, these progressively irregularly coiling species correspond to *G. charoides, G. gordialis,* and *G. irregularis,* respectively. Common.

Glomospira gordialis (Jones and Parker)
Plate 1, figures 14–15

Trochammina squamata Jones and Parker var. *gordialis* JONES and PARKER, 1860, p 304
Ammodiscus gordialis (Jones and Parker).—GRZYBOWSKI, 1896, figs. 44–45.
Glomospira gordialis (Jones and Parker).—BARKER, 1960, pl. 38, figs. 7–9.—GEROCH, 1960, pp 46–47, pl. 4, figs 1, 2, 5.—JURKIEWICZ, 1967, pp. 59–60, text-fig. 8, pl. 2, fig. 23.—WEBB, 1975, pl. 1, fig. 12.—RÖGL, 1976, pl. 2, fig. 28.—GRADSTEIN and BERGGREN, 1981, pl 3, figs 2–3.

Remarks: More irregularly coiled than *G. charoides.* Common.

Glomospira irregularis (Grzybowski)
Plate 1, figure 12

Ammodiscus irregularis GRZYBOWSKI, 1898, pl. 11, figs. 2–3.
Glomospira irregularis (Grzybowski).—GEROCH, 1960, pp. 47–48, pl. 4, figs 9–10.—JURKIEWICZ, 1967, pp. 61–62, text-fig. 9, pl. 2, figs. 18–19.—RÖGL, 1976, pl. 3, fig. 12.—GRADSTEIN and BERGGREN, in press, pl. 3, figs 1–4.

Remarks: Irregularly coiled. Common.

Glomospira serpens (Grzybowski)
Plate 1, figure 13

Ammodiscus serpens GRZYBOWSKI, 1898, pl 10, figs. 31–33.
Glomospira serpens (Grzybowski).—GEROCH, 1960, pp 47–48, pl. 4, fig. 13.—JURKIEWICZ, 1967, pp. 61–62, pl. 2, figs. 24–27.—WEBB, 1975, pl. 1, fig. 15.

Remarks: Oval, slightly irregularly coiling. Found only in cores 13 and 14. Rare.

Glomospirella sp.
Plate 1, figures 16–17

Remarks: Initial coil as in *G. irregularis,* later as in *Ammodiscus.* Common.

Subfamily TOLYPAMMININAE Cushman, 1928

Ammolagena clavata (Jones and Parker)
Plate 1, figure 9

Trochammina irregularis (d'Orbigny) var. *clavata* JONES and PARKER, 1860.
Ammolagena clavata (Jones and Parker).—WEBB, 1975, pl 10, figs. 31–33

Remarks: Bulbous test with elongate tubular neck found attached to *Ammodiscus cretaceus.* Rare.

Superfamily LITUOLACEA de Blainville, 1825
Family HORMOSINIDAE Haeckel, 1894
Subfamily HORMOSININAE Haeckel, 1894

Hormosina ovulum (Grzybowski)
Plate 1, figure 20

Reophax (*Rheophax*) *ovulum* GRZYBOWSKI, 1896, pl 8, figs. 19–21.
Hormosina ovulum (Grzybowski).—GEROCH, 1960, p. 43, pl. 2, figs 20–22, pl. 10, figs 8–9 —JURKIEWICZ, 1967, pp. 52–53, pl 1, fig. 28.—KRASHENINNIKOV, 1974, pl. 17, fig. 12 —WEBB, 1975, pl. 1, figs. 1–2.—RÖGL, 1976, pl. 2, figs. 24–25, pl. 3, fig. 36 —KRASHENINNIKOV, 1977, pl 1, figs 17–18 —GRADSTEIN and BERGGREN, in press, pl. 2, figs. 1–2.

Remarks: Found only as single chambers. Rare.

Hormosina globulifera Brady
Plate 1, figure 19

Hormosina globulifera BRADY, 1879, pl. 4, figs. 4–5.—RÖGL, 1976, pl. 3, fig. 29.

Remarks: Common.

Hormosina sp 3
Plate 1, figure 21

Remarks: Fine-grained, smaller-sized tests. Occur predominantly as multi-chambered individuals. Common.

"***Hormosina***" sp. 4
Plate 1, figure 18

Remarks: Found only as single chambers, and therefore may belong to the genus *Pelosina.* One or both "apertures" are often jagged, suggesting that whole individuals are uniserial. We assign these oval-chambered individuals to *Hormosina.* Common.

Reophax pilulifera Brady
Plate 1, figure 22

Reophax pilulifera BRADY, 1884, p 292, pl 30, figs. 18–20.
Reophax pilulifer Brady.—BARKER, 1960, pl. 30, figs. 18–20.—JURKIEWICZ, 1967, pp. 48–49, text-fig. 4, pl 1, fig. 24.—RÖGL, 1976, pl 4, fig 7 —GRADSTEIN and BERGGREN, 1981, pl 2, figs 10–11.

Remarks: Two or more overlapping, flattened chambers. Rare.

Reophax nodulosus Brady
Plate 1, figure 23

Reophax nodulosus BRADY, 1879, pl. 4, figs 7–8.—BARKER, 1960, pl. 31, figs. 1–9

Remarks: Rare.

Family RZEHAKINIDAE Cushman, 1933

Spirosigmoilinella compressa Matsunaga
Plate 2, figure 5

Spirosigmoilinella compressa MATSUNAGA, 1955, figs. 1–2.

Remarks: First described from the middle Miocene of Japan, occurs in the late middle to late Eocene of site 112 and the later Eocene of the North Sea associated with *Cyclammina amplectens.* Noted by Gradstein and Berggren (1981) as "*Rzehakina*" sp. 1. Evolute test is insoluble in HCl and lacks a tooth; it is therefore distinct from *R. epigona.* Rare.

Family LITUOLIDAE de Blainville, 1825
Subfamily HAPLOPHRAGMOIDINAE Maync, 1952

Haplophragmoides eggeri Cushman
Plate 2, figure 6

Haplophragmoides fontinense EGGER (not Terquem), 1910, p. 10, pl. 3, figs 16–18
Haplophragmoides eggeri CUSHMAN, 1926, p. 583, pl 15, fig. 1; 1946, p 20, pl 2, figs 9–10 —JURKIEWICZ, 1967, p 77, pl. 4, fig. 11.—GRADSTEIN and BERGGREN, 1981, pl. 6, figs. 1–4.

Remarks: Small, 5½ to 6½ chambers in last whorl. Fine-grained, smooth-walled as opposed to Cushman's coarser grained specimens. The interior of some specimens appears to be complex, and thus may be referred to *Aveolophragmium.* Occurs only in core 16. Common.

Haplophragmoides walteri (Grzybowski)
Plate 2, figure 7

Trochammina walteri GRZYBOWSKI, 1898, pl. 11, fig. 31.
Haplophragmoides walteri (Grzybowski).—GEROCH, 1960, pp 49–50, 127, pl. 5, fig. 5.—JURKIEWICZ, 1967, pp. 73–74, text-fig. 16, pl 3, fig 14; pl 4, fig 5.—GRADSTEIN and BERGGREN, in press, pl 6, figs. 5–7
(?)*Haplophragmoides excavata* CUSHMAN and WATERS, 1927, p. 82, pl 10, fig 3 —CUSHMAN and JARVIS, 1932, p 12, pl 3, fig. 1.

Remarks: Only flattened specimens found. Rare.

"**Haplophragmoides**" sp.
Plate 2, figure 8

Remarks: This species is similar to *H. eggeri,* except for more rounded chambers, incised sutures, and a tendency toward trochospiral (as in *Trochammina*) development. Common.

"**Cribrostomoides**" **subglobosus** (Sars)
Plate 2, figure 9

Lituola subglobosa G. SARS, 1872 (1871), p. 253.
Alveolophragmium subglobosum (G. Sars).—BARKER, 1960, pl 34, figs. 7, 8, 10.
Cribrostomoides subglobosus (G. Sars).—LEROY and HODGKINSON, 1975, p. 432, pl. 4, figs. 2–4 —RÖGL, 1976, pl. 4, fig. 21.
Haplophragmoides suborbicularis (Grzybowski).—GEROCH, 1960, pl. 5, fig. 1.—JURKIEWICZ, 1967, p. 77, pl. 4, figs. 12–13.—WEBB, 1975, p. 834, pl. 2, figs. 8–9.
"Cribrostomoides" subglobosus (G. Sars).—GRADSTEIN and BERGGREN, 1981, pl. 6, figs. 10–11.

Remarks: Ranges from flaring to narrow final chanber. Latter specimens approach *C. scitulus* (Brady) of Gradstein and Berggren (1981) and Barker (1960) in axial view but are more like *C. subglobosus* in plane view. Included here are specimens with and without visible aperture slit. Common.

Paratrochamminoides sp.
Plate 2, figure 10

Remarks: Large, irregularly coiled tests. Sutures may be poorly developed, therefore, some specimens are difficult to differentiate from *Glomospira irregularis.* Rare.

Recurvoides walteri (Grzybowski)
Plate 2, figure 11

Haplophragmoides walteri GRZYBOWSKI, 1898, pl. 10, fig. 24.
Recurvoides walteri (Grzybowski).—JURKIEWICZ, 1967, p. 78, pl. 4, fig. 14.
Thalmanammina walteri (Grzybowski).—HANZLIKOVA, 1972, p 44, pl. 7, fig. 6.
Recurvoides ex. gr. *walteri* (Grzybowski).—GRADSTEIN and BERGGREN, 1981, pl 8, figs 1–7

Remarks: In poorly preserved specimens, this streptospirally coiled species may be difficult to differentiate from *Cribrostomoides subglobosus.* Occurs only in cores 15 and 14-CC (*Cribrostomoides-Recurvoides* assemblage). Common.

Trochamminoides sp.
Plate 2, figure 18

Remarks: Planispiral, multi-chambered (greater than 10 chambers in last whorl), partially evolute form. Probably equivalent to *T. subtrullissatus* (Grzybowski).

Labrospira pacifica Krasheninnikov

Labrospira pacifica KRASHENINNIKOV, 1973, pl. 2, figs. 4–5; 1974, p. 638, pl. 3, figs. 1–2.

Remarks: One individual noted in core 16.

Subfamily CYCLAMMININAE Marie, 1941

Cyclammina amplectens Grzybowski
Plate 2, figure 16

Cyclammina amplectens GRZYBOWSKI, 1898, pl. 12, figs. 1–3.—MUYLAERT, 1966, p. 128, pl. 40, figs. 15–16.—JURKIEWICZ, 1967, pp. 82–85, pl. 10, fig. 10, text-fig. 17.—MJATLIUK, 1970, pp 90–91, pl 21, figs. 5, 11, 12; pl. 25, figs. 4–6; pl. 28, figs. 2–3.—GRADSTEIN and BERGGREN, 1981, pl. 7, figs. 13–17.

Remarks: Found in the middle to late Eocene in Site 112. Common.

Cyclammina trullissata Brady
Plate 2, figures 14–15

Trochammina trullissata BRADY, 1879, pl. 5, fig. 10.
Cyclammina bradyi CUSHMAN, 1910, fig. 174.
Cyclammina trullissata (Brady).—PARKER, 1952, p. 400.—BARKER, 1960, pl. 40, fig. 13.

Remarks: Fewer chambers (approximately 9) in last whorl than in *C. amplectens,* axial periphery more pinched, chambers more incised. Rare.

Family TROCHAMMINIDAE Schwager, 1877
Subfamily TROCHAMMININAE Schwager, 1877

Trochammina globigeriniformis (Parker and Jones)
Plate 2, figure 19

Lituola nautiloidea Lamarck var. *globigeriniformis* PARKER and JONES, 1865, p. 407, pl. 15, figs. 46–47, pl. 17, figs. 96–98.
Trochammina globigeriniformis (Parker and Jones).—CUSHMAN, 1946, p. 51, pl. 15, figs. 8, 10, 11 —JURKIEWICZ, 1967, p. 92, pl. 6, fig. 13.—HILTERMANN, 1972, pp. 643–652.—GRADSTEIN and BERGGREN, 1981, pl. 8, fig. 13

Remarks: This globigeriniform agglutinated species varies from quadrate with an umbilical aperture and a higher spire similar to *T. globigeriniformis* var. *altiformis* Cushman and Renz to a low-spired form with an umbilical-extraumbilical aperture as illustrated in plate 2, figure 19. Typically 8½ chambers in final whorl. Restricted to core 16. Common.

Trochammina sp.
Plate 2, figure 12

Remarks: Distinguishable from *Trochamminoides* sp. by greater number of chambers (greater than 10) in the last whorl and trochospiral development.

Cystammina globigerinaeformis (Krasheninnikov)
Plate 2, figures 13, 17, 21

Praecystammina globigerinaeformis KRASHENINNIKOV, 1973, p. 211, pl. 2, figs. 1–2; 1974, p. 641, pl. 6, figs. 1–3.—GRADSTEIN and BERGGREN, in press, pl. 9, figs. 11–15.

Remarks: This streptospiral, globigeriniform agglutinated species is characterized by its distinct, narrow areal aperture with prominent lip. Common.

Family TEXTULARIIDAE Ehrenberg, 1838
Subfamily SPIROPLECTAMMININAE Cushman, 1927

Spiroplectammina cubensis (Cushman and Bermudez)
Plate 2, figure 1

Spiroplectoides cubensis CUSHMAN and BERMUDEZ, 1937, pl 1, figs. 44–46.
Bolivinopsis cubensis (Cushman and Bermudez).—DOUGLAS, 1973, pl. 3, fig. 5.

Remarks: Two initial planispiral whorls; later biserial portion has a tendency to flare slightly. Rare.

Spiroplectammina spectabilis (Grzybowski)
Plate 2, figures 2–3

Spiroplecta spectabilis GRZYBOWSKI, 1898, pl 12, fig. 12.
Bolivinopsis spectabilis (Grzybowski) —HANZLIKOVA, 1972, p 48, pl. 10, fig. 8.—WEBB, 1975, pl. 3, figs. 1–3.—RÖGL, 1976, pl 3, figs. 23–25.
Spiroplectammina spectabilis (Grzybowski) —HILTERMANN, 1972, pp. 43–61, pls. 1–2.—JURKIEWICZ, 1967, pp. 88–90, text-fig. 19, pl. 5, figs. 12, 15.—GRADSTEIN and BERGGREN, in press, pl. 5, figs. 1–5

Remarks: Narrower cross section than Maestrichtian to Paleocene forms in the North Sea and Canadian margin. Both megalospheric and microspheric forms noted. Ranges throughout the Eocene. Abundant.

Family ATAXOPHRAGMIIDAE Schwager, 1877
Subfamily GLOBOTEXTULARIINAE Cushman, 1927

Dorothia sp.

Remarks: Rapidly inflating chambers. Rare.

Subfamily VALVULININAE Berthelin, 1880

Martinottiella sp
Plate 2, figure 4

Remarks: Uniserial portion not always preserved. Common.

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PLATE 1

"Simple" agglutinated foraminfera
Astorhizidae, Saccamminidae, Ammodiscidae and Hormosinidae

1 *?Bathysiphon* sp.
Sample 14-2, 102–113 cm, × 30.

2 *Rhizammina* sp.
Sample 15-CC, × 40.

3 *Saccammina complanata* (Franke)
Sample 16-1, 107–111 cm, × 112.

4 *Saccammina placenta* (Grzybowski)
Sample 14-2, 102–113 cm, × 74.

5 *Saccammina difflugiformis* (Brady)
Sample 16-1, 46–49 cm, × 93.

6 *Ammodiscus cretaceus* (Reuss)
Sample 15-CC, × 42.

7 *Ammodiscus rugosus* Schijfsma (not Terquem)
Sample 13-CC, × 93.

8 *Ammodiscus glabratus* Cushman and Jarvis
Sample 16-1, 78–84 cm, × 88.

9 *Ammolagena clavata* (Jones and Parker)
Sample 14-3, 141–144 cm, × 42.

10–11 *Glomospira charoides* (Jones and Parker)
10, sample 16-1, 78–84 cm, × 200; 11, sample 16-1, 46–49 cm, × 163.

12 *Glomospira irregularis* (Grzybowski)
Sample 16-1, 78–84 cm, × 70.

13 *Glomospira serpens* (Grzybowski)
Sample 14-CC, × 47.

14–15 *Glomospira gordialis* (Jones and Parker)
14, sample 16-1, 46–49 cm, × 151; 15, sample 16-1, 78–84 cm, × 153.

16–17 *Glomospirella* sp.
16, sample 16-1, 46–49 cm, × 130; 17, sample 16-1, 46–49 cm, × 128.

18 *"Hormosina"* sp. 4
Sample 16-1, 46–49 cm, × 70.

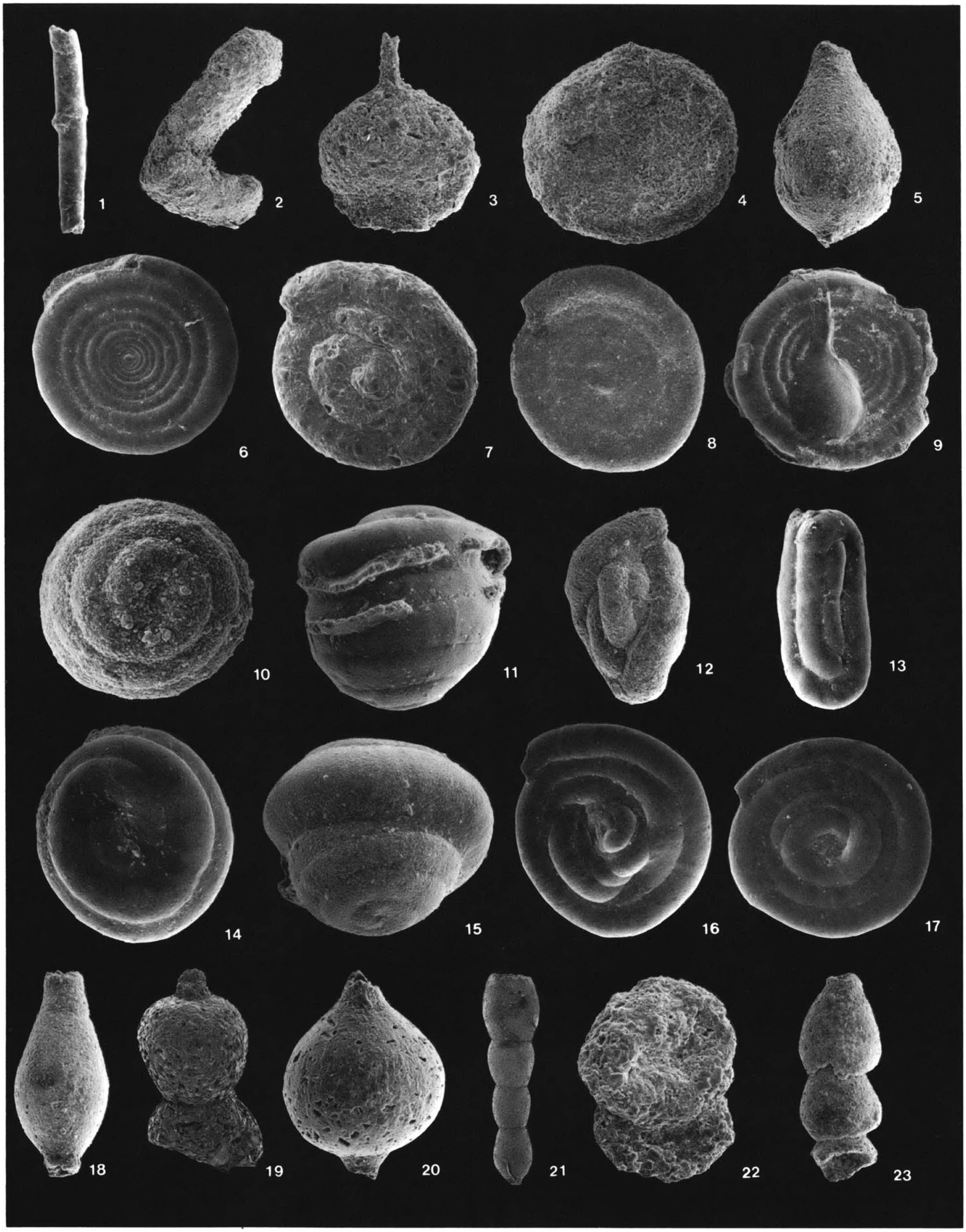
19 *Hormosina globulifera* Brady
Sample 14-2, 102–113 cm, × 53.

20 *Hormosina ovulum* (Grzybowski)
Sample 16-1, 46–49 cm, × 93.

21 *Hormosina* sp. 3
Sample 16-1, 107–111 cm, × 47.

22 *Reophax pilulifera* Brady
Sample 14-2, 102–113 cm, × 84.

23 *Reophax nodulosus* Brady
Sample 15-1, 12–25 cm, × 27.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MOORE, T. C., VAN ANDEL, TJ. H., SANCETTA, C., and PISIAS, N.,

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PLATE 2

"Complex" agglutinated foraminifera, Site 112
Lituolidae, Trochamminidae, Textulariidae and Ataxophragmiidae

1 *Spiroplectammina cubensis* (Cushman and Bermudez)
Sample 14-3, 141–144 cm, × 116.

2–3 *Spiroplectammina spectabilis* (Grzybowski)
2, sample 16-1, 78–84 cm, × 37; 3, sample 13-5, 146–149 cm, × 37.

4 *Martinottiella* sp.
Sample 14-CC, × 98.

5 *Spirosigmoilinella compressa* Matsunaga
Sample 13-2, 107–117 cm, × 88.

6 *Haplophragmoides eggeri* Cushman
Sample 16-1, 46–49 cm, × 130.

7 *Haplophragmoides walteri* (Grzybowski)
Sample 16-1, 78–84 cm, × 112.

8 *"Haplophragmoides"* sp.
Sample 16-1, 46–49 cm, × 56.

9 *"Cribrostomoides" subglobosus* (G. Sars)
Sample 14-CC, × 70.

10 *Paratrochamminoides* sp.
Sample 16-1, 78–84 cm, × 70.

11 *Recurvoides* ex. gr. *walteri* (Grzybowski)
Sample 16-1, 12–25 cm, × 70.

12 *Trochammina* sp.
Sample 14-CC, × 116.

13, 17, 21 *Cystammina globigerinaeformis* (Krasheninnikov)
13, sample 14-5, 144–147 cm, × 163; 17, sample 16-1, 46–49 cm, × 130; 21, sample 16-1, 78–84 cm, × 112.

14–15 *Cyclammina trullissata* (Brady)
14, sample 15-CC, × 60; 15, sample 15-CC, × 63.

16 *Cyclammina amplectens* Grzybowski
Sample 14-2, 102–113 cm, × 51.

18 *Trochamminoides* sp.
Sample 14-CC, × 88.

19 *Trochammina globigeriniformis* (Parker and Jones)
Sample 16-1, 107–111 cm, × 221.

20 *Dorothia* sp.
Sample 14-CC, × 112.

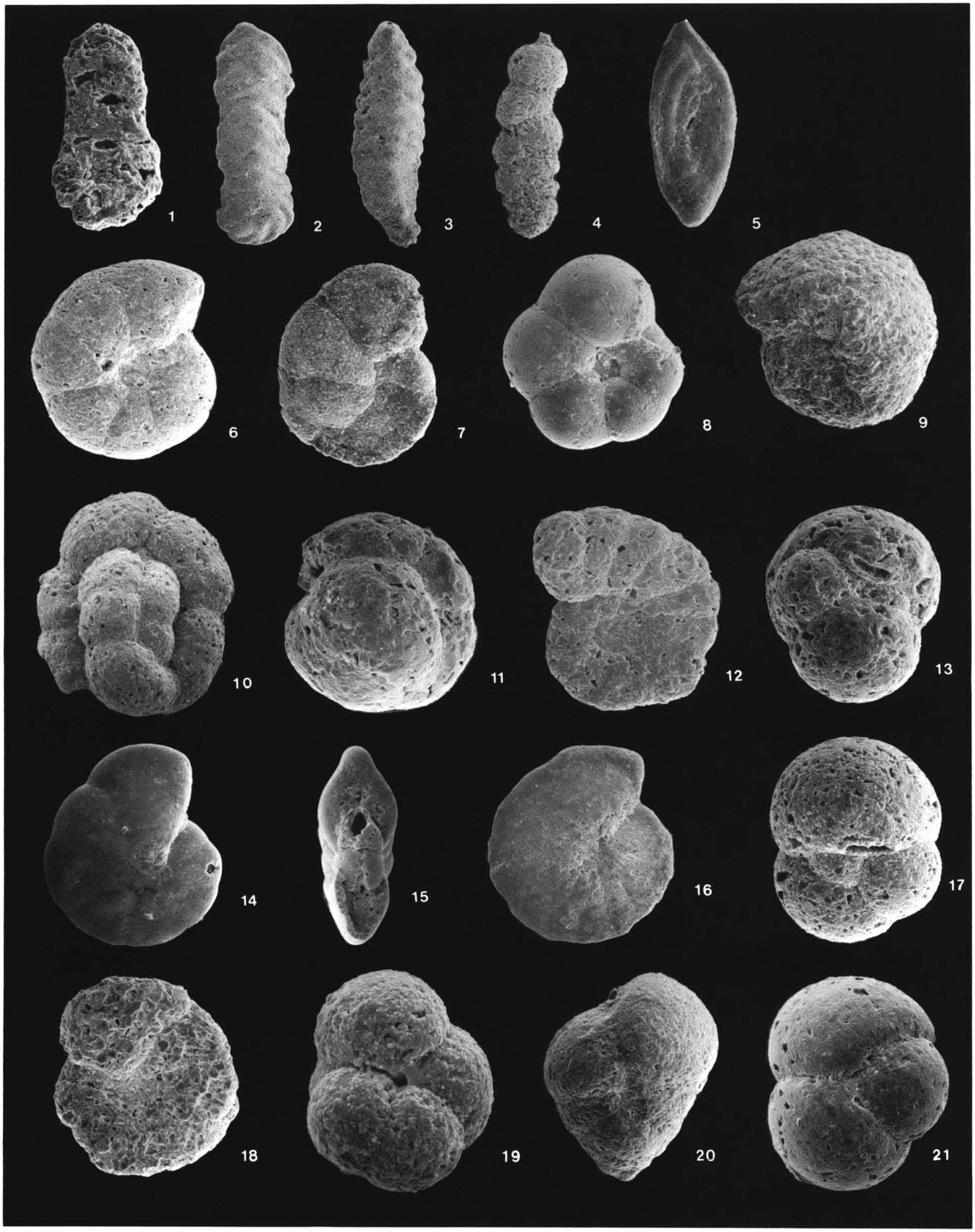
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PLATE 3

Calcareous foraminifera, Site 112

1 *Pleurostomella* sp.
Sample 13-4, 86–96 cm, × 44.

2 *Stilostomella aculeata* (Cushman and Jarvis)
Sample 15-CC, × 37.

3 *Stilostomella* sp.
Sample 14-3, 141–144 cm, × 140.

4 *Dentilina* sp.
Sample 12-1, 100–102 cm, × 34.

5 *Gyroidina* sp.
Sample 14-CC, × 120.

6 *Nuttallides truempyi* Nuttall
Sample 14-3, × 242.

7 *Pullenia quinqueloba* (Reuss)
Sample 14-3, 141–144 cm, × 140.

8 *Pullenia eocenica* Cushman and Siegfus
Sample 14-CC, × 140.

9 *Globobulimina* sp.
Sample 13-CC, × 74.

10 *Alabamina* sp.
Sample 14-3, 141–144 cm, × 237.

11–12 *Pseudohastigerina wilcoxensis* (Cushman and Ponton)
11, sample 15-CC, × 191; 12, sample 13-5, 146–149 cm, × 195.

13 *Acarinina bulbrooki* (Bolli)
Sample 15-CC, × 233.

14–15 *Subbotina linaperta* Finlay
14, sample 15-CC, × 112; 15, sample 15-CC, × 116.

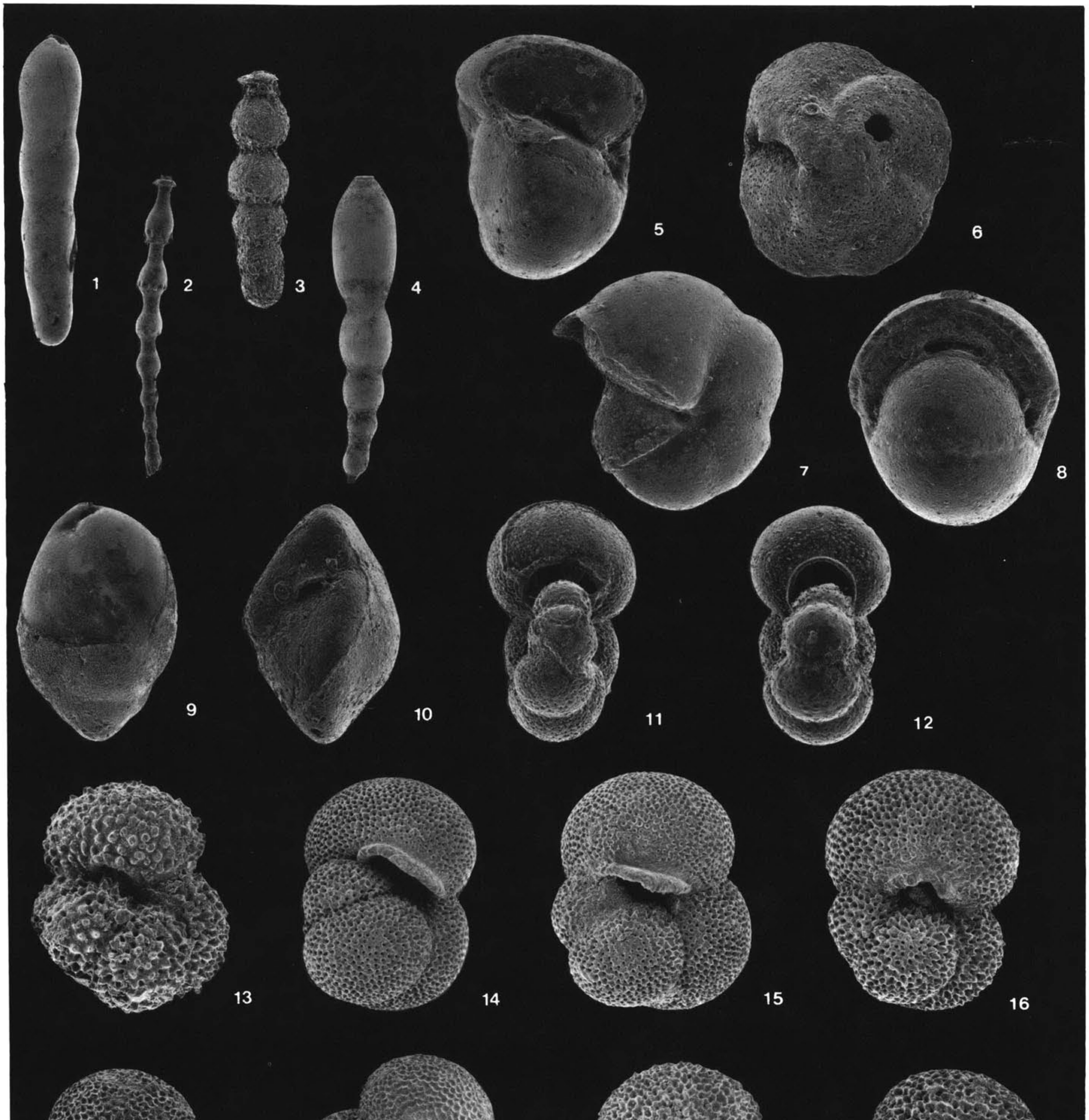
16 *Subbotina frontosa* Subbotina
Sample 15-CC, × 126.

17–18 *Dentoglobigerina galavasi* (Bermudez)
17, sample 14-2, 102–113 cm, × 116; 18, sample 14-2, 102–113 cm, × 177.

19 *Catapsydrax unicavus* Bolli
Sample 14-2, 144–146 cm, × 149.

20 *Catapsydrax perus* (Todd)
Sample 15-1, 12–25 cm, × 153.

PLATE 3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APPENDIX 2

Eocene to Oligo benthic foraminiferal isotopic record in the Bay of Biscay

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Eocene to Oligocene benthic foraminiferal isotopic record in the Bay of Biscay

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We present here oxygen and carbon isotopic records of Eocene to Oligocene benthic foraminifera from two Bay of Biscay Deep Sea Drilling Project (DSDP) sites (119 and 401). δ^{18}O of benthic foraminifera increases 1.9‰ from a middle Eocene minimum (Zones P10–P11) to an earliest Oligocene maximum (Zone NP21). Approximately 1.4‰ of the increase in benthic foraminiferal δ^{18}O occurs during the late Eocene to earliest Oligocene (Zones P15/16–NP21). Previous results from other North Atlantic DSDP sites (400A and 398) have significantly lower δ^{18}O values of benthic foraminifera, some by as much as 2‰ (refs 1–3). We believe that these differences result from diagenetic alteration of the sediments in the deeper-buried Sites 400A and 398.

Previous reports of Pacific and Southern Ocean DSDP sites noted that a 1.0–2.0‰ enrichment of ^{18}O occurs from the middle Eocene to early Oligocene in both planktonic and benthic foraminifera and that ~1‰ of this change occurs near the Eocene–Oligocene boundary[4–9]. Eocene to Oligocene δ^{18}O records from three DSDP locations depart from this general Pacific–Southern Ocean trend: (1) in the Philippine Sea (Site 292) the ^{18}O enrichment occurs only in benthic foraminifera[9]; (2) in the South Atlantic (Site 357), an isotopic enrichment occurs between the middle and late Eocene[10]; and (3) in the North Atlantic (Site 398) the magnitude of the isotopic increase is twice that noted in the Pacific and Southern Oceans and the enrichment occurs near the middle/late Eocene boundary[1,2].

The Eocene to Oligocene oxygen isotopic enrichment in the North Atlantic is poorly documented. Strong diagenetic effects on isotopic composition were noted in sites drilled in the Bay of Biscay (Site 400A) (refs 1, 3, 11) and off Portugal (Site 398) (refs 1, 2, 12, 13). Eocene benthic foraminiferal δ^{18}O values are low and calculated palaeotemperatures are correspondingly high (~17 °C)[1,2] relative to Pacific, Southern Ocean[4–9], and South Atlantic[10] δ^{18}O records. Such low δ^{18}O values and high inferred palaeotemperatures may result from diagenetic effects. It is necessary to distinguish between altered isotopic signals and original unaltered signals to determine abyssal palaeocirculation patterns.

We analysed the oxygen and carbon isotopic composition of early-middle Eocene to Oligocene benthic foraminifera at two North Atlantic DSDP sites (119 and 401). We chose these sites for the shallow burial depth of their Eocene sediments (~100–400 m) versus Sites 400A and 398 (~550–600 m). Late Eocene sediments are missing from Site 119. However, drilling at nearby Site 401 recovered a relatively complete Eocene section, but no Oligocene section[14]. Site 119 was drilled in 4,447 m of water on Cantabria Seamount[15]. Using the backtracking method of Berger and Winterer[16], we calculate that Site 119 was deeper than 3,000 m throughout the Eocene and Oligocene. Site 401 was drilled in the Bay of Biscay on the Armorican margin in 2,495 m of water[17]. Palaeobathymetric estimates place Site 401 near its present depth throughout the early Tertiary[18]. Biostratigraphic age control for the sites is taken from the *Initial Reports of the DSDP* Legs 12 (refs 15, 19–21) and 48 (refs 14, 17, 22); a biostratigraphic review and detailed palaeobathymetric estimates will be presented elsewhere (K.G.M., work in preparation).

Table 1 Oxygen and carbon isotopic data for Sites 119 and 401

Sample	Zonal age (Myr)	Taxa	δ^{18}O	δ^{13}C
Site 119				
12-CC	NP25 (24.0)	*Cibicidoides*	1.15	0.39
13-2 146–149 cm	NP25 (25.7)	*Cibicidoides*	1.40	0.30
13-CC	NP25 (26.3)*	*Cibicidoides*	1.54	0.53
14-CC	P21, NP24 (28.7)	*Cibicidoides*	1.36	−0.03
15-CC	P20/21, NP24 (31.0)*	*Cibicidoides*	1.61	−0.01
			1.52	−0.33
		Gyroidinoides	1.24	−0.47
			1.21	−0.46
			1.72	−0.68
		Catapsydrax	0.81	−0.59
16-2 119–121 cm	NP23 (31.6)	*Cibicidoides*	1.27	0.06
16-CC	NP23, P20/21 (32.3)*	*Cibicidoides*	1.81	0.56
			1.78	0.42
		Gyroidinoides	2.26	−0.06
			2.17	−0.23
		Catapsydrax	1.12	0.52
17-5 146–149 cm and 17-CC	NP22	*Catapsydrax*	0.84	0.71
17-CC	NP22 (34.7)	*Cibicidoides*	1.59	0.75
		Gyroidinoides	2.05	0.53
18-2 144–147 cm	NP21/22 (35.7)	*Cibicidoides*	1.35	0.47
18-CC	NP21 (37.0)	*Cibicidoides*	1.96	1.28
			2.02	1.27
		Gyroidinoides	2.56	0.76
			2.34	0.46
19-CC	Lower NP15 (47.0)	*Cibicidoides*	0.14	0.38
20-3 top	NP15 (47.4)	*Cibicidoides*	0.42	0.44
20-CC	NP15 (47.6)	*Cibicidoides*	0.43	0.53
		Gyroidinoides	0.79	0.05
21-CC	NP13 (49.3)	*Cibicidoides*	0.73	0.51
		Gyroidinoides	0.62	0.03
Site 401				
2-1 8–10 cm	*G. cerrolazulensis* Zone, NP21 (ref. 22) (37.5)	*Cibicidoides*	1.26	0.89
3-1 15–21 cm	P15/16 (38.9)	*Cibicidoides*	0.64	0.67
4-CC	P14 (41.0)	*Cibicidoides*	0.86	0.55
5-3 93–97 cm	P13 (42.8)	*Cibicidoides*	0.22	0.14
6-3 76–80 cm	P12 (43.9)	*Cibicidoides*	0.62	0.53
7-3 76–80 cm	P12 (45.3)	*Cibicidoides*	0.20	0.73
8-3 44–48 cm	P11 (46.6)	*Cibicidoides*	0.43	0.65
9-3 96–100 cm	P10 (47.7)	*Cibicidoides*	0.45	0.63
10-3 98–102 cm	P10 (49.0)	*Cibicidoides*	0.42	0.49

Nannoplankton zonations for Site 119 after ref. 20 and M.-P. Aubry, personal communication. Foraminiferal zonations for Site 119 after ref. 19 and K.G.M. work in preparation and for Site 401 after ref. 14.

* Discrepancies resulting from differences between zonal age and interpolated ages (indicated in parentheses in Myr) obtained by assuming constant sedimentation rate.

Isotopic analyses were performed on mixed species of the genus *Cibicidoides*. Interregional isotopic variations of this taxon in Holocene core tops reflect the distribution of temperature, δ^{18}O of seawater, and δ^{13}C of ΣCO_2 in the modern ocean[23–26]. Although *Cibicidoides* species apparently secrete their tests lower in δ^{18}O than values calculated from the Epstein *et al.*[27] palaeotemperature equation, the offsets seem to be constant[23–26]. In addition, we analysed paired samples of *Gyroidinoides* spp. and *Cibicidoides* spp. from Site 119; *Gyroidinoides* were ~0.3‰ enriched in ^{18}O relative to *Cibicidoides*. This offset is similar to differences observed between these taxa in studies of Holocene core-top sediments[26].

At Site 119, δ^{18}O values increase from ~0.5‰ in the early–middle Eocene (nannoplankton Zones NP13 to NP15) to ~2.0‰ in the earliest Oligocene (Zone NP21) (Fig. 1); δ^{13}C values correspondingly increase from ~0.4 to ~1.3‰ from the middle Eocene to earliest Oligocene. Unfortunately a 9-Myr

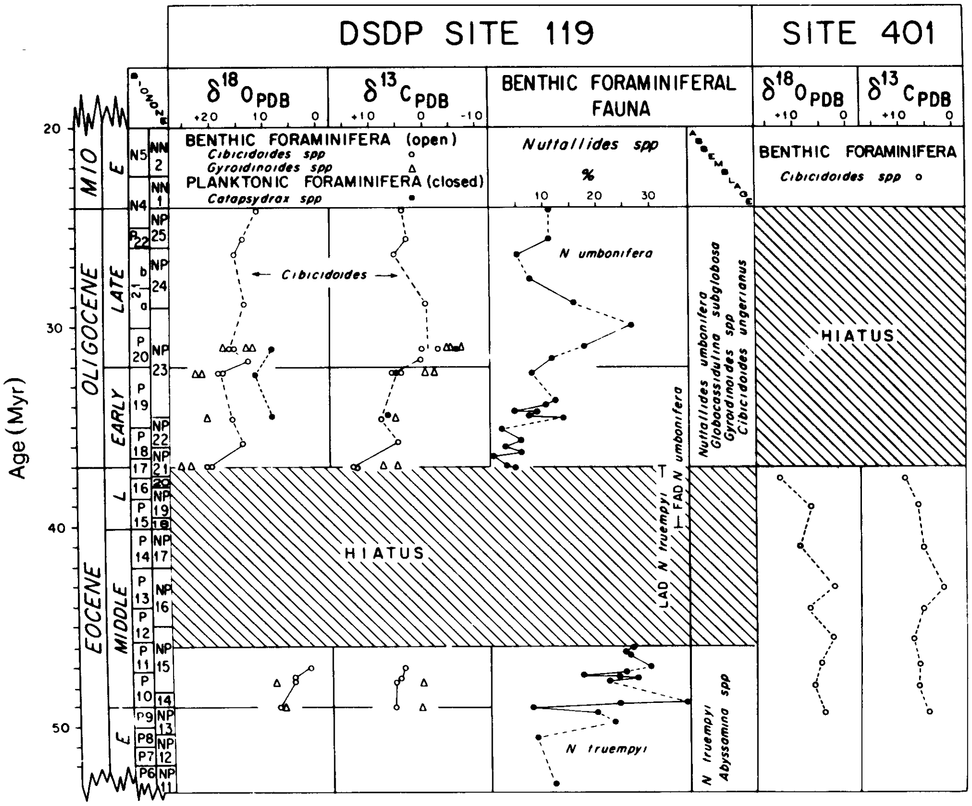


Fig. 1 Age versus foraminiferal isotopic composition for Sites 119 and 401 and faunal composition for Site 119. Last appearance datum (LAD) of *N. truempyi* and first appearance datum (FAD) of *N. umbonifera* after ref. 39.

hiatus precludes determination of the exact timing and nature of this isotopic change between the middle Eocene and earliest Oligocene in Site 119. In the Oligocene, δ^{18}O values decrease from ~2.0 to ~1.1‰. δ^{13}C values decrease to a minimum of ~0.0‰ in the middle Oligocene (Zones P20–P21), then increase again to ~0.5‰. At Site 401 (Fig. 1), δ^{18}O values are ~0.5‰ in the early–middle Eocene, similar to observed values in Site 119. δ^{18}O values increase ~0.4‰ in the late middle Eocene (Zone P14). A δ^{18}O increase of 0.6‰ occurs within the late Eocene (Zones P15/16–P17).

Combining the isotopic record of both sites (Fig. 2) suggests that a δ^{18}O increase of ~1.4‰ occurs from the late Eocene to earliest Oligocene in the North Atlantic. This change occurred within an interval that may be <1 Myr (that is within Zone NP21) or as much as 4 Myr (between Zones P15 and NP21; see biostratigraphic zonations Table 1). There are pitfalls in combining sites of different palaeodepths to obtain a complete record; however, δ^{18}O values obtained for Sites 119 and 401 are similar in the overlapping section (middle Eocene). The modern temperature difference between these water depths in the Bay of Biscay is <1 °C (ref. 28), equivalent to an oxygen isotopic difference of <0.25‰. These reasons justify combining the δ^{18}O records of Sites 119 and 401. Although the middle to late Eocene record may be found entirely within Site 401, the nature and timing of the δ^{18}O shift near the Eocene–Oligocene boundary involves comparing records from two different sites, and should be considered as preliminary. The Eocene–Oligocene interval in the North Atlantic is usually discontinuous[29,30], and a complete Eocene to Oligocene isotopic record from one site is not available at present.

The δ^{18}O records of Sites 119 and 401 apparently correlate with isotopic records from the Pacific and Southern Oceans[4–8] where an ^{18}O enrichment occurs in the late Eocene to early Oligocene. In detail, the curves may not correlate. Keigwin[9] showed that the ^{18}O enrichment occurs in the Pacific above the Eocene–Oligocene boundary, while in our record the change occurs from the late Eocene to earliest Oligocene. However, given the errors of the biostratigraphic age assignments, we consider that the enrichments may be synchronous. Further biostratigraphic studies are needed to determine if the enrichments are diachronous or synchronous.

Although the δ^{18}O values we report are similar to values obtained from sites in the Pacific and Southern Oceans[4–9], they are markedly higher than Eocene δ^{18}O values obtained from other sites in the North Atlantic[1–3] (Fig. 3). In the Eocene, benthic foraminiferal δ^{18}O in Sites 400A and 398 are as much as 2‰ lower than in Sites 119 and 401 (Fig. 3). Such implied differences in bottom water temperature (8 °C) between sites of similar palaeodepth (for example, Sites 400A and 119) within the small basin of the Bay of Biscay are oceanographically unreasonable. As the total range of δ^{18}O values for benthic

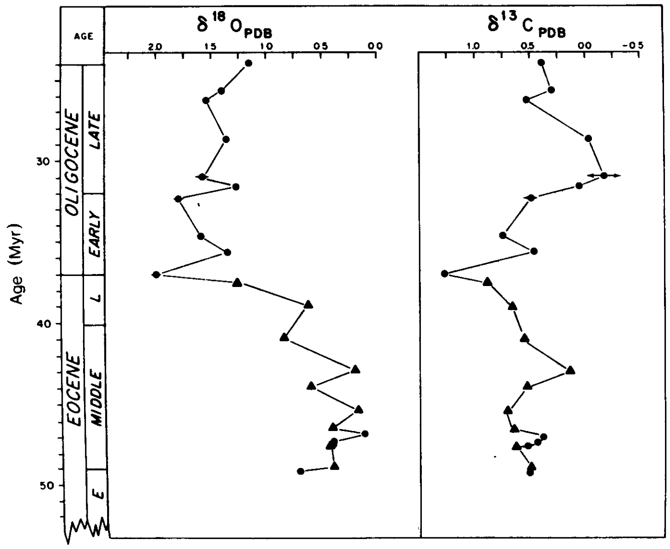
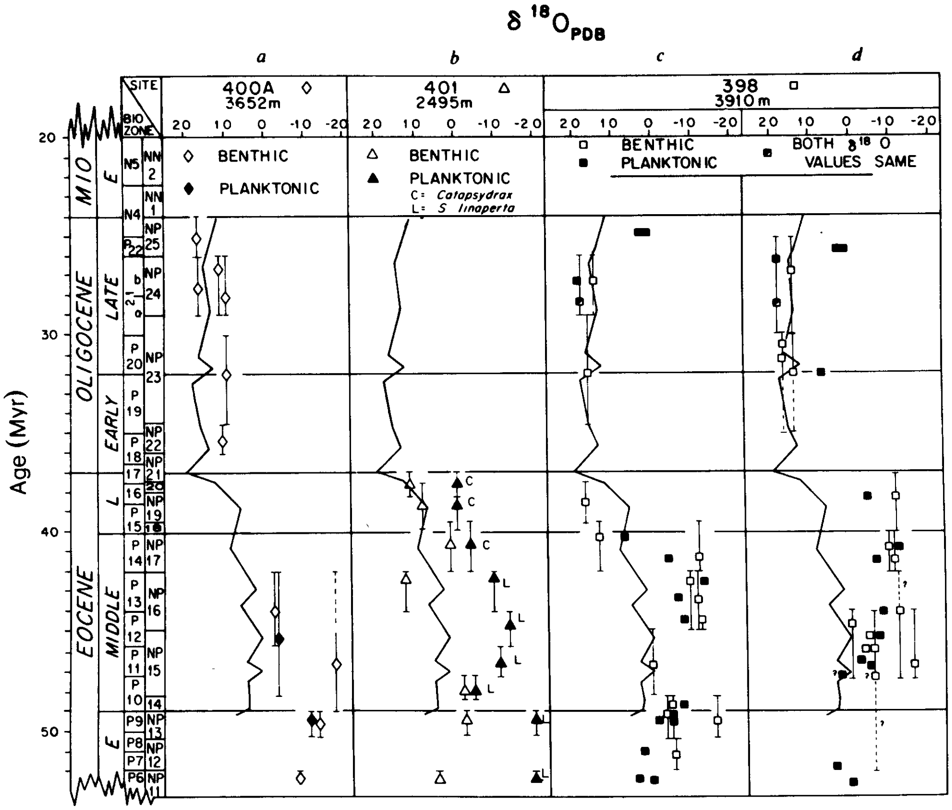


Fig. 2 Combined oxygen and carbon isotopic record obtained for *Cibicidoides* from Sites 119 (●) and 401 (▲). Where more than one value was obtained per sample, the mean value was plotted and the range indicated with arrows.

Fig. 3 Comparison of previously reported[1] Eocene to Oligocene δ^{18}O values for benthic and planktonic foraminifera from North Atlantic DSDP sites. Combined oxygen isotopic curve obtained from Sites 119 and 401 is superimposed on each (solid line). Columns *a* and *b* are from Bay of Biscay Sites 400A and 401, respectively. Columns *c* and *d* are from Site 398 off Portugal: *c* represents ages assigned using calcareous nannoplankton[37], *d* represents the same data with ages assigned using planktonic foraminifera[38]. Biostratigraphic ages for Sites 400A and 401 have been assigned using planktonic foraminifera[14] and nannoplankton[22]. Error bars indicate length of biostratigraphic zone(s) to which the sample has been assigned; they have been omitted for clarity from planktonic foraminiferal δ^{18}O values in columns *c* and *d*.



foraminifera within any Recent core-top sample[23-26] or Palaeogene sample[10] is <1‰, the 2‰ difference probably cannot result from sampling biases caused by selective dissolution or analysis of mixed benthic foraminiferal assemblages. However, the low values δ^{18}O observed can result from diagenetic alteration of the foraminiferal test at high temperatures associated with great burial depth[2,13]. While simple addition of calcite overgrowths of the test would be sufficient to lower δ^{18}O values, recrystallization of the foraminiferal calcite is necessary to produce both the similar δ^{18}O values of planktonic and benthic foraminifera in the Eocene of Sites 400A and 398 and the high bottom water temperatures (~17 °C) calculated for these sites[1-3]. At Site 401, δ^{18}O of planktonic foraminifera are substantially lower than benthic foraminifera[3], supporting the idea that this site is not substantially affected by diagenetic alteration. Similarly, planktonic foraminifera have substantially lower δ^{18}O composition than benthic foraminifera in the Oligocene of Site 119; due to dissolution, sufficient planktonic foraminifera were not available for isotopic analyses of the Eocene of Site 119.

Limits can be calculated for the abyssal temperature change in the Bay of Biscay. Shackleton and Kennett[6] estimated that δ^{18}O of seawater before the formation of the present ice sheets was −1.2‰ (PDB). Assuming that *Cibicidoides* precipitates $CaCO_3$ 0.65‰ lower than equilibrium[26] and that there was no significant glacial ice buildup, the increase in benthic foraminiferal δ^{18}O from Sites 119 and 401 corresponds to a maximum decrease of bottom-water temperatures from 8 to 2 °C between the middle Eocene and early Oligocene. Alternatively, assuming glacial buildup equivalent to the present ice sheets, the increase represents a minimum change in bottom-water temperature from 8 to 6 °C.

The benthic temperature drop in the late Eocene to early Oligocene in the Pacific and Southern Oceans has been attributed to the first formation of bottom waters in the Antarctic region[7]; however, the North Atlantic basins may have had a different source of cold bottom water. Seismic stratigraphical evidence from the northern North Atlantic Ocean indicates that northern sources of vigorously circulating bottom water began in the late Eocene to early Oligocene[30-32]. Miller and Tucholke[30] suggested that this bottom water was of Arctic origin and that it flowed through the Norwegian–Greenland Sea, the Faeroe–Shetland Channel, and possibly through the Denmark Straits into the North Atlantic.

The isotopic changes observed in Site 119 correlate with a change from a benthic foraminiferal assemblage dominated by *Nuttallides truempyi* to an assemblage dominated by *N. umbonifera* (Fig. 1); however, the hiatus prevents an exact determination of the timing of this faunal change. Within the Oligocene, *N. umbonifera* reaches a peak in abundance correlating with a δ^{13}C minimum. As *N. umbonifera* is negatively correlated with carbonate saturation in the modern ocean[33] and lower δ^{13}C values are often associated with older water masses[34-36], we speculate that the isotopic and faunal records of Site 119 reflect a change from younger, less corrosive water in the early Oligocene to older, more corrosive bottom water in the middle Oligocene. The change from lower carbon and oxygen values in the Eocene to higher carbon and oxygen values in the earliest Oligocene may reflect a change from older and warmer Eocene bottom water to colder, younger, more vigorously circulating bottom waters of northern origin in the early Oligocene.

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DEVELOPMENT OF CENOZOIC ABYSSAL CIRCULATION SOUTH OF THE GREENLAND-SCOTLAND RIDGE

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ABSTRACT

Seismic, lithostratigraphic, faunal, and isotopic evidence from the western and northern North Atlantic indicates that formation of northern sources for strongly circulating bottom water began in the late Eocene to early Oligocene. The widely distributed reflector R4 correlates with an unconformity eroded along basin margins at the Eocene/Oligocene boundary. This change in abyssal regime also correlates with a major benthic foraminiferal turnover in the deep southern Labrador Sea (DSDP Site 112) and with a faunal reorganization in the Bay of Biscay. The principal bottom-water source probably was of Arctic origin; it entered the Norwegian Sea following separation of Greenland and Spitsbergen and flowed south across the Greenland-Scotland Ridge through the Faeroe-Shetland Channel and possibly across a sill east of Greenland. This flow may have been supplemented by dense Arctic water entering the basin via Nares Strait and Baffin Bay, and by cooling and sinking of saline surface water south of the Greenland-Scotland Ridge and in the Labrador Sea. Current-controlled sedimentation and erosion, often of a chaotic nature, continued through the Oligocene above reflector R4, but the general intensity of abyssal circulation is thought to have decreased. Above reflector R2 (upper lower Miocene) current-controlled sedimentation became more coherently developed, and a major phase of sedimentary drift development was initiated. We interpret this to be a result of a further general reduction and especially a stabilization of the abyssal circulation, possibly linked with degeneration of numerous fracture-zone conduits that previously funnelled bottom water across the Reykjanes Ridge. The gross nature of the circulation has not changed substantially since the middle Miocene, although it has been punctuated by further climatic and tectonic events.

INTRODUCTION

Deep and bottom waters formed at high latitudes in the North Atlantic and in the Norwegian-Greenland Seas strongly affect the Holocene sedimentary record throughout the North Atlantic Ocean, especially along the western margins of the basins. It has been known for more than a decade that this current influence also had a significant effect on the sedimentary record well back into the Tertiary (e.g. Jones, et al., 1970). In recent years, an expanding data base of seismic reflection profiles, DSDP boreholes, and piston cores has allowed more extensive study of these current influences, but the definition of even such first-order events as the time of initiation of geologically significant deep circulation in the North Atlantic and the effects of potential tectonic barriers like the Greenland-Scotland Ridge still is in dispute.

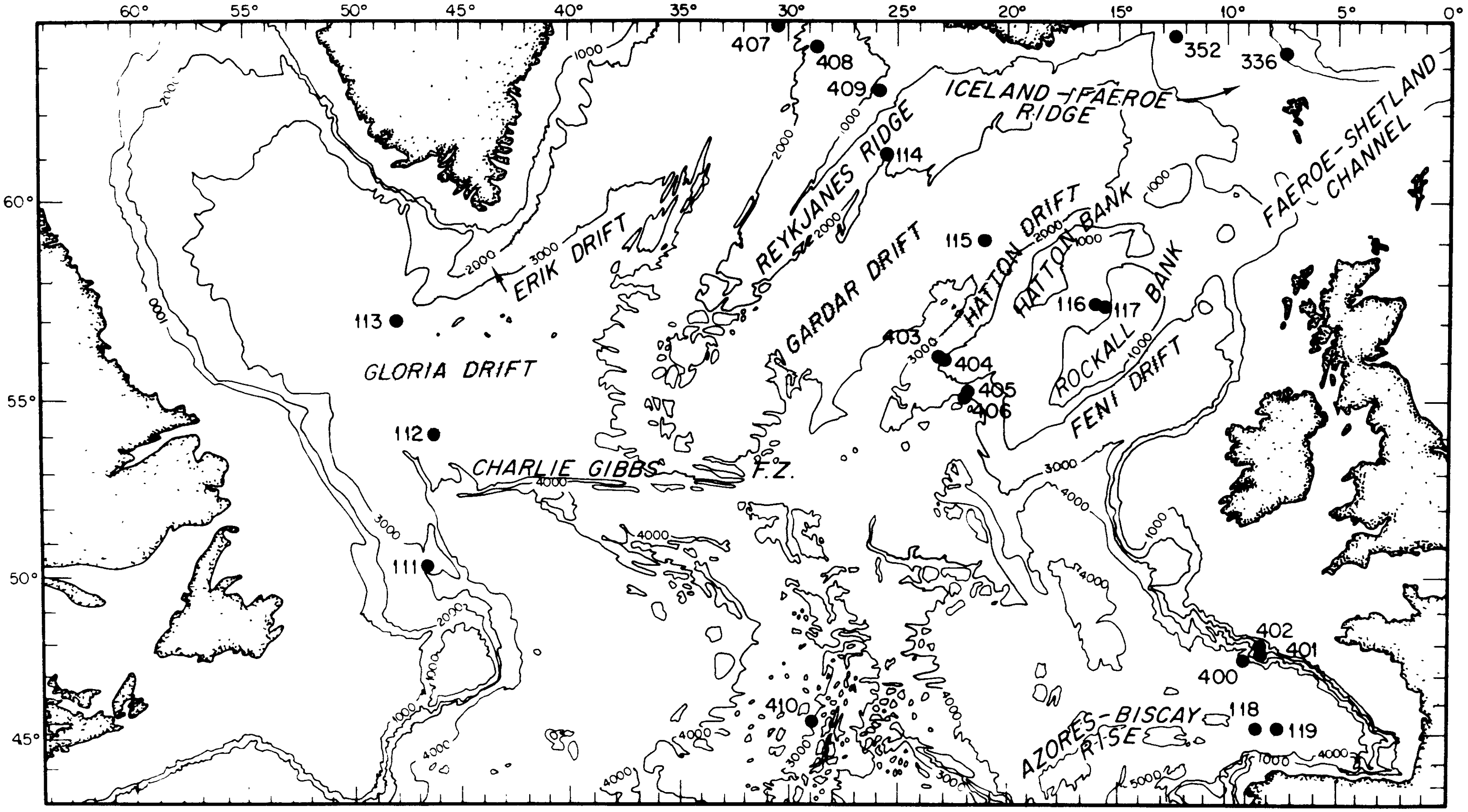
Part of the ambiguity is due to a lack of necessary data, but significant dispute also results from varying ways of interpreting existing data. For example, one form of evidence clearly documenting the presence of abyssal circulation is the development of large-scale current-controlled bedforms and sediment drifts. Such drifts were well developed by the early to middle Miocene in both the western and northern North Atlantic (Figures 1, 2). However, it is unclear exactly what change in abyssal circulation the first development of these drifts represents, and there are at least four possibilities:

1) Reduction in current intensity below the critical threshold between erosional/transportational and depositional conditions (especially likely if the drifts are developed above unconformities).
2) Stabilization of abyssal currents compared to earlier erratic flow, with consequent regular development of bedforms and accumulation patterns.
3) Interaction of newly developed flows with existing abyssal currents to form a stable depositional pattern.
4) Intensification of abyssal currents from near-zero flow to trigger sediment transport to areas of stable deposition.

Each of these may be equally plausible in a given situation, and other evidence usually is necessary to resolve the nature of the circulation event.

Perhaps the most convincing evidence for intensification of abyssal circulation is the development of regional deep-sea unconformities, especially along basin margins where currents are geologically most effective. These unconformities may or may not be accompanied by hiatuses in the central parts of ocean basins.

Additional evidence for changes in abyssal circulation also is found in shifts in lithology (e.g. texture and composition) and



60°
55°
50°
45°
40°
35°
30°
25°
20°
15°
10°
5°
0°
407
408
409
114
352
336
ICELAND-FAEROE RIDGE
REYKJANES RIDGE
FAEROE-SHETLAND CHANNEL
ERIK DRIFT
GARDAR DRIFT
HATTON DRIFT
HATTON BANK
ROCKALL BANK
FENI DRIFT
GLORIA DRIFT
CHARLIE GIBBS F.Z.
AZORES-BISCAY RISE
113
115
116
117
403
404
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111
410
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Figure 1. Bathymetric location map of the northern North Atlantic with locations of DSDP sites and major sedimentary drifts. Contour interval 1000 m.

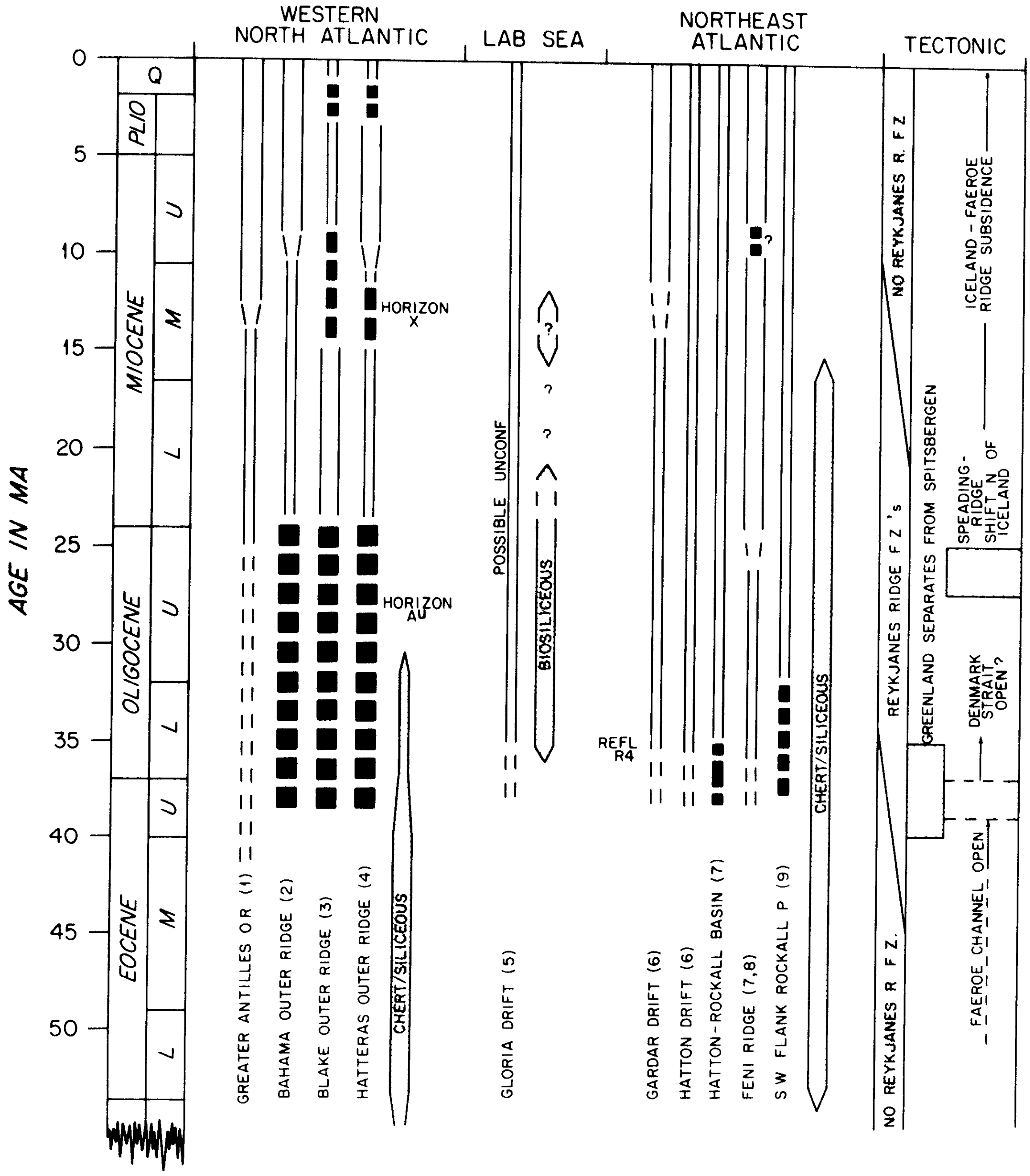


Figure 2. Summary of development of unconformities and current-controlled sedimentation in the North Atlantic, based on existing literature. Black bars - approximate time range of hiatuses, and open bars - current-controlled sedimentation (wider during major development of drifts); both dashed where time limits uncertain. Distribution of biosiliceous sediments based on DSDP data. References: 1) Tucholke and Ewing (1974); 2) Sheridan et al. (1974); 3) Ewing and Hollister (1972); 4) Tucholke and Laine (1981); 5) Egloff and Johnson (1975); 6) Ruddiman (1972); 7) Roberts (1975); 8) Jones et al. (1970); 9) Montadert and Roberts (1979). Tectonic events from Ruddiman (1972) and Talwani and Eldholm (1977).

benthic fauna (taxonomic and isotopic composition). These effects in themselves rarely are diagnostic of the nature of the circulation change, but they can provide important supporting evidence. For example, Berggren and Hollister (1974) suggested that increased early to middle Eocene deposition of biosiliceous sediments in the North Atlantic (Figure 2) correlates with the first formation, sinking, and upwelling of cold, deep waters in the northern North Atlantic. This correlation is reasonable, but it is possible that the biosiliceous sedimentation was stimulated by other mechanisms. Therefore, the lithologic change by itself is not definitive.

The purpose of this paper is to examine available evidence for the initiation and development of the abyssal circulation in the northern North Atlantic. Based upon evaluation of this evidence, we postulate a preliminary model to explain Eocene to Miocene bottom-water sources, including overflow across the Greenland-Scotland Ridge. Although the model is not definitive, it identifies several major unsolved and testable problems relating to the North Atlantic deep and bottom-water circulation.

PREVIOUS WORK

Over the past decade, there has been growing debate about when northern sources (North Atlantic, Norwegian-Greenland Sea, or Arctic Ocean) of bottom water began to form. As noted above, Berggren and Hollister (1974) suggested a correlation between the early Eocene deposition of biosiliceous sediments in the western North Atlantic basin and the formation of the first cold deep waters in the North Atlantic. They suggested that the source of bottom water was overflow across the Greenland-Scotland Ridge from the Norwegian-Greenland Sea. Jones et al. (1970) also felt that Norwegian Sea overflow affected sedimentation in Rockall Trough beginning in the late Eocene to early Oligocene, although the overflow was interpreted to influence other parts of the North Atlantic as early as the Paleocene. A widespread reflector in the northern North Atlantic variously termed "R" (Jones et al, 1970; Ruddiman, 1972), or "R4" (Roberts, 1975; Roberts et al., 1979), figures prominently in the interpretation of abyssal circulation history because it appears to divide current-influenced sedimentation above from largely pelagic sedimentation below (Figures 1, 2). This reflector is approximately late Eocene to early Oligocene in age (Ruddiman, 1972; Roberts, 1975; Laughton and Berggren, 1972; Roberts et al., 1979; Miller et al., in press). Vogt (1972) suggested that the apparent increase in abyssal circulation at this time correlated with bottom-water outflow from the Norwegian-Greenland Sea following initial subsidence of the Greenland-Scotland Ridge about 40 Ma.

The first clear evidence for strong bottom-current activity is the erosion of a prominent unconformity (Horizon A^u) along the continental rise of eastern North America. Tucholke (1979) and Tucholke and Mountain (1979) demonstrated that this erosion occurred at some point between the late Eocene and earliest Miocene. They suggested that the unconformity was caused by a precursor to the modern Western Boundary Undercurrent (Heezen et al., 1966) which presently contains a combination of Greenland-Scotland Ridge overflow water, Labrador Sea Water, and, in places, Antarctic Bottom Water.

Stratigraphic relations show that by the early Miocene, current-controlled deposition was well established in the Rockall Trough (Feni Drift; Roberts, 1975) and in the western North Atlantic (Blake Ridge; Ewing and Hollister, 1972; Tucholke and Mountain, 1979; Figure 2). By the middle to late Miocene, coherently deposited sediments also accumulated at significantly increased rates on such sedimentary drifts as the Greater Antilles, Bahama, Hatteras, and Gardar Ridges (Figure 2). Based on this observation and on the widespread distribution of a middle Miocene hiatus, Shor and Poore (1979) suggested that the deep circulation of the northeast Atlantic did not attain its present configuration until the early to middle Miocene; they related this to an inferred late Oligocene to early Miocene marine connection across the Iceland-Faeroe Ridge. Vogt (1972) also suggested a significant deepening of this marine connection in the middle to late Miocene (ca. 10-18 Ma) based upon a subsidence model; he related the presumed increase in Norwegian Sea overflow to a middle Miocene change in abyssal circulation noted by Ruddiman (1972) on the Gardar Ridge. Although there is some coherence in the sequence of events outlined above, the existing studies still show a general lack of agreement not only on the timing of initial development of the abyssal circulation, but also on the timing and influence of subsidence of the Greenland-Scotland Ridge.

Similar disagreements appear in interpretations of biostratigraphic, benthic faunal, and isotopic data. Schnitker (1979) inferred from the benthic faunal and isotopic records that a change in abyssal circulation occurred in the middle Miocene. He also (1980a, b) related this change to the first deep connection across the Greenland-Scotland Ridge as suggested by DSDP Leg 38 scientists (Talwani and Udintsev, 1976). Similarly, Blanc et al. (1980) interpreted carbon isotopic data to indicate that the production of oxygenated deep water in the North Atlantic began during the middle Miocene. In contrast, Miller et al. (in press) suggested that significant bottom-water formation began in the North Atlantic in the late Eocene to early Oligocene. Their interpretation was based on a marked shift from agglutinated to calcareous benthic foraminifera at DSDP Site 112 in the Labrador Sea.

Within the varying interpretations outlined above, there are generally well-defined patterns of current-influenced sedimentation (Figure 2). It is apparent that strong abyssal circulation eroded unconformities in the western North Atlantic and portions of the northern North Atlantic as early as the end of the Eocene. Current-influenced sedimentation, perhaps in rather unstable patterns, followed during the Oligocene to early Miocene, and rapid stable accretion of many sediment drifts was established by the middle to late Miocene. In the ensuing discussions, we attempt to refine the interpretation of the significance of these generalized patterns.

NORTHERN NORTH ATLANTIC OBSERVATIONAL DATA

Lithofacies/Age Distribution - DSDP Boreholes

In Figures 3 to 6 we have summarized age-versus-lithofacies data for the DSDP boreholes that recovered significant pre-Pliocene sedimentary records in the northern North Atlantic. Age assignments are taken from the appropriate volumes of "Initial Reports of the Deep Sea Drilling Project" with the exceptions noted below.

At Site 117, drilled along the eastern edge of the Hatton-Rockall Basin, Berggren and Aubert (1976) revised the original biostratigraphic zonation by removing core 1 from the early Miocene and placing it in the late Oligocene (Figure 4). This greatly improves correlations between Sites 116 and 117 as suggested by the seismic stratigraphy (Figure 7). We also suggest a revision to the original zonation at Site 116, drilled 40 km west of Site 117. An intra-Oligocene hiatus was inferred between cores 22 and 23 on the basis of the assignment of core 22 to the Sphenolithus ciperoensis (NP25) Zone and cores 23 to 25 to the Erisonia obruta (NP21) Zone (Perch-Nielson, 1972). Bukry (1972), on the other hand, placed core 25 in the Helicosphaera reticulata (NP22) Zone. Roberts (1975) noted this discrepancy and used Bukry's assignment to suggest that sedimentation was slow but continuous in the Oligocene section. Examination of the planktonic foraminiferal data supports the contention of continuous sedimentation. In addition, Berggren (1972) noted the last appearances of Globigerina ampliapertura and Chiloguembelina in cores 24 and 23, respectively, but he assumed that Perch-Nielson's nannoplankton assignment was correct and therefore thought that the ranges of these taxa were truncated by an intra-Oligocene unconformity. However, if Bukry's assignment is used, then these last appearances may be taken as the G. ampliapertura (top of Zone P20) and Chiloguembelina (lower Zone P21) datums. Thus, cores 19 to 22 may represent the late Oligocene, cores 23 and 24 the middle Oligocene, and core 25 the early Oligocene. The average Oligocene sedimentation rate, if interpreted in this fashion, is continuous at 9.5 m/Ma (Figure 4).

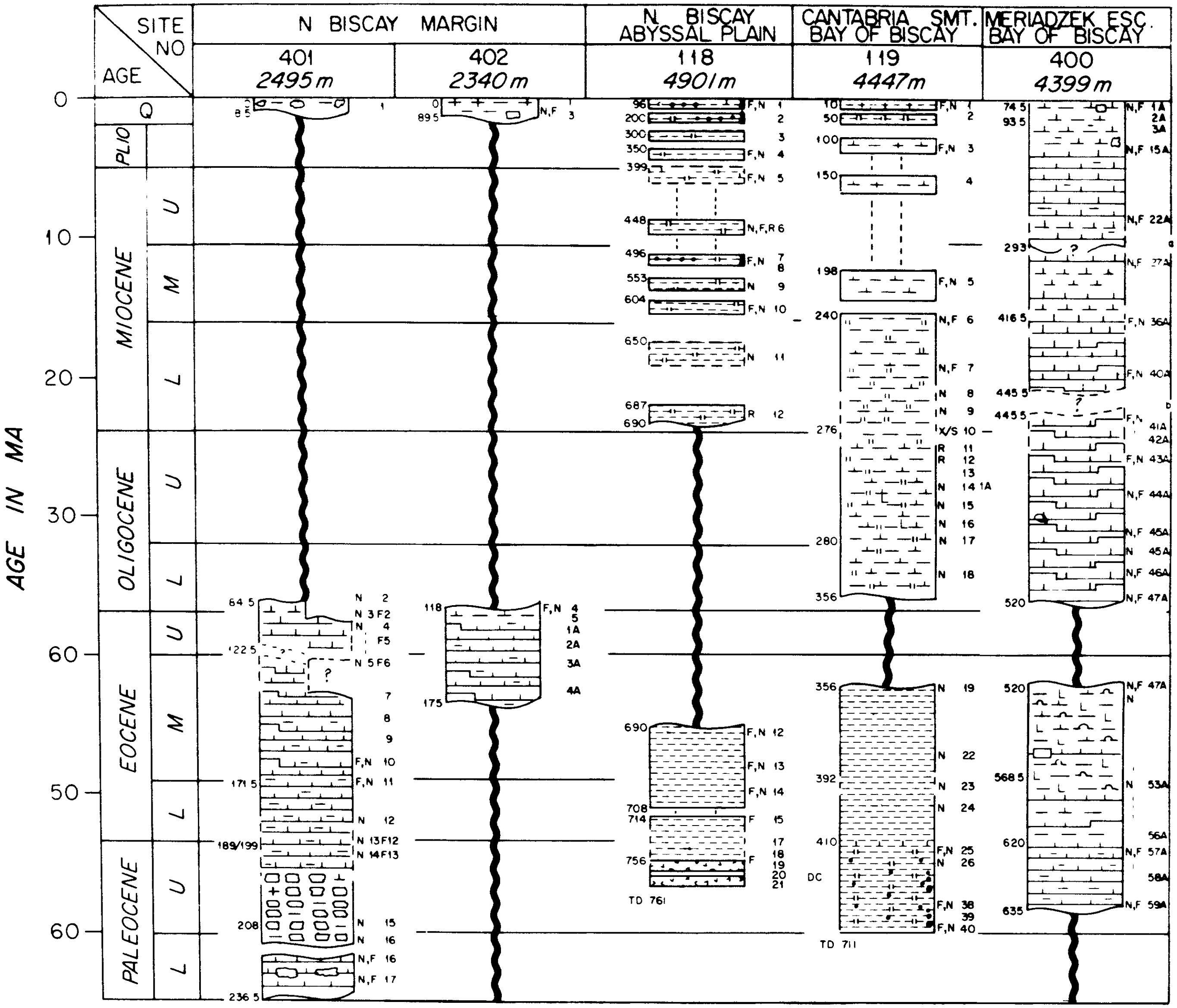
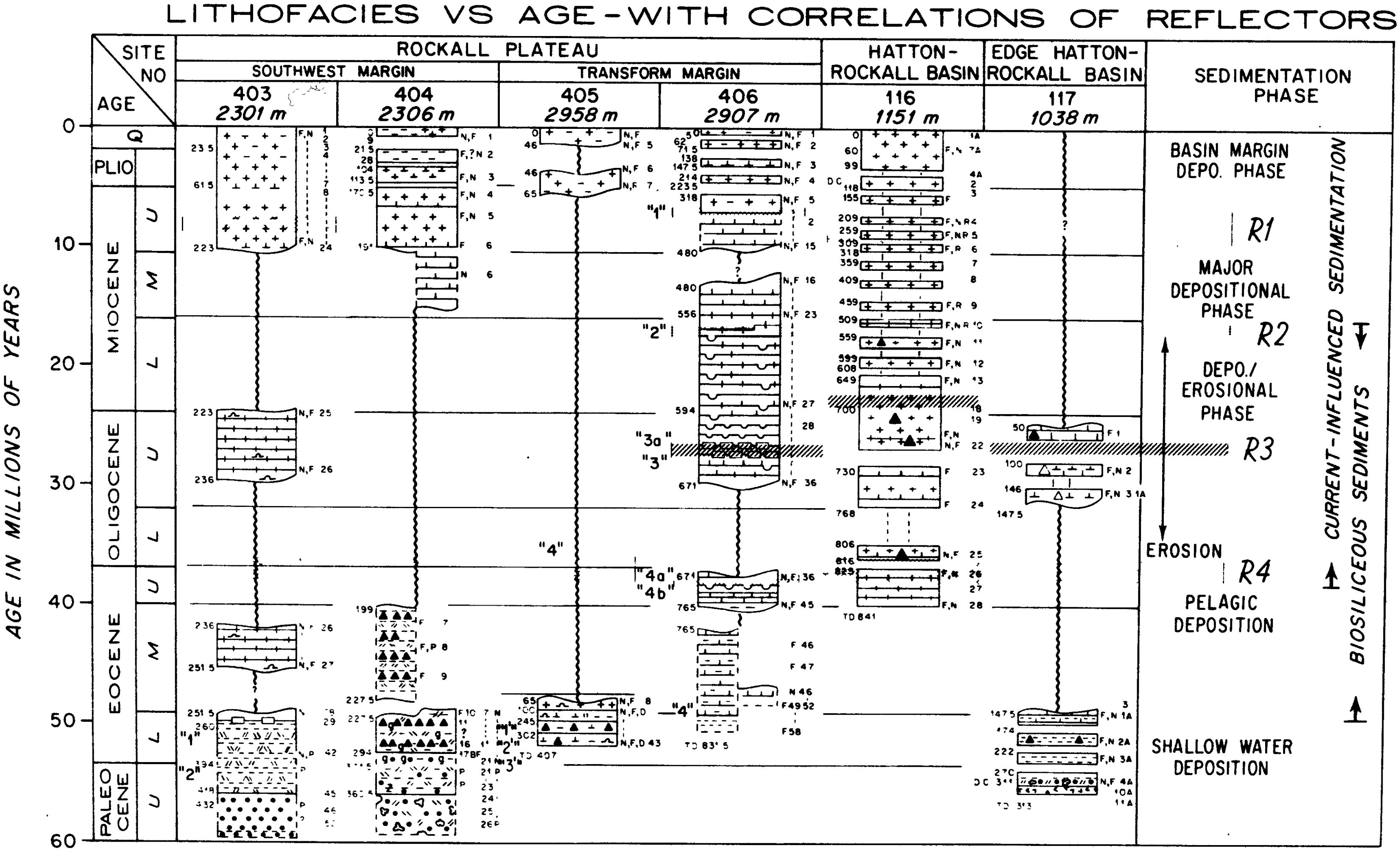


Figure 3. Age versus lithofacies for Tertiary sediments recovered at DSDP sites in the Bay of Biscay. Site locations in Figure 1. Lithologic symbols follow standard "Initial Reports" format. To left of columns are subbottom depths in meters (T.D.= total depth); to right are indicated core number and fossil group on which age is based (N = nannoplankton, F = foraminifers, R = radiolaria, P = palynmorphs, BF = benthic foraminifers, D = diatoms, X/S = extrapolated by sedimentation rate). Heavy wavy lines show hiatuses, dashed lines show inferred continuous sedimentation.



LITHOFACIES VS AGE-WITH CORRELATIONS OF REFLECTORS

Figure 4. Age versus lithofacies for DSDP sites in Rockall region with correlation of major reflectors. Explanation in Figure 3. Numbers in quotes at left of columns are reflector numbers used in individual site reports by Montadert and Roberts (1979).

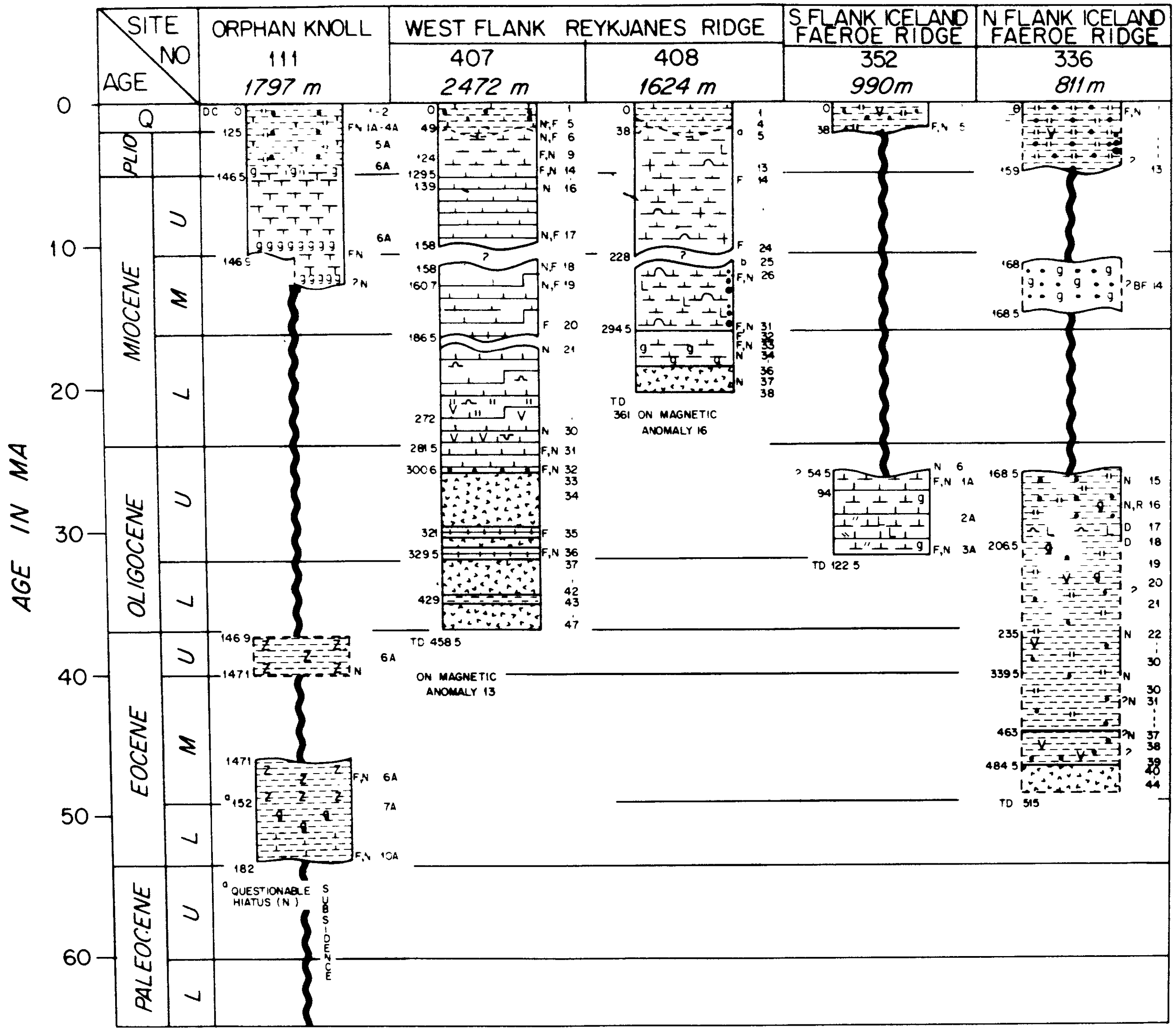


Figure 5. Age versus lithofacies for DSDP sites on Orphan Knoll and Reykjanes Ridge in western northern Atlantic and on Iceland-Faeroe Ridge. Explanation in Figure 3.

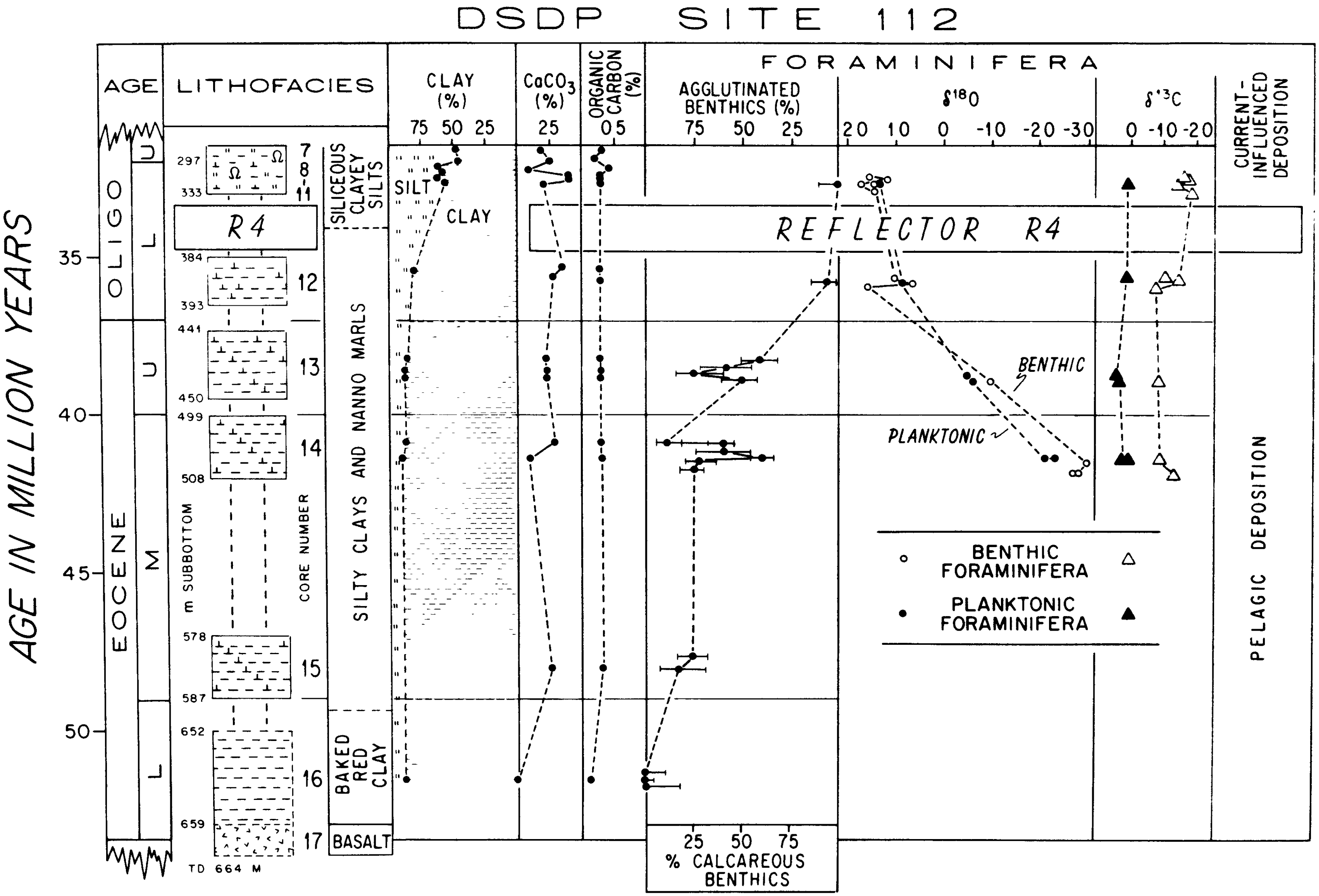


Figure 6. Summary of lithology and foraminiferal data at DSDP Site 112 in Gloria Drift, Labrador Sea. Isotopic values from Miller and Curry (in prep.). Explanation in text.

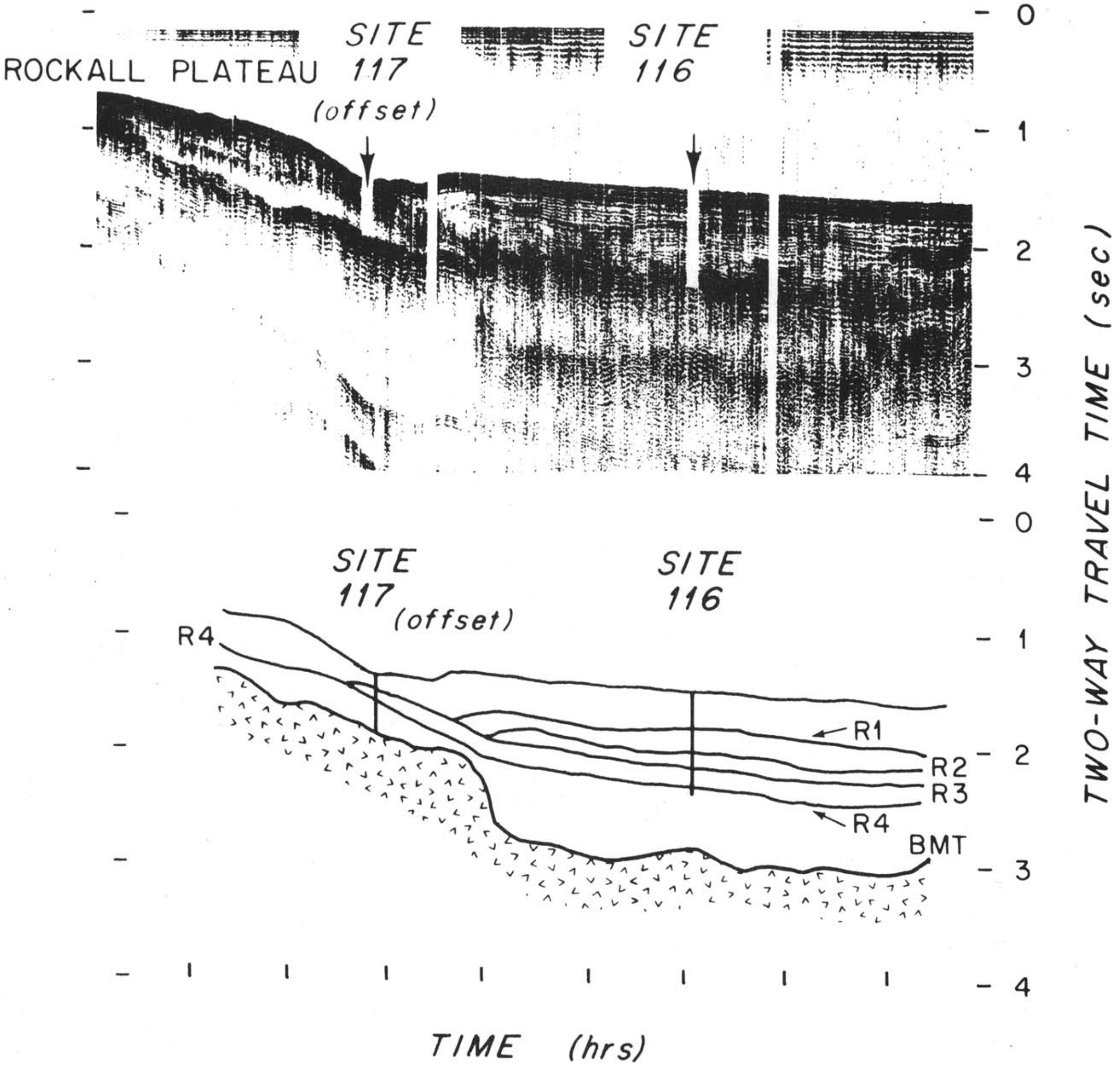


Figure 7. Seismic reflection profile and interpretation across DSDP Sites 116 and 117, Hatton-Rockall Basin.

Recognizing that many of the age assignments in Figures 3 to 6 still include uncertainties, there are nevertheless some striking correlations among the drillsites. The best defined is an unconformity at the Eocene/Oligocene boundary that occurs in every North Atlantic borehole except Sites 112 and 116. The lack of an unconformity in these two sites probably reflects their position away from basin margins. Site 112 is located in the central southern Labrador Sea on Gloria Drift, and Site 116, in contrast to adjacent Site 117, is located away from the margin of the Hatton-Rockall Basin (Figures 1, 7). All other sites are on or near areas where topographic boundary effects should have intensified the abyssal circulation. It is noteworthy that the hiatus appears to be mostly confined to the late Eocene (and earliest Oligocene?) at Sites 118, 119, and 400A; these sites are all in the Bay of Biscay, and they are the only drillsites at present water depths greater than 4000m. The missing time-slice common to all other drillsites (all less than 3000m water depth) is the latest Eocene to early late Oligocene. Thus, pronounced development of unconformities occurred throughout the late Eocene to early Oligocene North Atlantic, but erosion or non-deposition in the deep basin may have slightly preceded erosion at shallower depths.

A second unconformity occurs near the end of the middle Miocene in most drillsites, the notable exceptions being Site 116 in the Hatton-Rockall Basin, and the sites below 4000m water depth in the Bay of Biscay (Sites 118, 119, 400). The hiatus persists back into the middle to early Miocene at most sites, apparently reflecting the degree of sediment removal or the length of a period of non-deposition during this event. In contrast, the upper Miocene to Holocene sedimentary record is reasonably continuous at many sites. We interpret this distribution of hiatuses to indicate that a significant pulse of erosion by bottom currents occurred in the middle Miocene (see also Shor and Poore, 1979). However, less significant and/or local erosional episodes at other times are not precluded.

Lithofacies in the drillsites also show several broad, roughly chronostratigraphic changes (Figures 3-6). The first significant biosiliceous sedimentation occurs in the lower Eocene (e.g., Sites 117, 404, 405; Rockall Plateau region) and cherts also are developed in these largely non-calcareous sediments even though their present depth of burial generally is less than 300m. Biosiliceous detritus occurs patchily throughout the overlying Eocene to early Miocene sedimentary record in most of the northern North Atlantic. The only significant exception occurs in the Labrador Sea (Site 112) where abundant biogenic silica is confined to the Oligocene to Miocene section.

Excepting the siliceous/clayey sections noted above, calcareous sediments occur throughout most of the Tertiary and Quaternary record on the shallow Rockall Plateau. Elsewhere the occurrence of carbonates also tends to correlate with depth of the drillsite. Site 112, for example, has mostly low-carbonate (about 25%) sediments except in the Oligocene where a few values reach 40%. Site 118 in the Bay of Biscay is the deepest in the northern North Atlantic (4901 m) and has low carbonate throughout. Except for the Rockall drillsites, the Quaternary (and often the Pliocene) sedimentary record is largely non-calcareous.

Seismic Stratigraphy

The earliest clear effect of abyssal currents on the sedimentary record of the northern North Atlantic is in the form of sediment drifts (Figures 1, 2) overlying a subhorizontal sedimentary reflector, R4 (Roberts, 1975; = reflector R of Jones et al., 1970, and Ruddiman, 1972). Differential thickening of sediment that is typical of current-controlled deposition occurs above this horizon in the basins east of the mid-ocean ridge, whereas deposition conformable with basement and typical of pelagic accumulation occurs in most of the pre-R4 section (Figures 2, 7; Roberts, 1975; Ruddiman, 1972). West of the mid-ocean ridge and within Gloria Drift in the Labrador Sea, Egloff and Johnson (1975) suggest that "crenulation" of a strong reflector apparently correlative to R4 indicates current-control during or even preceding formation of this surface. However, it is not clear that the crenulation is not a seismic focussing effect created by the overlying intrasediment bedforms.

Reflector R4 is a continuous, interbasinal horizon which can be traced throughout the Rockall region (Roberts, 1975), the Iceland Basin (Ruddiman, 1972), and the Labrador Sea (Egloff and Johnson, 1975). The change in sedimentary regime associated with this reflector therefore must have a regional cause. Like Horizon A^u in the western North Atlantic, reflector R4 may truncate deeper horizons (Roberts, 1975), but the exact relation of reflector R4 to the Horizon-A complex in the western North Atlantic has not been established. The lack of a significant horizon of similar stratigraphic position in the South Atlantic indicates that the paleoceanographic event associated with reflector R4 and Horizon A^u had a North Atlantic genesis. Jones et al. (1970) determined that reflector R4 is pre-late Oligocene in age, and more recent reflector/borehole correlations indicate that it dates to the latest Eocene to early Oligocene (Figures 4, 6; Laughton and Berggren, 1972; Roberts, 1975; Montadert and Roberts, 1979; Miller et al., in press).

Ruddiman (1972) discussed a relatively continuous intermediate reflector (IR) above reflector R4 in the southwestern Iceland Basin. He dated the IR as early Miocene (17 Ma or younger, based on the age of pinchout on oceanic crust) and attributed it to an intensification of abyssal circulation, clearly differentiating this intensification from the initiation of current-controlled deposition at the level of reflector R4. Both Shor and Poore (1979) and Schnitker (1980b) interpreted Ruddiman to mean that current-influenced deposition began in the northern North Atlantic in the early to middle Miocene. However, differential thickening noted in the R4 to IR section (Ruddiman, 1972) clearly documents the initiation of current-controlled sedimentation at reflector R4.

This post-R4 current control also is suggested by the work of Roberts (1975) in the Rockall area. He discussed several intermediate horizons above reflector R4 which he termed 1 through 3 and which he noted were laterally and vertically impersistent. He also noted that internal unconformities present in the reflector 3 to R4 seismic interval suggest that the interval is unconformable with both reflectors 3 and R4. We have identified three horizons in the Iceland Basin/Rockall region above reflector R4 which we term reflectors R1 to R3, corresponding to reflectors 1-3 of Roberts (1975). Reflector R2 also corresponds to the intermediate reflector (IR) of Ruddiman (1972).

We have traced both R4 and the overlying reflectors and tied them to boreholes using seismic reflection data of Lamont-Doherty Geological Observatory and Woods Hole Oceanographic Institution (cruises V28-04, V27-06, C 09-13, and KN51), Challenger Legs 12 and 48, and previously published profiles and picks (Roberts, 1975; Montadert and Roberts, 1979). A convenient method for initial determination of subbottom depth of reflectors is to use velocity-regression equations derived from wide-angle sonobuoy measurements. These equations are of the form $V = Vo + KT$, where Vo is the initial velocity, K is the acceleration term, and T is half travel-time in seconds. This can be integrated to determine depth, $H = VoT + KT^2/2$. We used constants derived by Houtz (1980, and unpublished) for the Rockall area, and then modified the depth determination if correlation of the reflector to a nearby lithologic break could be improved, or if auxiliary velocity/density data from the sediments or borehole were available and justified a modification. In either case we restricted the resulting interval velocities to geologically reasonable values. In order to clarify some previously published correlations and to document our new correlations, distribution of reflectors at several drillsites is discussed below.

In the Labrador Sea near Site 112, reflector R4 lies at 0.41 seconds subbottom; Laughton and Berggren (1972) originally correlated the reflector with a hard layer encountered in core 10 (upper lower Oligocene; 315m subbottom). More recently Miller et al. (in press) correlated reflector R4 to the interval between cores 11 and 12 (333-384m) in the lower lower Oligocene section. This yields a more reasonable sediment interval velocity of 1.62 to 1.87 km/s. If a velocity as high as 1.95 km/s is allowed, reflector R4 lies very nearly at the Eocene/Oligocene boundary. The cause of reflector R4 in Site 112 is not clear, although it may represent the upward transition from indurated clays to siltier biosiliceous clays (Figure 6).

At Site 117 on the margin of Hatton-Rockall Basin, reflector R4 lies at 0.17 seconds and reflector R3 at 0.09 seconds subbottom. Reflectors R1 and R2 pinch out between Sites 117 and 116 (Figures

4,7). Using the velocity-regression equation with Vo=1.61 and K=2.28, reflector R4 lies at 147m, which correlates with the unconformity between the early Eocene and late Oligocene. Reflector R3 is close to 75m subbottom, correlating with an uncored interval between cores 1 and 2 (late Oligocene cherty limestones and oozes).

At Site 116, farther from the margin of the basin, four major sedimentary reflectors are present (Figs. 4, 7). Here, reflector R4 was first correlated with increased lithification (oozes to chalks) at 700m subbottom, corresponding to a late Oligocene to early Miocene age (Laughton and Berggren, 1972). Roberts (1975) recorrelated this horizon to the boundary between upper Eocene oozes/chalks and lower Oligocene cherts. At Site 116 reflector R4 lies at 0.79 to 0.81 seconds subbottom, reflector R3 at 0.70 seconds, reflector R2 at 0.60 seconds, and reflector R1 at 0.31 seconds. Although the constants in Houtz' velocity-regression equation yield good correlations of reflectors to lithostratigraphic breaks in the thin sedimentary section at Site 117, this is not the case for the thicker section at Site 116. The high acceleration term (2.28 km/s^2) for the Rockall region results in the placement of horizons below 0.3 to 0.4 seconds at anomalously great depths. We adjusted the acceleration term downward to 2.09 km/s^2 based on the sonic logs of the very similar section at Site 406 (south flank Rockall Plateau); this is within the standard error of estimate in the measurement of the acceleration term. As a result, depths computed using either acceleration term are very similar above 0.3 seconds (less than 8m difference) but diverge significantly at greater reflection times (71m difference at 0.8 seconds). The computed depth of reflector R4 then is 800-825m (late Eocene to early Oligocene) in agreement with Roberts' (1975) interpretation. Reflector R3 falls at 692m and correlates with the latest Oligocene upward change from cherty chalks to oozes, reflector R2 at 577m matches the top of early Miocene siliceous sediments, and reflector R1 at 284m falls within calcareous oozes in the lower upper Miocene section. This chronostratigraphic placement of horizons at Site 116 correlates well with their placement at Site 406 (Figure 4) where depths were estimated using synthetic seismograms computed from sonic logs (Montadert and Roberts, 1979). It also agrees with the late early Miocene age of reflector R2 (or IR) suggested by Ruddiman (1972).

At Site 406 on the southwest edge of Rockall Plateau, nine significant sedimentary reflectors were noted in multichannel seismic data obtained for DSDP Leg 48 drilling (Montadert and Roberts, 1979). Logging at Site 406 provides good control on the placement of seismic horizons, and our study of single-channel seismic data in this area agrees with these results. Our reflector R4 lies at 0.72 seconds subbottom, reflector R3 at 0.64 seconds, R2 at 0.60 seconds, and reflector R1 at 0.39 seconds. Using velocity-regression constants as for Site 116, the reflectors are at 715m, 622m, 577m, and 354m, respectively (Fig. 3). Reflector R4 is an intra-upper Eocene horizon; it occurs within or near an upward change from marly limestones to diatomites and chalks and is just below the unconformity at the Eocene/Oligocene boundary. This correlation is similar to that of Roberts et al. (1979) who matched R4 with the unconformity. Reflector R3 is an intra-upper Oligocene horizon, reflector R2 is an upper lower Miocene horizon at the upper boundary of silica-rich sediments, and reflector R1 falls within upper Miocene calcareous chalks.

In Site 406, our reflectors R4 to R1 as noted above correspond to reflectors "4a"/"4b", "3a"/3", "2", and "1" identified by Montadert and Roberts (1979) in their higher-resolution multichannel data. Their reflector "4", which corresponds to an unconformity separating upper Eocene from middle Eocene sediments, lies below our reflector R4. Montadert and Roberts (1979) computed synthetic seismograms using sonic logs and they predicted reflectors at 0.72/0.74, 0.64/0.66, 0.61, and 0.39 seconds which corresponded to subbottom depths of 690/710, 600/640, 570, and 350 m. These depths are in good agreement with the depths we obtained using the velocity-regression equation.

Correlation of reflectors is difficult at adjacent Site 405 because poor hole conditions prevented logging above 240m subbottom, and the post-R4 sedimentary record is too thin to rely on the velocity-regression equation. Reflectors R1 to R3 at Site 406 pinch out on R4 where traced to Site 405. A prominent horizon at Site 405 was correlated with the unconformity between upper Miocene and lower Eocene sediments (Montadert and Roberts, 1979) and is thought to be reflector R4. It occurs at less than 0.1 second subbottom and therefore corresponds approximately to the unconformity at 65m (Figure 4).

At Site 403, four reflectors were noted by Montadert and Roberts (1979). The uppermost of these ("1") occurred at 0.3 seconds subbottom and correlated with the top of an early Eocene tuff. We observe two significant reflectors above reflector "1". The first, occurring at 0.17 to 0.19 seconds (144-162m subbottom), falls in the upper Miocene section and it may correlate with reflector R1. The second, at 0.25 to 0.27 seconds (218 to 236 m subbottom), is due to an upward decrease in velocity (shown on the sonic log) associated with the unconformity separating upper Oligocene from middle Eocene strata. This horizon probably correlates with reflector R4. This interpretation agrees with that of Roberts et al. (1979) who correlated reflector R4 with the unconformity at Site 403.

Reflector R4 at Site 403 can be traced to a position 0.23 to 0.25 seconds subbottom (199-218m) at Site 404. This also correlates with the unconformity separating Miocene from Eocene strata as suggested by Montadert and Roberts (1979).

The foregoing correlations suggest that we can characterize each of the major reflectors by rather specific lithologic and age limits. Reflector R4 in the Rockall area shows a strong correlation to the widely distributed unconformity straddling the Eocene-Oligocene boundary (Figures 3-6; see also Roberts et al., 1979), except possibly at Site 406 where it may underlie the unconformity by up to 45m. Away from basin margins the unconformity is not present (e.g. Site 116; Roberts, 1975), and in those regions reflector R4 appears to correlate with the uppermost Eocene to lower Oligocene section. Similarly, the apparently correlative reflector in the Labrador Basin, which we also term R4, dates to the lower part of the lower Oligocene in an apparently continuously deposited hemipelagic clay sequence at Site 112. A reflector similar to R4 in stratigraphic position has been noted in the Bay of Biscay, where it is thought to correlate with the middle Eocene to early Oligocene unconformity (Montadert and Roberts, 1979); however, reflector R4 has not been traced seismically from the Rockall region into the Bay of Biscay.

Reflector R3 is observed only in the Rockall Plateau region where it dates to the latest Oligocene to earliest Miocene. Because the R3-R4 seismic interval commonly has high-amplitude, discontinuous reflectors it often obscures the underlying reflector R4; although reflector R3 has no consistent lithologic correlation, this seismic signature suggests it may cap a "lower" Oligocene sequence of (cherty) chalks variably altered by diagenesis (Roberts, 1975). In addition, the R3-R4 interval commonly wedges out toward basin margins, contorted internal reflectors often appear, and apparent sediment waves occur locally within the Gardar Drift. We interpret this as evidence for continued control of sedimentation by abyssal currents.

Reflector R2 (IR of Ruddiman, 1972, in the Gardar Drift) dates to the latest early Miocene in the reasonably continuous sedimentary sections in the Gardar Drift and at Sites 116 and 406 on Rockall Plateau. Elsewhere, excepting the Bay of Biscay and possibly Site 408 on the west flank of the Reykjanes Ridge, a hiatus is present in this time interval. It is thus possible that R2 is another widespread unconformity, and in many places it may have re-excavated R4. There is some support for this observation in the apparently unconformable attitude of Ruddiman's IR (R2) within the Gardar Drift. Reflector R2 also correlates approximately to the level below which siliceous debris is a significant sedimentary component. The R2-R3 interval lacks the discontinuous, high-amplitude reflectors found in the underlying seismic interval on Rockall, but it retains many of the convoluted reflectors and pinchouts that suggest continued current-controlled sedimentation and erosion.

Reflector R1 is important only in the Rockall Plateau region where it dates to the late Miocene and may represent a lithification boundary (e.g., chalk/ooze) in the calcareous sediments. It is noteworthy that the widespread upper middle Miocene unconformity documented by drilling falls between reflectors R2 and R1 and is not marked by a significant reflector.

The seismic interval between reflector R2 and the seafloor shows the most coherent pattern of current-controlled deposition in the northern North Atlantic, especially in the form of prominent development of sediment drifts and associated sediment waves. Along at least some basin margins (e.g., Sites 403-405) this depositional phase lagged until the late Miocene (about R1 time).

Summary of Evidence for Abyssal Circulation

We can summarize the evidence from DSDP borehole age/lithofacies data and from seismic reflection data in the following manner. Very near the Eocene/Oligocene boundary a widespread unconformity developed, apparently because of erosion by a newly developed system of vigorous abyssal boundary currents. Areas away from basin margins (e.g. Sites 112 and 116) were not eroded but did experience current-controlled deposition. The erosion may have occurred first (latest Eocene) in the deeper parts of the basin (i.e. 4000m, Bay of Biscay), and later (early Oligocene) it affected shallower areas that now are less than 3000m deep. A widespread early Oligocene hiatus also is developed on the continental shelves around the North Atlantic (e.g., northeast U.S. margin - Olsson et al., 1980; Canadian margin - Gradstein and Srivastava, 1980; Europe - Pomerol, 1973; Gulf of Mexico - Murray, 1961). However, there is no known relation between these shallow, sea level-induced hiatuses and the current-induced erosion in the deep basin. It also is not clear that the timing of the two events is exactly the same.

The late Eocene to early Oligocene timing for development of the abyssal circulation is supported indirectly by a major faunal change at the end of the Eocene in the deep Labrador Sea (DSDP Site 112) where predominantely agglutinated Eocene benthic foraminiferal assemblages were replaced by an Oligocene calcareous fauna (Figure 6; Miller et al., in press). It is possible that reflector R4 at Site 112 postdates the faunal exit by up to 2 Ma, but Miller et al. suggest that the changes in fauna and depositional conditions both are expressions of changes in hydrographic regime and development of vigorous abyssal circulation. Miller and Curry (in prep.; Figure 6) also note that a 4^{o}/oo shift in oxygen isotopic composition of benthic foraminifera coincides with the faunal change. Some of this isotopic signal may be due to diagenesis and perhaps to ice volume, but part of the shift probably represents a temperature drop which in turn can be attributed to the initiation of northern sources of bottom water. Similar supporting evidence is found in the Bay of Biscay (Site 400A) where a peak in benthic faunal turnover also occurs between the middle Eocene and the early Oligocene (Schnitker, 1979). An associated shift in oxygen isotopes also occurs in this region (Site 400A, Vergnaud-Grazzini et al., 1978, 1979; Site 119, Miller and Curry, in prep.); at Site 119 this shift represents a temperature drop of at least 2°C but less than 6°C.

Following the initial erosional phase associated with the developing abyssal circulation, strong abyssal currents continued to control sedimentation patterns throughout the North Atlantic. This is clearly demonstrated between reflectors R4 and R2 by the development of sediment waves and convoluted beds in the Gloria and Gardar Drifts, and possibly in the Rockall Basin, and larger-scale differential sedimentation patterns with pinchouts and possible erosional truncations throughout the basins east of the Mid-Atlantic Ridge (Ruddiman, 1972; Egloff and Johnson, 1975; Roberts, 1975). Except on the Gloria Drift in the Labrador Sea and locally on the Gardar Drift, there was little coherent development of large-scale bedforms, and unconformities may be widely but rather randomly developed. A coherent erosional pulse may have occurred at the level of reflector R2 in the late early Miocene, and another pulse almost certainly occurred in the late middle Miocene above R2; however, neither had much, if any, geologic effect in the Bay of Biscay. Coherently deposited sediment drifts and waves in the northeastern North Atlantic first appear above reflector R2. Because of the well-developed unconformities and associated effects of strong currents below reflector R2, we interpret this observation as a general weakening and stabilization of abyssal flow; the event also appears to mark the onset of near-modern conditions of current-controlled deposition (e.g., Shor and Poore, 1979).

POTENTIAL BOTTOM-WATER SOURCES

There are three possible bottom-water sources in the northern Atlantic that may have contributed to the late Eocene to early Oligocene development of strong abyssal circulation. These are Greenland-Scotland Ridge overflow, Arctic connection via Baffin Bay/Davis Straits, and bottom-water formation in the northern Atlantic south of the Greenland-Scotland Ridge.

Greenland-Scotland Ridge Overflow

Considerable debate, based on a variety of differing lines of evidence, exists as to when shallow and deep water connections were established across the Greenland-Scotland Ridge (Figure 8). Müller (1976) noted similarities in early Eocene nannoflora from the Norwegian-Greenland Sea and the North Atlantic. It is not clear whether this was a result of marine connection across the Greenland-Scotland Ridge or of migrations through epicontinental seas such as the North Sea, for Müller also noted a gradual floral differentiation extending from the North Atlantic into the Norwegian-Greenland Sea. This suggests that there was some paleogeographic restriction between the two basins because the early Eocene was a time of very reduced thermal gradients (Haq, in press). The presence of a continuous waterway from the North Atlantic to the Arctic by the middle Eocene also is supported by terrestrial faunal differentiation between Europe and North America/Greenland (McKenna, 1972; 1975). McKenna (1975; this volume) attributed this differention to the breaking of the "Thulean" land bridge between Greenland and Scotland. The continued presence of a waterway across the Greenland-Scotland Ridge into the early to middle Oligocene is supported by similar nannoflora north and south of the ridge (Haq and Lohmann, 1976; Müller, 1976).

The timing of a sea-level connection across the Greenland-Scotland Ridge (as opposed to around the ridge) has only been estimated by subsidence models. Vogt (1972) used a subsidence model based on a 60 Ma age for initial spreading to suggest that the first surface-water connection between Greenland and Scotland occurred in the late Eocene (37-40 Ma); he also estimated that no "deep" (several hundred meters) connection existed prior to the middle Miocene. Subsequent work by Talwani and Eldholm (1977) has shown that spreading began in this region during anomaly 24 time (53 Ma, time scale of Hailwood et al., 1979; 56.5 Ma, time scale of Hardenbol and Berggren, 1978). Vogt's original estimate of subsidence of the Greenland-Scotland Ridge below sea level by 40 Ma (late Eocene) therefore may be too old by several million years; this revision would place the first surface water connection in the latest Eocene to earliest Oligocene. In contrast, Talwani and Udintsev (1976) estimated from drilling results at Site 336 and from other geologic inferences that subsidence below sea level of even the oldest crust did not occur until 25 Ma; they suggested this as the time that the Thulean land bridge was breached.

The above models both assume that the entire Greenland-Scotland Ridge is oceanic and therefore can be backtracked along a normal age-versus-depth curve for oceanic crust (Sclater et al., 1971; Berger and Winterer, 1974) starting with the initiation of seafloor spreading at anomaly 24 time. However, in the region of the Faeroe Islands and southward, there are complexities that suggest a distinctly different history. Most important is the observation that continental crust probably underlies the "Faeroe Islands Block" (Casten, 1973, Bott et al., 1974, 1976; Casten and Nielson, 1975) and almost certainly underlies Rockall Plateau (Montadert and Roberts, 1979). These continental blocks presumably were separated from the Shetland margin during Mesozoic rifting (e.g. Vogt et al., 1981) thereby forming the Rockall Trough (Roberts, 1975) and possibly the Faeroe-Shetland Channel. This supposition is supported by recent multichannel seismic observations of thick sedimentary sections below "acoustic basement" (Paleocene tuffs and basalts) in the vicinity of the Faeroe-Shetland Channel, Faeroe Bank Channel, and northern Rockall Trough (Figure 9; Ridd, this volume; Roberts, this volume). Subsidence of the Faeroe Channel system therefore could significantly predate the Paleocene spreading phase, and the Faeroe Channel system may have provided a seaway to the Norwegian-Greenland Sea in the Late Mesozoic to Early Cenozoic. The entire region was overprinted by massive Paleocene basalt flows, which for a period may have provided a land bridge for terrestrial faunal migration. By no later than 50 Ma, this land bridge was severed (McKenna, this volume). We speculate that 1) the Faeroe Channel system may have been part of an epicontinental sea in the pre-Anomaly 24 interval, 2) during the initiation of spreading in the Norwegian-Greenland Sea, Paleocene basalts formed the Thulean land bridge (probably the Wyville Thompson Ridge system), and 3) subsidence of this region below sea level broke this terrestial faunal connection and allowed migration of marine flora between the Atlantic and Norwegian-Greenland Sea by early Eocene time (Figure 8). This model presumes that the Paleocene igneous overprinting did not completely "reset" the rate of crustal subsidence in this region. If it had, the sill in Faeroe Bank Channel, now at approximately 800-900 m depth (Figures 10, 11), would initially have been nearly 1500m above sea level; it would not have subsided below sea level until about 30 Ma, which is 20 Ma after McKenna's postulated breach in the Thulean land bridge. Clearly, the age and tectonic history of the Faeroe Channel system and Wyville Thompson Ridge deserve detailed study.

There also is a great deal of uncertainty about the Paleogene configuration and depth of the ridge between Iceland and Greenland (Figures 10, 11). On the edge of the Greenland margin just north-east of the ridge, prograded sediments at least 1 to 2 seconds thick occur. The east Greenland margin is still not well surveyed by seismic data, but recent work by H.C. Larsen, B. Larson (both this volume), and Hinz and Schlüter (1978) indicates that thick prograding sediments have caused both narrowing and shoaling of the Denmark Straits. Accurate age assignments for these sediments and "backstripping" in conjunction with a subsidence model will be necessary to fully evaluate the possibility of a Paleogene marine passage in this area, but it presently seems possible that a significant passage did exist. The area also has an uncertain plate tectonic history. Talwani and Eldholm (1977) suggested that at anomaly 7 time (27Ma) a westward jump of the spreading center occurred on the Greenland-Scotland Ridge, from a central position on the Ridge to a position near the Greenland margin. This particular jump disagrees with magnetic-anomaly identifications on the Ridge by Vogt et al. (1980), but a jump just north of the Ridge still is required. The depth and morphology of any existing passage adjacent to Greenland therefore may have been substantially changed in the late Oligocene by this spreading-center jump and by the subsequent formation of the Iceland Plateau.

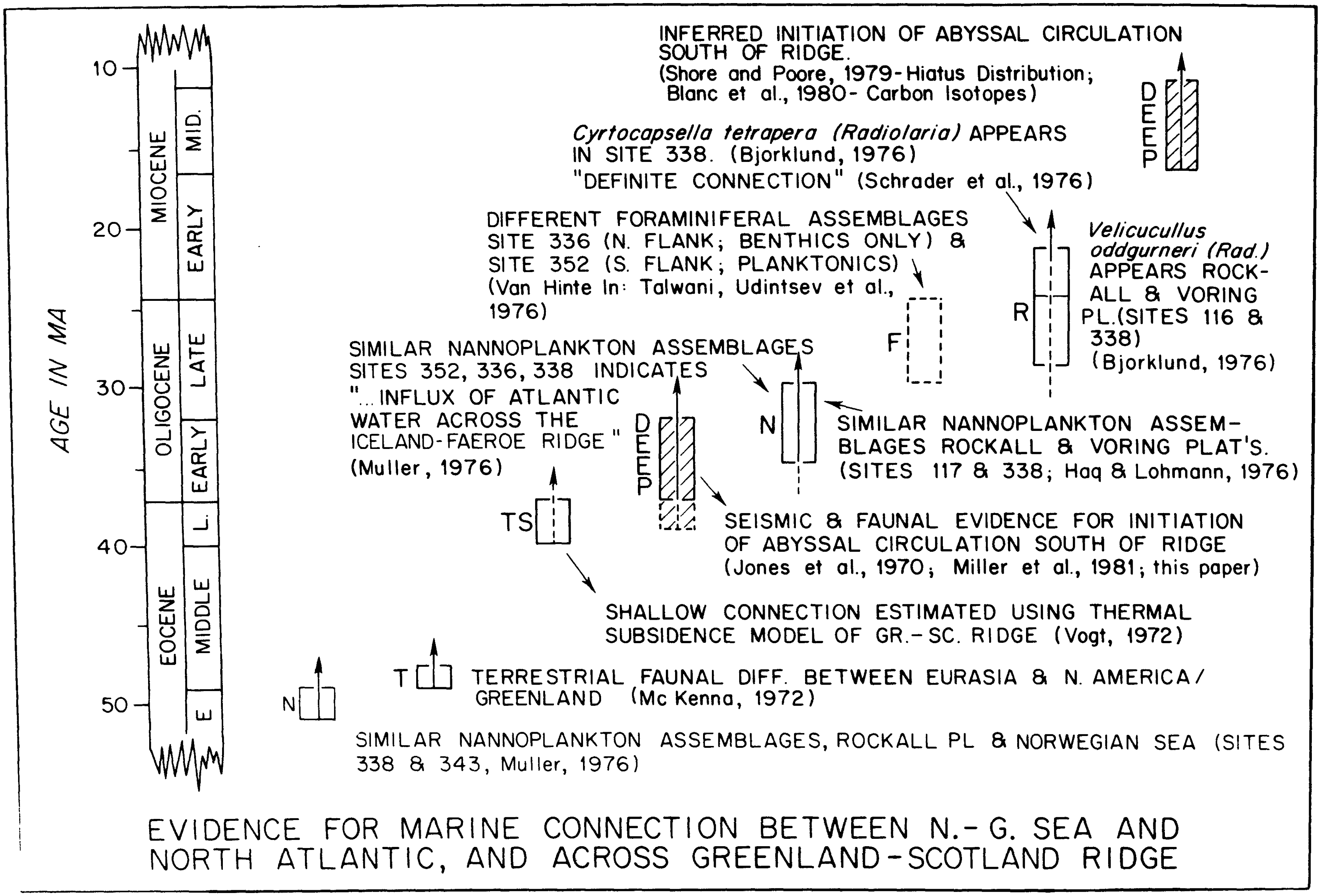


Figure 8. Summary of evidence for marine connection between North Atlantic and Norwegian-Greenland Sea, with reference to Greenland-Scotland Ridge.

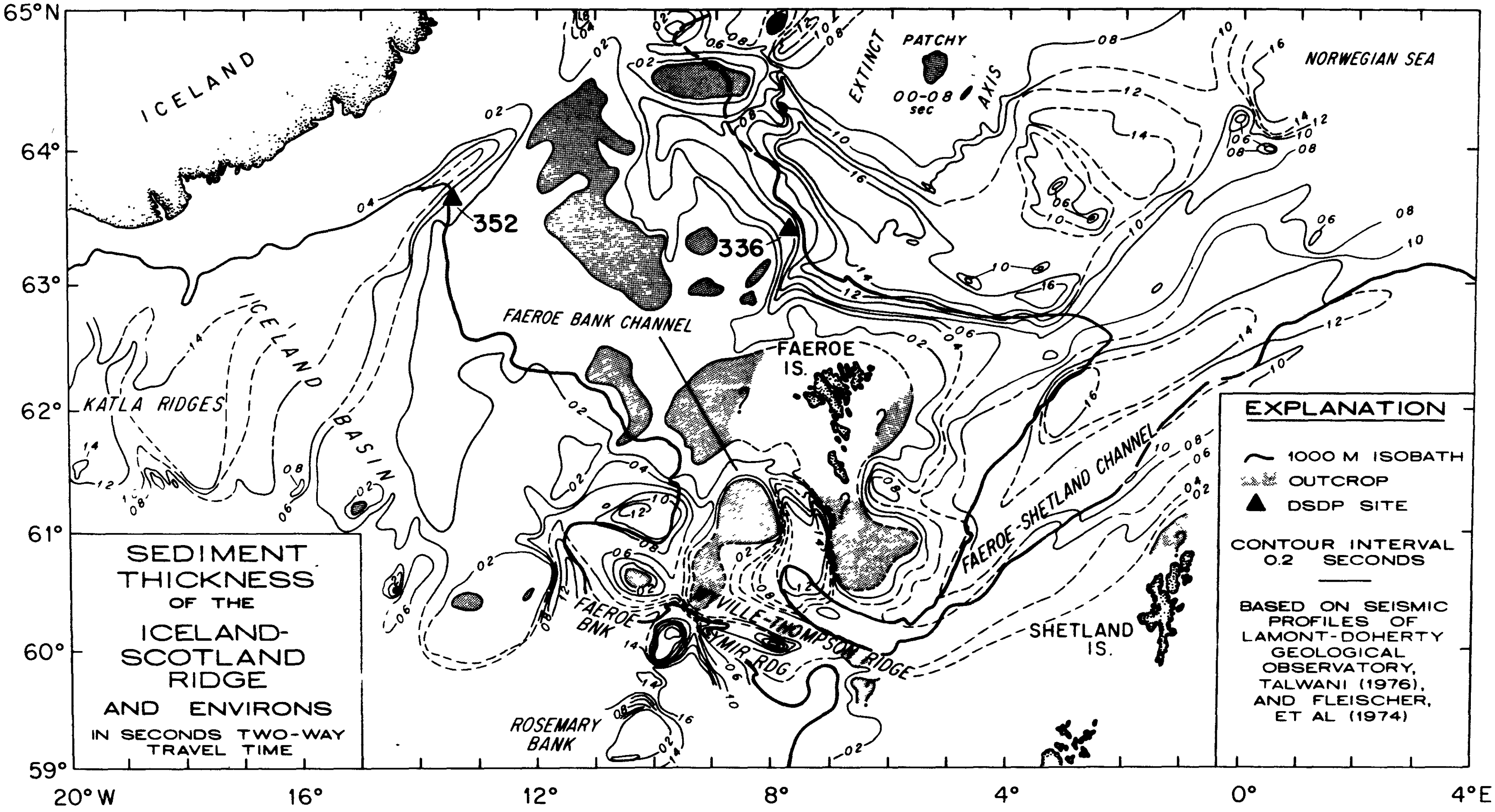

65°N
64°
63°
62°
61°
60°
59°
20° W
16°
12°
8°
4°
0°
4°E
ICELAND
ICELAND BASIN
KATLA RIDGES
352
336
FAEROE BANK CHANNEL
EXTINCT AXIS
PATCHY
00-08 sec
NORWEGIAN SEA
FAEROE IS.
FAEROE-SHETLAND CHANNEL
FAEROE BNK
WYVILLE-THOMPSON RIDGE
YMIR RDG
ROSEMARY BANK
SHETLAND IS.
SEDIMENT THICKNESS OF THE ICELAND-SCOTLAND RIDGE AND ENVIRONS IN SECONDS TWO-WAY TRAVEL TIME
EXPLANATION
1000 M ISOBATH
OUTCROP
DSDP SITE
CONTOUR INTERVAL 0.2 SECONDS
BASED ON SEISMIC PROFILES OF LAMONT-DOHERTY GEOLOGICAL OBSERVATORY, TALWANI (1976), AND FLEISCHER, ET AL (1974)

Figure 9. Sediment thickness above "acoustic basement" (including Paleocene basalts) on Iceland-Scotland Ridge, in seconds two-way travel time.

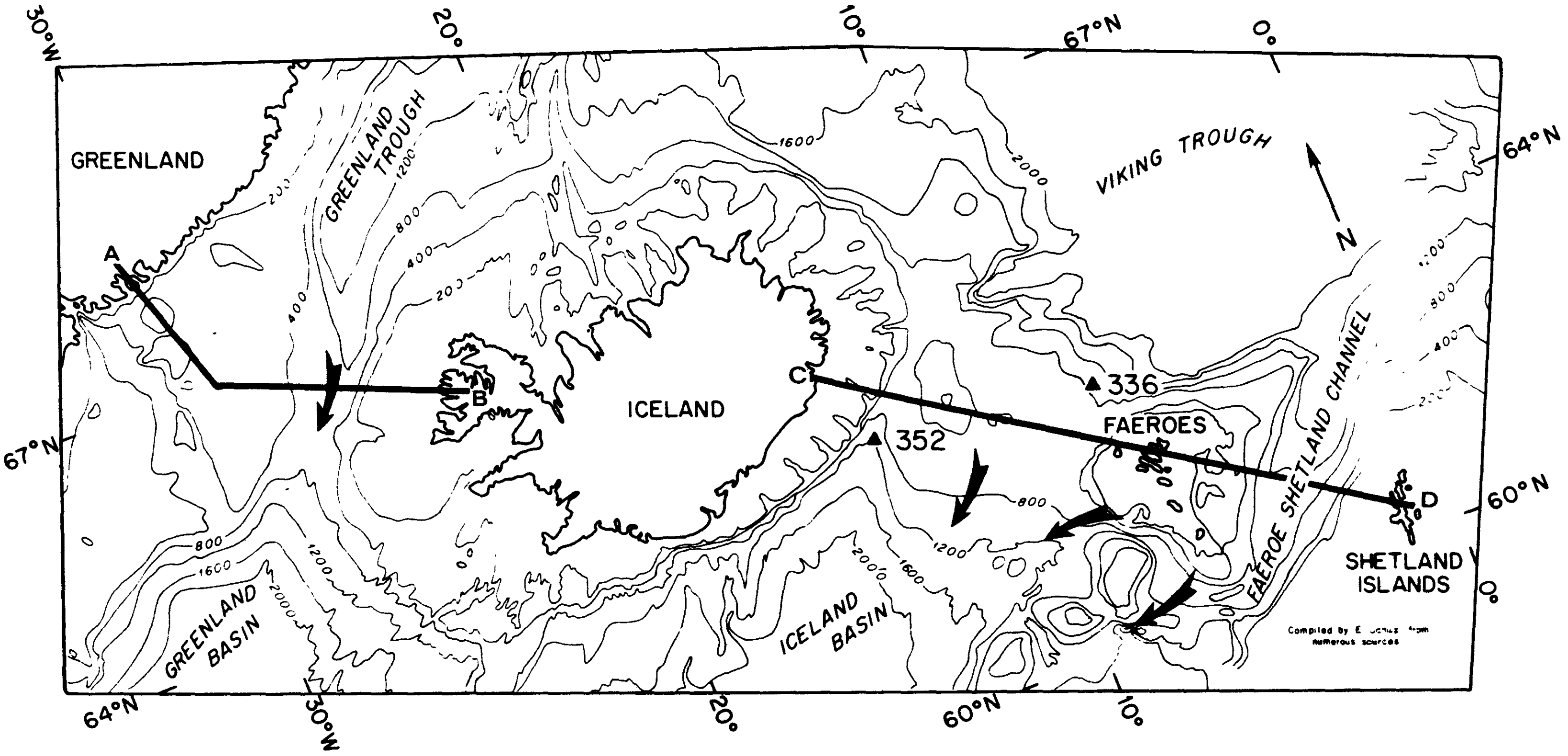

30°W
20°
10°
67°N
0°
64°N
GREENLAND
GREENLAND TROUGH
1200
800
400
200
1600
2000
VIKING TROUGH
N
A
B
C
D
ICELAND
336
352
FAEROES
FAEROE SHETLAND CHANNEL
67°N
60°N
SHETLAND ISLANDS
GREENLAND BASIN
ICELAND BASIN
Compiled by E. Uchupi from numerous sources
64°N
30°W
20°
60°N
10°
0°

Figure 10. Bathymetry of Greenland-Scotland Ridge, in meters (from Uchupi and Hays, unpublished data). Arrows indicate modern overflow routes.

All the above evidence for a Paleogene seaway between the North Atlantic and the Norwegian-Greenland Sea is in marked contrast with interpretation of DSDP Leg 38 drilling results (Talwani and Udintsev, 1976) which suggested isolation of the Norwegian-Greenland Sea from the North Atlantic during the late Oligocene (Sites 336, 352; Figure 10). This separation was invoked because of differences in the upper Oligocene benthic foraminiferal assemblages north and south of the Greenland-Scotland Ridge (van Hinte: in Talwani and Udintsev, 1976; Figure 8); however, this difference may be due simply to different paleobathymetry at the two sites (Berggren and Schnitker, this volume). The appearance of the radiolarian species *Velicucullus oddgurneri* (late Oligocene) and *Cyrtocapsella tetrapera* (early Miocene) in the Norwegian-Greenland Sea (Schrader et al., 1976) indicate a definite connection existed in the early Miocene. Because there also are faunal preservational problems in the Norwegian-Greenland Sea, the suggested late Oligocene isolation from the North Atlantic may be more apparent than real, or if real, it may have been a transient event.

In summary, surface-water connections between the northern Atlantic and the Norwegian-Greenland Sea probably existed more or less continuously from the early Eocene onward. Whether these connections were across the Greenland-Scotland Ridge will remain a matter of interpretation until subsidence history of the Ridge is better determined.

Baffin Bay - Davis Straits

Timing of potential circulation of cool, dense water from the Arctic into the northern Atlantic via Baffin Bay/Davis Straits is poorly constrained by both tectonic and paleobiogeographic data. Gradstein and Srivastava (1980) recently summarized this kind of data for the area and deduced that a seaway probably has linked the Arctic and Atlantic since the Late Cretaceous. They noted that faunal data indicate a northward incursion of warm-water masses in the early to middle Eocene, correlating with a climatic optimum; they also suggested that southward flow of cold surface water (Labrador Current) did not begin until at least the late Miocene. However, there is no data available to constrain the timing of possible southward flow of bottom water in deep off-shelf areas. Thus the possibility that cool polar waters entered the Paleogene abyssal Atlantic via this route cannot be ruled out.

Climatic Controls

The inferred initiation of bottom-water formation in the northern North Atlantic and/or Norwegian-Greenland Sea also may be related to or accentuated by the climatic history of this region. The Tertiary global climatic record shows that a major cooling

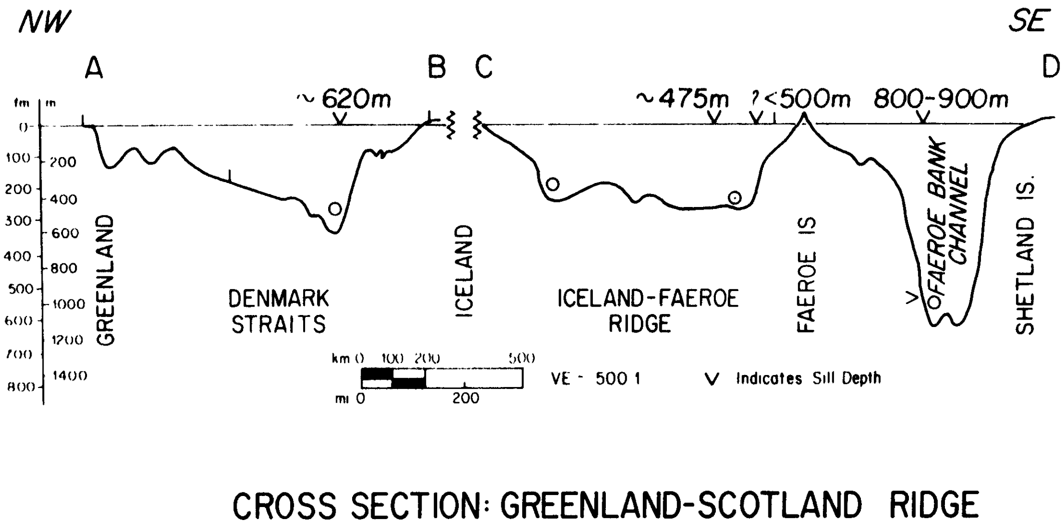

NW
SE
A
B
C
D
~620m
~475m
?<500m
800-900m
fm
m
0
100
200
300
400
500
600
700
800
200
400
600
800
1000
1200
1400
GREENLAND
DENMARK STRAITS
ICELAND
ICELAND-FAEROE RIDGE
FAEROE IS
FAEROE BANK CHANNEL
SHETLAND IS.
km 0 100 200 500
mi 0 200
VE - 500 1
V Indicates Sill Depth
CROSS SECTION: GREENLAND-SCOTLAND RIDGE

Figure 11. Bathymetric cross-section of the Greenland-Scotland Ridge (located in Figure 10) showing maximum present sill depths.

occurred in the middle Eocene to early Oligocene and resulted in an increase in the latitudinal thermal gradient. This cooling has been interpreted from the isotopic, planktonic faunal and floral, and terrestrial floral records (Table 1). A dramatic shift in oxygen isotopic composition of planktonic foraminifera near the Eocene/Oligocene boundary in the southern hemisphere is thought to reflect a major temperature drop at this time (Kennett and Shackleton, 1976). A similar shift in oxygen isotopes in benthic foraminifera has been interpreted as reflecting the initial formation of cold bottom water of southern origin (Kennett and Shackleton, 1976). However, the benthic isotopic shift in the North Atlantic may reflect formation of northern bottom-water sources (this study; Figure 6). It is also possible that both shifts may be attributable in part to ice volume and not to temperature (Matthews and Poore, 1980). In either case, the distribution of planktonic organisms clearly shows a significant cooling of North Atlantic surface waters in the middle Eocene to early Oligocene (Gradstein and Srivastava, 1980; Berggren, 1978; Haq et al., 1977, 1979).

In the modern ocean Norwegian Sea Overflow Water and Labrador Sea Water (which form the bottom and intermediate waters, respectively, in the North Atlantic) form by winter convection of high-salinity subtropical water that has been advected northward and cooled (Worthington, 1976; McCartney and Talley, in press). The source of this low latitude water is thought to be either the Gulf Stream/North Atlantic Current (Worthington, 1976) or Mediterranean Outflow Water (Reid, 1979). A similar situation may have developed in the late Eocene. At that time, the configuration of continents

TABLE 1. GLOBAL EVIDENCE OF MIDDLE EOCENE TO EARLY OLIGOCENE CLIMATIC COOLING

Author	Location
Isotopes	
Dorman (1966)	Australia.
Devereaux (1967)	New Zealand.
Savin et al. (1975)	Equatorial Pacific, DSDP Sites 44 (1428m, present depth of site), 167 (3166m), 171 (2283m).
Shackleton & Kennett (1975)	Southern Ocean, DSDP Site 277 (1214m).
Kennett & Shackleton (1976)	
Boersma & Shackleton (1977)	Mid-latitude South Atlantic, DSDP Site 357 (2086m).
Buchardt (1978)	North Sea.
Vergnaud-Grazzini et al. (1978, 1979)	Eastern North Atlantic, DSDP Sites 398 (3910m), 400A (4399m), 401 (2495m).
Miller & Curry (in prep.)	Labrador Sea, DSDP Site 112 eastern North Atlantic, DSDP Site 119 (4447m).
Distribution of Calcareous Nannoplankton	
Edwards & Perch-Nielson (1975)	Southern Ocean.
Haq et al. (1977; 1979)	Atlantic.
Distribution of Planktonic Foraminifera	
Jenkins (1968, 1975)	New Zealand.
Hornibrook (1971)	
Kennett (1977, 1978)	Antarctic Ocean.
Berggren (1978)	North Atlantic.
Gradstein & Srivastava (1980)	Labrador Sea.
Paleobotanical Evidence	
Northern Hemisphere:	
Dorf (1964)	Western United States.
Wolfe & Hopkins (1967)	
Addicott, (1969 1970)	
Wolfe (1978)	
Tanai & Huzioka (1967)	Japan.
Southern Hemisphere	
Kemp (1978)	Antarctic Ocean, Antarctica, and Australia.
Kemp & Barrett (1975)	

was similar to that of today (Figure 12A), although the basins were slightly smaller and the continental latitudes were 5-10° lower (Sclater et al., 1977). Northward advection of warm, presumably high-salinity water is thought to have occurred in an anticyclonic subtropical gyre (proto-North Atlantic Current). Another possible low-latitude source of warm, high-salinity water was the proto-Mediterranean (Tethys) Sea which lay at latitudes 15° to 30°N in the Eocene/Oligocene (Berggren and Hollister, 1974); such latitudes are the present sites of formation of high-salinity surface and intermediate water in the North Atlantic (Worthington, 1976). Cooling of these high-salinity waters in the northern North Atlantic and/or Norwegian-Greenland Sea presumably occurred during the Eocene/Oligocene phase of climatic cooling and increased latitudinal thermal gradients; based upon the modern analog, the likely sites of most intense cooling (with possible convection and bottom-water formation) would have been in regions where polar air was delivered from the large land masses, i.e. east of North America and Greenland. Thus, bottom-water formation could have occurred not only in the Norwegian-Greenland Sea, but also south of the Greenland-Scotland Ridge and in the Labrador Sea. Unfortunately it is not presently possible to determine whether such northerly advection and cooling generated significant amounts of deep and bottom water in the northern Atlantic. It seems unlikely that the shift in climatic pattern alone was responsible for the dramatic development of abyssal circulation at the Eocene/Oligocene boundary. However, it may have accentuated bottom-water formation, especially in the Labrador-Baffin and Norwegian-Greenland Seas.

MODEL OF CIRCULATION

Figure 12A summarizes our interpretation of the abyssal circulation at the beginning of the Oligocene. The widespread distribution of reflector R4 and its characteristic correlation with an unconformity indicates that strong abyssal circulation affected the North Atlantic basins both east and west of the mid-ocean ridge. For abyssal circulation to have affected so strongly the sedimentary record in the Rockall region and in the Bay of Biscay, we consider it necessary that the eastern North Atlantic had a major source of bottom water, probably from the Norwegian-Greenland Sea via the Faeroe-Shetland and Faeroe Bank Channels. This timing is coincident with the opening of a deep-water passage between Greenland and Spitsbergen that connected the Arctic Ocean and the Norwegian Sea (Talwani and Eldholm, 1977). The implication therefore is that the bottom water had an Arctic source, possibly supplemented in the Norwegian Sea and south of the Greenland-Scotland Ridge by convective overturn of cooled, saline surface waters. The timing coincidentally agrees with the marine connection through the Faeroe Bank Channel suggested by Vogt (1972). However, as discussed earlier, the passage could have been deeper than Vogt supposed. Thus, once the Greenland-Spitsbergen connection was severed, Arctic water presumably flowed unimpeded into the North Atlantic to form the first strong abyssal circulation in the basin (see Berggren and Hollister, 1974). This "tectonic-threshold" control on introduction of bottom water to the North Atlantic is the most satisfactory explanation of the suddenness and intensity of erosion in the Atlantic basins.

The margin-intensified, anticlockwise circulation in the northeastern Atlantic was contained by the topographic barriers of the continental margin and the mid-ocean ridge (Figure 12A). Bottom water that flowed south along the flank of the mid-ocean ridge probably was turned north along the Azores-Gibraltar Ridge and Azores-Biscay Rise into the Bay of Biscay. Along the first of these, minimum depths presently are about 4300m and maximum depths in narrow passages are 4900m. Backtracking along a North Atlantic empirical age-depth curve (Tucholke and Vogt, 1979) indicates these depths were on the order of 3500m and 4000m, respectively, in the early Oligocene. The same analysis applies to the Azores-Biscay Rise which probably was an even shallower barrier at 3000-3500 m. Thus these ridges must have precluded significant bottom-water exchange with more southerly parts of the eastern North Atlantic at depths below 3000-3500m. By the same token, they blocked any significant input of bottom water to the Bay of Biscay from possible southern sources. Although Schnitker (1980a) preferred such Antarctic sources for bottom water in the northern North Atlantic, it is unlikely that they could have been significant even in the absence of these barriers, given the minor influence of Antarctic Bottom Water in this region under modern, more glaciated conditions.

The circulation probably was most intense along the basin margins, as observed by unconformity development in DSDP boreholes. It is also possible that the first development of bottom currents affected only the deepest part of the basins (3500m) as indicated by the possible diachrony in unconformity development below 4000 m and above 3000m as noted earlier. Continued flooding of the basin by dense bottom waters could then have affected progressively shallower seafloor.

Bottom water from the eastern basin probably entered the western basins through major fracture zones in the mid-ocean ridge (e.g. Charlie Gibbs F.Z.) as well as through numerous smaller fracture-zone conduits in the Reykjanes Ridge (Ruddiman, 1972; Figure 12A). Magnetic anomaly patterns suggest that these fracture-zone conduits did not develop until the late Eocene to early Oligocene (Figure 2; Ruddiman, 1972). The abyssal circulation west of the mid-ocean ridge may also have been augmented by overflow across the Greenland-Scotland Ridge adjacent to Greenland and possibly by bottom water derived from the area of Baffin Bay/Nares

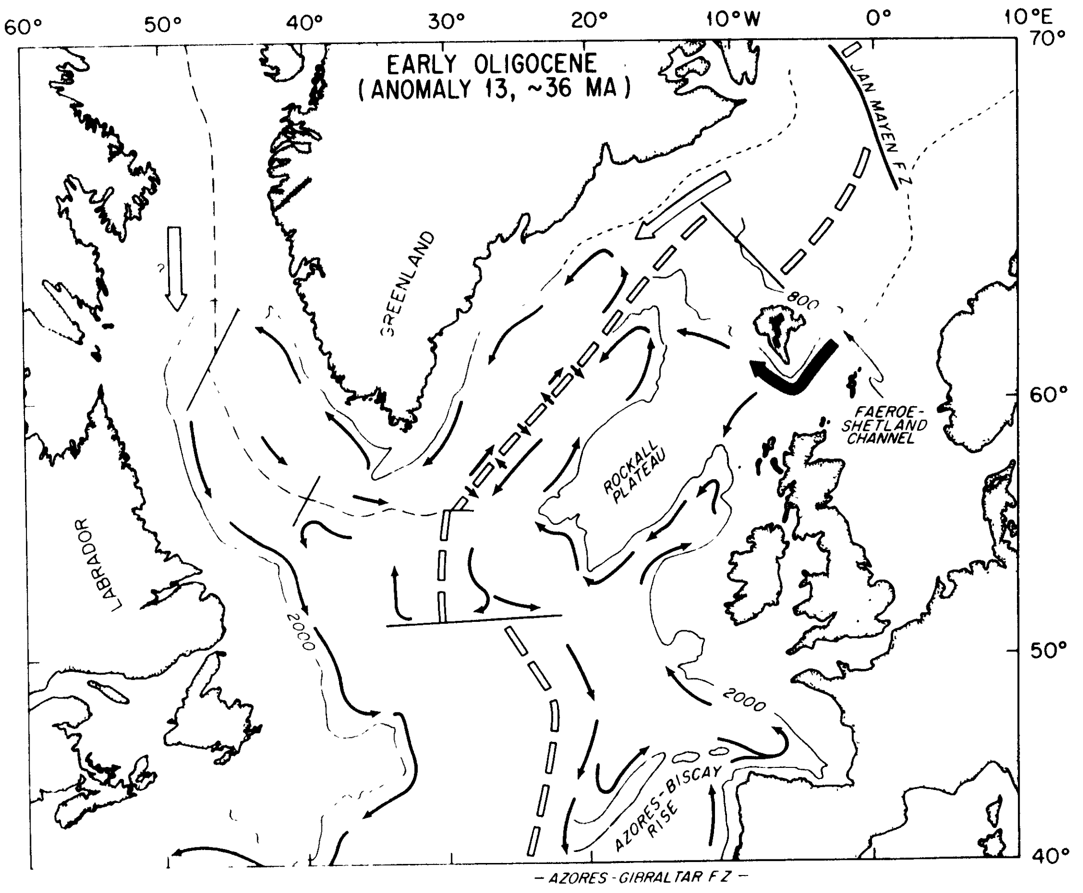

60°
50°
40°
30°
20°
10°W
0°
10°E
70°
60°
50°
40°
EARLY OLIGOCENE
(ANOMALY 13, ~36 MA)
GREENLAND
JAN MAYEN F.Z.
800
?
FAEROE-SHETLAND CHANNEL
ROCKALL PLATEAU
LABRADOR
2000
2000
2000
AZORES-BISCAY RISE
- AZORES-GIBRALTAR F Z -

Figure 12A. Model of abyssal circulation in the North Atlantic in the early Oligocene (Anomaly 13 time). Solid and open bold arrows - probable and possible bottom-water sources; solid arrows - inferred abyssal circulation (length and spacing approximately proportional to intensity of circulation). Active spreading centers indicated by open rectangles (i.e with fracture-zone conduits) or continuous parallel lines (no conduits). Major fracture zones are solid lines. Extinct spreading axes are long dashed lines, and approximate ocean-continent boundaries are short dashed lines. Present 2000m bathymetric contour shown for reference. Tectonic data and plate geometries from Ruddiman (1972), Talwani and Eldholm (1977), Srivastava (1978), and Vogt et al. (1980).

Strait (Figure 12A). We suggest that the combined flow from these sources formed the intense abyssal boundary current that eroded the Horizon A^u unconformity in the western North Atlantic, probably in the early to middle Oligocene.

By the late Oligocene (reflector R3), subsidence of the Faeroe Bank Channel permitted increased flow from the Norwegian-Greenland

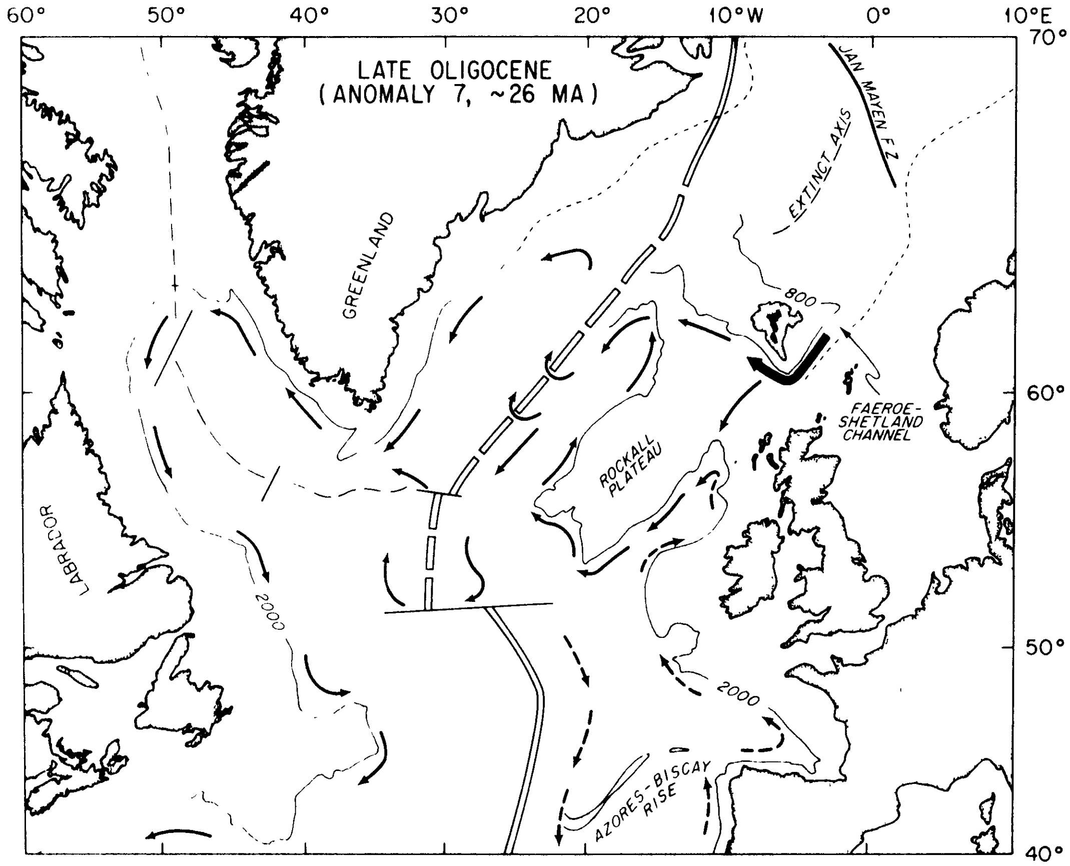

LATE OLIGOCENE
(ANOMALY 7, ~26 MA)
60°
50°
40°
30°
20°
10°W
0°
10°E
70°
60°
50°
40°
GREENLAND
LABRADOR
ROCKALL PLATEAU
EXTINCT AXIS
JAN MAYEN F.Z.
FAEROE-SHETLAND CHANNEL
AZORES-BISCAY RISE
800
2000
2000

Figure 12B. Model of circulation in the late Oligocene (Anomaly 7 time). Explanation in Figure 12A.

Sea (Figure 12B) into the Rockall region, probably accounting for continued development of unconformities and chaotic current-controlled deposition. However, abyssal-current influence on sedimentation in the Bay of Biscay apparently declined as the Azores-Biscay Rise and Azores-Gibraltar Ridge subsided and failed to divert currents into the Bay of Biscay.

Reykjanes Ridge fracture zones continued to provide conduits for bottom water to enter the western basins, but north of the Greenland-Scotland Ridge a spreading-ridge jump to the Greenland margin may have blocked any significant flow from the Norwegian-Greenland Sea directly into the northwestern basins (Figure 12B). The implied decrease in abyssal boundary flow in the western North Atlantic correlates with current-controlled deposition above the Horizon A^u unconformity in the late Oligocene to early Miocene (Figure 2).

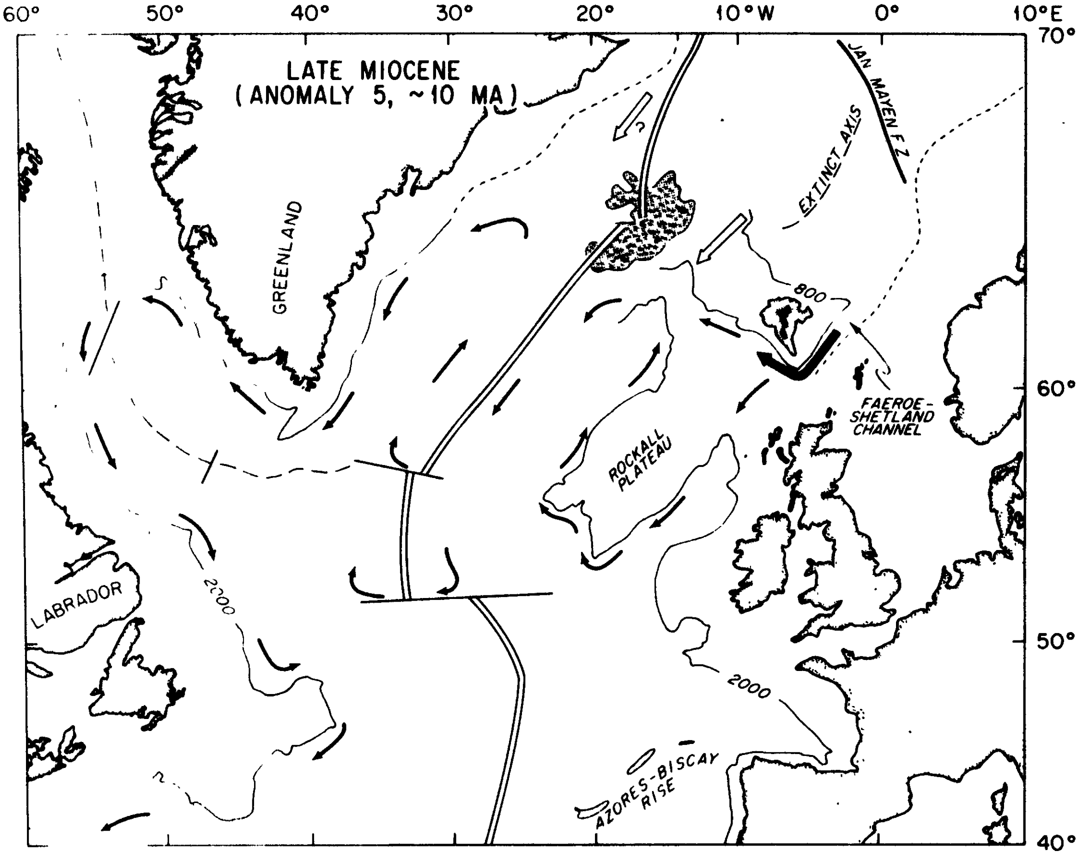

60°
50°
40°
30°
20°
10°W
0°
10°E
70°
60°
50°
40°
LATE MIOCENE
(ANOMALY 5, ~10 MA)
GREENLAND
JAN MAYEN F.Z.
EXTINCT AXIS
?
800
FAEROE-SHETLAND CHANNEL
ROCKALL PLATEAU
LABRADOR
2000
2000
AZORES-BISCAY RISE

Figure 12C. Model of circulation in the late Miocene (Anomaly 5 time). Explanation in Figure 12A.

During the early to middle Miocene the fracture zone conduits across the Reykjanes Ridge ceased to be developed, and, in agreement with Ruddiman (1972) and Shor and Poore (1979), we correlate this trend toward east-west basin isolation with the coherent development of sediment drifts and waves in the northern and western North Atlantic (Figures 2, 12C). However, we interpret this coherent development to be a result of a general decrease and stabilization of the abyssal circulation. This interpretation is based on the fact that the preceding sedimentary record, characterized by chaotic sedimentation and prominent unconformities, must have required much higher current velocities.

It is not clear how the deterioration of the Reykjanes Ridge fracture zones may have attenuated the abyssal circulation, aside from a general restriction of possible flow conduits, nor is it clear that this was the sole factor responsible for reduced deep circulation. For example, Shackleton and Kennett (1975), Savin et al. (1975), and Woodruff et al. (1981) noted a dramatic shift in oxygen isotopic composition of benthic foraminifera that began in the middle Miocene. Although they attributed this shift primarily to buildup of the Antarctic ice cap, Woodruff and Douglas (1981) suggested that at least some of the shift represents a major bottom-water cooling and a "thickening of Antarctic Bottom Waters." Such an implied increase of Antarctic sources of bottom water may have resulted in reduced influence of bottom water derived from northern sources.

There is some independent circumstantial evidence for reduced abyssal flow in the North Atlantic at this time. Gradstein and Srivastava (1980) observed apparently decreasing poleward advection of warm Atlantic surface waters in the Labrador Sea during the Oligocene and Miocene; this implies decreasing production and exodus of bottom water from the northern Atlantic. If such a correlation is valid, then late Miocene development of the cold southward-flowing Labrador surface current (Gradstein and Srivastava, 1980) also may be indicative of reduced abyssal-current flow at this time.

Our overall interpretation of the development of abyssal circulation in the North Atlantic is 1) a rapid increase in current strength during the latest Eocene to early Oligocene that created strongly erosional conditions, 2) a general decrease in intensity of flow during the Oligocene to early Miocene resulting in increased coherence of geological effects, and 3) a further decrease and stabilization of abyssal flow in the middle to late Miocene with coherent development of sediment drifts that has continued more or less unaltered to the present.

Clearly, any long-term trend in abyssal circulation can be punctuated by tectonic and climatic events, several of which we have discussed. Two additional events are the possible unconformity at reflector R2 and the widespread upper middle Miocene unconformity between reflectors R2 and R1. Either or both of these could be a response to temporarily increased flow caused by such factors as subsidence of the Greenland-Scotland Ridge (e.g. between Iceland and Faeroe Islands, Figure 12C) or to northern hemisphere effects of the middle Miocene climatic cooling. In addition, the Plio-Pleistocene glacial cycles must have had "fine-scale" effects in the uppermost portion of the sedimentary record. At present, however, speculation on anything more detailed than the gross kinds of changes noted above probably is not justified.

CONCLUSIONS

In the foregoing discussion we presented a model for the development of the abyssal circulation in the northern North Atlantic that is far from being completely developed or correct in detail. However, it does explain the lithostratigraphic, seismic stratigraphic, and floral/faunal data that must be considered in any discussion of the general development of the deep circulation. It also emphasizes the important control that abyssal currents have had on sedimentation patterns since the Eocene.

A better understanding of the abyssal circulation history obviously can be developed by the acquisition of additional, carefully placed geological (borehole) and geophysical data. However, within the province of existing information, we feel that there are three kinds of analyses that will significantly improve our perception of the abyssal circulation history:

1) Basin-wide mapping of the seismic stratigraphic framework in quasi-chronologic intervals, to provide clearer definition of the intra-basin geometries of current effects in relation to tectonic barriers and tectonic threshold events.

2) Improved subsidence models of the Greenland-Scotland Ridge, taking into account such factors as the age of rifting of apparently continental crust beneath the Faeroe Bank Channel, sill depths with appropriate sediment overburden removed, and spreading history and ridge jumps. This will allow more specification of the timing of shallow and deep marine connections across the ridge.

3) Continued benthic faunal, planktonic floral and faunal, and isotopic studies of existing DSDP cores in order to clarify the Tertiary patterns of paleocirculation, the potential northern-hemisphere climatic effects on bottom-water formation, independent of the southern hemisphere climatic and ice-volume record, and the history of surface and deep water connections of the Norwegian-Greenland Sea with the North Atlantic.

We look forward to these and comparable studies as worthwhile tests of our interpretations of the North Atlantic abyssal-circulation history.

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APPENDIX 4

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Late Paleogene (Eocene to Oligocene) Paleoceanography of the Deep Bay of Biscay: Benthic Foraminiferal Evidence

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Draft for submission to Marine Micropaleontology

ABSTRACT

A major change in benthic foraminiferal assemblages occurred in the deep Bay of Biscay (> 3 km water; DSDP Sites 119 and Site 400A) between early middle Eocene and earliest Oligocene. Predominant Eocene deep-sea taxa (*Nuttallides truempyi*, *Clinapertina* spp., *Abyssamina* spp.) and associated rarer species become extinct in this interval. These extinctions are followed by an increase in abundance of bathymetrically wide-ranging and stratigraphically long-ranging taxa: *Globocassidulina subglobosa*, *Oridorsalis* spp., *Gyroidinoides* spp., and the *Cibicidoides ungerianus* plexus. The extinctions cannot be dated precisely from the stratigraphic record recovered to date in the Bay of Biscay; however, the replacement of the *N. truempyi*-dominated assemblage has been noted previously in the deep South Atlantic/Caribbean as occurring within the middle Eocene. No major faunal changes are noted within the Eocene at the shallower Site 401 (~ 2 km water) in the Bay of Biscay. During the Oligocene, *Nutallides umbonifera* replaces the Eocene species *N. truempyi* as the predominant deep-sea benthic foraminifera, reaching peak abundance in the middle Oligocene at Sites 119 and Site 400A. In the modern oceans, the abundance of *N. umbonifera* is positively correlated with increased corrosiveness of bottom water, while at Site 119 the abundance of *Nutttallides* spp. is negatively correlated with $\delta^{13}C$ values in benthic foraminifera. As lower $\delta^{13}C$ values are often associated with older water masses, large numbers of *Nuttallides* spp. are thought to reflect older, and more corrosive bottom water. The faunal data and oxygen and carbon isotopic data are compared with a circulation model derived from North Atlantic seismic stratigraphic studies to show that old, warm, corrosive, and sluggish Eocene bottom water is replaced by younger, colder, less corrosive, more vigorously circulating bottom water of northern origin by the early Oligocene. Faunal and isotopic data suggest that bottom water became older and more corrosive again in the middle Oligocene, reflecting a reduction in circulation that can also be inferred from the seismic record in the nearby Rockall Plateau region.

INTRODUCTION

Foraminifera are the major group of fossil benthic organisms found in deep-sea sediments, and therefore constitute the main paleontological monitor of deep-sea environmental changes. Several studies (Streeter, 1973; Schnitker, 1974; Lohmann, 1978, among others) have shown that the distribution of modern deep-sea benthic foraminifera can be correlated with water masses. These water mass-foraminiferal relationships have been extrapolated into the fossil record in order to determine changes in ancient water masses and circulation patterns, both for the Quaternary (Streeter, 1973; Schnitker, 1974; Lohmann, 1978; Corliss, 1979) and the Tertiary (Schnitker, 1979; Tjalsma and Lohmann, 1982; Douglas and Woodruff, 1982).

Changes in the taxonomic or isotopic composition of benthic foraminifera may signal changes in circulation, but are not necessarily indicative of the nature of circulation changes. For example, a major change in deep-sea benthic foraminifera has been reported to have occurred near the Eocene/Oligocene boundary (Douglas, 1972; Boersma, 1977; Schnitker, 1979). This change was associated with an increase in $\delta^{18}O$ in benthic foraminifera interpreted as a drop in deep-sea paleotemperatures (Shackleton and Kennett, 1975, Savin et al., 1975; Boersma and Shackleton, 1977; Vergnaud-Grazzinni et al., 1978, 1979). In contrast, Corliss (1979, 1981) recongnized more gradual middle Eocene to Oligocene benthic faunal changes in the Southern Ocean, and Tjalsma and Lohmann, (1982) suggested that a series of sequential benthic foraminiferal changes occurred between the middle Eocene and early Oligocene in the South Atlantic/Caribbean. Schnitker (1979, 1980a,b) ascribed Eocene/Oligocene benthic foraminiferal changes in the Bay of Biscay to the inital formation of cold bottom water derived from the Southern Ocean. Miller et al. (1982), on the other hand, attributed faunal changes in Eocene to Oligocene strata of the Labrador Sea to initial formation of northern sources of cold bottom water. These conflicting interpretations illustrate potential pitfalls in using benthic foraminiferal changes (taxonomic or isotopic composition) as diagnostic of the nature and timing of changes in abyssal circulation.

The seismic stratigraphic record provides less ambiguous evidence for changes in abyssal circulation. In the deep North Atlantic, the flow of deep and bottom waters formed in high latitudes have influenced sediment distribution well back into the Tertiary (Jones et al., 1970; Ruddiman, 1972; Roberts, 1975; Miller and Tucholke, in press). Miller and Tucholke summarized the seismic stratigraphic and lithostratigraphic evidence from this region and concluded that northern sources (i.e. the Norwegian-Greenland Sea or Arctic Ocean) for vigorously circulating bottom water began to form in the late Eocene to early Oligocene in the North Atlantic. Based upon seismic sequence analyses, they presented a model for the development of Tertiary abyssal circulation in the North Atlantic which suggested that 1) an erosional sequence, representing the most vigorous bottom water flow, occurred in the latest Eocene through early Oligocene, and 2) subsequent depositional sequences, representing decreasing bottom water flow, occurred in the late Oligocene through Miocene. I will use this independent evidence of changes in abyssal circulation in the North Atlantic to evaluate changes in the taxonomic and isoptopic composition of benthic foraminiferal assemblages Bay of Biscay.

This study compares benthic foraminiferal assemblages from Deep Sea Drilling Project (DSDP) Sites 119 (early to middle Eocene and Oligocene), 401 (Eocene), and 400A (Oligocene) in the Bay of Biscay. The assemblages changes are interpreted in view of the circulation model derived from the seismic stratigraphic record as supplemented by oxygen and carbon isotopic studies made at these sites.

PREVIOUS WORK

Tertiary deep-sea benthic foraminifera have been the subject of several recent papers that delineate the taxonomy, biostratigraphy, paleobiogeography, and paleoecology of this hitherto poorly-known group (Berggren, 1972; Douglas, 1973; Resig, 1976; Boersma, 1977; Proto-Decima and Bolli, 1978; Schnitker, 1979; Corliss, 1979, 1981; Tjalsma and Lohmann, 1982; Douglas and Woodruff, 1982). These studies have concentrated on bathyal and abyssal benthic foraminifera recovered by the DSDP. Prior to the DSDP, Tertiary deep-sea benthic foraminifera were known primarily from classic exposures in Mexico, Trinidad, the Dominican Republic, and Barbados (Cushman, 1925, 1926; Cushman and Jarvis, 1928, 1929, 1932; Cushman and Stainforth, 1945; Cushman and Renz, 1946, 1947, 1948; Bermudez, 1949; Beckmann, 1953, among others). Partly because of the sparsity of material prior to the DSDP, but also due to a weak taxonomic base, the taxonomy of deep-sea benthic foraminifera remained in a state of disarray. More recently, however, Tjalsma and Lohmann (1982) have made a detailed revision of the taxonomy of bathyal and abyssal benthic foraminifera from the South Atlantic and Caribbean. The establishment of this taxonomic base allows the placement of further studies into a standardized framework in which paleoenvironmental interpretations can be made.

Berggren (1972) noted high-diversity Eocene and Oligocene abyssal benthic foraminiferal assemblages at Site 119 in the Bay of Biscay that have marked affinities to assemblages found in Barbados; the Site 119 assemblages have not been described in detail. Schnitker (1979) described the Eocene and Oligocene benthic foraminiferal assemblages recovered from the Sites 400A (4399m) and 401 (2495m) on the Armorican margin in the Bay of Biscay (Fig. 1). He determined that a "peak in faunal turnover" occurred at the Eocene/Oligocene boundary. Regrettably, the late middle and late Eocene section is missing in Site 400A and the Oligocene is missing in Site 401 (Montadert, Roberts et al., 1979; Krasheninnikov, 1979; Muller, 1979). Thus, Schnitker's conclusions are based upon a comparison of different time intervals from different paleodepths and do not clearly document the timing of the taxonomic changes.

Due to differences in species identification and the lack of published percentage data, it is not possible to directly compare these studies with that of Tjalsma and Lohmann (1982). In the South Atlantic and Caribbean, Tjalsma and Lohmann (1982) have shown that a sequence of faunal changes take from the middle Eocene to Oligocene. These changes are 1) a change from an early to middle Eocene *Nuttallides truempyi* assemblage to a late Eocene *Globocassidulina subglobosa*, *Gyroidinoides* spp., *Cibicicdoides ungerianus*, and *Oridorsalis umbonatus* assemblage, 2) the extinction of *N. truempyi* in the latest Eocene, and 3) the appearance of *Nuttallides umbonifera* in the late Eocene. In addition, Tjalsma (1982 and personal communication) noted 4) the continued dominance of *N. umbonifera* and the "late Eocene" assemblage (*G. subglobosa*, *Gyroidinoides* spp., *C. ungerianus*, *O. umbonatus*)

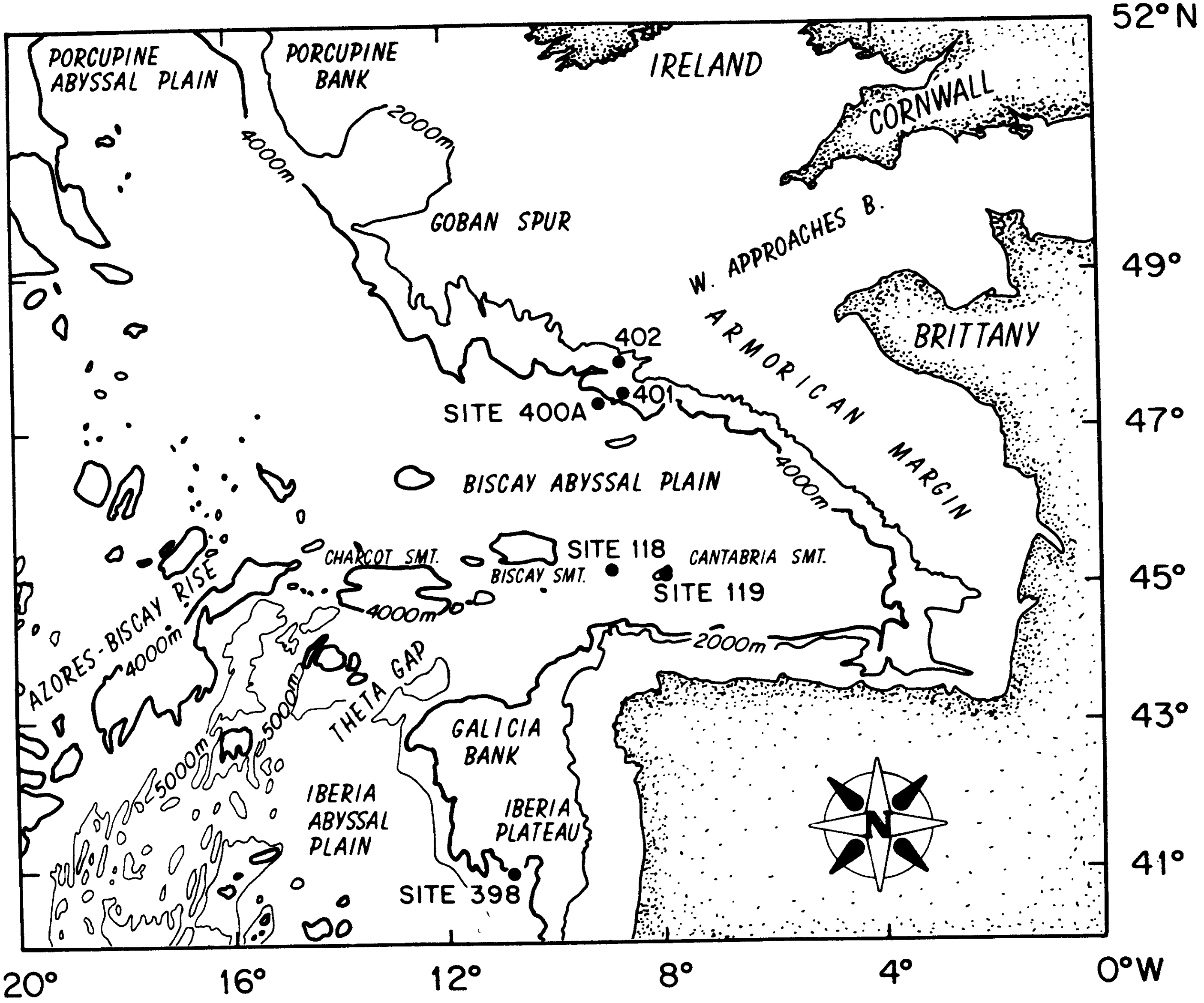
52° N
PORCUPINE ABYSSAL PLAIN
PORCUPINE BANK
IRELAND
CORNWALL
2000m
4000m
GOBAN SPUR
W. APPROACHES B.
49°
ARMORICAN MARGIN
BRITTANY
402
401
SITE 400A
47°
BISCAY ABYSSAL PLAIN
4000m
SITE 118
CHARCOT SMT.
BISCAY SMT.
CANTABRIA SMT.
SITE 119
45°
4000m
2000m
AZORES-BISCAY RISE
4000m
THETA GAP
5000m
5000m
43°
GALICIA BANK
IBERIA ABYSSAL PLAIN
IBERIA PLATEAU
SITE 398
N
41°
20°
16°
12°
8°
4°
0°W

Figure 1. Location map, Bay of Biscay.

across the Eocene/Oligocene boundary into the Oligocene. The most dramatic of these changes is the diachronous replacement of the *N. truempyi* assemblage by the *G. subglobosa*-*Gyroidinoides*-*C. ungerianus*-*O. umbonatus* assemblage in the early middle Eocene (shallow sites) and middle/late Eocene (deeper sites). Tjalsma and Lohmann concentrated on Caribbean and South Atlantic samples; they did not examine benthic foraminifera from North Atlantic sites in detail. I have tried to determine when the events they observed took place in the North Atlantic. I have used their taxonomy to define the assemblages found in Site 119, to compare the Site 119 assemblages with the Site 401 assemblages, to compare the Bay of Biscay distribution data with their data, and to evaluate Schnitker's conclusion that a major faunal change occurred at the Eocene/Oligocene boundary in the Bay of Biscay.

METHODS

Thirty-seven samples from the Eocene and Oligocene sediments of Site 119 were examined for benthic foraminifera (Table 1). In addition, nine samples from the middle to late Eocene section of Site 401 were examined (Table 2). In the course of the study, *N. umbonifera* occurred in high abundance in the Oligocene of Site 119. This species was not identified in previous studies of Sites 119 (Berggren, 1972) or 400A (Schnitker, 1979); therefore, fourteen samples from the latest Eocene to Oligocene of Site 400A were examined (Table 5) in order to confirm the presence of *N. umbonifera* in Site 400A.

The samples were washed through a 63 μm sieve with hydrogen peroxide and sodium carbonate or a calgon solution. Aliquots from the > 149 μm size-fraction (the same size-fraction used by Tjalsma and Lohmann, 1982) were picked and mounted on a reference slide. Approximately 300-500 specimens were picked whenever possible; however, several samples contained fewer than 300 specimens (Table 1,2). Sample weight is approximately 25 to 50 grams; exact sample weight is not available for Site 119 as these were taken shipboard.

The benthic foraminifera were identified using the taxonomy outlined by Tjalsma and Lohmann (1982) which is further described in the Taxonomic Notes. Using computer programs provided by Lohmann (1980 and personal communication), I have directly compared my census values from Sites 119 and 401 with the species distribution (percent) data of Tjalsma and Lohmann (1982). In addition, I computed loadings of my Site 119 and 401 samples onto Tjalsma and Lohmann's Eocene Q-mode principal components in order to determine how these samples compare with the assemblages they defined. These loadings are components of the Site 119/401 sample vectors projected onto the coordinate system defined by Tjalsma and Lohmann's principal component analyses of their Eocene samples. Only samples with greater than 100 specimens were used in these comparisons.

Backtracking and paleodepths estimates are given in Appendix I.

RESULTS

Biostratigraphy and Lithology, Site 119

Site 119 was drilled in 4447m water depth on Cantabria Seamount in the Bay of Biscay (Fig. 1). The entire Paleocene to Oligocene section (cores 6 to 27; 249 to 433m subbottom), was continuously cored. Basement was not penetrated (Laughton, Berggren et al., 1972).

The base of the cored section consists of Paleocene indurated calcareous gray turbidites with alternating red clay bands. Nannofossils and planktonic

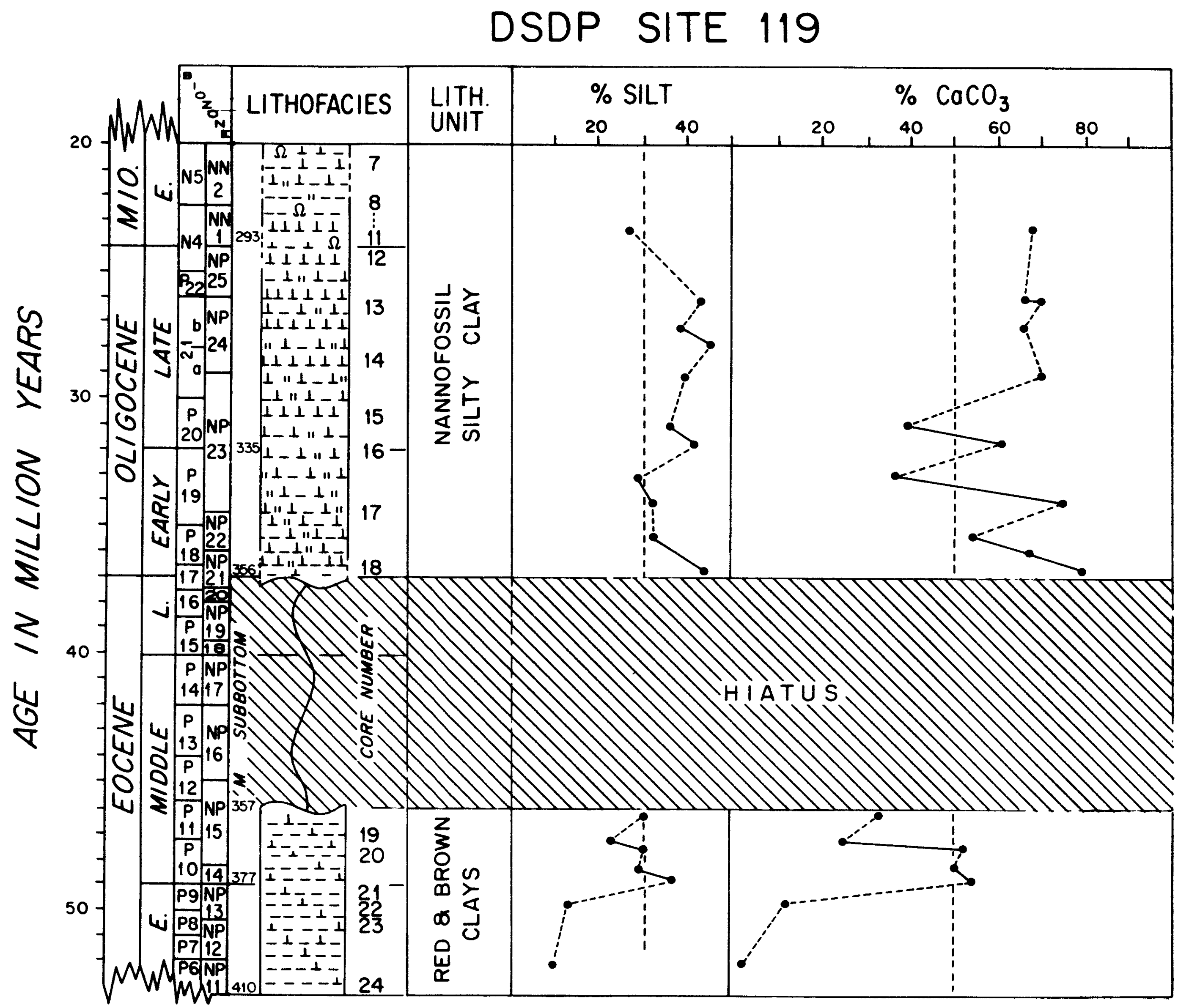

Figure 2. Lithofacies, percent calcium carbonate, and grain size versus age Site 119. Data from Laughton, Berggren et al., 1972.

foraminifera were originally used to place cores 19 to 24 in the early to middle Eocene and cores 12 to 18 in the Oligocene. The Eocene strata consist of red and brown pelagic clays, while the Oligocene strata consist of nannofossil silty clays that are readily distinguished by their coarser grain size, color, and greater calcium carbonate content (~ 60 °/° versus < 50 °/°; Fig. 2; fig. 10 in chapt. 10, Laughton, Berggren et al., 1972). A major unconformity, representing ~ 9 my, separates the lower middle Eocene pelagic clays from the lowermost Oligocene nannofossil silty clays (Laughton, Berggren et al., 1972).

Preservation of planktonic foraminifera is, in general, poor in the early to middle Eocene. Oligocene planktonic foraminifera are better preserved but are primarily low-diversity, high latitude assemblages (Berggren, 1972). These conditions preclude reliable assignments of the standard planktonic foraminiferal biozonations (Fig. 3, Table 3).

Despite poor preservation of calcareous nannoplankton, these forms have been more useful in zoning Site 119 samples (Bukry, 1972; Perch-Nielsen, 1972; Aubry, personal communication). The various zonal age assignments made on the basis of nannofossils and planktonic foraminifera have been summarized in Figure 3 and Table 3. Independent estimates of sample age have been obtained assuming two separate constant sedimentation rates for the Eocene and Oligocene, respectively (Fig. 3). Although this method yields reasonable agreement with the biostratigraphic age assignments (Fig. 3), it should be noted that assumption of constant sedimentation is an approximation, and that the true age uncertainty is equivalent to the length of the zone (in my) to which the sample has been assigned. It should be noted also that the biostratigraphic age control is especially poor in cores 15 and 16, and the possibility exists that a minor undetected hiatus occurs in this middle Oligocene section.

DSDP SITE 119

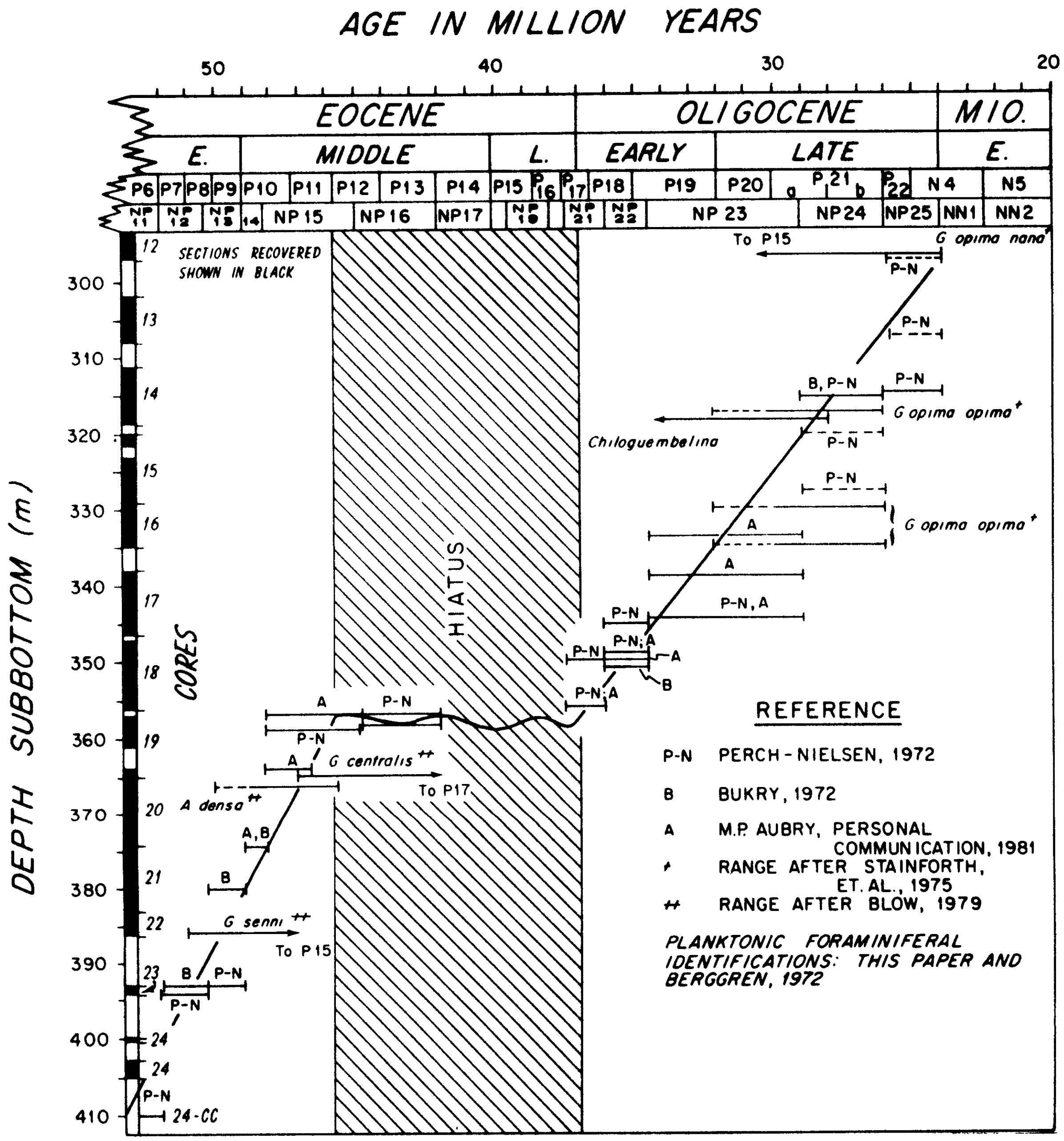

Figure 3. Zonal age assignments and estimated average sedimentation rate Site 119.

The Eocene to Oligocene section of Site 119 shows evidence of dissolution and the benthic foraminifera are often pitted and corroded (Plate 2, fig. 8). The low numbers of planktonic foraminifera in these sections reflect preferential dissolution of these less resistant forms. Dissolution also may have altered the composition of the benthic foraminiferal assemblages. However, the close correspondence of assemblage composition and percentages of abundant taxa between Site 119 and the study of Tjalsma and Lohmann (1982) suggest that the major faunal patterns are preserved.

The percentage of the dissolution-resistant taxon *Gyroidinoides* may be taken as an indicator of dissolution-induced biases. Corliss and Honjo (1981) have shown that *Gyroidinoides* is more resistant to dissolution than *Cibicidoides*, *N. umbonifera*, and *Oridorsalis*. The percentages of *Gyroidinoides* in Site 119 samples are approximately the same as samples from similar depths in Tjalsma and Lohmann's (1982) study. The lack of covariance of the abundance of this dissolution-resistant taxon and dissolution in Site 119 can be shown further by the fact that the greatest percentages of *Gyroidinoides* occur in the earliest Oligocene when the Calcite Compensation Depth (CCD) was deepest and carbonate preservation was at a maximum (Heath, 1969; Berger, 1973; van Andel et al., 1975; Kennett and Shackleton, 1976).

Benthic Foraminiferal Assemblages, Site 119

The early to early middle Eocene assemblages are dominated by *Nuttallides truempyi*, *Abyssamina* spp. (almost exclusively *A. poagi*), and *Clinapertina* spp. (mostly *C. inflata*; Fig. 4), all of which become extinct by the end of the Eocene (Tjalsma and Lohmann, 1982). *Globocassidulina subglobosa*, *Gyroidinoides* spp., the *C. ungerianus* plexus, and *Oridorsalis* spp. (mostly *O.*

DISTRIBUTION OF DOMINANT EOCENE TAXA, SITE 119

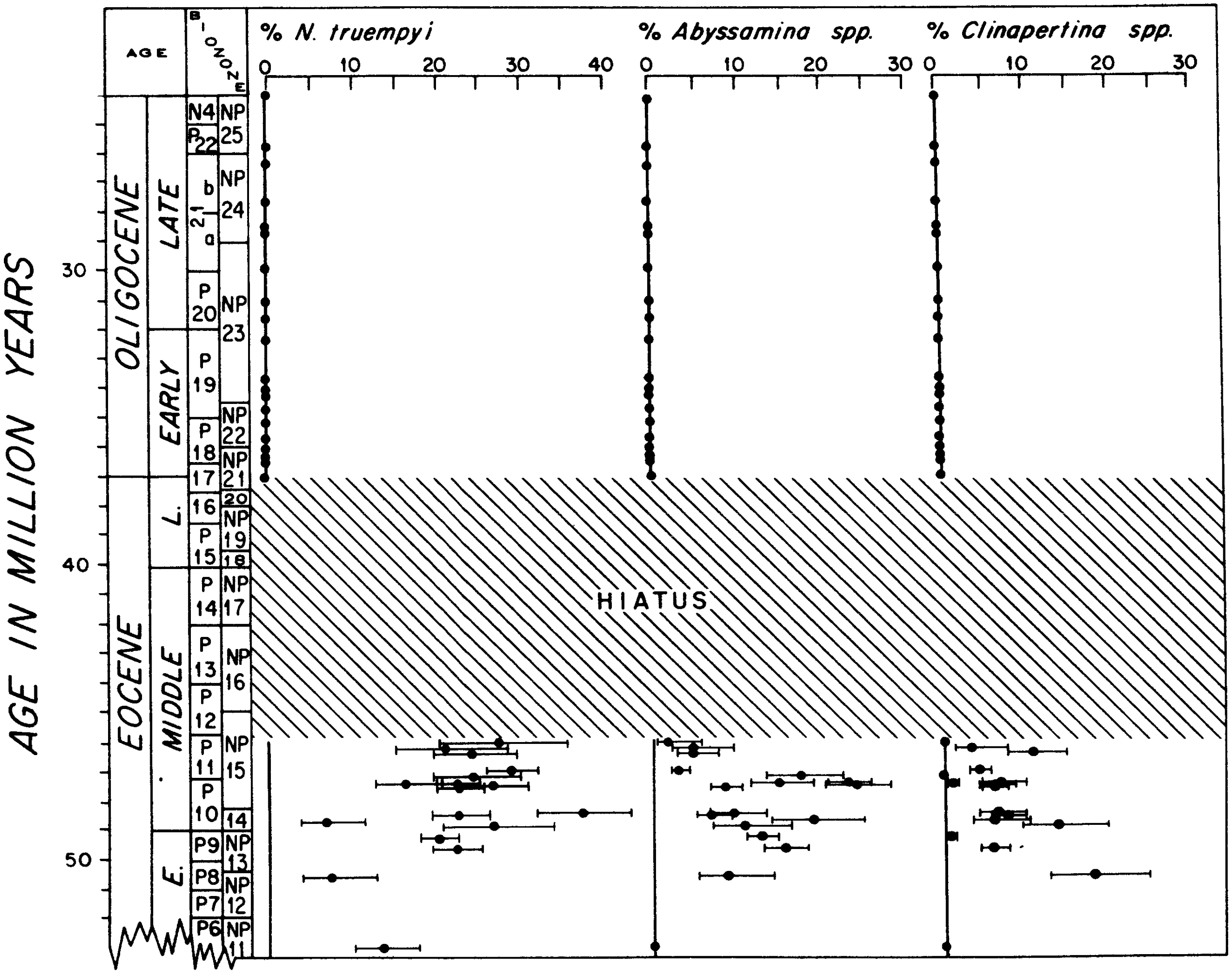

Figure 4. Distribution of dominant Eocene taxa at Site 119. 80°/∘ confidence interval is indicated.

umbonatus; Fig. 5) are also abundant in the early to middle Eocene. These seven taxa dominate the early to middle Eocene, but *N. truempyi* is consistantly most abundant (Table 1). The only exception to this pattern occurs in the early Eocene, where *N. truempyi* is less abundant and *Clinapertina* spp. dominates. However, the low percentages of calcium carbonate found in the early Eocene (Fig. 2) suggest that these samples may be biased by differential dissolution; this is supported by the low abundances of individuals in cores 23 and 24 (Table 1) and the corrosion of specimens in core 24.

The Oligocene is dominated by *N. umbonifera*, *G. subglobosa*, *Gyroidinoides* spp., the *C. ungerianus* plexus (mostly *C. ungerianus* sensu stricto), and *Oridorsalis* spp. (mostly *O. umbonatus*; Fig. 5). The Oligocene differs from the early to early middle Eocene primarily by abundance of *N. umbonifera* and the absence of *N. truempyi*, *Abyssamina* spp., and *Clinapertina* spp. The greater abundance of *G. subglobosa*, *Gyroidinoides* spp., *C. ungerianus*, and *Oridorsalis* spp. in the Oligocene may simply reflect the absence of the Eocene taxa.

Tjalsma and Lohmann (1982) noted that the replacement of the *N. truempyi* assemblage by the *G. subglobosa*-*Gyroidinoides*-*C. ungerianus*-*O. umbonatus* assemblage occurred near the middle/late Eocene boundary in deep sites (>3km) in the South Atlantic and Caribbean, and that this assemblage dominated the deep late Eocene. This replacement occured during the hiatus at Site 119; their "late Eocene" assemblage dominates the Oligocene of this site. In the South Atlantic, Tjalsma (1982 and personal communication) showed that the *G. subglobosa*-*Gyroidinoides*-*C. ungerianus*-*O. umbonatus* assemblage continued from the late Eocene into the Oligocene, and that no major abyssal benthic foraminiferal faunal turnover was associated with the Eocene/Oligocene boundary. However, as a result of the incomplete record recovered at Site

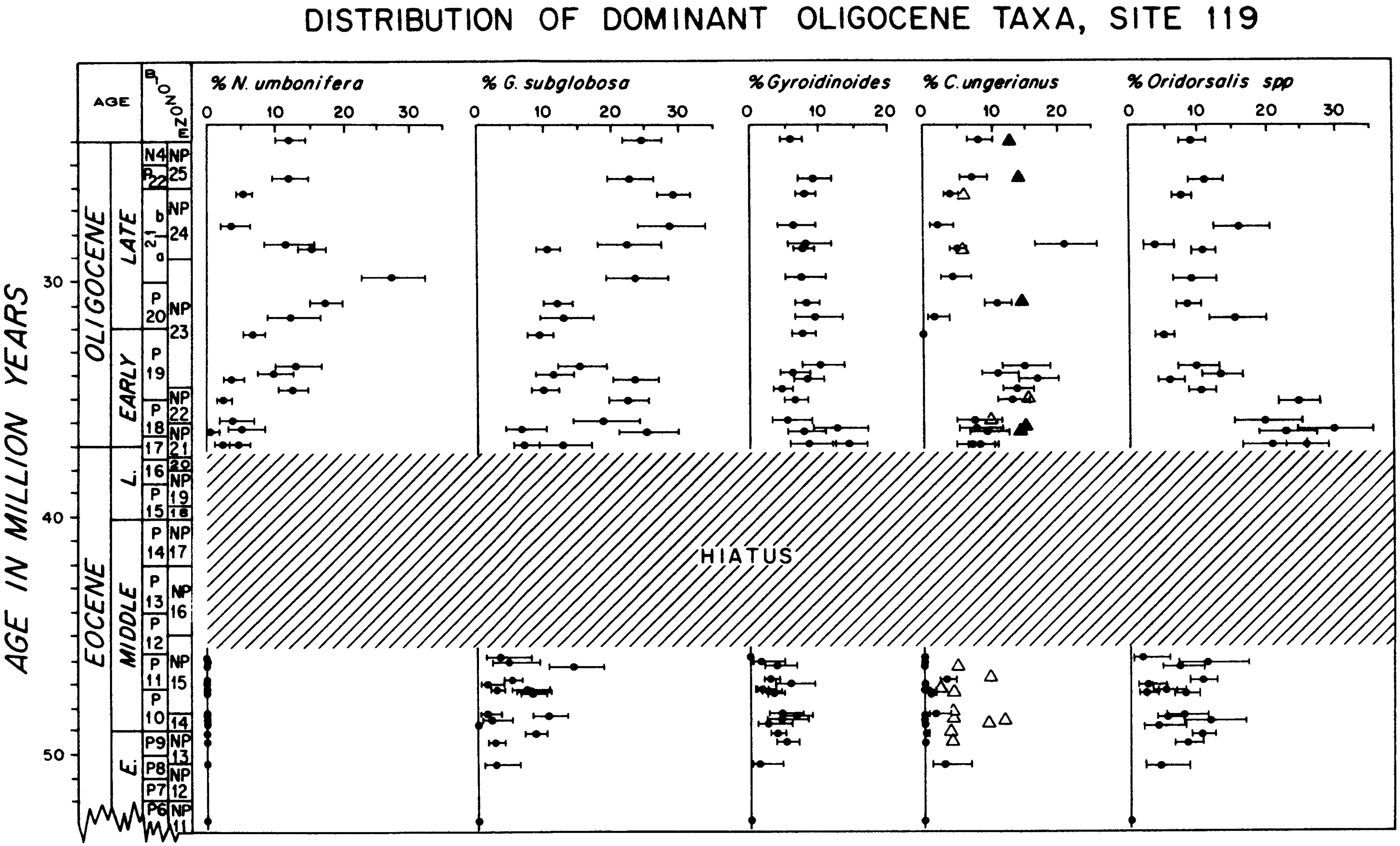

Figure 5. Distribution of dominant Oligocene taxa at Site 119. 80°/∘ confidence interval is indicated. Triangles indicate percent _C. ungerianus_ plexus; dots indicate percent _C. ungerianus_ sensu stricto.

119, it is not possible to date more precisely this change in assemblages in the Bay of Biscay, nor to comment on the nature of abyssal benthic foraminiferal changes near the Eocene/Oligocene boundary in this region.

The benthic faunal assemblages noted in the Eocene of Site 119 are typical of "deep" (>3km) water sites studied by Tjalsma and Lohmann (1982). The sparsity to absence of *Lenticulina* spp., *Uvigerina* spp., *Buliminids* (except *Buliminella* cf. *grata*), and the higher relative abundance of *N. truempyi*, *Abyssamina* spp., and *Clinapertina* spp. all indicate paleodepths greater than 3 km. In addition, the Site 119 Oligocene assemblages are typical for "deep" (>3km) water Oligocene sites studies by Tjalsma (personal communication), characterized by low numbers of *Siphonina* spp., *Stilostomella curvatura*, and *Planulina renzi*, the presence of both *Cibicidoides havanensis* and *C. grimsdalei*, and the greater numbers of *Pullenia eocenica*.

The stratgigraphic ranges for the benthic foraminifera of Site 119 have been compiled (Fig. 6). Although the majority of the taxa are long-ranging stratigraphically, the similar ranges obtained for taxa in Sites 119 and for taxa studied by Tjalsma and Lohmann (1982) show that abyssal benthic foraminifera can be useful in zonation of sediments, especially when planktonic organisms are rare or absent.

Biostratigraphy and Lithology, Site 401

Site 401 was drilled in 2495m water depth on the Meriadzek Terrace on the northern margin of the Bay of Biscay (Fig. 1). Core 1 recovered Pleistocene sediments. A coring cap of 85.5m separates core 1 from the latest Eocene sediments of core 2. Cores 2 to 10, containing the middle to late Eocene section, were continuously cored.

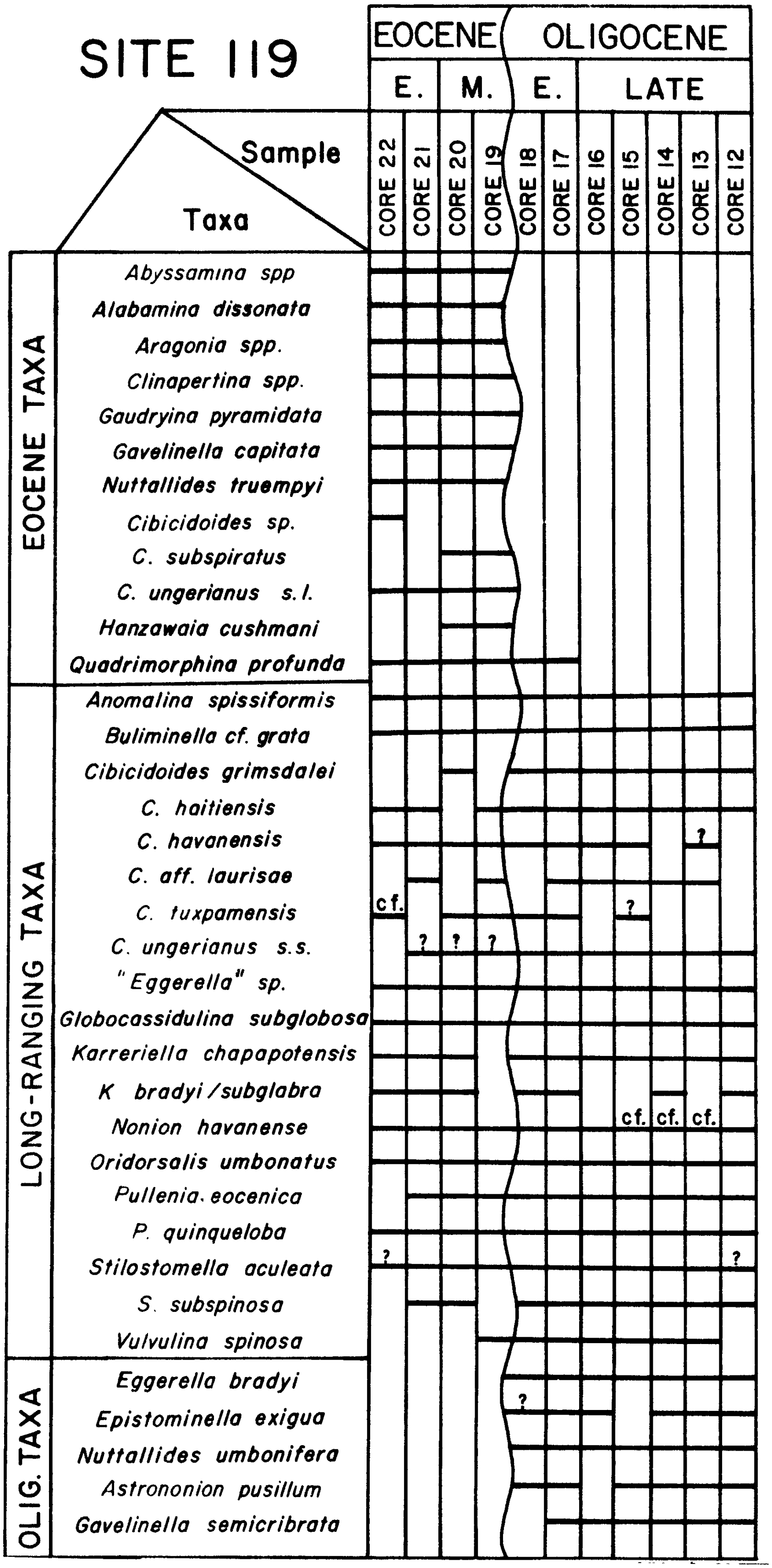

Figure 6. Range chart for Site 119. Clinapertina includes C. inflata, C. complanata, and C. subplanispira; Gaudryina pyramidata includes both s.s. and s.l. forms; Abyssamina includes A. poagi and A. quadrata.

The middle to late Eocene sediments are nannofossil chalks with abundant (~ 20 °/°) biosiliceous components. Cores 2 to 4 (latest middle to late Eocene) are distinguished from the underlying cores 5 to 10 (middle Eocene) by their lower carbonate and higher silica content (Fig. 7; Montadert, Roberts et al., 1979).

A hiatus of ~ 2 my may occur in the late middle Eocene. Published planktonic foraminiferal studies place a hiatus between Zones P14 and P11 (between cores 5 and 6; Montadert, Roberts et al., 1979). Published calcareous nannoplankton studies place a hiatus between Zones NP18 and NP16 (between cores 4 and 5; Montadert, Roberts et al., 1979). Krasheninnikov (1979), on the other hand, identified every middle to late Eocene planktonic foraminiferal biozone; he recognized that a hiatus may occur between the *T. rohri* (=P14) and *O. beckmanni* (=P13) biozones (between cores 4 and 5), but that poor recovery makes a delineation of this hiatus difficult. In any case, the duration of this hiatus, if present, is minor.

The zonal assignments of Krasheninnikov (1979), as updated herein (Table 4) and their correlations to the Hardenbol and Berggren (1978) time scale were used to place Site 401 samples into a chronostratigraphic framework. Of critical importance to this study and to the correlation of the isotopic results from Site 401 (Vergnaud-Grazzini et al., 1979; Miller and Curry, 1981, in press; Goldstein and Anderson, 1981) is the age of the uppermost sediments found in core 2. Muller (1979) placed this core in the earliest Oligocene based upon its assignment to Zone NP21; subsequent examination of core 2 agrees with the assignment to Zone NP21 (M.-P. Aubry, personal communication). It should be noted, however, that this zone straddles the Eocene/Oligocene boundary (Hardenbol and Berggren, 1978; for further discussion as to correlation and possible diachrony of this zone see Van

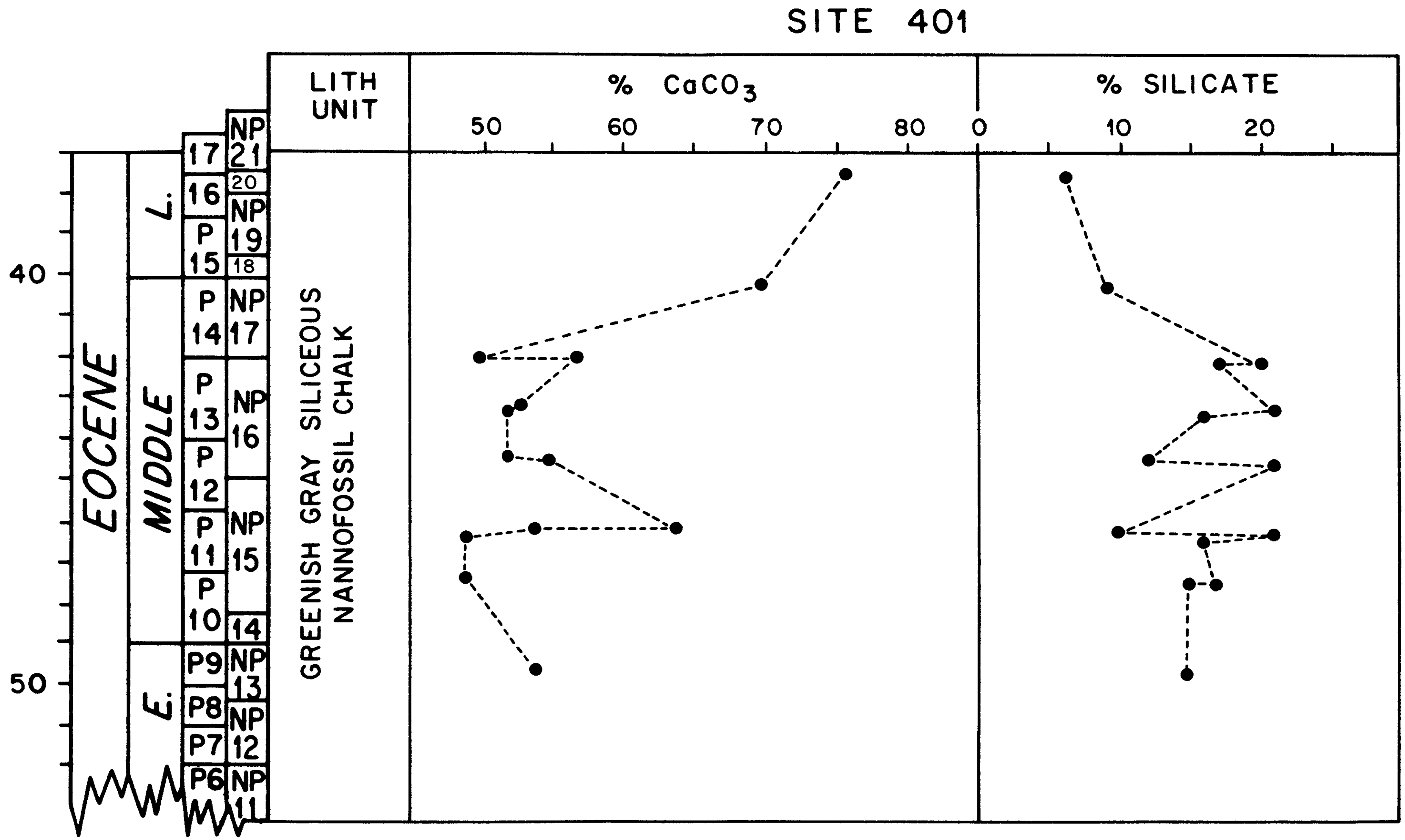

Figure 7. Lithology, percent calcium carbonate, and percent silica versus age at Site 401. Data from Montadert, Roberts et al. (1979) and Auffret and Pastouret (1979).

Couvering et al., in press), so that this zonal assignment does not preclude a latest Eocene age. Planktonic foraminiferal studies assign core 2 to Zone P17 (Montadert and Roberts, et al., 1979; Schnitker, 1979) of Blow (1969) which also apparently straddles the Eocene/Oligocene boundary, and to the late Eocene *G. cocoaensis* Zone of Bolli (1957a,b; Krasheninnikov, 1979). I have used the zonal scheme of Toumarkine and Bolli (1970) to assign sample 2-1, 8-10cm to the latest Eocene *Globorotalia cerroazulensis cunialensis* Zone based upon the presence of the nominate subspecies (Plate 5, figs. 7,8). This assignment poses no discrepancy with the other zonal assignments, agreeing with other authors' assignments to Zones P17 and NP21. Core 2 is therefore taken to be latest Eocene (within the last 0.5 my of that epoch).

Also of critical interest to faunal and isotopic studies is the placement of the middle/late Eocene boundary. Muller (1979) and M.-P. Aubry (personal communication) used nannoplankton to place sample 4-CC in Zone NP18 (earliest late Eocene) and the top of core 5 in Zone NP16 (late middle Eocene). Published foraminiferal studies (Montadert, Roberts et al., 1979) are not clear as to the placement of the middle/late boundary. Their figure 10 showed that Zone P14 (late middle Eocene) occurs in the base of core 4, but reported the middle/late Eocene boundary (mistakenly reported as the early/middle Eocene boundary, p. 88) as occurring between cores 4 and 5. Krasheninnikov (1979) assigned core 4-CC to the middle Eocene *T. rohri* (=P14) and core 3 to the late Eocene *G. semiinvoluta*/*G. cocoaensis* Zones (=P15/16). I note the presence of *Truncorotaloides rohri* and *Morozorella spinulosa* in core 4-CC (both LAD P14; Stainforth et al., 1975) which supports Krasheninnikov's assignments. I have taken the middle/late Eocene boundary as occurring between cores 4 and 5. As this still results in a significant age discrepancy with calcareous nannoplankton age assignments, this correlation should be considered tentative.

Benthic Foraminiferal Assemblages, Site 401

The middle to late Eocene assemblages of Site 401 are taxonomically distinct from both the Eocene and Oligocene assemblages of Site 119. Site 401 is dominated by *Oridorsalis* spp., *G. subglobosa*, and *N. truempyi* (Fig. 8). *Gyroidinoides* spp., *Anomalinoides* spp., *Nonion havanense*, the *C.ungerianus* plexus, buliminids, and *Osangularia* spp. are also important taxa. The Site 401 assemblage is distinguished from the Site 119 Eocene assemblages by its lower percentage of *N. truempyi*, the low numbers of *Alabamina dissonata* and *Abyssamina* spp., and the absence of *Clinapertina* spp., *Quadrimorphina profunda*, and *Aragonia* spp.; it is distinguished from Site 119 Oligocene assemblages by the dominance of *N. truempyi*, absence of *N. umbonifera*, and lower numbers of *C. ungerianus* sensu stricto. The higher abundance of buliminids (inlcuding *B. semicostata* and *B. impendens*), *Osangularia* spp. (mostly *O. mexicana*), and *Vulvulina* spp. (mostly *V. spinosa*) distinguish the shallower Site 401 from both the deeper Eocene and Oligocene of Site 119. These faunal differences result from both paleodepth ("intermediate" versus "deep") and age (Eocene versus Oligocene) differences between the two sites.

Comparison of Sites 119 and 401 with Tjalsma and Lohmann's (1982) Data

The close affinities of the Site 119 and 401 assemblages with Tjalsma and Lohmann's (1982) can be demonstrated. I computed loadings of Site 119 and 401 samples on their Eocene Q-mode principal components (Fig. 9). This procedure neglects taxa that are present only at Site 119 or 401. This should not substantially bias the analysis, for their Q-mode analyses emphasize abundant species. All of the abundant Eocene species at Site 119 are represented in

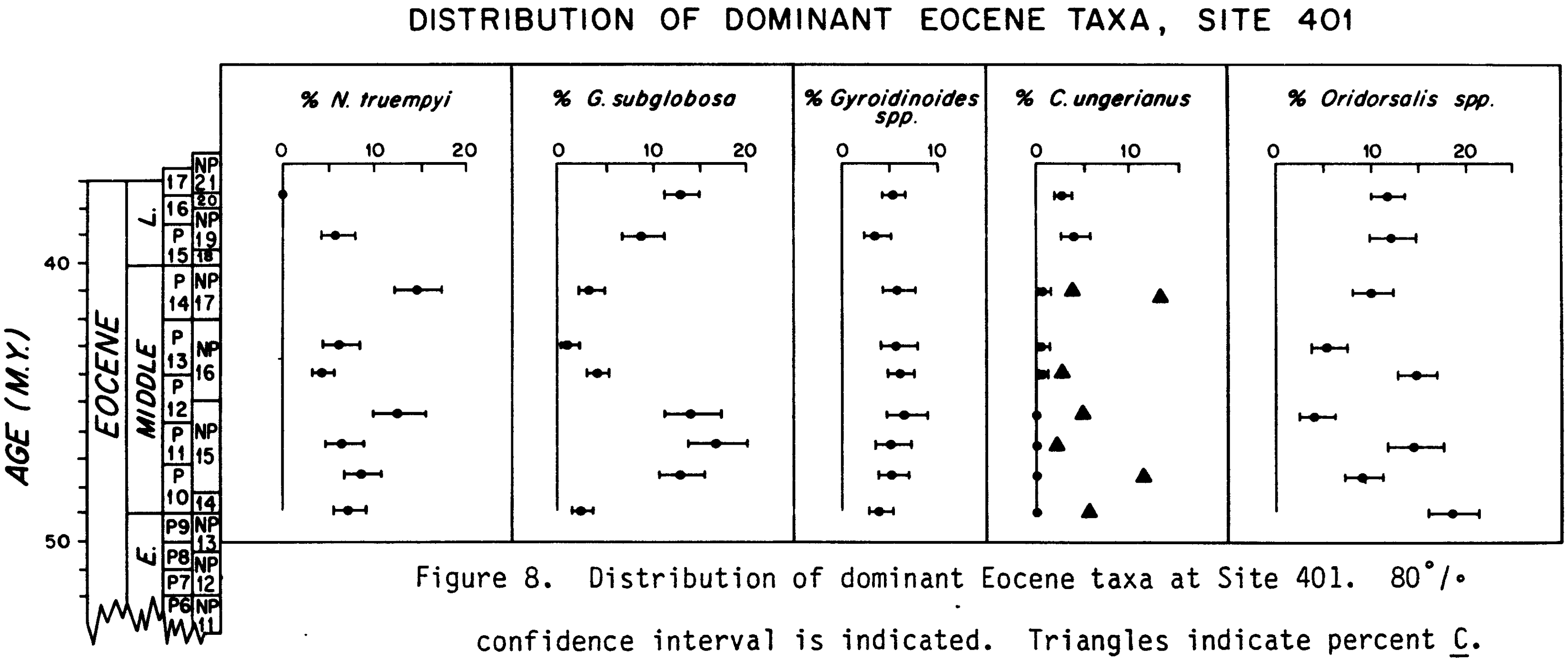

Figure 8. Distribution of dominant Eocene taxa at Site 401. 80°/o confidence interval is indicated. Triangles indicate percent _C. ungerianus_ plexus; dots indicate percent _C. ungerianus_ sensu stricto.

their samples, although several species which are abundant in the Oligocene of Site 119 (especially *N. umbonifera*) are not. The exclusion of these species de-emphasizes differences between samples of the two studies.

Tjalsma and Lohmann's (1982) first 3 principal components explain 71 °/° of the total observed variation accounted for in the estimation of loadings for the Site 119 and 401 samples. Their principal component 2 explains 26 °/° of the total observed variation. Principal component two represents the dichotomy between the *N. truempyi* assemblage (positive loadings, fig. 50; Tjalsma and Lohmann, 1982) and the *O. umbonatus*-*Globocassidulina subglobosa*-*Gyroidinoides*-*Cibicicdoides ungerianus*-*Lenticulina* assemblage (negative loadings). Site 119 Oligocene samples have high negative loadings on principal component 2 (Fig. 9), while Site 119 Eocene samples have moderate to high positive loadings (Fig. 9). Site 119 Oligocene samples have high positive loadings on their principal component 3. Principal component 3 represents the dichotomy between the *G. subglobosa*-*Gyroidinoides*-*C. ungerianus*-*Oridorsalis umbonatus* assemblage (positive loadings, fig. 53; Tjalsma and Lohmann, 1982) and the *Lenticulina*-*Bulimina*-*Osangularia* assemblage (negative loadings).

As with the Oligocene of Site 119, the Site 401 samples give high negative loadings on Tjalsma and Lohmann's (1982) principal component 2 (Fig. 9), corresponding to their *G. subglobosa*-*Gyroidinoides*-*C. ungerianus*-*O. umbonatus* assemblage. The dominance of *Oridorsalis* and *G. subglobosa*, together with the lower abundance of *N. truempyi*, and the absence of deep-Eocene taxa (*Clinapertina*, *Abyssamina*, etc.) make the Site 401 Eocene assemblages more similar to the deep-Oligocene than to the deep-Eocene assemblages. Schnitker (1979) came to a similar conclusion. Using cluster analysis, he noted that 3 Eocene to Oligocene assemblages occurred in Bay of Biscay: a deep-Eocene, a

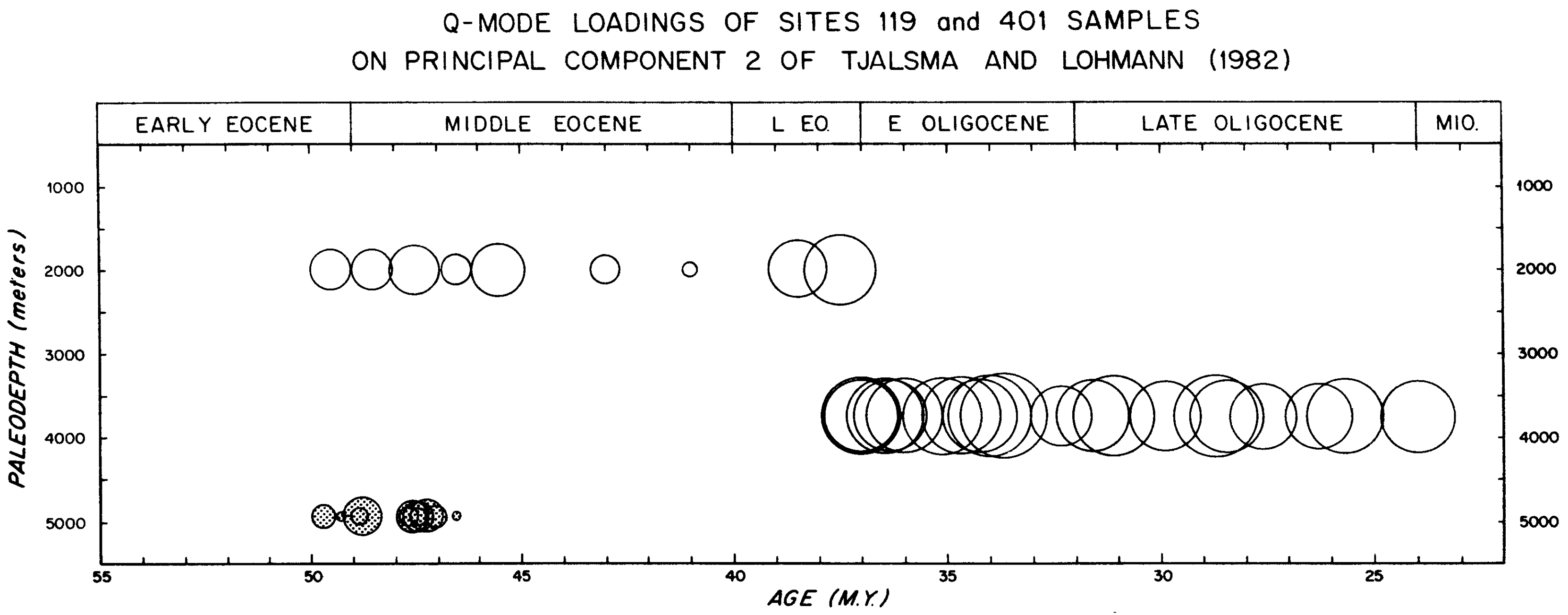

Figure 9. Distribution of loadings of Site 119 and 401 samples on Principal Component 2 of Tjalsma and Lohmann (1982). Positive loadings indicated with stipples. Size of circle is proportional to loadings; largest circle corresponds to loadings of 0.72

deep-Oligocene (both Site 400A), and a shallower-Eocene (Site 401) assemblage. Based upon their hierachial rankings, he suggested that the Oligocene assemblage was most closely related to the shallower-Eocene assemblage. Nevertheless, it should be remembered that the Site 401 assemblages are taxonomically distinct, both in species abundance and composition, from both the deep-Eocene and Oligocene assemblages.

The apparent similarity between the the deep Oligocene and shallower Eocene can be attributed to the continued dominance of the wide-ranging (bathymetrically) and long-ranging (stratigraphically) taxa including Oridorsalis spp., G. subglobosa, Gyroidinoides spp., and the C. ungerianus plexus. These taxa are abundant in all three assemblages. Thus the similarity cannot be attributed to migration of shallower Eocene species to the deep Oligocene as suggested by Schnitker (1979), for the mentioned species were also present in and important faunal constituents of the deep-Eocene assemblages. Instead, such wide-ranging and long-ranging taxa expanded in abundance in the early Oligocene following the demise of the deep-Eocene assemblage.

Comparing percentage distribution data of Tjalsma and Lohmann (1982) with Site 119 and 401 data shows that G. subglobosa and Gyroidinoides spp. apparently occur in greatest abundance at 2000-3000 m in the Eocene; by the Oligocene, these taxa increase in abundance below 3000 m (Figures 10, 11). Despite taxonomic problems in the identification of species in the C. ungerianus plexus, this group appears to follow a similar pattern (Fig. 12). This faunal pattern probably represents the increase in abundance of these taxa in the deepest sites due to the extinction of the deep-water Eocene fauna. The depth-time plot of Oridorsalis spp. (Fig. 13) is less clear, for it appears that it is abundant throughout the range of paleodepths (1-5km) and ages (Eocene-Oligocene) considered.

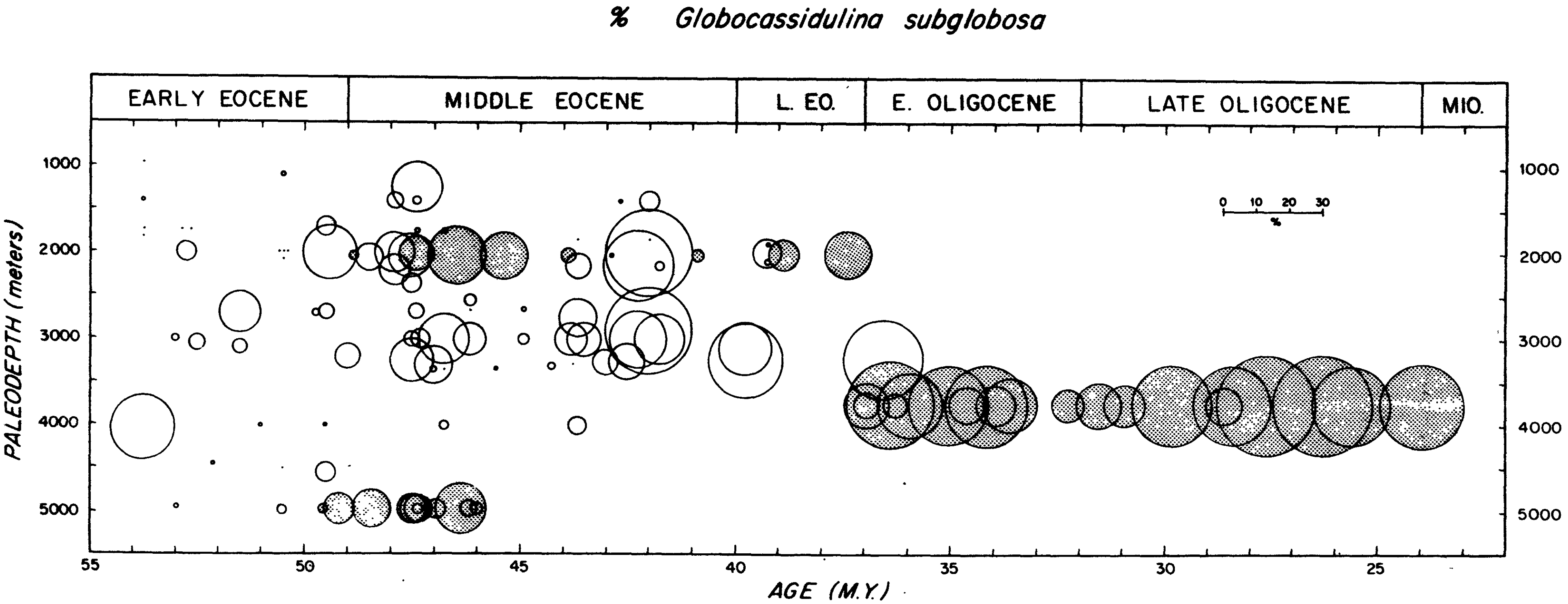

Figure 10. Distribution of Globocassidulina subglobosa. Open circles from Tjalsma and Lohmann (1982); stippled circles from Sites 119 and 401.

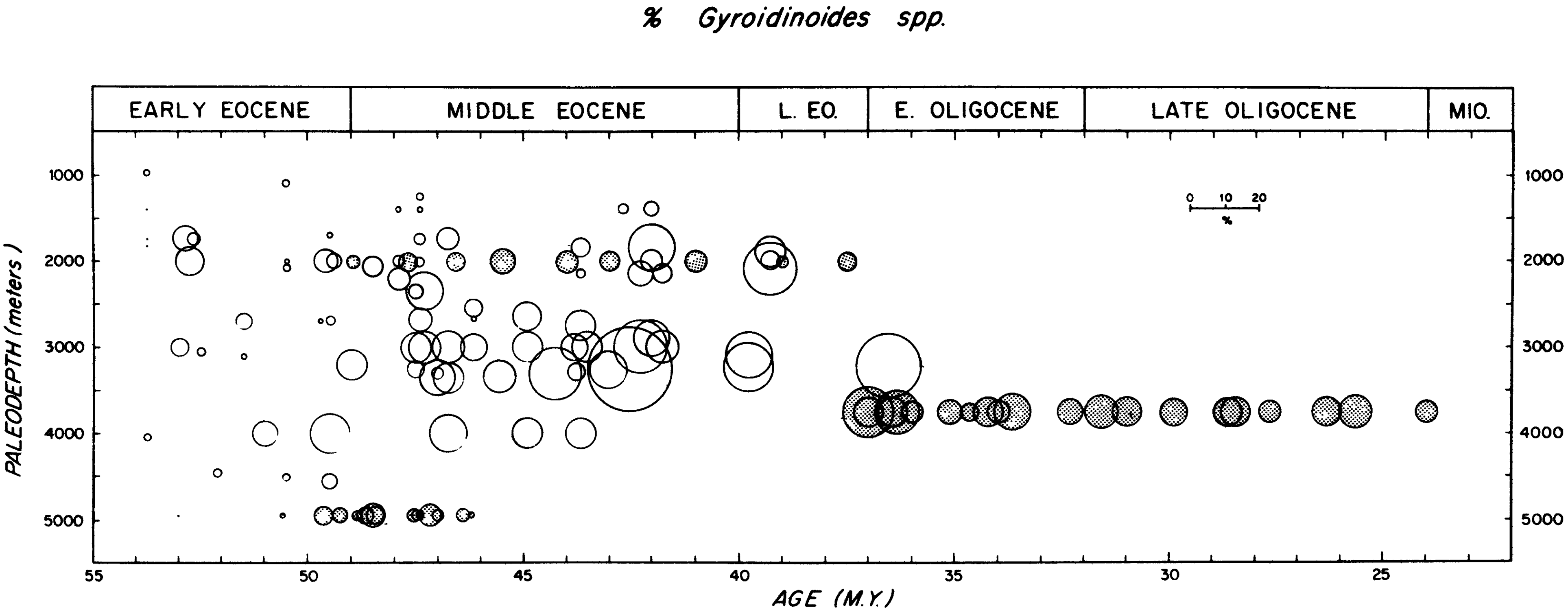

Figure 11. Distribution of Gyroidinoides spp. Open circles from Tjalsma and Lohmann (1982); stippled circles from Sites 119 and 401.

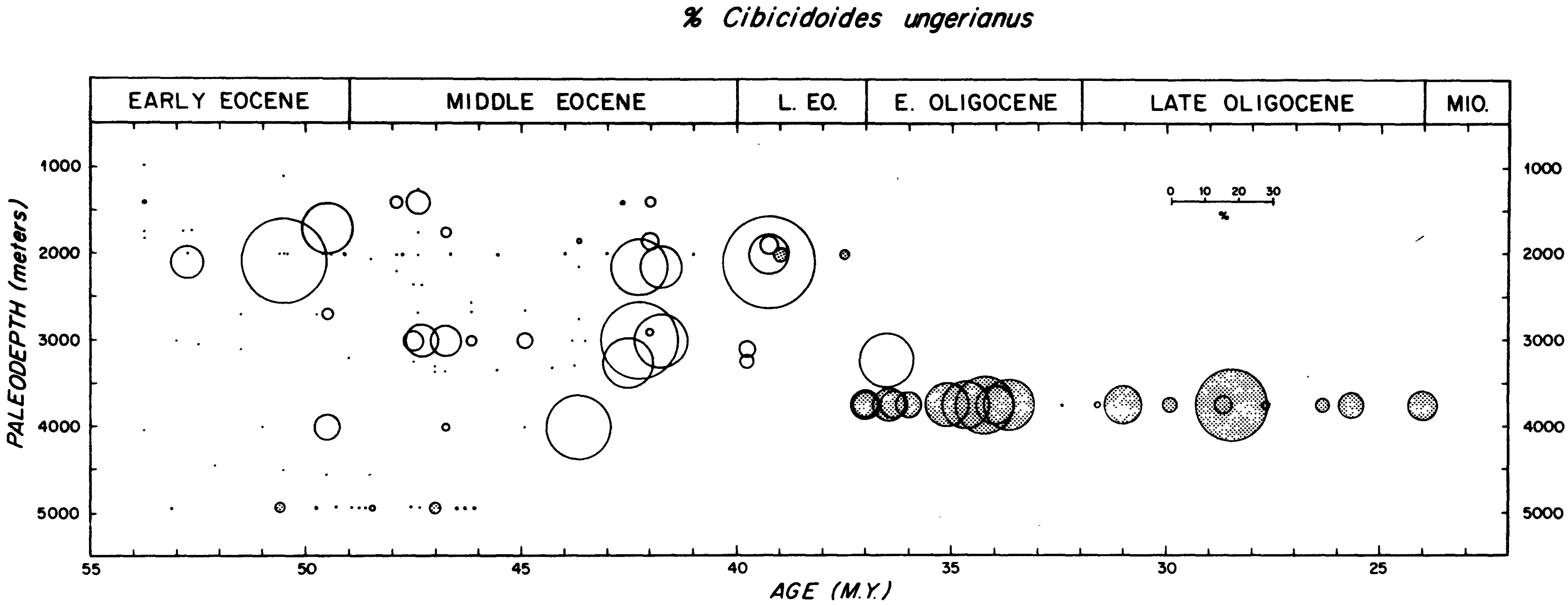

Figure 12. Distribution of Cibicidoides ungerianus. Open circles from Tjalsma and Lohmann (1982); stippled circles from Sites 119 and 401.

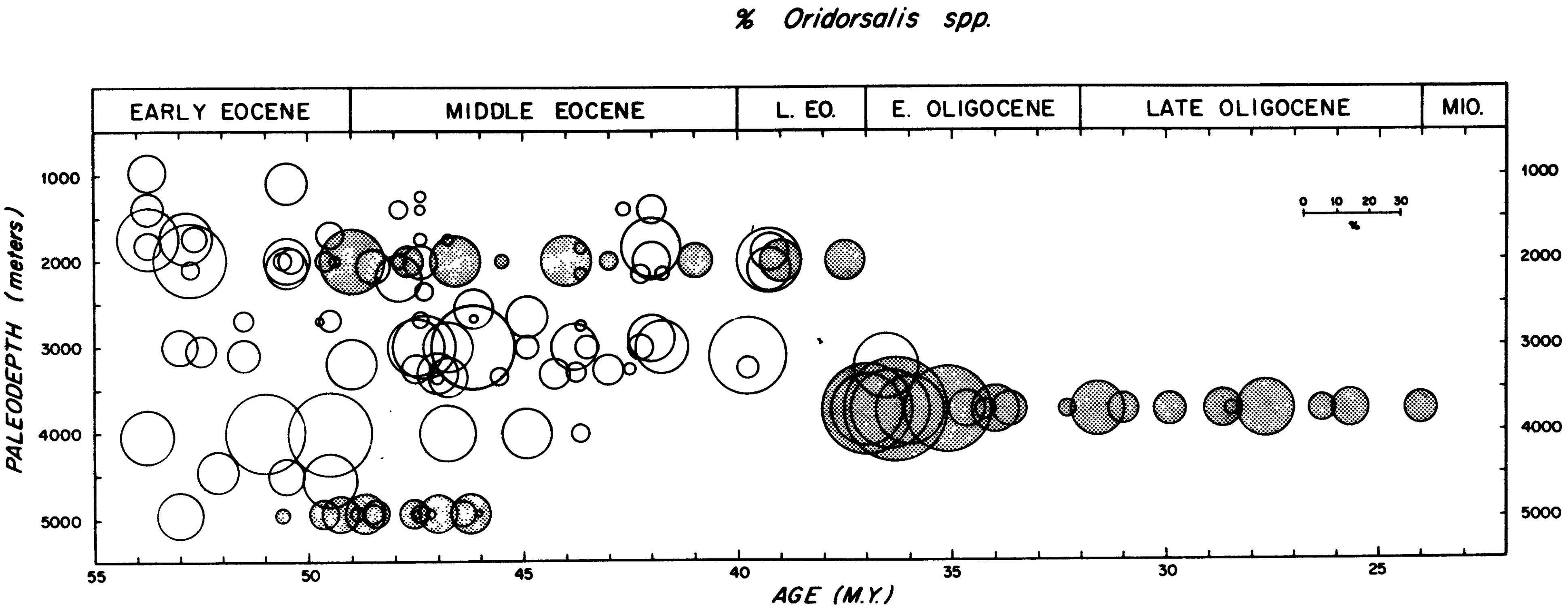

Figure 13. Distribution of Oridorsalis spp. Open circles from Tjalsma and Lohmann (1982); stippled circles from Sites 119 and 401.

DISCUSSION

Distribution of *Nuttallides* spp.: Correlations with Isotopic Data

The core-top distribution of *N. umbonifera* in the modern ocean has been correlated with decreasing temperature, increasing alkalinity, silica, nutrients, depth, and the distribution of modern Antarctic Bottom Water (Schnitker, 1974; Lohmann, 1978; Corliss, 1978; Bremer and Lohmann, in press). Bremer and Lohmann have shown, however, that the only interregionally consistent correlation is between the abundance of this species and corrosiveness of bottom water. Tjalsma and Lohmann (1982) extrapolated a similar water mass relationship for *N. truempyi* and noted that the abundance of this species is negatively correlated with the frequency of hiatuses in the Atlantic (Moore et al., 1978). From this extrapolation and correlation, they speculated that the *N. truempyi* assemblage appeared and expanded as a response to sluggish deep-ocean circulation. The extrapolation of *N. umbonifera*'s environmental relationship to *N. truempyi* seems reasonable, for *N. umbonifera* appears to have expanded to replace *N. truempyi* as the predominant deep (> 3 km) benthic foraminifera in the Oligocene of the Bay of Biscay (Figures 4, 5, and 14).

There is a strong negative correlation between $\delta^{13}C$ of benthic foraminifera and the abundance of *Nuttallides* spp. (Fig. 14). The lowest $\delta^{13}C$ values noted (early middle Eocene, ~ 0.4 °/°° PDB; middle Oligocene, ~ 0.0 °/°°) are associated with the highest percentages of *N. truempyi* and *N. umbonifera*, respectively. Conversely, the lowest percentages of *Nuttallides* spp. (0-5.2 °/°) occur in the early Oligocene when $\delta^{13}C$ values are highest (Fig. 14), the CCD is deepest, carbonate preservation

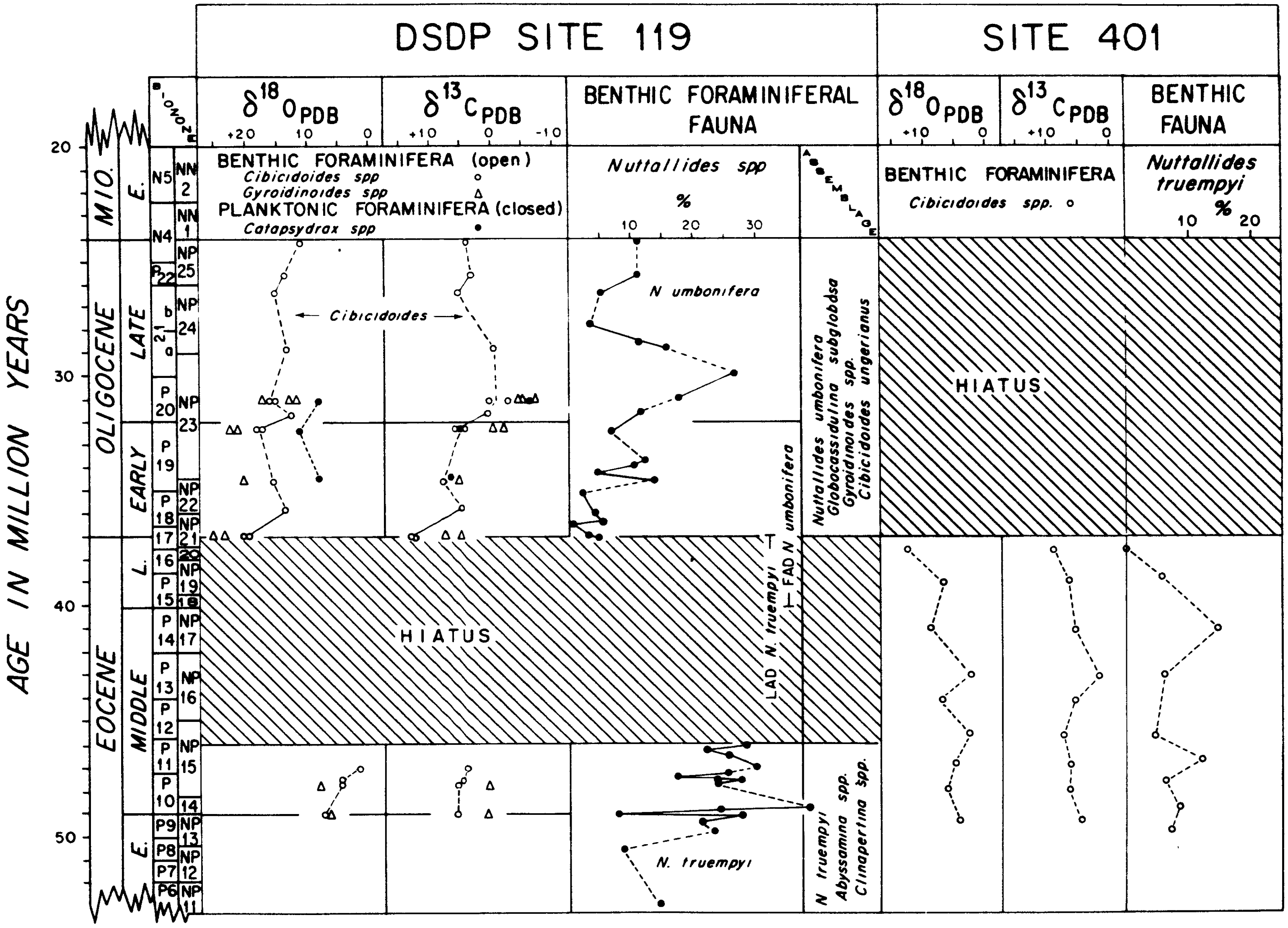

Figure 14. Isotopic composition (after Miller and Curry, in press) and distribution of Nuttallides spp. at Sites 119 and 401.

is best (Heath, 1969; Berger, 1973; van Andel et al., 1975; Kennett and Shackleton, 1976), and bottom-water circulation is vigorous in the North Atlantic (Miller and Tucholke, in press).

The correlation of abundant *Nuttallides* spp. with low $\delta^{13}C$ values is expected if *Nuttallides* is associated with the oldest, most corrosive, and sluggish bottom water as Tjalsma and Lohmann (1982) suggested. Lower $\delta^{13}C$ values are associated with older water masses (Kroopnick et al., 1972; Kroopnick, 1974, 1980), and are strongly correlated with greater dissolution (Lohmann and Carlson, 1981). Still, abyssal variations in $\delta^{13}C$ are not attributable solely to variations in age of water masses, for lower $\delta^{13}C$ values can result from increased input of organic matter from greater terrigenous input during sea level lowstands (Shackleton, 1977).

Nuttallides umbonifera reached acme in the middle Oligocene of Site 119 (Fig. 14). This acme is associated with minimum $\delta^{13}C$ values, but oxygen isotopic composition of benthic foraminifera remains constant. At least two interpretations of this phenomenon are possible:

1). The increase in *N. umbonifera* and decrease in $\delta^{13}C$ are associated with a general decrease in bottom water flow from the early Oligocene, and therefore result from older, more corrosive bottom water.

2). The decrease in $\delta^{13}C$ occurs in response to greater input of light terrestrial organic matter into the oceanic system as a result of a major lowstand of sea level in the middle Oligocene (Vail et al., 1977), while *N. umbonifera* may have increased in response to a number of factors including increased organic matter which may have increased carbonate solubility.

The latter hypothesis is less likely given the lack of a concomitant change in $\delta^{18}O$ in the middle Oligocene. In general, the sea level curve of Vail et al. (1977) shows a good correspondence to the $\delta^{18}O$ curve for the

Tertiary (Vail, 1981). This is expected since lowered sea level results in increased continentality, hence less equable climates. In addition, the $\delta^{18}O$ curve reflects the volume of continental ice buildup. At the present time there is no known mechanism of lowering or raising sea level at the rates and magnitudes required by Vail et al. (1977) and Hardenbol et al. (1981) other than the waxing and waning of continental ice sheets (Vail and Hardenbol, 1979, 1981). Thus, a major sea level lowering, such as the middle Oligocene, should be reflected in a major enrichment of ^{18}O.

Olsson et al. (1980) suggested that the sea level curve of Vail et al. (1977) is miscorrelated, and that the major sea level drop occurred near the Eocene/Oligocene boundary. In fact, a major ^{18}O enrichment occurs near this boundary (Shackleton and Kennett, 1975; Savin et al., 1975; Keigwin, 1980; Miller and Curry, in press), supporting independent indicators of cooling climates (see Table 1 in Miller and Tuckolke, in press) and a suggested buildup of continental ice sheets (Matthews and Poore, 1980). The $\delta^{18}O$ increase occurs in Site 119 between the early middle Eocene and earliest Oligocene (Fig. 14). Combining the isotopic records of Site 401 and Site 119 suggests that this temperature drop occurred in the Bay of Biscay in the late Eocene to earliest Oligocene (Miller and Curry, in press). This enrichment represents a bottom-water temperature drop of as much as 6°C (ice-free world) or as little as 2°C (assuming buildup of early Oligocene ice volume equivalent to modern ice volume; Miller and Curry, in press). In any case, the major sea level drop of Vail at al. (1977) is not associated with an ^{18}O enrichment, while the minor drop in sea level near the Eocene/Oligocene boundary is associated with a major increase in $\delta^{18}O$. The lack of correspondence between the the late Eocene to middle Oligocene sea level curve and the $\delta^{18}O$ record is striking.

Further documentation is needed to determine whether the acme of *N. umbonifera* was widespread in the Bay of Biscay. The percentage of *N. umbonifera* was computed from fourteen samples from the latest Eocene to Oligocene section of nearby Site 400A (Table 5). Site 400A was drilled in 4399m water depth at the foot of the Meriadzek escarpment (Montadert, Roberts, et al., 1979; Fig. 1). Although no Oligocene hiatuses were originally reported at this site, examination of the biostratigraphic data (Table 5) suggests that a hiatus of ~ 3 my may be present the middle Oligocene within Zone NP23 (with Zone NP24/25 occuring in sample 45-3 and Zone NP22 found in sample 45-CC; Muller, 1979; M.-P. Aubry, personal communication). This hiatus is also apparent when plotting biostratigraphic datums for Site 119 and Site 400A on a depth-depth diagram (Shaw, 1964) . Nevertheless, as with Site 119, the biostratigraphic resolution in the middle Oligocene section is poor, and the detection of any middle Oligocene hiatus is tentative. A maximum of *N. umbonifera* occurs in the middle Oligocene of Site 400A, but the peak is less well defined and the distribution more sporadic than at Site 119 (Fig. 15). Also of interest is the high percentage of *N. umbonifera* in sample 47-2 which correlates with similar percentages of *N. umbonifera* in the latest Eocene of Site 277 on the Campbell Plateau (Corliss, 1979); however, drilling disturbance makes any results involving samples 46-6 through core 47 tenuous (Table 5). Further biostratigraphic, isotopic, and paleoceanographic studies from other regions are needed to determine whether the middle Oligocene and latest Eocene peaks in the abundance of *N. umbonifera* are correlateable, interregional phenomena resulting from changes in abyssal circulation or are local perturbations.

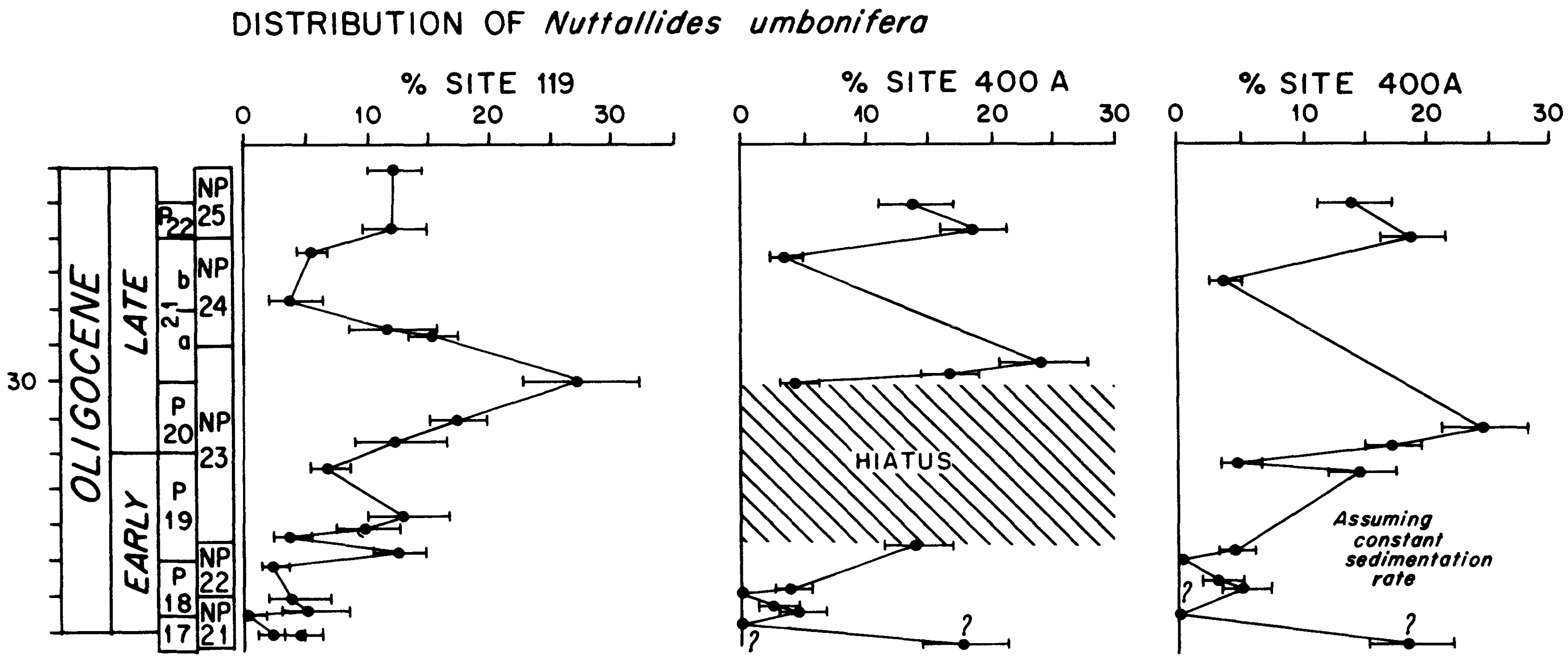

Figure 15. Distribution of Nuttallides umbonifera from the latest Eocene and Oligocene of Sites 119 and 400A. 80°/∘ confidence interval is indicated. The second plot assumes a hiatus in Site 400A between core 45-5 and 45-CC; the third assumes a constant sedimentation rate.

Timing and Nature of Faunal Change

The faunal changes from Sites 119 and 401 can be summarized as follows: 1) between the early middle Eocene and the earliest Oligocene a N. truempyi assemblage was replaced by an assemblage dominated by long- and wide-ranging taxa, Oridorsalis spp., G. subglobosa, Gyroidinoides spp., and C. ungerianus, 2) N. umbonifera appeared between the middle Eocene and earliest Oligocene, 3) N. umbonifera increased in the Oligocene, attaining a well-documented middle Oligocene maximum, and 4) N. truempyi became extinct in the late Eocene.

The major faunal differences between the assemblages studies herein (shallower middle-late Eocene, Site 401; deep early-middle Eocene, Site 119; deep Oligocene, Site 119) are attributable to differences in paleodepth (or properties which tend to covary with depth) and age. The major faunal change is the extinction of important elements of the deep-Eocene assemblage as Schnitker (1979) and Tjalsma and Lohmann (1982) have previously concluded. Unfortunately, the late middle to latest Eocene hiatus in Site 119 (and a similar hiatus in Site 400A) prevents a determination of the timing of the replacement of the N. truempyi-dominated assemblage or the appearence of N. umbonifera in this region. Schnitker suggested that the deep-Eocene assemblage subsequently was replaced by newly evolving species or by migration of taxa from shallower depths. This is not strictly the case, for although some major Oligocene faunal constituents originated when the deep-Eocene assemblage disappeared (e.g. N. umbonifera and perhaps Epistominella exigua), the deep-Oligocene fauna consisted primarily of long-ranging and wide-ranging taxa which are important in both the shallower and deeper Eocene environments.

Tjalsma and Lohmann (1982) noted that the replacement of the N. truempyi assemblage had taken place by the late Eocene in several Caribbean and South

Atlantic localities and that the replacement coincided with an enrichment in benthic foraminiferal ^{18}O in the South Atlantic (Site 357) noted by Boersma and Shackleton (1977). However, the late Eocene to early Oligocene record was poorly recovered at Site 357, with only one late Eocene sample available; thus the timing of the ^{18}O enrichment is not well constrained in the South Atlantic. The replacement of the N. truempyi assemblage in the South Atlantic and Caribbean does not correlate with the ^{18}O enrichment in the Pacific/Southern Ocean (Shackleton and Kennett, 1975; Savin et al., 1975) or Bay of Biscay (Miller and Curry, in press). This raises the questions of whether the deep faunal change (> 3 km) correlates with the ^{18}O enrichment in each ocean basin and whether it is diachronous between ocean basins.

To approach these questions in the Bay of Biscay, Miller and Curry (in press) combined isotopic records at Sites 119 and 401, and Schnitker (1979) combined benthic faunal records at Site 400A and 401 in order to obtain complete stratigraphic records. They concluded that the changes in isotopic composition and fauna, respectively, occurred near the Eocene/Oligocene boundary. Combining records in this manner appears to be permissible for the isotopic record, but not for the faunal record. Both sites yield the same oxygen and carbon isotopic values in the sections of stratigraphic overlap (Miller and Curry, in press) and the modern temperature-salinity values of these locations are similar (< 1°C; < 0.5 ‰ salinity; Fuglister, 1960; corresponding to < 0.50 ‰ $\delta^{18}O_{PDB}$). However, different assemblages are found in the shallow and deep sites, both presently (Pujos-Lamy, 1975) and in the Eocene (this paper). Therefore, one should not combine the faunal records in this manner.

No major assemblage changes were observed in the shallower Site 401. However, N. truempyi became extinct between Zones P15/16 and the G.

cerroazulensis *cunialensis* Zone (latest Eocene). The absence of Oligocene sediments at Site 401, prevents a documentation of Eocene to Oligocene changes at a shallower site. The recent recovery of relatively complete Eocene through Oligocene sections from shallow sites in the Bay of Biscay by DSDP/IPOD Leg 80 may provide this documentation.

Paleoceanographic Synthesis

During the early Eocene climatic optimum (Haq, 1981), bottom water flow was at an ebb in the North Atlantic. As a result, the CCD remained high, $\delta^{13}C$ values remained low, and *N. truempyi* dominated the benthic foraminiferal assemblages in the Bay of Biscay. General climatic cooling began in the middle to late Eocene, culminating in a sharp cooling in the latest Eocene to earliest Oligocene. Climatic coolong was accompanied by the replacement of the *N. truempyi* assemblage and the ultimate demise of *N. truempyi* and several other Eocene taxa. The first formation of cold, vigorously circulating bottom water of northern (Miller and Tucholke, in press) and southern (Kennett and Shackleton, 1976) origin was associated with this climatic cooling. The influence of southern bottom water on the deep Bay of Biscay was probably minimal due to its inhibition by the Azores-Biscay Rise (Fig. 1), Azores-Gibraltar Fracture Zone/Rise, and the Madeira-Torre Rise. The invigorated global bottom-water circulation resulted in more rapid oceanic turnover (Kennett and Shackleton, 1976), which in turn, increased carbonate preservation and enriched ^{13}C in benthic foraminifera.

The early Oligocene was characterized by widespread erosion throughout the North Atlantic. The unconformable sedimentary and seismic sequences of the northern North Atlantic can be traced to their correlative conformities (= sequence boundary, sensu Vail et al., 1977) in the late Eocene to earliest Oligocene in the Rockall Plateau region (Site 406 and 116) and in the early Oligocene in the Labrador Sea (Site 112; Miller and Tucholke, in press). In the Bay of Biscay, the sequence may become conformable near the Eocene/Oligocene boundary (ie. the overlapping section between Sites 119 and 401 is Zone NP21). Associated with this sequence boundary elsewhere in the northern North Atlantic is a horizon that separates pelagic deposition below from current-influenced deposition above, reflector R4. However, reflector R4 has not been traced from the Rockall Plateau region into the Bay of Biscay. In any case, reflector R4 and the overlying sequence probably represents the initiation of strong bottom water flow in the northern North Atlantic including the Bay of Biscay (Miller and Tucholke, in press).

Miller and Tucholke (in press) noted that the reflector R3 to R4 seismic sequence in the Rockall Plateau region is predominantely erosional in nature, containing chaotic reflectors with numerous unconformities (Roberts, 1975), and dated reflector R3 as middle to late Oligocene. The reflectors of the overlying R3 to R2 (late Oligocene to early Miocene) sequence are more conformable. This is interpreted as reflecting a decrease in bottom water flow in the North Atlantic begining at the time of reflector R3 (Miller and Tucholke, in press). The middle Oligocene reduction in bottom water flow inferred from the seismic record of the Rockall region correlates with the increase in age of bottom water, indicated by the faunal and isotopic record in the Bay of Biscay.

CONCLUSIONS

1. Three benthic foraminiferal assemblages were noted in Sites 119 and 401: shallower middle-late Eocene, deep early-middle Eocene, and deep Oligocene.

2 A major faunal change occurs between the early middle Eocene and earliest Oligocene at two deep (> 3 km) sites in the Bay of Biscay. *Nuttallides truempyi*, *Clinapertina* spp., and *Abyssamina* spp., which dominated the Eocene deep-sea benthic assemblage, are replaced by an increase in abundance of the wide- and long-ranging taxa *Oridorsalis* spp., *Globocassidulina subglobosa*, *Gyroidinoides*, and *C. ungerianus* plexus.

3. The faunal replacement is similar to that noted by Tjalsma and Lohmann (1982) in the Caribbean and South Atlantic by the late middle Eocene. However, the stratigraphic record recovered in the deep Bay of Biscay is insufficient to resolve further the timing of the replacement in this region.

4. *Nuttallides umbonifera* first appeared in the earliest Oligocene of Site 119 above the middle Eocene to lowermost Oligocene unconformity. This species reaches maximum abundance in the middle Oligocene. This is interpreted to reflect increased corrosiveness of bottom water.

5. The extinction of *N. truempyi* is the only major faunal change noted within the Eocene of the shallowest (~ 2 km) site in the Bay of Biscay. Unfortunately, the absence of Oligocene sediments there and the compositional differences between the deep and shallow assemblages prevent further determination of the timing of faunal changes.

6. The abundance of *Nuttallides* spp. is negatively correlated with $\delta^{13}C$ in benthic foraminifera. This is though to reflect the covariance of older water with lower $\delta^{13}C$, older water with increased corrosiveness, and corrosiveness with abundance of *Nuttallides* spp.

7. The faunal, isotopic, and seismic evidence indicate that:

a. Old, warm, corrosive, and sluggish bottom water of the Eocene was replaced by younger, colder, less corrosive, and more vigorously circulating bottom water by the early Oligocene. The source of this bottom water was from the north (Norwegian-Greenland Sea or Arctic Ocean).

b. During the Oligocene, bottom water circulation was reduced, and the age and corrosiveness of bottom water increased.

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Figure 16. Uplift of Site 119.

APPENDIX I

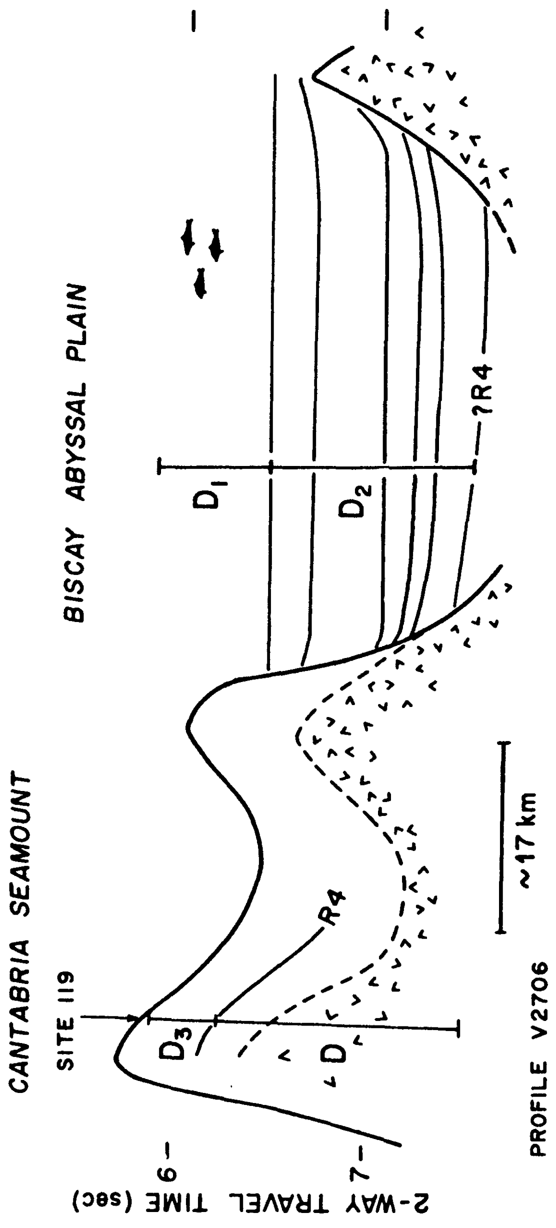
Backtracking and Paleodepths

Berger and Winterer (1974) and Sclater et al. (1977) summarized the use of empirical age-versus-subsidence curves (Sclater et al., 1971; Parsons and Sclater, 1977) to determine the paleodepth of oceanic crust. This backtracking method assumes simple thermal subsidence. However, Cantabria, Charcot, and Biscay Seamounts and other portions of the Azores-Biscay Rise (Fig. 1), were uplifted above surrounding seafloor as a result of the Pyrenean Orogeny (Laughton, Berggren et al., 1972). The uplift is apparently correlative with the middle to latest Eocene hiatus at Site 119 (fig. 17 in chapter 10, Laughton, Berggren et al., 1972), although termination of turbidite deposition in cores 24-25 suggests that the uplift of Cantabria may have begun by the early Eocene.

Even though Cantabria was uplifted in the Eocene, limits can be calculated for paleodepths. Site 119 lies within the Cretaceous Magnetic Quiet Zone close to anomaly 34 (Schouten, personal communication); crustal age therefore is 90 my ± 10 my. Assuming subsidence along the empirical North Atlantic age-depth curve for crust of 90 my (Tucholke and Vogt, 1979) until uplift (~ 49 to 37 my), early Eocene paleodepths are ~ 4700m, uncorrected, or ~ 4950m when corrected for 410m post-early Eocene sediments. Due to the exponential shape of the age-subsidence curve, the age uncertainty of ~ 20 my does not result in a major depth uncertainty. However, this method assumes that the crust formed at ~ 2700m of water; this is reasonable due to the low residual depth anomalies calculated for this region (Sclater et al., 1975).

It is possible to obtain estimates of the uplift of Site 119, which presently lies ~ 450m above the adjacent abyssal plain. The turbidites in Site 119 were deposited contiguous with abyssal plain turbidites (i.e. prior to uplift). The difference between the depth of these two surfaces represents the minimum uplift. A prominent seismic horizon that correlates with the middle to late Eocene unconformity (0.4 sec. subbottom, Site 119; Fig. 16) is approximates the top of the turbidites, although the top may actually underlie this horizon by 50m (0.05 sec). This horizon correlates with reflector R4, a horizon which separates current-controlled sedimentation above from pelagic sedimentation below in the northern North Atlantic (Roberts, 1975; Miller and Tucholke, in press). Examination of seismic profiles of Lamont-Doherty Geological Observatory and Laughton, Berggren et al. (1972) show that reflector R4 probably lies at ~ 1.07 seconds subbottom under the adjacent abyssal plain; the unconformity at Site 119 therefore lies ~ 1200m above the correlative horizon (Fig. 4). Thus, Oligoene paleodepth is estimated to be ~ 3750m. Due to assumptions involved, these estimates may be in error by several hundred meters; still, faunal comparisons with Tjalsma and Lohmann (1982) and Tjalsma (personal communication) show that Site 119 assemblages are, in general, found below ~ 3000m and that this site probably was situated in abyssal depths throughout the Eocene and Oligocene.

Site 401 was drilled in 2495m water on continental crust of the Meriadzek Terrace (Montadert et al., 1979), and therefore cannot be "backtracked" in the manner of normal oceanic crust. However, constraints can be placed upon the paleodepth. Coral indicative of shallow-water conditions were identified in the Porlandian/Kimmeridgian section of Site 401. By the upper Albian, deep-water calcareous oozes were deposited (Montadert, Roberts, et al., 1979). The Campanian-Maestrichtian section contains chalks where planktonic foraminifera constitute 99 °/° of the foraminiferal assemblages; these chalks have been interpreted as lower bathyal (> 1000m). If a simple thermal cooling model is assumed, then the subsidence rate would decay exponentially, although not necessarily with the same constants as oceanic crust; such an assumption seems valid for the Armorican margin (Montadert, et al., 1979). Assuming simple thermal subsidence, water depths of near sea level in the Kimmeridgian, of ~ 1000m in the Maestrictian, and of 2500m presently, the Eocene-Oligocene section would be ~ 2000m. This agrees well with the paleodepth estimates of Schnitker (1979) who concluded that Paleogene benthic foraminiferal faunas of Site 401 compared well with faunas from other DSDP sites with paleodepths between 1800 and 3000m. Faunal comparisons with Tjalsma and Lohmann (1982) also show that the Site 401 assemblages are found from ~ 2000 to 2500m.



$$D = D_1 + D_2 - D_3$$

$$D = (0.6\text{ sec})(1.53\text{ km/sec})/2 + (1.07\text{ sec})(2.19\text{ km/sec})/2 - (0.4\text{ sec})(2.19\text{ km/sec})/2$$

$$D \approx 1200\text{ m}$$

Figure 16. Uplift of Site 119.

APPENDIX II

TAXONOMIC NOTES

For a more detailed description and illustation see Tjalsma and Lohmann, 1982.

Abyssamina poagi Schnitker and Tjalsma. 1980, p. 237, pl. 1, figs. 7-9. Tjalsma and Lohmann, 1982, pl. 19, figs. 3a-4c.

Restricted to the Eocene. Very rare in Site 401. Comprises 2 to 24 °/° of the benthic foraminiferal fauna in the Eocene of Site 119.

Abyssamina quadrata Schnitker and Tjalsma. 1980, p. 237, pl. 1, figs. 1-6. Tjalsma and Lohmann, 1982, pl. 4, fig.5, pl. 19, figs. 1a-2c.

This straight-sutured form occurs rarely in the early-middle Eocene of Site 119.

Alabamina dissonata (Cushman and Renz). *Pulvinulinella atlantisae* Cushman var. *dissonata* Cushman and Renz, 1948, p. 35, pl. 7, figs. 11-12. Tjalsma and Lohmann, 1982, pl. 17, figs. 3a-b, pl. 20, fig. 5.

Restricted to the Eocene. Very rare in Site 401. Rare in the early Eocene of Site 119; up to 6 °/° in the middle Eocene.

Anomalina spissiformis Cushman and Stainforth. *Anomalina alazanensis* Nuttall var. *spissiformis* Cushman and Stainforth, 1945, p. 71, pl. 14, figs. 5a-c. Tjalsma and Lohmann, 1982, pl. 20, figs. 4a-c.

Occurring in all 3 assemblages with highest abundances in the shallow Eocene of Site 401 (up to 6 °/°) and deep Oligocene of Site 119 (up to 9 °/°).

Aragonia capdevilensis (Cushman and Bermudez). *Bolivina capdevilensis* Cushman and Bermudez, 1937, p. 14, pl. 1, figs. 49-50. Tjalsma and Lohmann, 1982, pl. 11, figs 3a-b.

Aragonia semireticulata (LeRoy). *Bolivina semireticulata* LeRoy, 1953, p. 20, pl. 8, fig. 26. Tjalsma and Lohmann, 1982, pl. 11, figs. 1a-b.

Aragonia spp. are restricted to the early-middle Eocene of Site 119, where they constitute up to 7 °/° of the fauna

Astrononion pusillum Hornibrook. 1961, p. 96, pl. 12, figs. 229, 236.

Restricted to the Oligocene of Site 119. Rare in the early Oligocene; up to 5 °/° of the late Oligocene fauna. Early Oligocene forms tend to have more chambers (> 7) than the type, and are assigned to cf. *pussillum*. My specimens differ from the illustated type specimen in that in axial view the early chambers of the last whorl are more pinched relative to the final chambers and the umbilicus tends to be more open.

Bolivina huneri Howe. 1939, p. 66, pl. 9, figs. 3-4. Tjalsma and Lohmann, 1982, pl. 11, figs 5a-b.

Very rare in Eocene of Site 119; absent from Eocene of Site 401. Constitutes up to 3 °/° of the Site 119 Oligocene fauna.

Bulimina cf. *alazanensis* Cushman. *Bulimina* cf. *alazanensis* Cushman, 1927, p. 161, pl. 25, fig. 4. Tjalsma and Lohmann, 1982, pl. 14, fig. 4.

Occurs in all three assemblages. Specimens from Site 119 differ from the typical form in that they are weakly triangular to rounded in sections and more elongated. The lack of a distinctly triangular test may be an abyssal variation. Rare.

Bulimina glomarchallengeri Tjalsma and Lohmann. 1982, pl. 13, figs. 8-12c.

Previously reported only from high latitude South Atlantic by Tjalsma and Lohmann (1982), this species is restricted to the Eocene of Site 401. Rare (< 1 °/°).

Bulimina impendens Parker and Bermudez. 1937, p. 514, pl. 58, figs. 7a-c, 8. Tjalsma and Lohmann, 1982, pl. 14, figs. 2a-b.

Not differentiated from *B. trinitatensis*. Rare in the early-middle Eocene of Site 119; forms up to 10 °/° of the Site 401 Eocene fauna.

Bulimina jarvisi Cushman and Parker. 1936, p. 39, pl. 7, figs. 1a-c. Tjalsma and Lohmann, 1982, pl. 13, figs. 4-5b.

This species is very difficult to differentiate from *B. semicostata*, for in samples 401: 3-1 and 4-CC (late middle to early late Eocene), *B. jarvisi* tends to be triangular in section. Found only in P14 to latest Eocene of Site 401 where it forms up to 6 °/° of the fauna.

Bulimina macilenta Cushman and Parker. 1936, p. 42, pl. 7, figs. 7a-8c. *Bulimina macilenta* Cushman and Parker, 1939 (new name), p. 93. Tjalsma and Lohmann, 1982, pl. 14, fig. 3.

Restricted to middle Eocene of Site 401 where it comprises up to 3 °/° of the fauna.

Bulimina semicostata Nuttall. 1930, p. 285, pl. 23, figs. 15-16. Tjalsma and Lohmann, 1982, pl. 13, figs. 1-3.

Restricted to Site 401 Eocene where it forms up to 6 °/° of the fauna. Forms assigned to *B.* cf. *semicostata* (see Tjalsma and Lohmann, 1982, pl. 13, figs. 6-7) were noted in sample 401: 6-3.

Bulimina trinitatensis Cushman and Jarvis. 1928, p. 102, pl. 14, figs. 12a-b. Tjalsma and Lohmann, 1982, p. 10, pl. 3, figs. 3-4; pl. 14, fig. 1

Bulimina tuxapamensis Cole. 1928, p. 212, pl. 32, fig. 23. Tjalsma and Lohmann, 1982,

Found in the middle Eocene of Sites 119 and 401. Rare.

Buliminella grata Parker and Bermudez, 1937, p. 515, pl. 59, figs. 6a-c.

B. grata s.s. (Tjalsma and Lohmann, 1982, pl. 12, figs. 7a-b) occurs along with *B. grata spinosa* (Tjalsma and Lohmann, 1982, pl. 12, figs. 8a-b) in Site 401.

Buliminella cf. *grata* Parker and Bermudez. *Buliminella grata* Parker and Bermudez, 1937, p. 515, pl. 59, figs. 6a-c. Tjalsma and Lohmann, 1982, pl. 12, figs. 9-10.

As in Tjalsma and Lohmann's *B.* cf. *grata*, ours often have only 3 chambers in the last whorl (although most are 4 chambered) and much less incised sutures in the last whorl than *B. grata* s.s. although they lack the coarse wall of their specimens. *B* cf. *grata* occurs in the Eocene and Oligocene of Site 119, where it constitutes up to 5 °/° of the fauna.

Cassidulina havanensis Cushman and Bermudez. 1936, p. 36, pl. 6, fig. 11.

Occurs in the late middle to late Eocene of Site 401 where it constitutes up to 5 °/° of the fauna.

Cibicidoides sp. 1. *Cibicidoides sinstralis* (Coryell and Rivero). Schnitker, 1979, pl. 12, figs. 4-6.

Occurs in the middle Eocene of Site 401 and the early Eocene of Site 119. This species looks like an *Osangularia* in that it has a sharp, pinched periphery and a prominent keel. It differs from this genus in its typical *Cibicides/Cibicidoides* aperture and from Coryell and Rivero's species by its more distinct keel and greater number of pores and less distinct and more oblique sutures on the dorsal side. In dorsal aspect this species is reminiscent of *C. havanensis*.

Cibicidoides grimsdalei (Nuttall). *Cibicides grimsdalei* Nuttall, 1930, p. 291, pl. 25, figs. 7-8, 11. Schnitker, 1979, pl. 11, figs. 1-3. Tjalsma and Lohmann, 1982, pl. 18, figs. 2a-c; pl. 22, figs. 6-7.

Rare to absent in the Eocene of Site 119 and 401; more abundant in the deep Oligocene of Site 119.

Cibicidoides haitiensis (Coryell and Rivero). *Cibicides robertsonianus* (Brady) var. *haitiensis* Coryell and Rivero, 1940, p. 335, pl. 44, figs. 4a-c, 5-6. Tjalsma and Lohmann, 1982, pl. 17, figs. 6a-b.

Found in all 3 assemblages. Most of the specimens in Site 119 have 5-6 chambers in the final whorl, and are < 250 μm in diameter. The original description noted 9-11 chambers in the last whorl, while Tjalsma and Lohmann noted more tightly coiled (7-9 chambers in last whorl) Eocene specimens. The number of chambers may simply be a function of the size, for larger specimens (> 250 μm) in Sites 119 and 401 typically have 7-9 chambers in the last whorl.

Cibicidoides havanensis (Cushman and Bermudez). *Cibicides havanensis* Cushman and Bermudez, 1937, p. 28, pl. 3, figs. 1-3. Tjalsma and Lohmann, 1982, pl. 22, figs. 4a-c.

Rare in the deep Oligocene. More common in the deep Eocene (up to 6 °/°). Absent from Site 401 except for the uppermost sample.

Cibicidoides aff. *laurisae* (Mallory). *Cibicides laurisae* Mallory, 1959, p. 267, pl. 24, figs. 8a-c. Tjalsma and Lohmann, 1982, pl. 17, figs. 2a-c.

Occurring in all 3 assemblages. Rare in Eocene of Sites 119 and 401. Comprises up to 10 °/° of benthic foraminifera in the deep Oligocene of Site 119. My specimens are plano-convex to very slightly concavo-convex, having a more convex spiral side and less pinched periphery than Mallory's specimens. As Tjalsma and Lohmann noted in their material, the dorsal side has a greater number of pores than noted by Mallory. Some specimens exhibit umbilical infilling noted by Mallory, but absent in Tjalsma and Lohmann's material.

Cibicidoides subspiratus (Nuttall). *Cibicides subspirata* Nuttall, 1930, p. 292, pl. 25, figs. 9-10, 14. Tjalsma and Lohmann, 1982, pl. 18, figs. 5a-b; pl. 22, figs. 5a-b.

Restricted to the middle Eocene of Site 119 (Zone NP15). Rare.

Cibicidoides tuxpamensis (Cole). *Cibicides tuxpamensis* Cole, 1928, p. 219, pl. 1, figs. 2-3; pl. 3, figs. 5-6. Tjalsma and Lohmann, 1982, pl. 18, figs. 3a-4c; pl. 22, figs. 1a-3c.

Cibicidoides ungerianus (d'Orbigny). *Rotalina ungeriana* d'Orbigny, 1846, p. 157, pl. 8, figs. 16-18. Tjalsma and Lohmann, 1982, pl. 18, figs. 1a-c; pl. 21, figs. 5-6.

Found in all three assemblages, although the sensu stricto form is rare in the Eocene of Sites 401 and 119.

Cibiciodoides ungerianus s.l.

Although the end members of *C. tuxapamensis* and *C. ungerianus* are distinct, apparently transitional forms occur in the Eocene of Sites 119 and 401 and have been included in *C. ungerianus* s.l. This transitional form is higher conical than sensu stricto forms. *C. ungerianus* sensu stricto and sensu lato have been lumped into the *C. ungerianus* plexus which is plotted on Figures 7 and 9 as triangles. Also included in this plexus is a small, many-chambered form found in site 119.

Clinapertina inflata, Tjalsma and Lohmann, 1982, pl. 16, figs. 3a-b; pl. 20, figs. 1a-2c.

Clinapertina complanata, Tjalsma and Lohmann, 1982, pl. 20, figs. 3a-c.

Clinapertina subplanispira, Tjalsma and Lohmann, 1982, pl. 19, figs. 7a-c.

Clinapertina spp. are restricted to the deep-Eocene assemblage of Site 119, where they comprise up to 18 °/° of the fauna. *C. inflata* is most abundant, while *C. complanata* and *C. subplanispira* are rare.

Eggerella bradyi (Cushman). *Verneulina bradyi* Cushman, 1911, p. 54, text-fig. 86a-b. *Eggerella bradyi* (Cushman), Cushman, 1933, p. 33, pl. 34, figs. 1a-b. Phleger, et al., 1953, pl. 5, figs. 8,9. Barker, 1960, pl. 47, figs. 4-7.

Much more rapidly inflating chambers than noted in the type and in the illustration of Barker. Similar to Phleger et al.'s fig. 9.

"*Eggerella*" sp.

Occurring in all 3 assemblages. Most abundant in upper sections of Site 401 (up to 6 °/°). This unusual, large (> 500 µm) agglutinated form is triserial in latter stages, with the earlier stages indeterminable. However, it apparently has a looped-shaped aperture, and thus may be related to *Makarskiana trochoidea* van Soest illustrated by Beckmann (pl. 17, fig. 11).

?*Eggerella* sp.

Restricted to the Eocene of Site 119, where it occurs in abundances up to 7 °/°. These very coarse grained agglutinant are uniformly triserial in the final whorl; however, all the specimens are small (< 250 µm), and may therefore be juveniles of *G.* aff. *laevigata*.

Epistominella exigua (Brady). *Pulvinulinella exigua* Brady, 1884, p. 696, pl. 103, figs. 13-14. Phleger et al., 1953, pl. 9, figs. 35-36.

Restricted to the Oligocene of Site 119. Comparison with Recent material shows that Recent specimens tend to have more inflated final chambers.

Gaudryina cf. *laevigata* Franke. *Gaudryina laevigata* Franke, 1914, p. 431, pl. 27, figs. 1-2. Tjalsma and Lohmann, 1982, pl. 8, figs. 5a-b.

Except for one questionable occurrence in site 401 (sample 6-3), restricted to the deep Eocene of Site 119 where it occurs in abundances up to 7 °/°.

Gaudryina pyramidata Cushman. *Gaudryina laevigata* Franke var. *pyramidata* Cushman, 1926, p. 587, pl. 16, figs. 8a-b. Tjalsma and Lohmann, 1982, pl. 8, figs. 1a-b.

Restricted to the deep Eocene of Site 119, where it constitutes up to 7 °/° of the fauna.

Gaudryina cf. *pyramidata* Cushman. *Gaudryina laevigata* Franke var. *pyramidata* Cushman, 1926, p. 587, pl. 16, figs. 8a-b. Tjalsma and Lohmann, 1982, pl. 8, figs. 2a-b.

As noted by Tjalsma and Lohmann, this form has a quadrate outline; in general, it is larger in size (> 500 µm) than the typical form. Restricted to the deep Eocene of Site 119. Rare.

Gavelinella capitata (Guembel). *Rotalia capitata* Guembel, 1868, p. 653, pl. 2, figs. 92a-c. Tjalsma and Lohmann, 1982, pl. 16, figs. 4a-5b.

Occurs only in the Eocene of Site 119 and 401. Rare.

Gavelinella micra (Bermudez). *Cibicides micrus* Bermudez, 1949, p. 302, pl. 24, figs. 34-36. Tjalsma and Lohmann, 1982, pl. 16, figs. 7a-b; pl. 20, fig. 7.

Found only in the Eocene of Site 401. Rare.

Gavelinella semicribrata (Beckmann). *Anomalina pompilioides* Galloway and Heminway var. *semicribrata* Beckmann, 1953, p. 400, pl. 27, fig. 3; text-figs. 24-25. Tjalsma and Lohmann, 1982, pl. 16, figs. 6a-b.

Found only in the deep Oligocene of Site 119 and samples 6-3 to 2-1 in Site 401 (from within zone P12 to latest Eocene). Tjalsma and Lohmann noted the appearence of typical forms in P12, agreeing with the chronostratigraphic distribution noted in the Bay of Biscay. Rare.

Globocassidulina subglobosa (Brady). *Cassidulina subglobosa* (Brady). Brady, 1884, p. 430, pl. 54, figs. 17a-c. Tjalsma and Lohmann, 1982, pl. 16, fig. 9.

Found in all three assemblages with highest (up to 29 °/°) abundance in the deep Oligocene of Site 119. In general, the deep Eocene forms of Site 119 are uniformly smaller than 250 µm, while the shallow Eocene and deep Oligocene have a wider size range. Corliss (1979) attributed the predominance of smaller-sized specimens to be indicative of greater dissolution or lesser availability of carbonate.

Hanzawaia cushmani (Nuttall). *Cibicides cushmani* Nuttall, 1930, p. 291, pl. 25, figs. 3, 5-6. Tjalsma and Lohmann, 1982, pl. 17, figs. 1a-c.

Restricted to the deep Eocene of Site 119. Rare.

Karreriella bradyi (Cushman). *Gaudryina bradyi* Cushman, 1911, p. 67, fig. 107. *Karreriella bradyi* (Cushman), Cushman, 1937, p. 135, pl. 16, figs. 6-11. Barker, 1960, pl. 46, figs. 1-4. *Karreriella subglabra* (Guembel), Tjalsma and Lohmann, 1982, pl. 9, figs. 1a-b.

Found in all three assemblages. Comparison of specimens of Recent *K. bradyi* with the Bay of Biscay material shows little difference between this species and Eocene *Karreriella* which Tjalsma and Lohmann assigned to *K. subglabra*; it is possible then that these two species are conspecific.

Karreriella chapapotensis (Cole). *Textularia chapapotensis* Cole, 1928, p. 206, pl. 33, fig. 9. Tjalsma and Lohmann, 1982, pl. 9, figs. 2a-3.

Occurring in all 3 assemblages. Included here are specimens assignable to *K. chapapotensis* (Cole) var. *monumentensis* Mallory (1959, pl. 5, figs. 3a-c).

Nonion havanenese Cushman and Bermudez. 1937, p. 19, pl. 2, figs. 13-14. Tjalsma and Lohmann, 1982, pl. 7, figs. 6a-b.

More abundant in the shallow Eocene of Site 401, where it comprises up to 8 °/° of the fauna.

Nuttallides truempyi (Nuttall). *Eponides truempyi* Nuttall, 1930, p. 287, pl. 24, figs. 9, 13-14. Tjalsma and Lohmann, 1982, pl. 6, figs. 4a-b, pl. 17, figs, 4a-5b, pl. 21, figs. 1a-4c.

Restricted to the Eocene of Site 119 and 401, this species becomes extinct in the shallower site in the late Eocene (between Zones P15/16 and the *G. cerroazulensis cunialensis* Zone)

Nuttallides umbonifera (Cushman). *Pulvinulinella umbonifera* Cushman, 1933, p. 90, pl. 9, figs. 9a-c. *Epistominella (?) umbonifera* (Cushman), Phleger et al., 1953, pl. 9, figs. 33-34. *Eponides bradyi* Earland, Barker, 1960, pl. 95, figs. 9-10. "*Epistominella*" *umbonifera* (Cushman), Lohmann, 1978, pl. 13, figs. 1-3.

Restricted to the Oligocene of Sites 119 and 400A, where it forms up to 27 °/° of the fauna. Phleger et al. noted the affinities of *Epistominella* (?) *umbonifera* with the genus *Nuttallides* Finlay, especially with respect to apertural characteristics (viz. small notch in the plane of coiling). Comparison with Recent material shows that the Oligocene forms tend to have a less rounded, more pinched axial view; however, I believe that Oligocene forms are conspecific with Recent *N. umbonifera* (cf. Pl. 1, figs. 1-3 with pl. 3, figs. 1-3 of Lohmann).

Oridorsalis umbonatus (Reuss). *Rotalina umbonata* Reuss, 1851, p. 75, pl. 5, fig. 35. Tjalsma and Lohmann, 1982, pl. 6, figs. 8a-b.

Found in all three assemblages. Predomiant *Oridorsalis* sp. found in all 3 assemblages.

Osangularia mexicana (Cole). *Pulvinulinella culter* (Parker and Jones) var. *mexicana* Cole, 1927, p. 31, pl. 1, figs. 15-16. Tjalsma and Lohmann, 1982, pl. 20, fig. 6.

Restricted to the shallow Eocene of Site 401, where it comprises up to 8 °/° of the fauna.

Pleurostomella acuta Hantken. 1875, p. 44, pl. 13, fig. 18. Braga et al., 1975, pl. 3, figs. 7a-b.

Pullenia eocenica Cushman and Siegfus. 1939, p. 31, pl. 7, figs. 1a-b. Tjalsma and Lohmann, 1982, pl. 16, fig. 1.

Found in all three assemblages. Rare in deep Eocene; most abundant in deep Oligocene of Site 401 (up to 10 °/°) and shallow Eocene of Site 401 (up to 5 °/°).

Pullenia quinqueloba (Reuss). *Nonionina quinqueloba* Reuss, 1851, p. 71, pl. 5, figs. 31a-b. Tjalsma and Lohmann, 1982, pl. 16, fig. 2.

I have followed Bermudez (1949) and Tjalsma and Lohmann (1982) in including 4-chambered forms; in addition, I have included 6 chambered forms. It becomes difficult to differentiate this species from *P. eocenica* in the deep site 119, especially in the Oligocene. Here, the outline is often rounded and poorly lobated, while *P. eocenica* becomes compressed and can have a pentagonal outline. However, it is often possible to differentiate these species in axial view, for *P. quinqueloba* has more pinched early chambers relative to the final chamber.

Quadrimorphina profunda Schnitker and Tjalsma. p. 239, pl. 1, figs. 16-21. Tjalsma and Lohmann, 1982 pl. 19, figs. 5a-6c.

Occurs only in the early-middle Eocene and early Oligocene of Site 119 (to within Zone NP23). Rare in Oligocene; forms up to 7 °/° of the Eocene assemblage.

Stilostomella aculeata (Cushman and Renz). *Ellipsonodosaria nuttalli* Cushman and Jarvis var. *aculeata* Cushman and Renz, 1948, p. 32, pl. 6, fig. 10. Tjalsma and Lohmann, 1982, pl. 14, fig. 12.

Found in all three assemblages. Slightly more common in Site 119.

Stilostomella gracillima (Cushman and Jarvis). *Ellipsonodosaria nuttalli* var. *gracillima* Cushman and Jarvis, 1934, p. 72, pl. 10, fig. 7. Tjalsma and Lohmann, 1982, pl. 14, figs. 13-15.

Rare.

Stilostomella modesta (Bermudez). *Ellipsonodosaria modesta* Bermudez, 1937, p. 238, pl. 20, fig. 3. *Siphonodosaria modesta* (Bermudez), Bermudez, 1949, p. 225, pl. 14, fig. 29. Beckmann, 1953, pl. 21, fig. 32.

Rare.

Stilostomella subspinosa (Cushman). *Ellipsonodosaria subspinosa* Cushman, 1943, p. 92, pl. 16, figs. 6-7b. Tjalsma and Lohmann, 1982, pl. 14, figs. 16-17.

Found in all three assemblages. Rare in deep Eocene; most common (up to 7 °/°) in deep Oligocene.

Turrilina alsatica Andreae. 1884, p. 120, pl. 8, figs. 18a-c.

Restricted to the Oligocene. Rare.

Turrilina robertsi (Howe and Ellis). *Bulimina robertsi*, Howe and Ellis, 1939, p. 63, pl. 8, figs. 32-33. Tjalsma and Lohmann, pl. 14, fig. 5.

Restricted to the Eocene. Rare.

Uvigerina rippensis Cole. 1927, p. 11, pl. 2, fig. 16. Tjalsma and Lohmann, pl. 14, figs. 6-7.

Restricted to the Eocene. Rare.

Vulvulina spinosa Cushman. p. 111, pl. 23, fig. 1. Tjalsma and Lohmann, 1982, pl. 10, figs. 1a-5b.

Found in all three assemblages. Occurring in greatest abundances in the early Oligocene (up to 8 °/°) and shallow Eocene (up to 9 °/°).

TABLE 1

Benthic Foraminifera, Site 119

SAMPLE	DEPTH (cm)	# counted	*Nuttallides*	*G. subglobosa* spp.[c]	*Gyroidinoides*	*C. ungerianus*[a] spp.	*Oridorsalis*
12-CC	--	354	12.1	24.6	6.2	8.2/13.0	9.3
13-2	146-149	250	12.0	22.8	9.6	7.2/14.0[b]	11.2
13-CC	--	575	5.4	29.3	8.3	4.0/5.9	7.8
14-2	147-150	135	3.7	28.8	6.6	2.2/2.2[b]	16.3
14-5	100-105	129	11.6	22.5	8.5	20.9/20.9[b]	3.9
14-CC	--	511	15.3	10.8	8.0	5.0/5.6[b]	11.0
15-3	147-150	139	27.3	23.7	7.9	4.3/4.3[b]	9.3
15-CC	--	403	17.4	12.2	8.5	10.9/14.6	8.7
16-2	119-121	122	12.3	13.1	9.8	1.6/1.6	15.6
16-CC	--	394	6.8	9.6	7.9	0/2.8	5.1
17-2	149-152	165	13.0	15.5	10.6	14.9/14.9	9.9
17-3	150-151	215	9.8	11.6	6.5	11.0/11.0	13.5
17-4	100-102	267	3.7	23.6	8.6	16.8/16.8	6.0
17-CC	--	385	12.5	10.3	4.9	13.8/13.8	10.6
18-1	50-52 & 145-148	337	2.4	22.6	6.8	13.1/15.1	24.9
18-3	146-149	105	3.8	19.0	5.7	7.6/9.5	20.0
18-4	144-148	115	5.2	7.0	13.0	7.8/14.8	30.0
18-5	50-52	161	0.5	25.4	8.1	9.3/14.2	23.0
18-6	146-148	124	2.4	12.9	8.9	7.3/10.5	21.0
18-CC	--	326	4.6	7.4	14.7	8.3/8.3	26.0
				HIATUS			
19-1	143-146	55	29.0	3.6	0	0	1.8
19-2	146-149	62	22.6	4.8	1.6	0	11.3
19-3	143-146	124	25.8	14.5	4.0	0/4.8	7.3
19-CC	--	385	30.5	5.4	3.1	3.3/9.5	10.7
20-1	147-150	115	26.0	1.7	6.1	0	2.6
20-2	146-149	152	17.8	7.7	1.9	0/2.6	4.5
20-3	0-5	434	24.0	3.0	1.8	0/2.5	5.3
20-3	93-96	205	28.3	8.3	2.9	0.5/2.4	2.4
20-CC	--	364	24.2	8.5	3.6	0.8/4.1	8.2
21-2	100-102	126	38.9	1.6	4.8	1.6/4.0	7.9
21-2	147-150	248	24.2	10.9	6.9	0/4.0	5.6
21-3	146-149	85	8.2	2.3	4.7	0/11.8	11.8
21-4	145-148	74	28.4	0	2.7	0/9.6	4.1
21-CC	--	533	21.8	9.0	4.1	0.2/3.6	10.7
22-CC	--	317	23.9	2.8	5.4	0/3.8	8.5
23-CC	--	68	8.8	2.9	1.5	2.9/2.9	4.4
24-CC	--	137	15.0	1.5	15.3	*	16.8

TABLE 1 (cont.)

* Preservation poor, percentages not computed.

[a] First number indicates percent *Cibicidoides ungerianus* sensu stricto, second number indicates percent *C. ungerianus* plexus. Samples indicated with [b] have some uncertainty attached to this differentiation in that specimens were picked for isotopic analyses prior to final taxonomic revision.

[c] *Nuttallides* spp. consists of *N. umbonifera* (cores 12-18) and *N. truempyi* (cores 19-24).

TABLE 2

Benthic Foraminifera, Site 401

SAMPLE	DEPTH (cm)	# counted	*N. truempyi*	*G. subglobosa*	*Gyroidinoides* spp.	*C. ungerianus*[a]	*Oridorsalis* spp.
2-1	8-10	556	0	13.1	5.4	2.7[b]	11.7
3-1	15-21	283	5.7	8.9	3.5	3.9[b]	12.1
4-CC	—	312	14.4	3.5	5.8	0.64/3.8	10.0
5-3	93-97	244	6.1	1.2	5.7	0.4/0.4	5.3
6-3	76-80	512	4.3	4.3	6.1	0.6/2.7	14.8
7-3	76-80	227	12.3	14.1	6.6	0/4.8	4.0
8-3	44-48	234	6.4	16.7	5.1	0/2.1	14.5
9-3	96-100	330	8.5	13.0	5.2	0/11.2	9.1
10-3	98-102	382	7.1	2.6	3.9	0/5.6	18.6

[a] First number indicates percent *Cibicidoides ungerianus* sensu stricto, second number indicates percent *C. ungerianus* plexus. Samples indicated with [b] have some uncertainty attached to this differentiation in that specimens were picked for isotopic analyses prior to final taxonomic revision.

TABLE 3
Zonal Age Assignments, Site 119

SAMPLE	AGE
12-CC	?*S. ciperoensis* Zone (NP25)[1]; Late Oligocene, within range of *G. opima nana*[4,5].
13-2 146-149cm	?*S. ciperoensis* Zone (NP25)[1].
13-CC	?*S. ciperoensis* Zone (NP25)[1].
Cores 12-13	*S. ciperoensis* Zone (NP25)[2*].
Cores μ12 to 14-4	*S. ciperoensis* Zone (NP25)[1].
14-5 148-150cm	*S. distentus* Zone (NP24)[2].
14-CC	P21[4]; *G. opima opima* (P21) Zone[5]; *S. distentus* Zone (NP24)[1].
15-CC	P21[4]; *G. ampliapertura*/*G. opima opima* Zones (P20/21)[5]; μ*S. distentus* Zone (NP24)[1].
Cores 14-15	*S. distentus* Zone (NP24)[2*].
Cores 14-4 to μ15-CC	*S. distentus* Zone (NP24)[1].
Cores 16-2 119-121cm	*S. praedistentus* Zone (NP23)[1].
16-CC	P21[4]; *G. ampliapertura*/*G. opima opima* Zones (P20/21)[5]; *S. praedistentus* Zone (NP23)[1]NP23[3].
17-1 122-126cm	NP23[3].
17-5 60cm	NP23[3].
Cores 16 to 17-5 70cm	*S. praedistentus* Zone (NP23)[1].
17-5 70cm to 18-2 50cm	?*H. reticulata* Zone (NP22)[1].
Cores 16-17	*S. praedistentus* Zone (NP23)[2*].
18-2 144-147cm	NP22[3]; *E. subdisticha* Zone (NP21)[1].
18-3 142-143cm	*H. reticulata* (NP22)[2,3]; *E. subdisticha* Zone (NP21)[1].
18-CC	*E. subdisticha* Zone (NP21)[1], NP21[3].
Core 18	*H. reticulata* Zone (NP22)[2*].
	HIATUS
19-1 107-111cm	NP15[3]; NP16[1].
19-1 140-141cm	*C. quadratus* Zone (NP15)[2]; NP16[1].
19-6 127-131cm	NP15[3]; *N. fulgens* Zone (NP15)[1].
Cores 19 to 20	*C. quadratus* Zone (NP15)[2*].
21-1 65-69cm	NP14[3]
21-1 143-144cm	*D. sublodoensis* Zone (NP14)[2].
21-4 149-150cm	*D. lodoensis* Zone (NP13)[2].
Cores 19-3 20 cm to 22	*N. fulgens* Zone (NP15)[1].
Core 22	*D. lodoensis* Zone (NP13)[2*].
23-1 141-142cm	*T. orthostylus* Zone (NP12)[2].
23-1 0 to 70cm	*D. lodoensis* Zone (NP13)[1].
23-1 70 to 23-cc	*M. tribrachiatus* (NP12)[1].
Core 24	*D. binodosus* (NP11)[1].

References: [1] Perch-Nielsen, 1972; [2] Bukry, 1972; [3] Aubry, personal communication; [4] Berggren, 1972; [5] this study; * Full age documentation not given.

TABLE 4, Miller

Zonal Age Assignments, Site 401

SAMPLE	DEPTH	ZONAL AGE (Adopted Age in Ma)
2-1	8-10cm	*G. cerrolazulensis cunialensis* Zone[1]; *G. cocoaensis* Zone[2]; P17[3,4]; NP21[5,6] (37.5)
3-1	15-21cm	G. semiinvoluta/G. cocoaensis Zone (P15/16)[2]; P15/16[3]; NP18[5]; NP19[6] (39.0)
4-CC	--	*T. rohri* Zone (P14)[1,2]; P14-15[3a]; P16[4]; NP18[5,6] (41.0)
5-3	93-97cm	*O. beckmanii* Zone (P13)[2]; P13/14[3a]; NP16[5,6] (43.0)
6-3	76-80cm	*G. lehneri* Zone (P12)[2]; P11-12[3a]; NP16[5,6] (44.0)
7-3	76-80cm	*G. lehneri* Zone (P12)[2]; P11[3]; P12[4]; NP16[5,6] (45.4)
8-3	44-48cm	*G. kugleri* Zone (P11)[2]; P11[3]; NP15[5], NP16[6] (46.6)
9-3	96-100cm	*H. aragonensis* Zones (P10)[2]; P10[3]; NP15[5], NP16[6] (47.7)
10-3	98-102cm	*H. aragonensis* Zones (P10)[2]; P10[3,4]; NP15[5], NP16[6] (49.0)

[1] This study; [2] Krasheninnikov, 1979; [3] Site reports, Montadert, Roberts et al., 1979; [3a] Contradictions as to planktonic foraminiferal zonal age assignment occurring within site reports; [4] Schnitker, 1979; [5] Muller, 1979; [6] M.-P. Aubry, personal communication.

Age discrepancies between zonal age assignment and adopted age in Ma result from assumption of constant sedimentation rate in computing adopted age.

TABLE 5, Miller

Zonal Age Assignments and Benthic Foraminifera, Site 400A.

SAMPLE	DEPTH	ZONAL AGE (Adopted age in Ma)	# Counted	% *N. umbonifera*
43-3	77-81cm	*G. ciperoensis* Zone (P22)[1] P19-21[2]; NP24/25[5](25.0/25.0)	213	14.0
43-5	68-72cm	*G. opima opima* Zone (P21)[1]; P19-21[2]; NP24/25[4,5](25.7/26.0)	350	18.8
44-1	70-74cm	*G. opima opima* Zone (P21)[1] P19-21[2]; P22[3]; NP24/25[4,5] (26.5/27.1)	337	3.6
45-3	95-99cm	*G. opima opima* Zone (P21)[1] NP23[4,5]; P20[3](29.3/31.3)	234	24.3
45-4	82-86cm	*G. opima opima* Zone (P21)[1] NP23[4,5](29.7/31.8)	444	16.9
45-5	69-73cm	NP23[4,5](30.0/32.2)	280	4.6
		Possible hiatus		
45-CC	--	NP22[4,5]; P19[3](34.5/34.6)	259	14.3
46-3	146-150cm	NP21[4,5](35.7/34.8)	309	4.2
46-4	65-70cm	NP21[4,5](35.9/35.4)	--	0
46-5	98-102cm	NP21[4,5](36.2/35.7)	175	5.0
		Drilling disturbance below		
46-6	21-27cm	NP21[4,5](36.3/36.3)	207	4.8[a]
47-1	17-20cm	µmiddle Eocene[c];NP21[4] P18[3](36.7/37.32)	--	0[b,c]
47-1	23-25cm	same as 47-1, 17-20cm	--	0[b,c]
47-2	147-150cm	NP21[2](37.25/37.25)	203	18.0[a,b]
		HIATUS		

Samples 47-3 and below are middle Eocene

First adopted age assumes middle Oligocene hiatus; second assumes aconstant sedimentation rate.

[1] Krasheninnikov, 1979; [2] Site reports, Montadert, Roberts et al., 1979; [3] Schnitker, 1979; [4] Muller, 1979; [5] M.-P. Aubry, personal communication.

[a] Possible reworking/core disturbance; identification of *N. umbonifera* and tentative corroded *N. truempyi*. [b] Definite core disturbance noted during sampling. [c] Core disturbance results in deep Eocene benthic foraminiferal assemblage (*N. truempyi*, *Clinapertina complanata*, *Abyssammina poagi*, and *Alabamina dissonata*), associated with nannoplankton indicative of Zone NP21[b].

PLATE 1

Rows 1 and 2: *Nuttallides* spp.

1,2,3 *Nuttallides umbonifera* (Cushman)
1, sample 119: 12-CC, x(330)
2, sample 119: 12-CC, x(340)
3, sample 119: 16-CC, x(320)

4,5,6 *Nuttallides truempyi* (Nuttall)
4, sample 401: 5-3, 93-97cm, x(230)
5, sample 119: 20-CC, x(300)
6, sample 119: 20-CC, x(335)

Rows 3 ə 4: Deep Eocene taxa, Site 119

7 *Nuttallides truempyi* (Nuttall)
Sample 119: 19-CC, x(230)

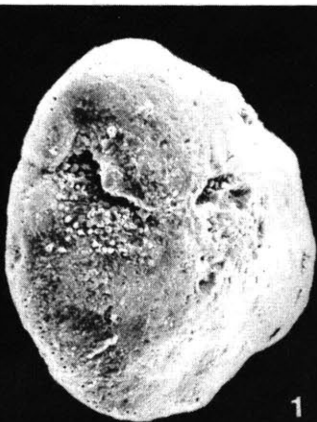
8 *Alabamina dissonata* (Cushman and Renz)
Sample 119: 19-CC, x(240)

9 *Abyssamina poagi* Schnitker and Tjalsma
Sample 119: 19-CC, x(225)

10 *Aragonia capdevilensis* (Cushman and Bermudez)
Sample 20-3, 0-5cm, x(315)

11 *Clinapertina inflata* Tjalsma and Lohmann
Sample 119: 19-CC, x(230)

12 *Hanzawaia cushmani* (Nuttall)
Sample 119: 19-CC, x(225)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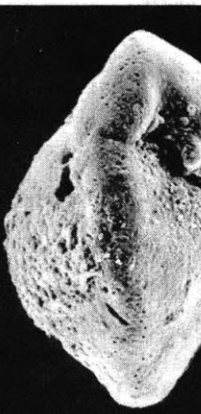
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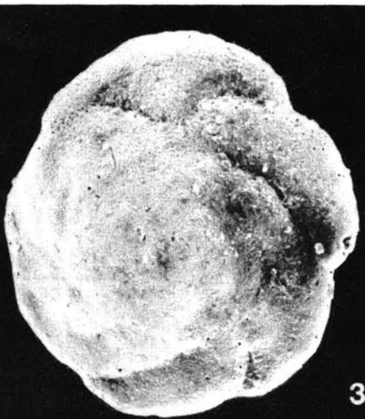
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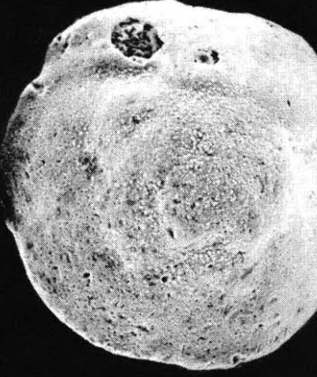
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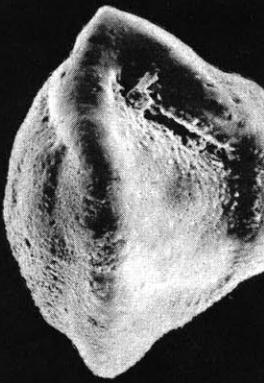
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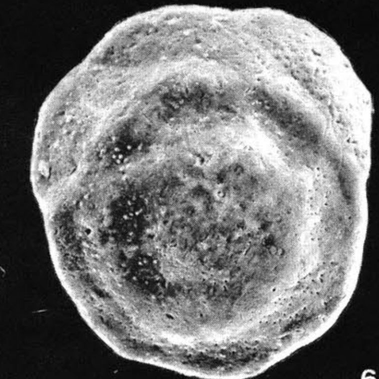
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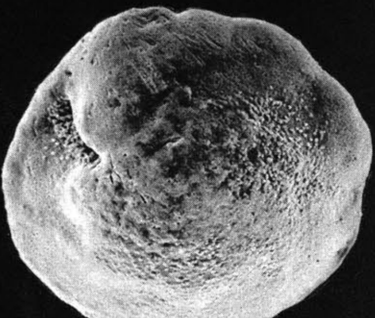
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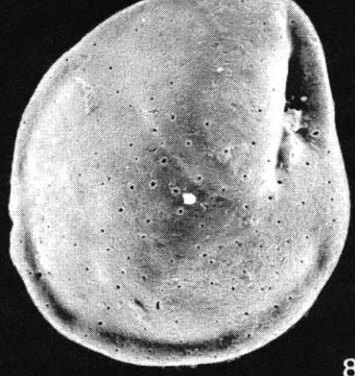
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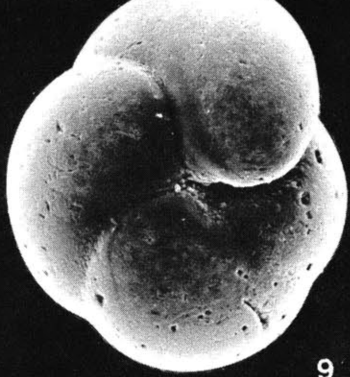
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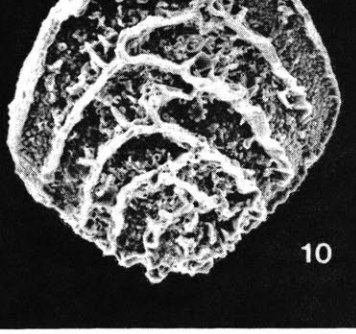
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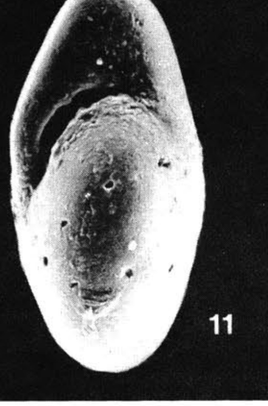
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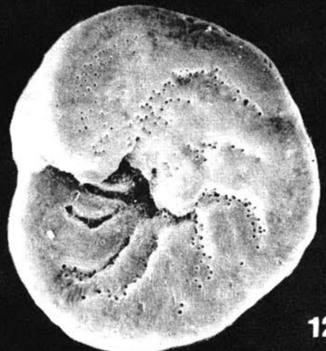
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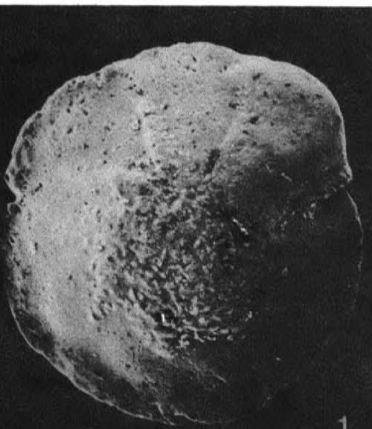
PLATE 1

PLATE 2

Cibicidoides spp.

1,2	*Cibicidoides* sp. 1 1, sample 401: 6-3, 76-80cm, x(210) 2, sample 401: 6-3, 76-80cm, x(200)
3,4	*Cibicidoides* aff. *laurisae* (Mallory) 3, sample 119: 14-CC, x(145) 4, sample 119: 14-CC, x(240)
5	*Cibicidoides haitiensis* (Coryell and Rivero) Sample 119: 15-CC, x(295)
6,7	*Cibicidoides ungerianus* (d'Orbigny) 6, sample 119: 14-CC, x(245) 7, sample 119: 14-CC, x(175)
8	*Cibicidoides grimsdalei* (Nuttall) Sample 119: 17-3, 150-151cm, x(130) Note dissolution pitting of pores.
9,10	*Cibicidoides havanenses* (Cushman and Bermudez) 9, sample 119: 20-CC, x(185) 10, 119: 20-CC, x(165)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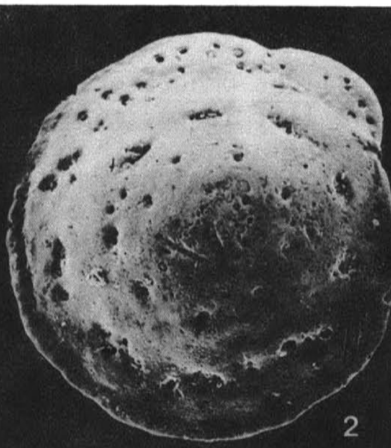
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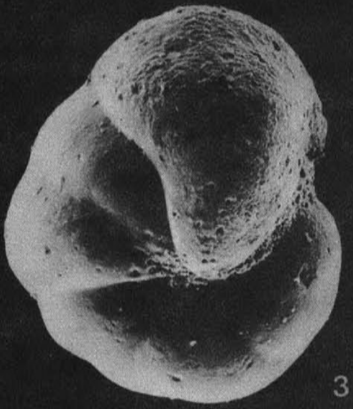
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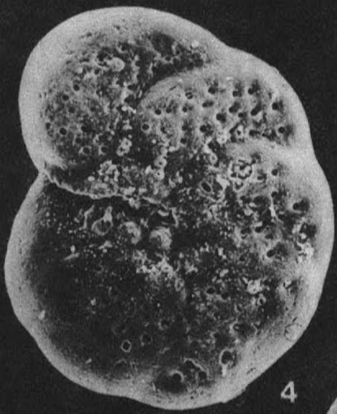
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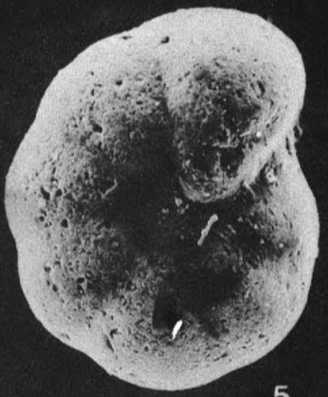
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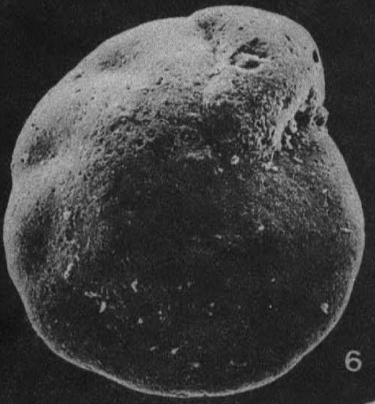
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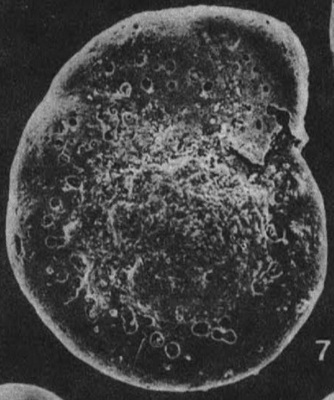
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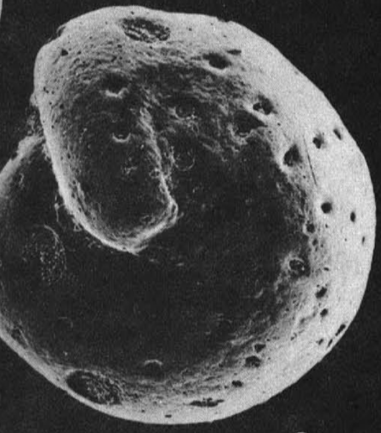
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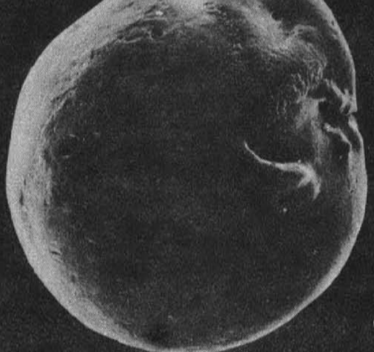
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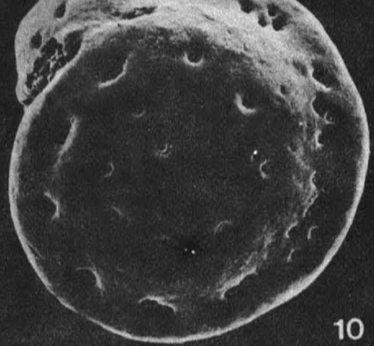
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PLATE 2

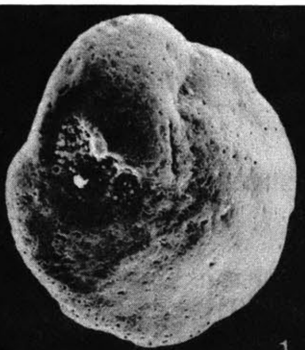
PLATE 3

Rows 1 and 2: Oligocene Taxa

1	*Nuttallides umbonifera* (Cushman) Sample 119: 13-CC, x(355)
2,5	*Gavelinella semicribrata* (Beckmann) 2, sample 119: 13-CC, x(405) 5, sample 119: 17-4, 100-102cm, x(150)
3	*Astrononion pusillum* Hornibrook Sample 119: 14-CC, x(265)
4	*Epistominella exigua* (Brady) Sample 119: 13-CC, x(305)
6	*Turrilina alsatica* Andreae Sample 119: 14-CC, x(320)

Rows 3-5: Shallow Eocene Taxa

7	*Bulimina trinitatensis* Cushman and Jarvis Sample 401:5-3, 93-97cm, x(130)
8	*Bulimina jarvisi* Cushman and Parker Sample 401: 3-1, 15-21cm,x(130)
9	*Turrilina robertsi* (Howe and Ellis) Sample 7-3, 76-80cm, x(330)
10	*Uvigerina rippensis* Cole Sample 401: 2-1, 8-10cm, x(210)
11	*Bulimina glomarchallengeri* Tjalsma and Lohmann Sample 401: 8-3, 44-48cm, x(320)
12	*Bulimina semicostata* Nuttall Sample 401: 3-1, 15-21cm, x(280)
13	*Bulimina* cf. *alazanensis* Cushman Sample 401: 5-3, 93-97cm, x(340)
14	*Osangularia mexicana* (Cole) Sample 401: 4-CC, x(180)
15,16	*Gavelinella micra* (Bermudez) 15, sample 401: 2-1, 8-10cm, x(290) 16, sample 401: 2-1, 8-10cm, x(235)



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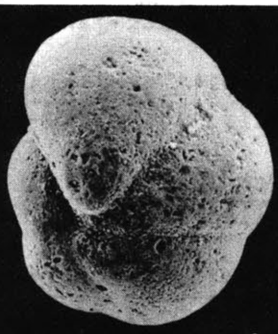

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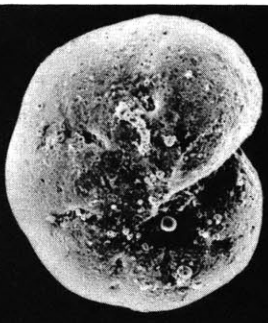

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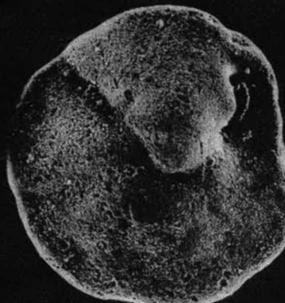

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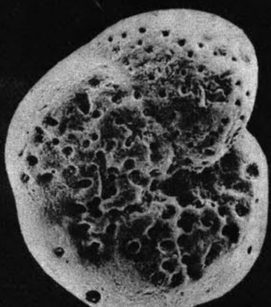

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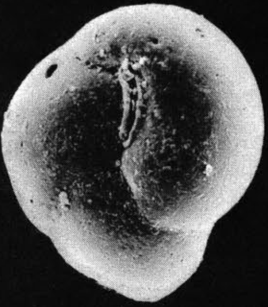

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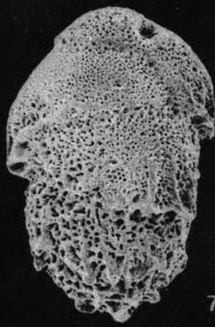

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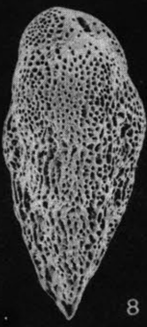

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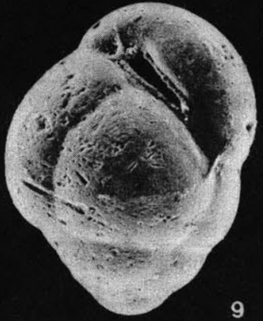

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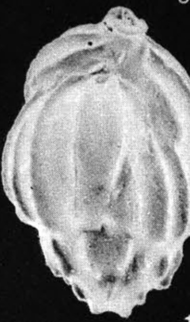

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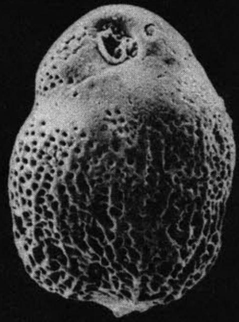

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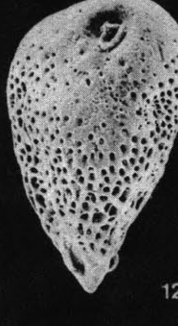

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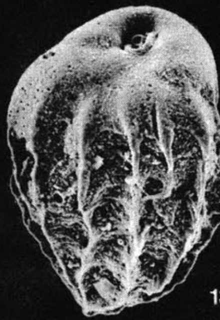

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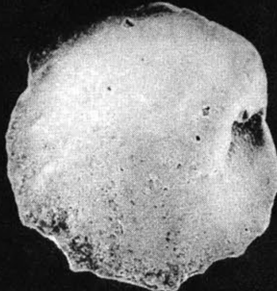

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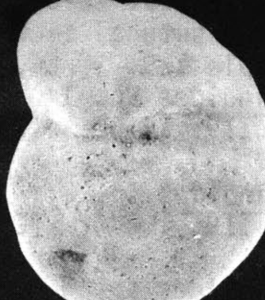

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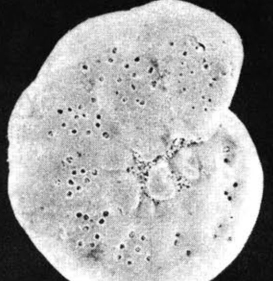

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PLATE 3

PLATE 4

Miscellaneous Taxa

1 *Stilostomella aculeata* (Cushman and Jarvis)
Sample 119: 17-4, 100-102cm, x(70)

2 *Stilostomella subspinosa* Cushman
Sample 119: 14-CC, x(100)

3 *Stilostomella gracillima* (Cushman and Jarvis)
Sample 401: 5-3, 93-97cm, x(140)

4 *Stilostomella modesta* (Bermudez)
Sample 401: 6-3, 76-80cm, x(130)

5 *Orthomorphina havanesis* (Cushman and Bermudez)
Sample 401: 5-3, 93-97cm, x(100)

6 *Bolivina huneri* Howe
Sample 119: 17-3, 150-151cm, x(210)

7 *Pleurostomella acuta* Hantken
Sample 119: 16-2, 119-121cm, x(205)

8 *Ellipsodimorphina* sp.
Sample 119: 17-4, 145-149cm, x(66)

Rows 1-3: Extant Taxa

9,10 *Pullenia quinqueloba* Cushman and Siegfus
9, sample 119: 13-CC, x(345)
10, sample 119: 12-CC, x(250)

11 *Pullenia eocenica* (Reuss)
Sample 119: 14-2, 147-150, x(305)

12 *Globocassidulina subglobosa* (Brady)
Sample 119: 12-CC, x(330)

13 *Buliminella* cf. *grata* Parker and Bermudez
Sample 119: 15-CC, x(320)

14,15 *Oridorsalis umbonatus* (Reuss)
14, sample 119: 16-CC, x(200)
Showing sutural secondary aperture.
15, sample 119: 16-CC, x(165)
16 *Gyroidinoides* sp. 1
Sample 119: 16-CC, x(245)

17 *Gyroidinoides* sp. 2
Sample 119: 16-CC, x(230)

18 *Gyroidinoides* sp. 3
Sample 119:15-CC, x(395)

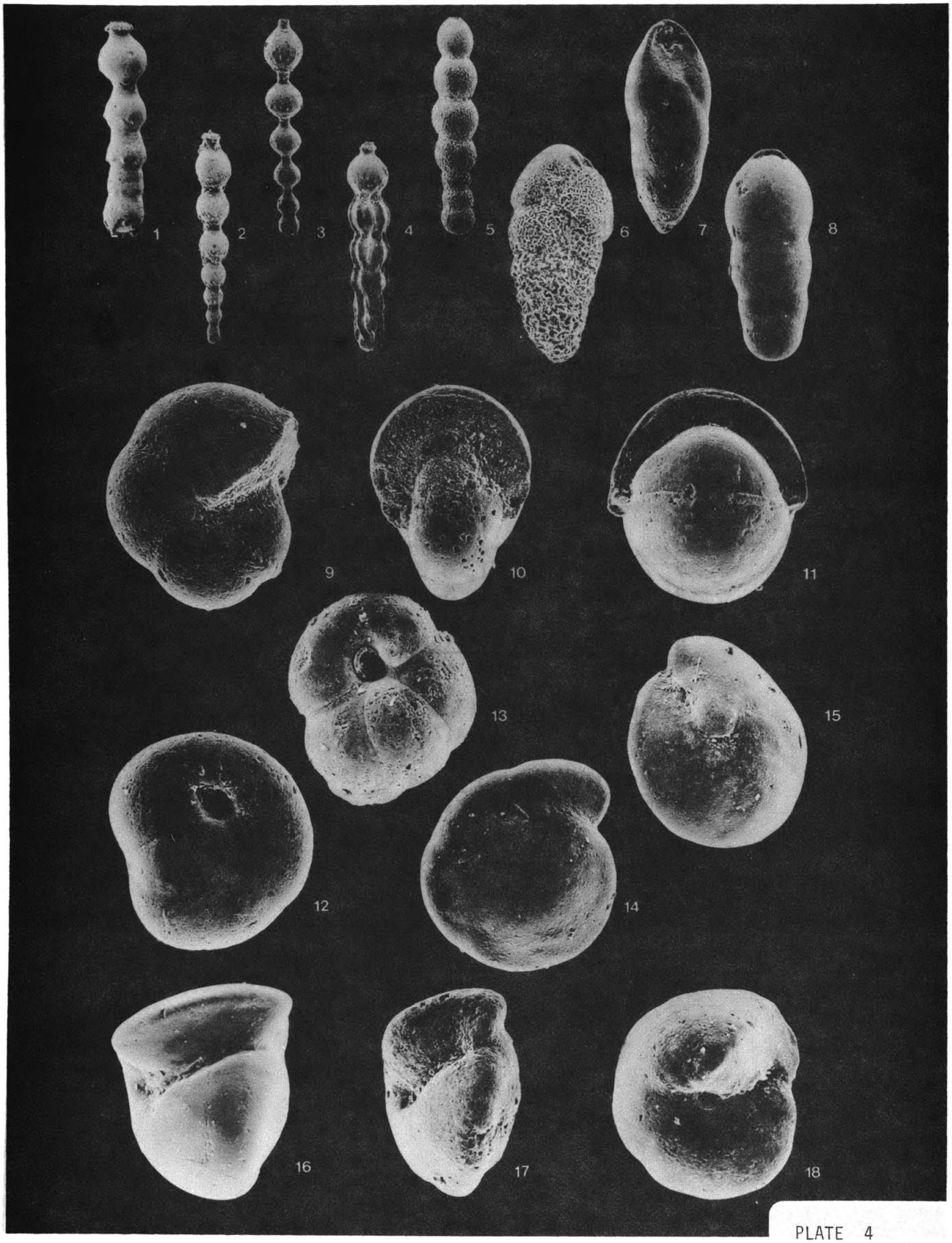
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PLATE 4

PLATE 5

Rows 1 and 2: Agglutinated Taxa

1 *Karreriella bradyi* (Cushman)/*subglabra* (Guembel)
Sample 119: 14-CC, x(175)

2 *Gaudryina pyramidata* Cushman
Sample 119: 20-CC, x(180)

3 *Gaudryina* cf. *pyramidata* Cushman
Sample 119 21-CC, x(110)

4,7 *Vulvulina spinosa* Cushman
Sample 119: 15-CC, x65

5 *Eggerella bradyi* (Cushman)
Sample 18-CC, x(455)

6 ?*Eggerella* sp.
Sample 119: 19-CC, x(200)

Rows 3 and 4: Planktonic Foraminifera

7,8 *Globorotalia cerroazulensis cunialensis* Toumarkine and Bolli
7, sample 401: 2-1, 8-10cm, x(370)
8, sample 401: 2-1, 8-10cm, x(335)

9 *Hantkenina alabamensis* Cushman
Sample 2-1, 8-10cm, x(200)

10 *Catapsydrax perus* (Todd)
Sample 119: 15-CC, x(235)

11 *Globorotalia opima opima* Bolli
Sample 119: 14-CC, x(280)

12 *Globorotalia opima nana* Bolli
Sample 119: 12-CC, x(385)

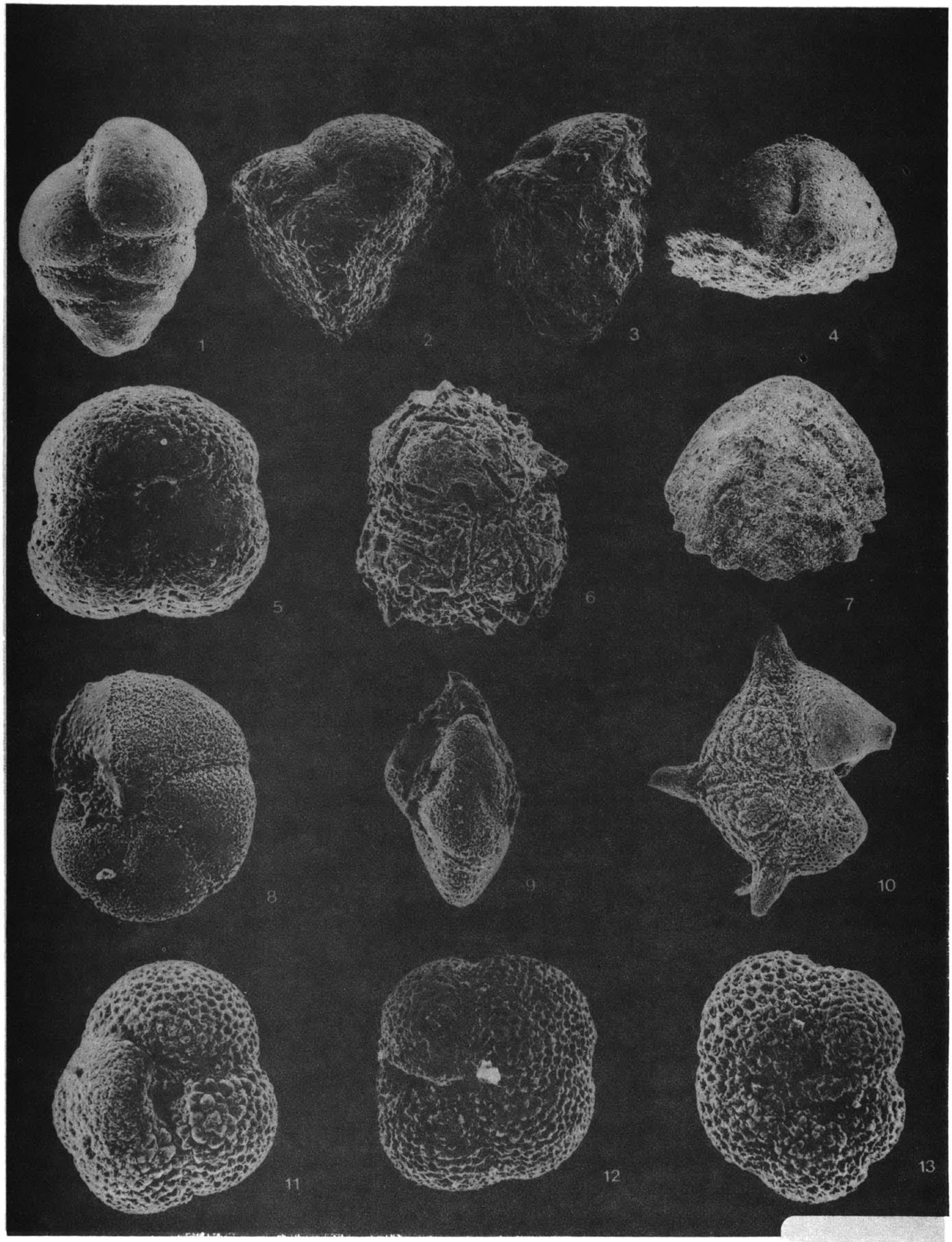
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PLATE 5

WOODS HOLE OCEANOGRAPHIC INSTITUTION
WOODS HOLE, MASSACHUSETTS 02543

Pre-print

LATE PALEOGENE (EOCENE TO OLIGOCENE) BENTHIC FORAMINIFERAL PALEOCEANOGRAPHY OF THE GOBAN SPUR REGION, DSDP LEG 80

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ABSTRACT

Major benthic foraminiferal changes occur in the late Eocene of Site 549. A *Nuttallides truempyi*-dominated assemblage is replaced by a buliminid assemblage (~ 38.5-40 Ma). This assemblage is replaced, in turn, by an assemblage dominated by stratigraphically long-ranging and bathymetrically wide-ranging taxa just below the Eocene/Oligocene boundary (~ 37.5 Ma). A series of late Eocene to earliest Oligocene first and last appearances of taxa accompany these abundance changes. Similar faunal abundance changes occur at Site 548; however, the incomplete record recovered at Site 548 prevents a firm dating of the changes. No major benthic foraminiferal changes are associated with the Eocene/Oligocene boundary (denoted by the extinction of *Hantkenina*, *Cribrohantkenina*, and *Globorotalia cerroazulensis*); instead, benthic foraminiferal abundance changes, last appearances (both extinctions and local disappearances), and first appearances occur throughout the late Eocene to earliest Oligocene interval (~ 36-40 Ma).

Benthic foraminiferal $\delta^{18}O$ increases ~ 1.0 $^{o}/oo$ in the late Eocene to earliest Oligocene (~ 38-36.5 Ma) at Site 549. Most (~ 0.7 $^{o}/oo$) of this increase occurs simultaneously with a 0.6 $^{o}/oo$ increase in benthic foraminiferal $\delta^{13}C$ as a geologically rapid event in the earliest Oligocene (~ 36.5 Ma). The major benthic foraminiferal abundance changes (~ 40-38 Ma) pre-date the major isotopic enrichments. A prominent seismic horizon, reflector R4, has been noted in noted in the Labrador Sea, Rockall, and Biscay regions and has been dated as latest Eocene to early Oligocene. This horizon marks the onset of increased intensity of abyssal circulation associated with the initial entry of bottom water from the Norwegian-Greenland Sea and/or Arctic Ocean into the North Atlantic. The age of reflector R4 may encompass

both the faunal and isotopic changes; the best estimate for the age of this horizon and the associated circulation change suggests that it correlates with the $\delta^{18}O$ increase. These data are combined in a scenario that suggests a temperature drop, decrease in age of bottom water, and an increase in itensity of abyssal circulation occurred in the late Eocene to earliest Oligocene of the North Atlantic.

INTRODUCTION

The recovery of numerous Tertiary deep-sea sections by the Deep Sea Drilling Project (DSDP) has led to a rapidly expanding data base for the study of Tertiary deep-sea benthic foraminiferal assemblages which, in turn, has led to renewed interest in this group (Berggren, 1972a; Douglas, 1973; Berggren and Aubert, 1975; Resig, 1976, Boersma, 1977, Proto-Decima and Bolli, 1978; Schnitker, 1979; Boltovskoy, 1980; Corliss, 1979a, 1981; Tjalsma and Lohmann, 1982; Douglas and Woodruff, 1982; Miller, in press; Berggren and Schnitker, in press). Comparison among these studies has been difficult due to disagreements in taxonomy (Boltovskoy, 1980) and differences in size fraction studied. However, Tjalsma and Lohmann (1982) have made a detailed taxonomic revision of deep-sea benthic foraminifers which now allows the placement of Early Paleogene taxa into a standarized framework.

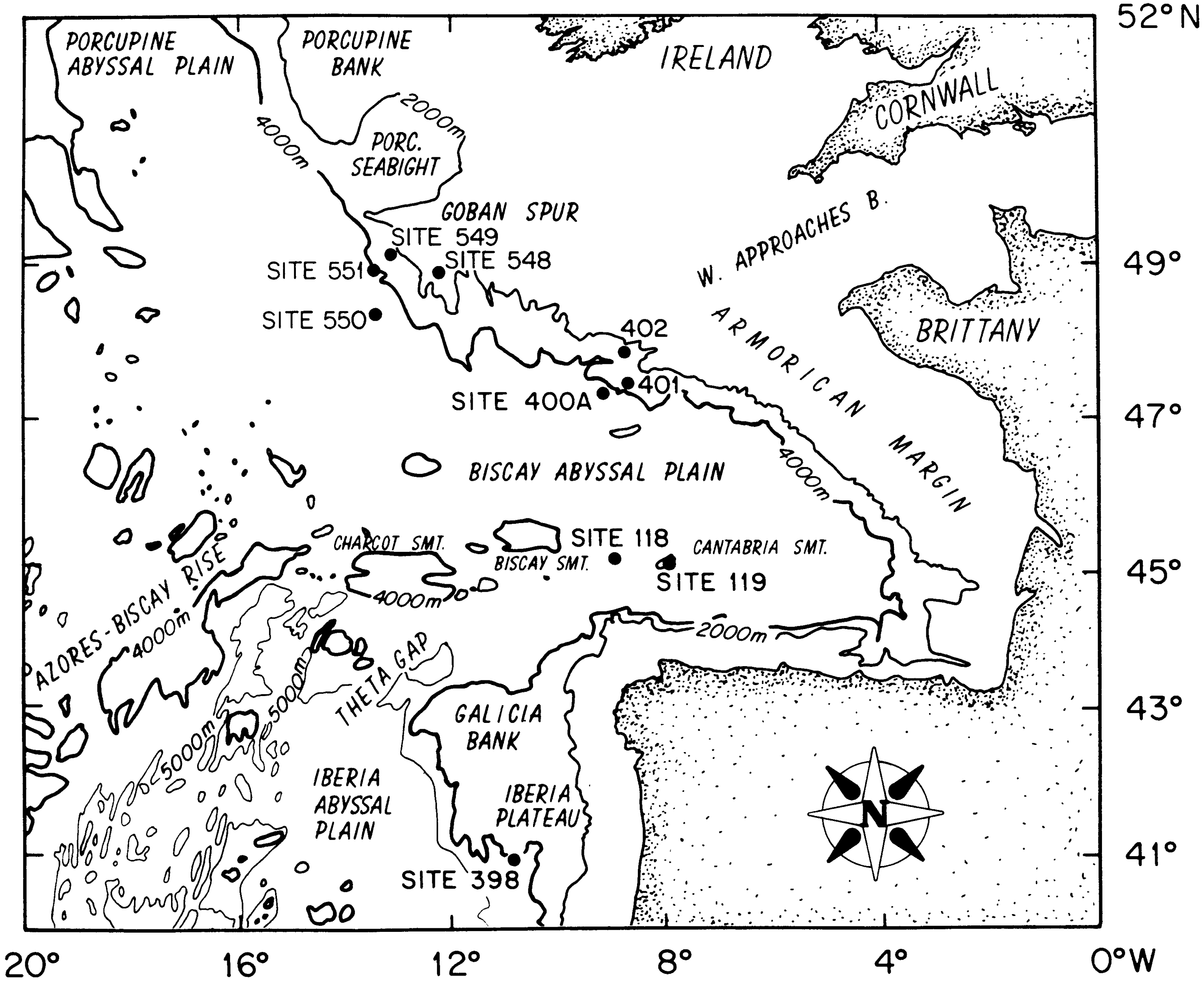
Distributions of modern deep-sea benthic foraminifers correlate with water masses and properties that co-vary with water masses (Streeter, 1973; Schnitker, 1974; Lohmann, 1978; Corliss, 1979b; Bremer and Lohmann, in press). Interpretations of water-mass/foraminiferal relationships have been extended to the Tertiary fossil record in order to determine changes in paleocirculation (Schnitker, 1979; Tjalsma and Lohmann, 1982; Douglas and

Woodruff, 1982). However, paleoceanographic interpretations of changes in deep-sea benthic foraminiferal assemblage or isotopic composition often conflict. This is illustrated by disputes as to the nature, timing, and causes of Late Paleogene benthic foraminiferal changes. Douglas (1973), Boersma (1977), and Schnitker (1979) suggested that a major benthic foraminiferal turnover occurred near the Eocene/Oligocene boundary, while Corliss (1979, 1981), Tjalsma (1982), and Tjalsma and Lohmann (1982) suggested more gradual/sequential changes. Benthic foraminiferal isotopic and assemblage data from both the early to middle Miocene (Blanc et al., 1980; Schnitker, 1979, 1980a,b) and the Late Paleogene (Miller and Curry, 1982; Miller, in press) have been cited as evidence for the first significant influx of bottom water into the North Atlantic from northern sources (Arctic and/or Norwegian-Greenland Sea).

The seismic stratigraphic record provides less ambiguous evidence for changes in abyssal circulation. Miller and Tucholke (in press) presented a model for the development of abyssal circulation in the North Atlantic derived from seismic stratigraphic studies of the Rockall region and southern Labrador Sea (Jones et al., 1970; Ruddiman, 1972; Roberts, 1975; Miller and Tucholke, in press). This method allows an independent determination of the nature of changes in abyssal circulation that are equivocal when interpreted from benthic foraminiferal changes alone. The model suggests that an erosional episode, representing the most vigorous bottom-water flow, occurred in the latest Eocene through early Oligocene, and that subsequent depositional sequences, representing generally decreasing bottom water flow (occasionally punctuated by erosional events), occurred in the late Oligocene through Neogene.

Miller (in press) used Tjalsma and Lohmann's (1982) taxonomic base to

Figure 1



52° N
49°
47°
45°
43°
41°
20°
16°
12°
8°
4°
0°W
PORCUPINE ABYSSAL PLAIN
PORCUPINE BANK
IRELAND
CORNWALL
PORC. SEABIGHT
2000m
4000m
GOBAN SPUR
SITE 549
SITE 548
SITE 551
SITE 550
W. APPROACHES B.
ARMORICAN MARGIN
BRITTANY
402
401
SITE 400A
BISCAY ABYSSAL PLAIN
4000m
SITE 118
CANTABRIA SMT.
SITE 119
CHARCOT SMT.
BISCAY SMT.
4000m
2000m
AZORES-BISCAY RISE
4000m
5000m
5000m
THETA GAP
GALICIA BANK
IBERIA ABYSSAL PLAIN
IBERIA PLATEAU
SITE 398
N

Location map, Bay of Biscay. (Modified after Miller, in press).

identify Late Paleogene bathyal to abyssal assemblages in the Bay of Biscay (DSDP Legs 12 and 48), and interpreted changes in assemblages in view of the abyssal circulation model. Miller and Curry (1982) attempted to date major benthic foraminiferal ^{18}O and ^{13}C enrichments in the Bay of Biscay and to relate them to changes in abyssal circulation and assemblages. Unfortunately, the stratigraphic record recovered in the Bay of Biscay previous to Leg 80 was not complete enough to resolve firmly the timing of major faunal iso- topic and assemblage changes and their relation to the circulation model.

This study analyzes changes in benthic foraminiferal assemblages from the Eocene through Oligocene of Site 548 (1256m present depth), Site 549 (2533m present depth), and Site 550 (4432m present depth), and presents detailed benthic foraminiferal $\delta^{18}O$ and $\delta^{13}C$ analyses from Site 549. The Eocene to Oligocene chronostratigraphic record recovered in the North Atlantic previous to Leg 80 is punctuated by numerous hiatuses (Moore et al., 1978; Miller and Tucholke, in press) that prevented delineating isotopic and faunal changes in this ocean. The relatively complete record recovered at Site 549 allows a determination of the faunal and isotopic changes across the Eocene/Oligocene boundary. Here, a major faunal change associated with a ^{13}C enrichment occurs in the early late Eocene; subsequent isotopic and faunal changes occur in the late Eocene, culminating in a sharp increase in $\delta^{18}O$ and $\delta^{13}C$ near the Eocene/Oligocene boundary.

PREVIOUS WORK

Schnitker (1979) described Paleocene to Oligocene benthic foraminiferal assemblages recovered from Sites 400A (4399m) and 401 (2495m) on the Armorican margin in the Bay of Biscay (Fig. 1). He determined that a "peak in faunal turnover" occurred at the Eocene/Oligocene boundary, agreeing with Kennett and Shackleton (1976) who suggested that a crisis in deep-sea benthic foraminifers occurred at this time. Subsequent workers disputed this conclusion. The late Eocene section is missing in Site 400A and the Oligocene is missing in Site 401 (Montadert, Roberts et al., 1979; Krasheninnikov, 1979; Muller, 1979); therefore, Schnitker's conclusions are based upon a comparison of different, time intervals from different paleodepths and do not clearly document the timing of the taxonomic changes (Miller, in press). Corliss (1979a; 1981), Tjalsma (1982), and Tjalsma and Lohmann (1982) proposed that no catastrophic changes occurred in response to the late Eocene to early Oligocene paleoceanographic changes, but that benthic foraminifers gradually or sequentially changed from the middle Eocene to the early Oligocene. As a means of testing this model, Corliss (1981) evaluated Schnitker's (1979) Site 400A data and concluded that in the Bay of Biscay there were two periods with an increased number of first and last appearences of taxa (Zones P12 and P18/19). However, hiatuses and drilling disturbance encountered in Site 400A make it likely that these periods of first and last appearances are artifacts of sampling. A more complete, undisturbed section was needed to determine Late Paleogene faunal changes in the Bay of Biscay.

Miller (in press) noted that a major faunal change occurred between the early middle Eocene and earliest Oligocene at two abyssal (> 3 km) sites in the Bay of Biscay (Sites 119 and 400A). Nuttallides truempyi, Clinapertina spp., and Abyssamina spp., which dominated the Eocene abyssal benthic assemblage, were replaced by increasingly abundant, bathymetrically wide-ranging and stratigraphically long-ranging taxa: Oridorsalis spp., Globocassidulina subglobosa, Gyroidinoides, and the Cibicidoides ungerianus plexus. Many of abyssal taxa become extinct before the early Oligocene, although a late Eocene hiatus encountered in Sites 119 and 400A prevented dating the timing of these extinctions. The faunal replacement noted in Site 119 and 400A is similar to that noted by Tjalsma and Lohmann (1982). In the shallower Bay of Biscay (Site 401; ~ 2000m paleodepth), Miller (in press) noted that an early late Eocene decrease in abundance and extinction of *N. truempyi* was the major faunal change. However, compositional differences between the deep and shallow benthic foraminiferal assemblages prevented further determination of the faunal changes.

Eocene to Oligocene deep-sea benthic foraminiferal changes are better defined in the South Atlantic and Caribbean due to the more complete reocrd recovered there. Here, Tjalsma and Lohmann (1982) showed that a sequence of faunal changes occurred during the middle Eocene to Oligocene: 1) a change from an early and middle Eocene *Nuttallides truempyi* assemblage to a late Eocene *Globocassidulina subglobosa*, *Gyroidinoides* spp., *Cibicicdoides ungerianus*, and *Oridorsalis umbonatus* assemblage; 2) the extinction of *N. truempyi* in the latest Eocene; and 3) the appearance of *Nuttallides umbonifera* in the late Eocene. In addition, Tjalsma (1982 and personal communication) noted that: 4) *N. umbonifera*, *G. subglobosa*, *Gyroidinoides* spp., *C. ungerianus*, *O. umbonatus* dominated in the Oligocene; and 5) no major changes in deep-sea benthic foraminiferal assemblages occurred across the Eocene/Oligocene boundary. The most dramatic of the middle Eocene to early Oligocene changes is the replacement of the *N. truempyi* assemblage by the *G. subglobosa*-*Gyroidinoides*-*C. ungerianus*-*O. umbonatus* assemblage. Based upon the replacement of *N. truempyi* in Site 357 (Rio Grande Rise) near the early/middle Eocene boundary, Tjalsma and Lohmann suggested that this replacement was diachronous, occurring in the early middle Eocene in shallow sites and near the middle/late Eocene boundary in deeper sites.

Conflicting paleoceanographic interpretations have been attributed to Late Paleogene benthic foraminiferal changes. Schnitker (1979, 1980a,b) ascribed the Eocene/Oligocene benthic foraminiferal changes in the Bay of Biscay to the initial formation of cold bottom water derived from the Southern Ocean. Miller and Tucholke (in press) noted that the influence of any southern bottom water on the deep Bay of Biscay probably was minimal due to its inhibition by topographic barriers: the Azores-Biscay Rise (Fig. 1), Azores-Gibraltar Rise, and Madeira-Tore Rise. Based upon this and the abyssal circulation model presented by Miller and Tucholke (in press), Eocene to Oligocene faunal (Miller, in press) and isotopic (Miller and Curry, 1982) changes in the Bay of Biscay were attributed to initial input of northern (viz. the Norwegian-Greenland Sea and/or Arctic Ocean) sources of cold bottom water.

Miller (in press) noted that during the Oligocene, Nuttallides umbonifera replaced the Eocene species N. truempyi as the predominant abyssal benthic foraminifer, reaching peak abundance in the middle Oligocene at Sites 119 and Site 400A. In the modern oceans, the abundance of N. umbonifera is positively correlated with increased corrosiveness of bottom water (Bremer and Lohmann, in press); at Site 119 the greatest abundances of Nuttallides spp. are associated with the lowest $\delta^{13}C$ values in benthic foraminifers (Miller and Curry, 1982). Lower $\delta^{13}C$ values are often associated with water masses that are, in turn, more corrosive to carbonate (Kroopnick et al., 1972; Kroopnick, 1974, 1980; Lohmann and Carlson, 1981) Thus, the early-middle Eocene and middle Oligocene of Site 119 were interpreted as reflecting older, more corrosive bottom water (Miller and Curry, 1982; Miller, in press).

Miller and Curry (1982) reported on oxygen and carbon isotopic compostion of benthic foraminifers from Sites 119 and 401. They found that $\delta^{18}O$ values increased ~ 1.9 $^{o}/oo$ and $\delta^{13}C$ values increased ~ 0.8 $^{o}/oo$ between the early middle Eocene (Zones NP13-NP15) and the earliest Oligocene (Zone NP21) of Site 119. They combined isotopic records for Sites 119 and 401 and concluded that ~ 1.4 $^{o}/oo$ of the $\delta^{18}O$ increase occurred in the late Eocene to earliest Oligocene. Based on biostratigraphic zonations, they could not resolve firmly the exact timing of the enrichment, which may have occurred within an interval less than one million year (within Zone NP21) or as long as 4 million years (between Zones P15 and NP21).

The isotopic results of Vergnaud-Grazzinni et al. (1978, 1979b) from Site 400A (Bay of Biscay) and Vergnaud-Grazzini (1979) from Site 398 (off Portugal)(Fig. 1) contrast with those of Miller and Curry (1982). However, Miller and Curry (1982) suggested that the low Eocene benthic foraminiferal $\delta^{18}O$ values noted in Sites 400A and 398 (as much as 2 $^{o}/oo$ lower than at Sites 119 and 401) resulted from diagentic alteration of foraminiferal tests in these deeper-buried sites.

METHODS

Benthic foraminifers were examined from the Eocene through Oligocene sections recovered at Site 549 (25 samples), Site 548 (13 samples), and Site 550 (14 samples). The samples were washed through a 63 μm sieve with hydrogen peroxide and sodium carbonate or a Calgon solution. Aliquots from the > 149 μm size-fraction (the size-fraction used by Tjalsma and Lohmann, 1982, and Miller, in press) were picked and mounted on a reference slide. Approximately 300-500 specimens were picked whenever possible (Tables 1-2).

The benthic foraminifers were identified using the taxonomy outlined by Tjalsma and Lohmann (1982) and Miller (in press). The census data is presented as percent of total benthic foraminifers counted (Tables 1,2).

Using computer programs provided by Lohmann (1980 and personal communication), we made principal component analyses and compared census values from Sites 548 (1256m present depth) and Site 549 (2533m present depth) with previous results from Sites 119 (4447m present depth) and 401 (2495m present depth)(both Bay of Biscay) and with the species distribution (percent) data of Tjalsma and Lohmann (1982).

We computed Q-mode principal components for Site 549 and Site 548 samples with > 100 specimens and computed loadings of the Bay of Biscay samples onto Tjalsma and Lohmann's (1982) Eocene Q-mode principal components. Tjalsma and Lohmann (1982) presented census data for 48 Eocene taxa that were used in all these computations. Although they did not record some taxa that were abundant in this study (mainly Astrononion pusillum), their 48 taxa account for approximately 70 $^{0}/o$ of the total benthic foraminifers counted for Sites 549 and 548 (except sample 549A 7-6; Tables 1,2). Tjalsma and Lohmann's (1982) 48 taxa also accounted for ~ 70 $^{0}/o$ of the total fauna counted within their data set (table 6 in Tjalsma and Lohmann, 1982).

Isotopic analyses were performed on monogeneric samples of the benthic foraminiferal taxon Cibicidoides. Samples for isotopic studies were ultrasonically cleaned in distilled water, lightly crushed in methyl alcohol, and heated at 370°C in a vacuum to remove organic matter. The $CaCO_3$ was reacted with H_3PO_4 at 50°C. Evolved gas was frozen in a liquid nitrogen cold finger as it was produced. CO_2 and water were separated by a series of distillations, and the purified CO_2 was analyzed in an on-line VG Micromass 602E massspectrometer at Woods Hole Oceanographic Institution. Results are presented as per mil differences from PDB. Based upon replicate analyses of NBS-20, analytical precision is ~ ±0.1 $^{0}/oo$ for both $\delta^{18}O$ and $\delta^{13}C$. Duplication of three benthic foraminiferal samples yielded results which differed by 0.16, 0.3, and 0.04 $^{0}/oo$ for $\delta^{18}O$ and 0.03, 0.04, and 0.11 $^{0}/oo$ for $\delta^{13}C$.

RESULTS AND DISCUSSION

Paleobathymetry

Sites 548 and 549 were drilled on continental crust and therefore cannot be "backtracked" in the standard manner (Berger and Winterer, 1974; Sclater et al., 1977). However, Site 550 was drilled on oceanic crust and empirical age-versus-subsidence curves for oceanic crust (Sclater et al., 1971; Parsons and Sclater, 1977) can be used to "backtrack" this site; its subsidence can be used to estimate the subsidence of Sites 548 and 549. Backtracking Site 550 (4432m present depth) along an empirical age-subsidence curve for the North Atlantic (Tucholke and Vogt, 1979), assuming a crustal age of early Albian (~ 107 Ma) and correcting for isostatic loading of sediments, suggests Eocene paleodepths of ~ 4300m and Oligocene paleodepths of ~ 4400m. These estimates agree with faunal estimates of paleodepth of > 3km. The benthic foraminifers at Site 550 are similar to assemblages noted at Site 119 and Site 400A (Miller, in press) and to abyssal sites of Tjalsma and Lohmann (1982)(for a full description see Miller, in press).

Assuming that the elevation difference of basement between Sites 550 and Site 549 remained constant (~ 1929m; de Graciansky, Poag et al., this volume), seafloor at Site 549 (2538m present depth) was at ~ 2400m in the Eocene and ~ 2500m in the Oligocene. Similarly, assuming a constant elevation difference between the basement at Site 548 and Site 550 (presently ~ 3300m difference), Eocene paleodepths are ~ 1000m and Oligocene paleodepths are ~ 1100m.

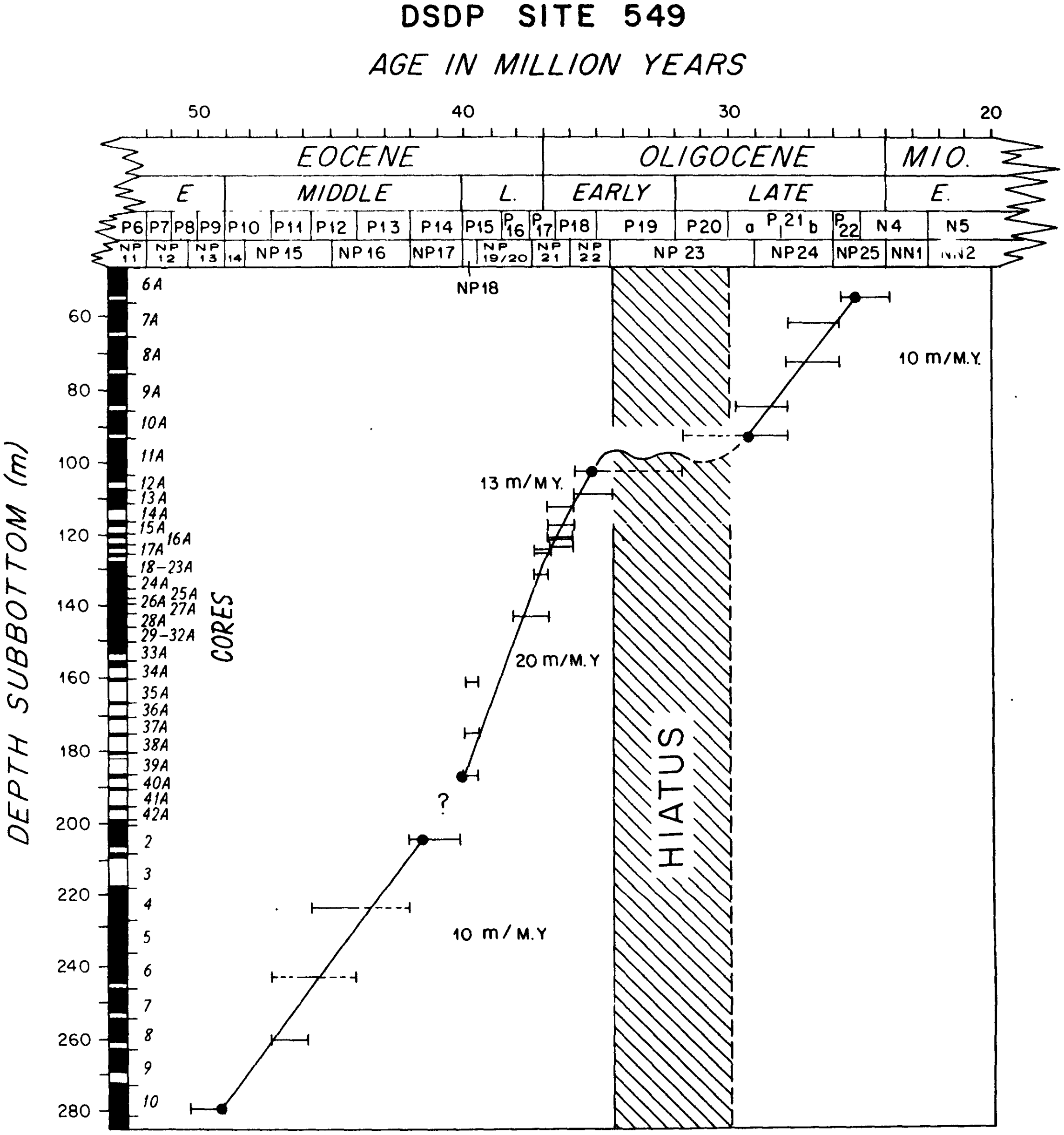
Faunal paleobathymetric evidence supports these crude estimates. Tjalsma and Lohmann (1982) and Tjalsma (personal communication) compiled benthic foraminiferal census data for DSDP sites on oceanic crust that had been backtracked. Benthic foraminifers at Site 549 are typical for intermediate sites when compared with Tjalsma and Lohmann's (1982) Eocene data. The absence of *Abyssammina* spp. and *Clinapertina* spp., low abundance of *Aragonia*, *Cibicidoides grimsdalei*, greater abundance of *C. tupamensis*, *Hanzawaia cushmani*, and *Osangularia* spp. are typical for sites between 2000 and 2500m. Similarly, comparison with Oligocene data (Tjalsma, personal communication) shows that the low abundance of *Lenticulina*, *Bulimina alazanensis*, *Siphonina*, *Nuttallides umbonifera*, and high abundance of *Astrononion pusillum* also are typical for these paleodepths.

The Site 548 assemblages compare well with those from Site 516 (Rio Grande Rise, South Atlantic), which has been backtracked to ~ 1.2km (Tjalsma, personal communication). These assemblages are characterized by the very high abundance of *Lenticulina* and *B. alazanensis*, high abundance of *Planulina* spp. and *Siphonina* spp., and lower abundance of pleurostomellids.

Biostratigraphy, Site 549

Two holes were cored at Site 549. Hole 549A was hydraulic piston cored from cores 1 through 42 (Pleistocene to late Eocene). Hole 549 (middle Eocene and older) was rotary drilled. The samples we obtained for quantitative benthic foraminiferal studies were also examined for qualitative planktonic foraminiferal content; in addition, nannofossil zonations for the samples were provided by M.-P. Aubry (personal communication)(Table 3). The zonations presented here (and those for Site 548 and 550) agree well with the zonations

Figure 2



Zonal age assignment and estimated average sedimentation rate Site 549. Error bars show the length of the biozone(s) to which the sample have been assigned (Table 3). Solid dots indicate ties for interpolating sedimentation rate and age estimates.

presented in the site report; although some disagreements between planktonic foraminiferal zonations occur in the middle Oligocene, the major stage and epoch boundaries are coincident.

The placement of the Eocene/Oligocene boundary in Site 549 warrants further discussion. The boundary was placed at the level of the extinction of *Globorotalia cerroazulensis cunialensis*, *G. cerroazulensis cocoaensis*, *Hantkenina alabamensis*, and *Cribrohantkenina* (Plate 6). This occurs within the 22m interval we assigned to Zone NP21, which is defined by the presence of *Ericsonia formosa*, *Discoaster tani nodifer*, *Isthmolithus recurvus*, and *Chiasmolithus oamaruensis* after the extinction of *D. saipanensis* and *D. barbadiensis* (M.-P. Aubry, personal communication). This agrees with the relationships noted by Van Couvering et al. (1982); they noted that the extinction of planktonic foraminifers used to identify the top of the Eocene occurs prior to the extinction of the taxa used to define the top of Zone NP21. Thus, Zone NP21 straddles the Eocene/Oligocene boundary (see also Hardenbol and Berggren, 1978). Average sedimentation rate for the late Eocene to early Oligocene section is ~ 20 m/m.y; there is no evidence for a hiatus at the Eocene/Oligocene boundary, although core recovery was incomplete (Fig. 2). Although the latest early Eocene through early Oligocene section is probably continuous, a major hiatus (~ 6 my) occurs in the middle Oligocene of Site 549 (between cores 11-6, 143-148cm and 10-6, 143-148cm)(Fig. 2). This hiatus correlates with a similar gap in Site 548.

The middle Oligocene hiatus delineated in Figure 2 may be revised in light of future biostratigraphic studies. The early Oligocene age (upper Zone NP21 to Zone NP22) of cores 17-1 through 11-6 is well documented (Table 3). However, Berggren (personal communication) notes that the last appearance of *Pseudohastigerina* occurs not at the top of Zone P19/20 and near the Zone

NP23/NP24 boundary as he suggested previously (Berggren, 1971, 1972b; Hardenbol and Berggren, 1978), but at the NP22/NP23 boundary. The overlying sample (10-6) is assigned to Zone NP23 (M.-P. Aubry, personal communication; Table 3); unfortunately, the planktonic foraminiferal age assignment is eqivocal (probably Zone P21A; possibly Zone P19/20; Table 3). Sample 9-6 is late Oligocene (Zone P21A; Zone NP24/NP25, M.-P. Aubry, personal communication). Zone NP23 (~ 5.5 my in duration; Hardenbol and Berggren, 1978) is therefore constrained to an interval less than 20m thick (average acumulation rate < 4 m/my), suggesting that a disconformity may exist. Revisions in the Paleogene time scale and zonations (Berggren et al., in preparation) will not substantially alter the length of Zone NP23; therefore, it is probable that a disconformity exists between samples 9-6 and 11-6.

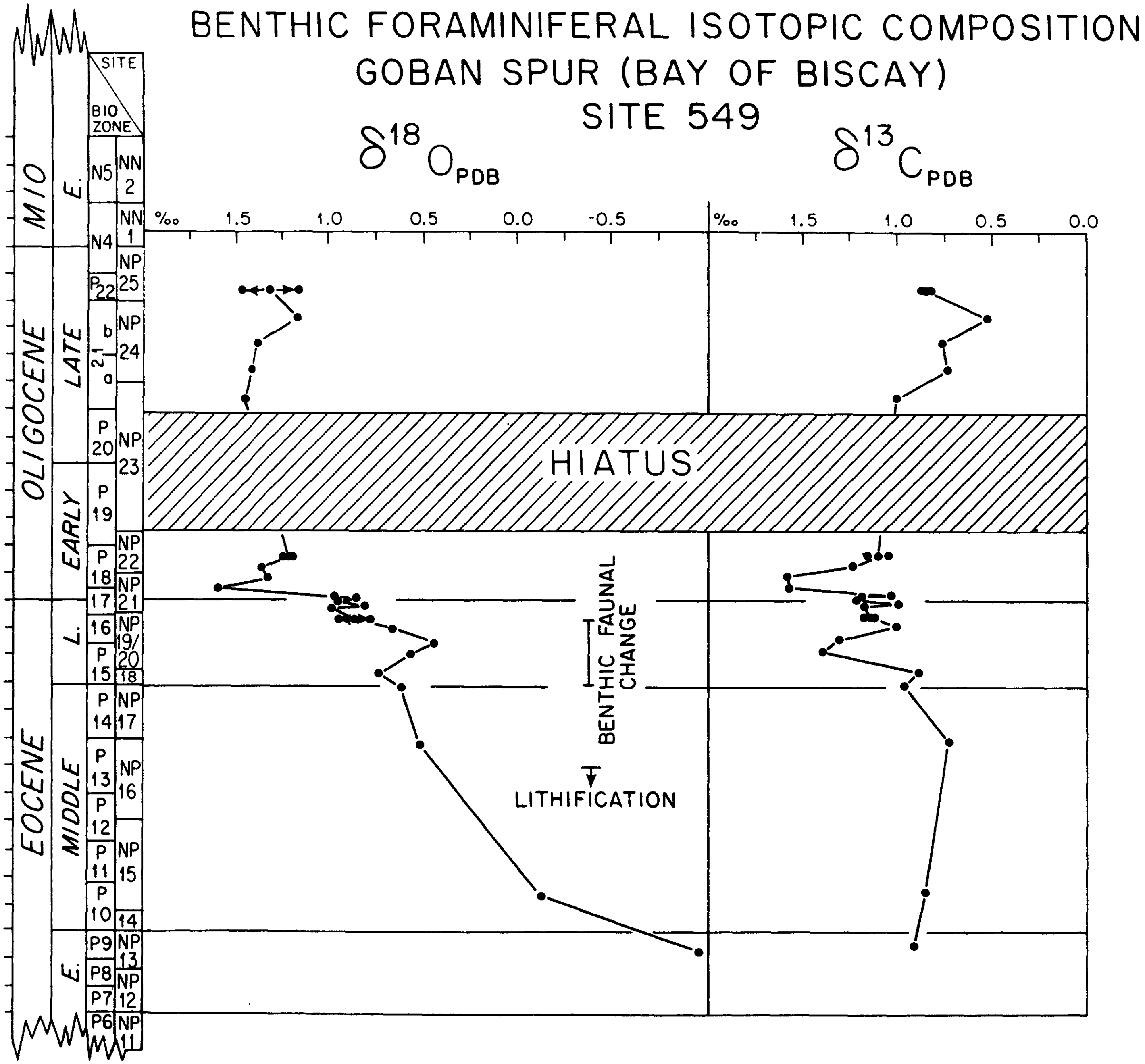
The zonal schemes outlined in Tables 3 and 4 are dated using the time scale of Hardenbol and Berggren (1978). There is considerable debate as to the age of the Eocene/Oligocene boundary, with one school (Odin, 1978; Odin et al., 1978; Glass and Crosbie, 1982) suggesting an age of 32 Ma and another school which maintains an age of ~ 37 Ma (Berggren, 1972; Hardenbol and Berggren, 1978; Berggren et al., in preparation). We have used the latter. Revisions or differences in the time scale used could substantially alter the "absolute timing" of the benthic events dated here. Still, the events presented are correlated to planktonic zonations and are therefore well established (both biostratigraphically and biochronologically) relative to the various stage/epoch boundaries. The chronostratigraphic relationships of these events should not change substantially.

Benthic Foraminiferal Isotopic Record, Site 549

Isotopic analyses were performed on monogeneric samples of the benthic foraminiferal taxon Cibicidoides (Plates 1-2). Interregional isotopic variations of this taxon in Holocene core tops reflect the distribution of temperature, $\delta^{18}O$ of seawater, and $\delta^{13}C$ and CO_2 of seawater in the modern oceans (Belanger et al., 1981; Graham et al., 1981). Cibicidoides species secrete their tests lower in $\delta^{18}O$ than equlibrium values calculated from the Epstein et al. (1953) paleotemperature equation; however, this offset appears to be constant (~ 0.65 lower, Shackleton and Opydyke, 1973; Duplessy et al., 1980; Woodruff et al., 1980; Belanger et al., 1981, Graham et al., 1981).

In sampling, we noted an increase in lithification in Hole 549, cores 3 to 8. Benthic foraminiferal recovery and preservation was poor in these cores. We therefore suspected diagnentic alteration of the sediments recovered in Hole 549 (as was noted in Site 400A, and Site 398), even though the burial depths for these cores are only ~ 220-280m. Insufficient benthic foraminifers were available for analysis through most of this section. Oxygen isotopic values for cores 549-8 and 549-10 are low (~ 0.0 and -1.0 $^{o}/oo$ PDB, respectively)(Fig. 3; Table 5) relative to other early to early middle Eocene values from Site 119 and to early through middle Eocene values from Site 401 (both ~ 0.5 $^{o}/oo$ PDB; Miller and Curry, 1982). These low values are similar to those noted from the diagnetically altered sections of Sites 400A and 398 (Vergnaud-Grazzini et al., 1978, 1979; Vergnaud-Grazzini, 1979). Miller and Curry (1982) noted that such differences within the Bay of Biscay are oceanographically unreasonable and are too great to be explained by sample biases (selective dissolution, analyses of mixed versus monogeneric

Figure 3



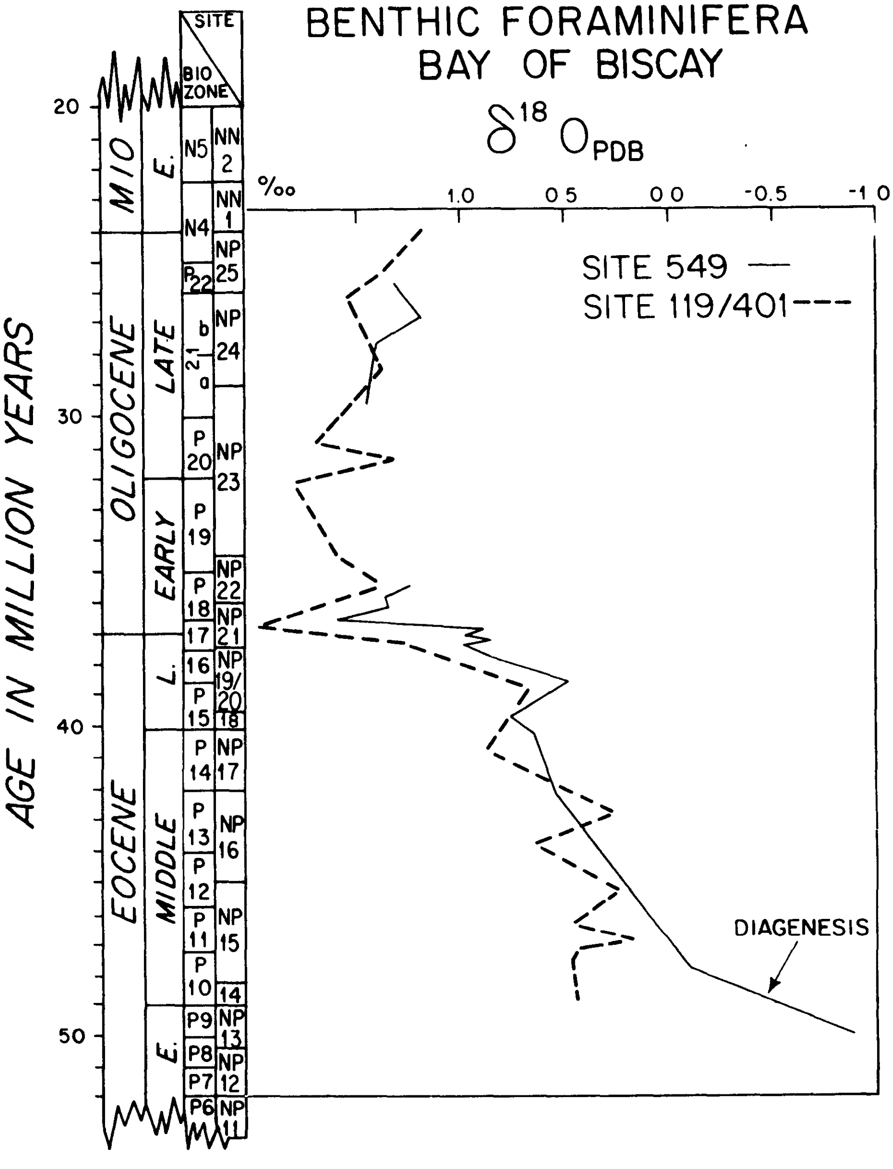
Age versus benthic foraminiferal isotopic compostion, Site 549. Faunal change indicated represents the change from a Nuttallides truempyi-Bulimina spp. assemblage to an assemblage dominated by stratigraphically long- and bathymetrically wide-ranging forms.

assemblages) alone. We therefore interpret these values as resulting from diagnetic alteration.

Samples from Hole 549A (and 549-2) show less lithification and better preservation and recovery of benthic foraminifers, while oxygen and carbon isotopic values are similar (Figs. 3,4) to the apparently unaltered isotopic records at Site 401 and Site 119 (Miller and Curry, 1982). Oxygen isotopic composition is relatively constant across the middle/late Eocene boundary (~ 0.6 $^{0}/oo$)(Fig. 3); values increase ~ 0.3 $^{0}/oo$ in the late Eocene and ~ 0.7 $^{0}/oo$ in the earliest Oligocene for a total late Eocene to earliest Oligocene increase of ~ 1.0 $^{0}/oo$. Little $\delta^{18}O$ change is seen in the Oligocene following the earliest Oligocene maximum. $\delta^{13}C$ increases ~ 0.8 $^{0}/oo$ in the early late Eocene, decreases ~ 0.4 $^{0}/oo$ in the latest Eocene, and increases ~ 0.6 $^{0}/oo$ in the earliest Oligocene (Fig. 3). The magnitude of the late Eocene to earliest Oligocene ^{18}O enrichment is greater in the Site 119/401 composite than in Site 549 (~ 1.5 $^{0}/oo$ versus ~ 1.0 $^{0}/oo$, respectively). This may reflect a greater temperature drop in the deep Site 119 relative to the shallower Site 549 (the Oligocene portion of the composite record is from Site 119). The carbon isotopic record shows greater detail in the late Eocene to earliest Oligocene of Site 549, but does not show the distinct middle Oligocene $\delta^{13}C$ minimum noted in Site 119 (Miller and Curry, 1982), probably owing to a hiatus (see above).

The $\delta^{18}O$ values we report from Site 549 are similar to values obtained in the Southern and Pacific Oceans: Sites 277 (Campbell Plateau; 1214m water depth; Kennett and Shackleton, 1976; Keigwin, 1980) and Site 298 (Phillipine Sea; 2943m water depth; Keigwin, 1980). Keigwin noted that ^{18}O enrichments of ~ 1.0 $^{0}/oo$ and ~ 1.5 $^{0}/oo$ occurred in Sites 277 and 292, respectively, just above the level of the Eocene/Oligocene boundary (following the

Figure 4

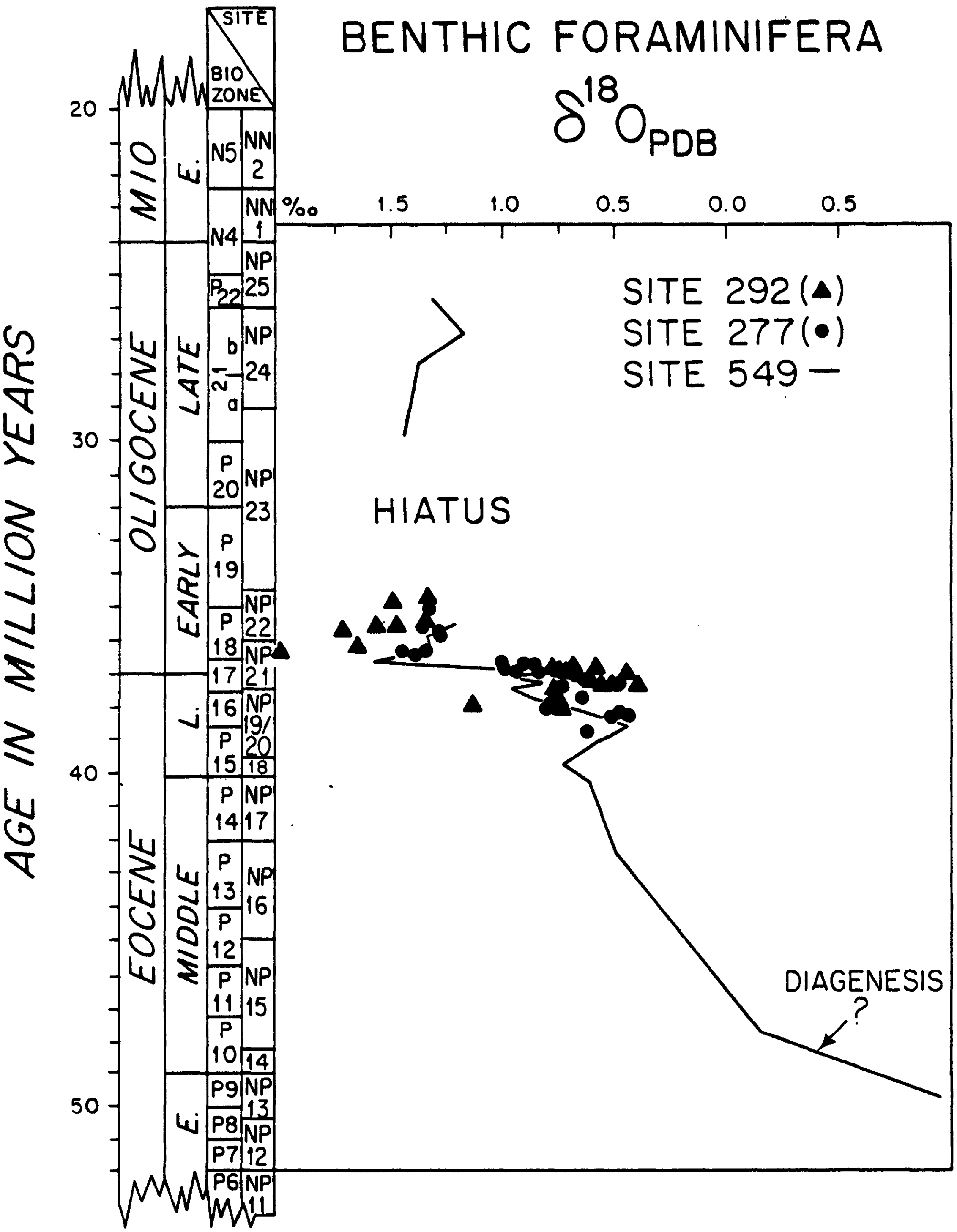


Comparison of Site 549 benthic foraminiferal oxygen isotopic composition with composite record from Bay of Biscay Sites 119 and 401. (Site 119 and Site 401 data after Miller and Curry, 1982).

extinction of Eocene planktonic foraminifers; Keigwin, 1980). Comparison of Keigwin's $\delta^{18}O$ records with the benthic foraminiferal $\delta^{18}O$ record at Site 549 (Fig. 5) shows that the enrichments at all three location may be considered synchronous. This conclusion assumes that the extinctions of the planktonic foraminifers used to defined the Eocene/Oligocene boundary in all three cases (Site 549, this paper; Site 277, Kennett, Houtz et al., 1974; Site 277, Karig, Ingle et al., 1975) are synchronous. The magnitude of the ^{18}O enrichment is the same at both Site 549 and Site 277. The magnitude of the enrichment at Site 292 is greater by ~ 0.5 $^{o}/oo$, but is similar to that noted in the Site 119/401 composite (Fig. 4).

The ^{18}O enrichment in benthic foraminifers may represent a change in the isotopic composition of seawater or a drop in bottom-water temperature or both. Early workers suggested that the major buildup of continental ice in the Tertiary did not occur until the middle Miocene, and therefore, that the enrichment noted near the Eocene/Oligocene boundary represented a major cooling (Shackleton and Kennett, 1975; Savin et al., 1975; Kennett and Shackleton, 1976; Savin, 1977). Matthews and Poore (1980), on the other hand, suggested that the enrichment may represent the first major buildup of continental ice on Antarctica. Their model requires that the enrichment in ^{18}O of both benthic and tropical planktonic foraminifers is globally synchronous, unless affected by local temperature-salinity perturbations. Presently two exceptions to this synchronous enrichment apparently occur: 1) in Site 292 the enrichment occurs mainly in benthic foraminifers (1.5 $^{o}/oo$ versus only 0.3 $^{o}/oo$ enrichment in planktonic foraminifers)(Keigwin, 1980); and 2) in the South Atlantic (Site 357) the isotopic enrichment occurs between the middle and late Eocene (Boersma and Shackleton, 1977). However, the record recovered at Site 357 is incomplete, and Boersma and Shackleton (1977)

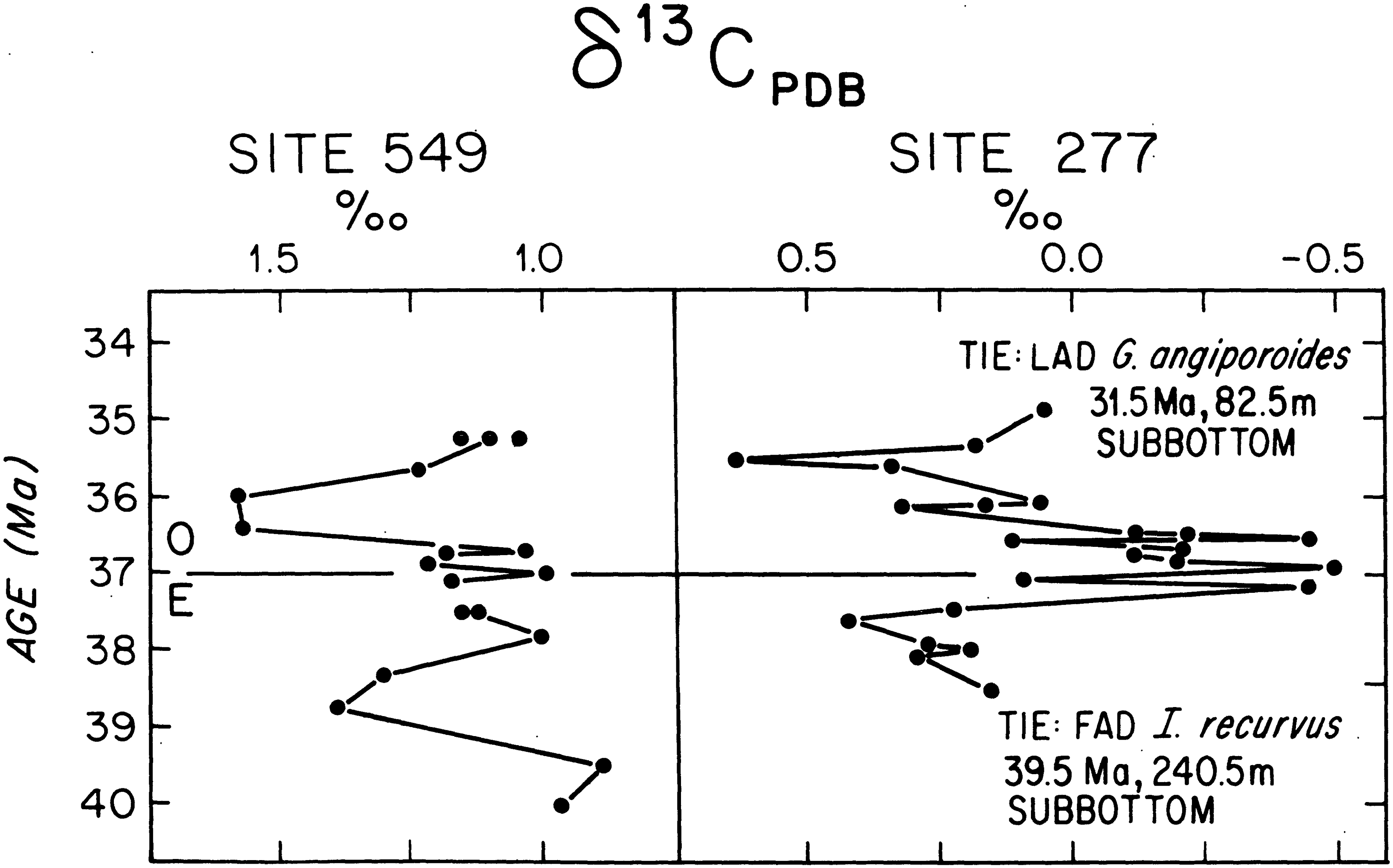
Figure 5



Comparison of Site 549 benthic foraminiferal oxygen isotopic composition with records from Site 277 (Campbell Plateau) and Site 292 (Phillipine Sea). Site 277 age control taken by interpolating between tie points at 31.5 Ma (82.5m subbottom, LAD S. angiporoides, Kennett, Houtz et al., 1974), 37.0 (188.5m, Eocene/Oligocene boundary, Keigwin, 1980), and 39.5 Ma (240.5m, FAD I. recurvus, Kennett, Houtz et al., 1974) which results in a constant sedimentation rate of 20 m/m.y. for the late Eocene to middle Oligocene section. Site 292 age control taken by interpolating between 30.0 Ma (225m subbottom, base of Zone P21, Karig, Ingle et al., 1975) and 37.0 Ma (Eocene/Oligocene boundary, Keigwin, 1980) and extrapolating the sedimentation rate to the bottom of the section. (Site 277 and Site 292 isotopic data after Keigwin, 1980).

reported only one value from the late Eocene to earliest Oligocene (P15-P18) for this site. Keigwin argued that the $\delta^{18}O$ increase in benthic foraminifers represented mostly a temperature decrease because of the lack of covariance at tropical Site 292, although this could be attributed, in part, to a local surface-water warming or decrease in salinity in the Phillipine Sea. Assuming buildup of glacial ice from an ice-free Eocene world to a fully-glaciated Oligocene world (with ice volume equal to present-day ice volume), Miller and Curry (1982) argued that the $\delta^{18}O$ increase must represent at least a 2°C temperature drop.

The late Eocene to earliest Oligocene $\delta^{13}C$ records at Site 549 and Site 277 show remarkably similar patterns, although they are ofset, differing in value by ~ 1.0 ‰ due, in part, to the different species analysed (Cibicidoides spp., Site 549; Oridorsalis spp., Site 277, Keigwin, 1980)(Fig. 6). It might be hypothesized that the $\delta^{13}C$ records are correlative, and that they are a function of global changes in carbon, reflecting either sea level (hence increased terrestrial-carbon or shelf-carbon input; Shackleton, 1977; Broecker, 1982) or global circulation changes. However, we believe that the offsets in time shown in the $\delta^{13}C$ records (Fig. 6) are real. Assuming that the peaks in $\delta^{13}C$ are correlative results in diachronous (relative to the $\delta^{13}C$ changes) placement of the $\delta^{18}O$ changes and the diachronous extinction of the planktonic foraminifers used to identify the Eocene/Oligocene boundary. The lack of a similar $\delta^{13}C$ record at Site 292 argues against global changes in the carbon budget; the coincidence of the $\delta^{18}O$ and planktonic foraminiferal extinctions at all three sites argues that these are isochronous and that the changes in $\delta^{13}C$ at Site 277 and Site 549 are, in fact, slightly diachronous.



Comparison of Site 549 benthic foraminiferal carbon isotopic composition with record from Site 277 (Campbell Plateau). Age control for Site 277 same as Figure 5.

Benthic Foraminiferal Assemblages, Site 549

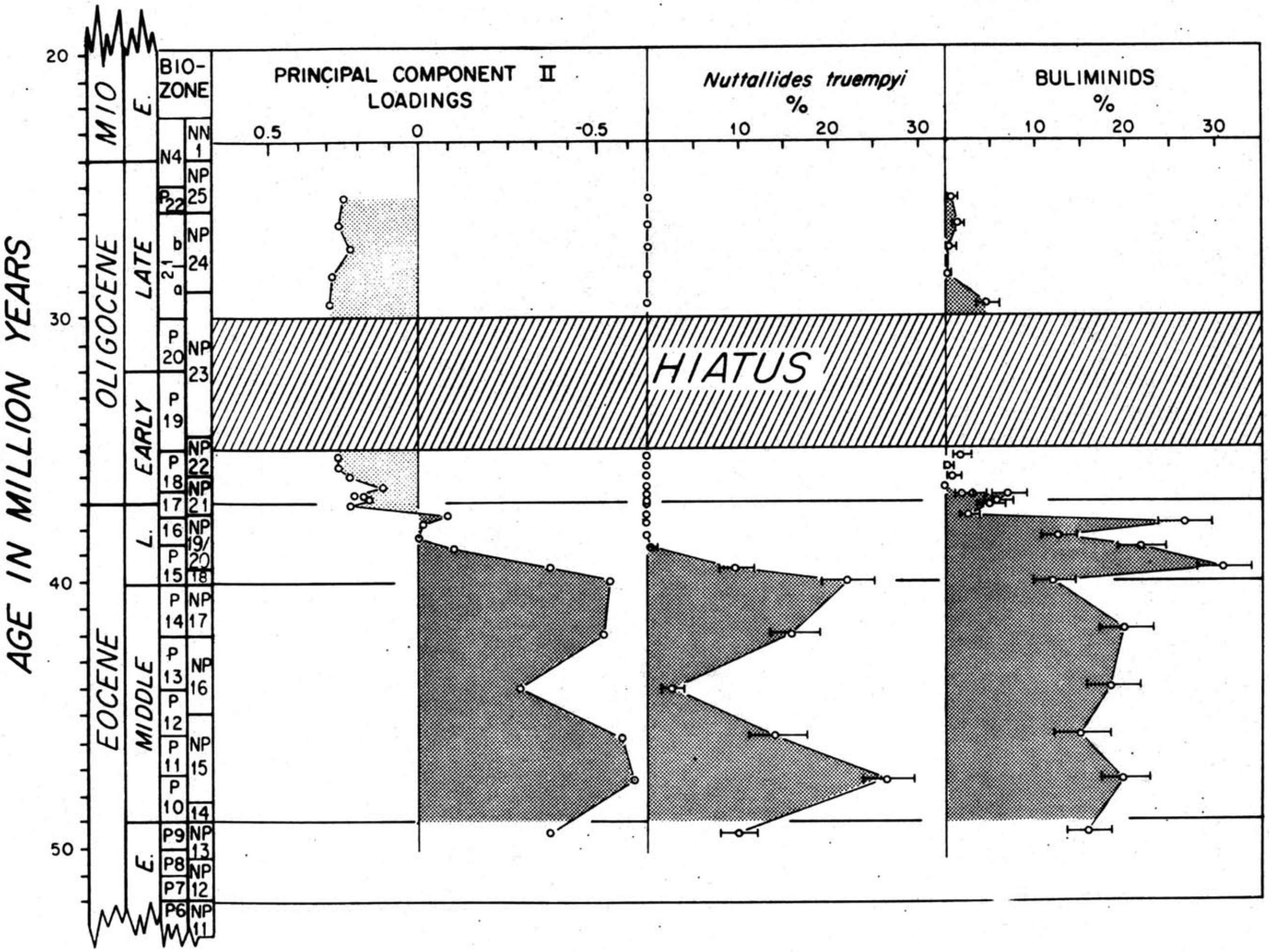
The early to middle Eocene assemblages are dominated by Nuttallides truempyi (Plate 3) and buliminids (Bulimina glomarchallengeri, B. trinitatensis/impendens, B. semicostata, with B. jarvisi (Plate 4) becoming important in the late middle Eocene)(Figs. 7,8). Lenticulina spp., Cibicidoides sp. 1, and Osangularia spp. (Fig. 8) are also more abundant in the early to middle Eocene. Oridorsalis spp. (mostly O. umbonatus), Gyroidinoides spp., Globocassidulina subglobosa, Cibicidoides ungerianus, Pullenia eocenica, and Stilostomella subspinosa are important, stratigraphically long-ranging and bathymetrically wide-ranging taxa that are abundant throughout the Eocene and Oligocene (Fig. 9).

Nuttallides truempyi decreases in abundance and becomes extinct in the earliest late Eocene (between ~ 38.5 and 40 Ma). The buliminids subsequently increase in abundance, only to decrease in abundance just below the Eocene/Oligocene boundary (~ 37.5 Ma).

The replacement of the N. truempyi assemblage and the increase and subsequent decrease of the buliminids represents the most important source of faunal variation in the Eocene to Oligocene section (Fig. 7). We made a Q-mode principal component analysis of the 25 Site 549 samples. Principal component 2 explains 28.8 $^{o}/o$ of the faunal variation (for the 48 taxa used by Tjalsma and Lohmann, 1982 and Miller, 1982; Table 1) in the Eocene to Oligocene section of Site 549. The N. truempyi assemblage is represented by high negative loadings on principal component 2 (Fig. 7); this assemblage also contains higher than average abundance of buliminids, Osangularia spp., and Lenticulina spp. High positive loadings on principal component 2 represent the latest Eocene through Oligocene assemblage dominated by the long- and

Figure 7

SITE 549



Benthic foraminiferal assemblage composition, Site 549. Principal component analysis was performed on 25 samples from this site. Error bars on percentages indicate 80 o/o confidence interval.

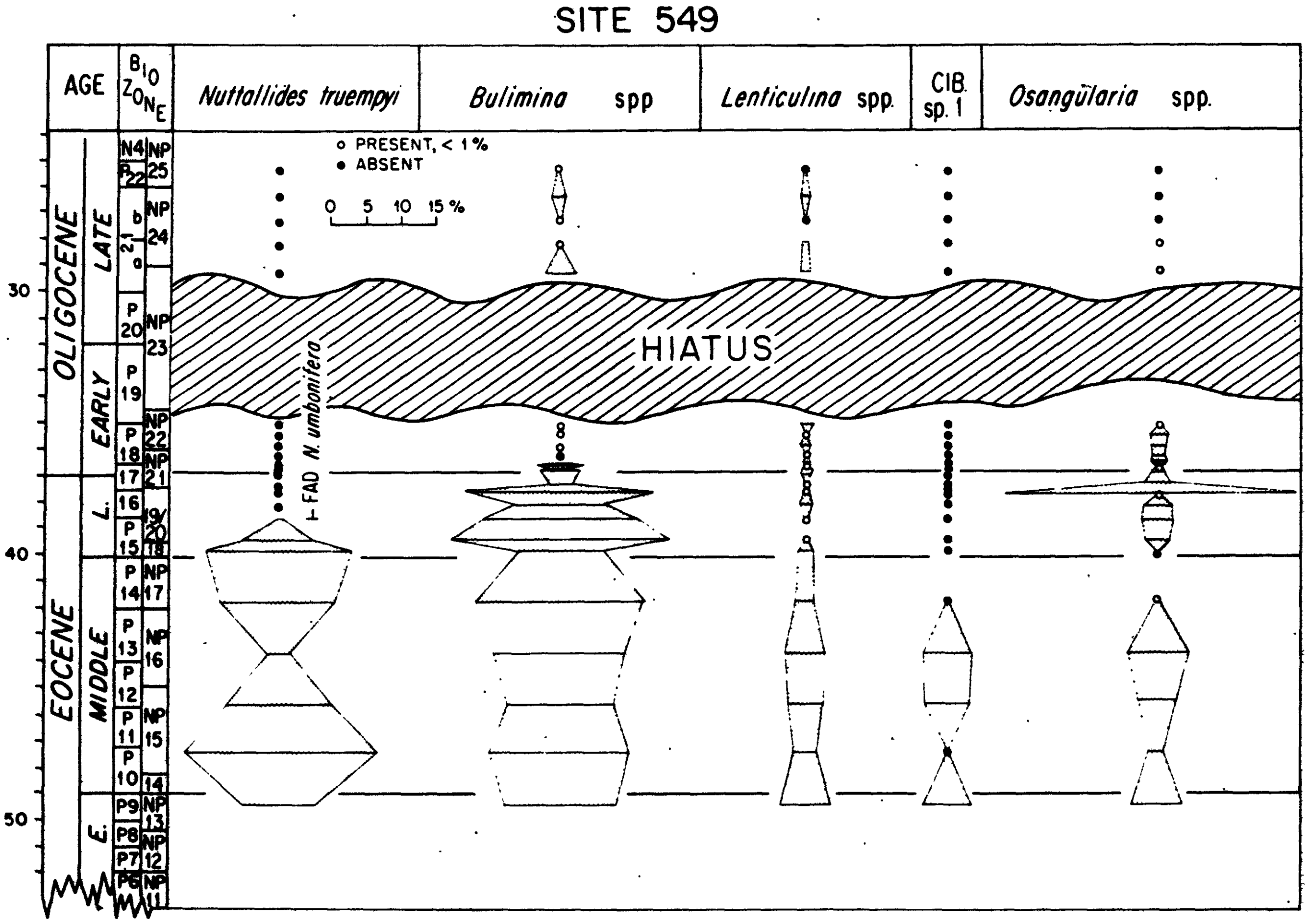
wide- ranging taxa (*G. subglobosa*, *C. ungerianus*, *Gyroidinoides* spp., *P. eocenica*, and *S. subspinosa*). Low positive loadings on principal component 2 in the late Eocene (37.5 to 38.5 Ma) represent the dominance of the buliminids after the extinction of *N. truempyi*.

Our assemblages compare closely with those identified by Tjalsma and Lohmann (1982) We estimated loadings of Site 549 samples onto Tjalsma and Lohmann's (1982) principal components in order to determine how these samples compare with the assemblages that they defined. These loadings are components of the sample vectors projected onto the coordinate system defined by Tjalsma and Lohmann's principal component analyses of their Eocene samples. The early to earliest late Eocene samples from Site 549 give low to moderate positive loadings on their Q-mode principal component 2, and negative loadings on their principal component 3. This corresponds to their *Lenticulina*-*Bulimina*-*Osangularia* assemblage. The late Eocene through Oligocene samples from Site 549 have high positive loadings on Tjalsma and Lohmann's principal component 3 and high negative loadings on principal component 2; this corresponds to their *G.subglobosa*-*Gyroidinoides*-*C. ungerianus*-*Oridorsalis* assemblage.

The replacement of the *Nuttallides truempyi*-dominated assemblage occurs near the middle/late Eocene boundary in deeper sites in the South Atlantic (Tjalsma and Lohmann, 1982), correlating with its replacement in Site 549. Elsewhere in the Bay of Biscay, a similar timing for the replacement is noted at Site 401 (Fig. 10)(~ 2.0- 2.5km paleodepth; Miller, in press). The records recovered at the shallower Site 548 and the deeper (> 3km) Site 119 and Site 400A are more incomplete (Fig. 10; fig. 11 in Miller, in press); however, the timing of the faunal replacement at these sites does not contradict the timing observed in the better-recovered records at Site 401 and Site 549.

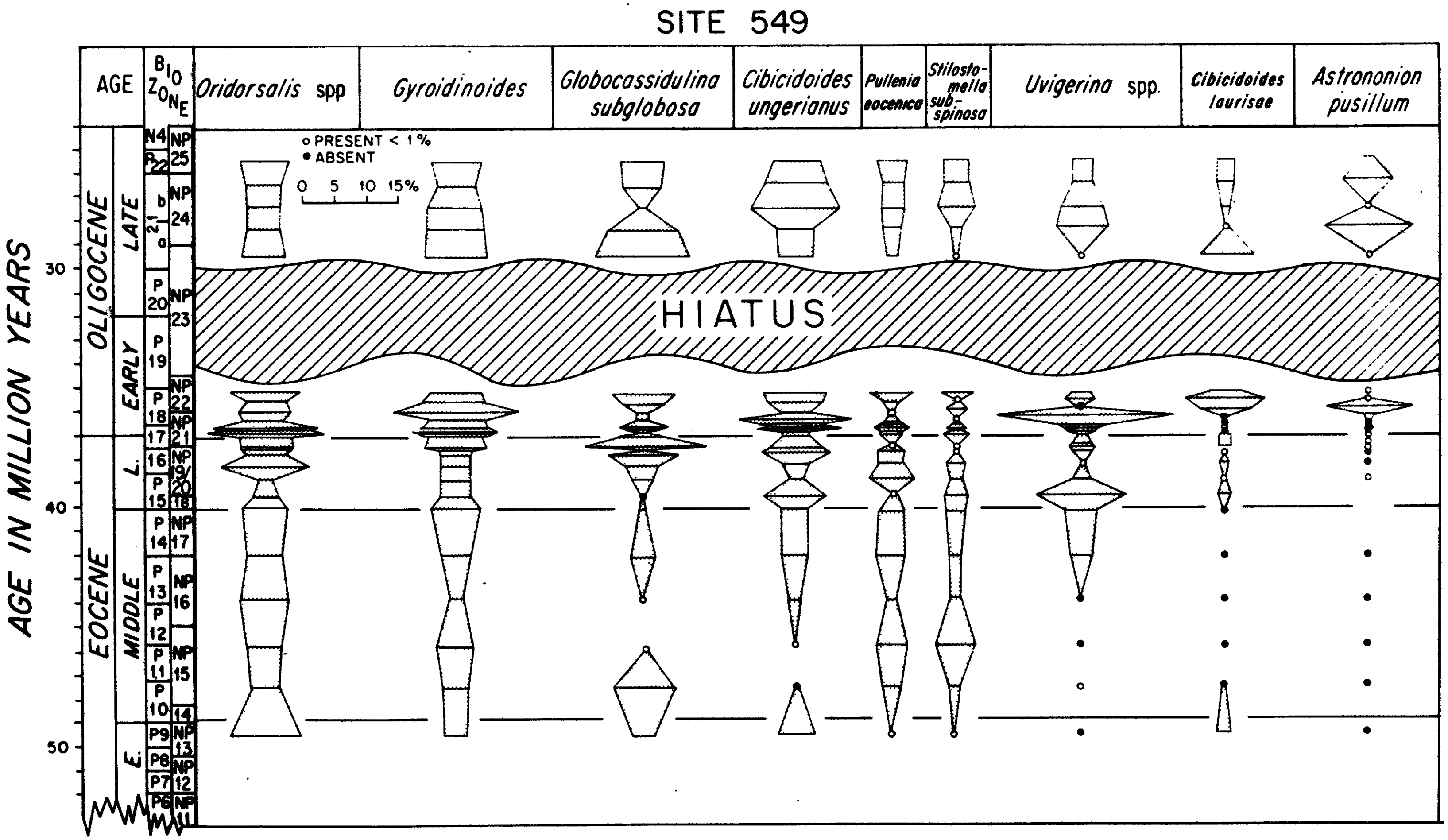
Tjalsma and Lohmann (1982) suggested that the replacement of the

Figure 8

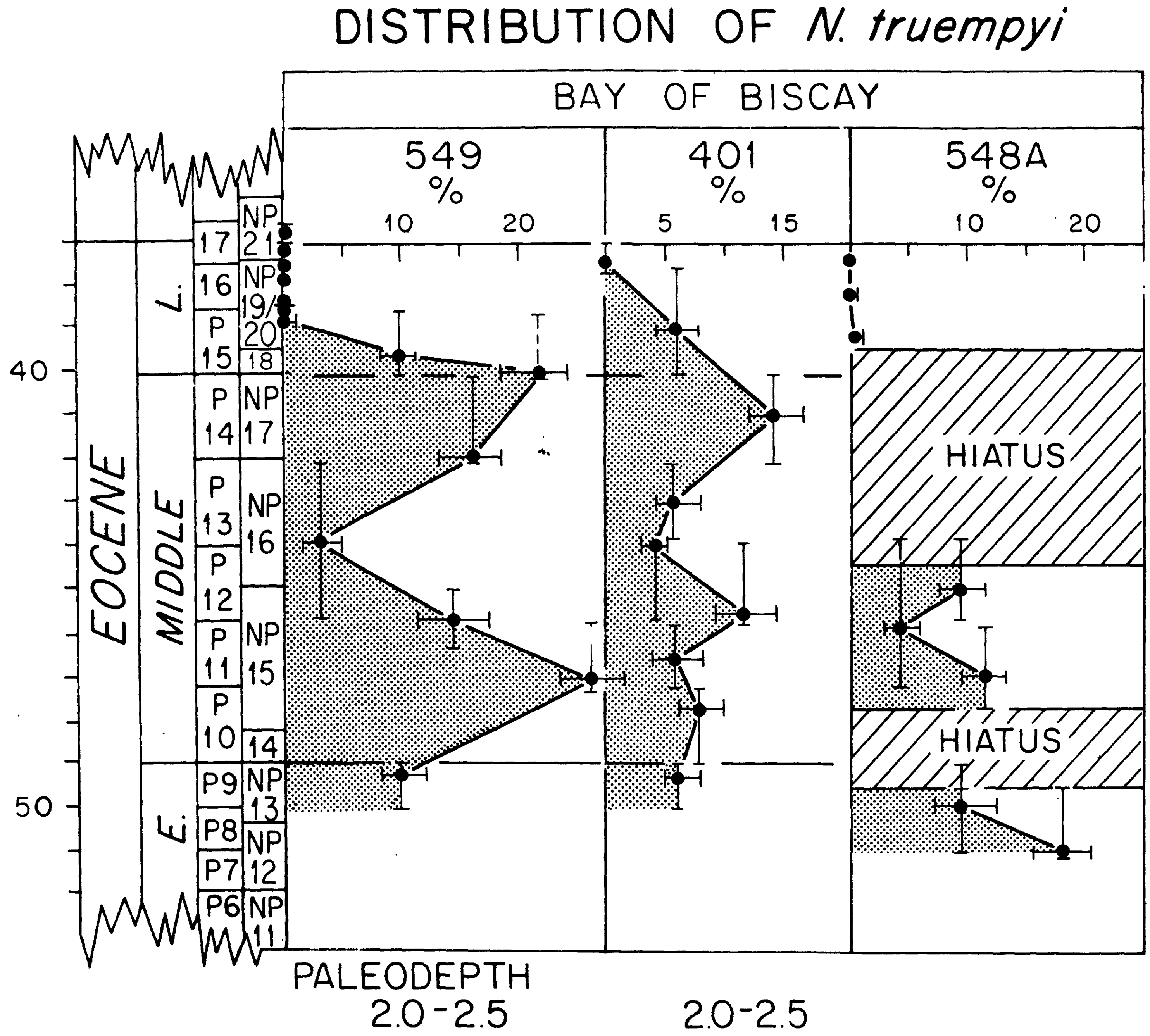


Distribution of dominant benthic foraminifers, Site 549. Shows distribution of forms more abundant in the Eocene.

Figure 9

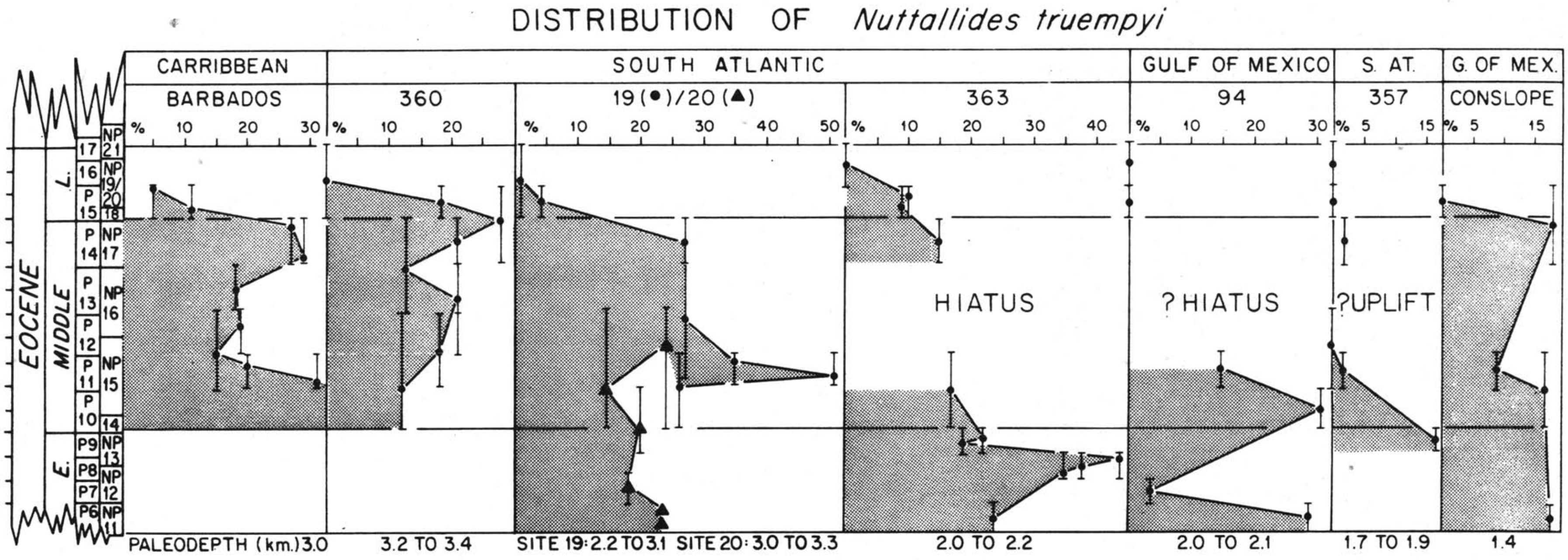


Distribution of dominant benthic foraminifers, Site 549. Shows distribution of stratigraphically long-ranging and bathymetrically wide ranging taxa that are abundant throughout the Eocene and Oligocene and forms that are more abundant in the Oligocene.



Distibution of Nuttallides truempyi in the Bay of Biscay, Sites 549, 401, and 548. Error bars on percentages indicate 80 °/o confidence interval.

Figure 11



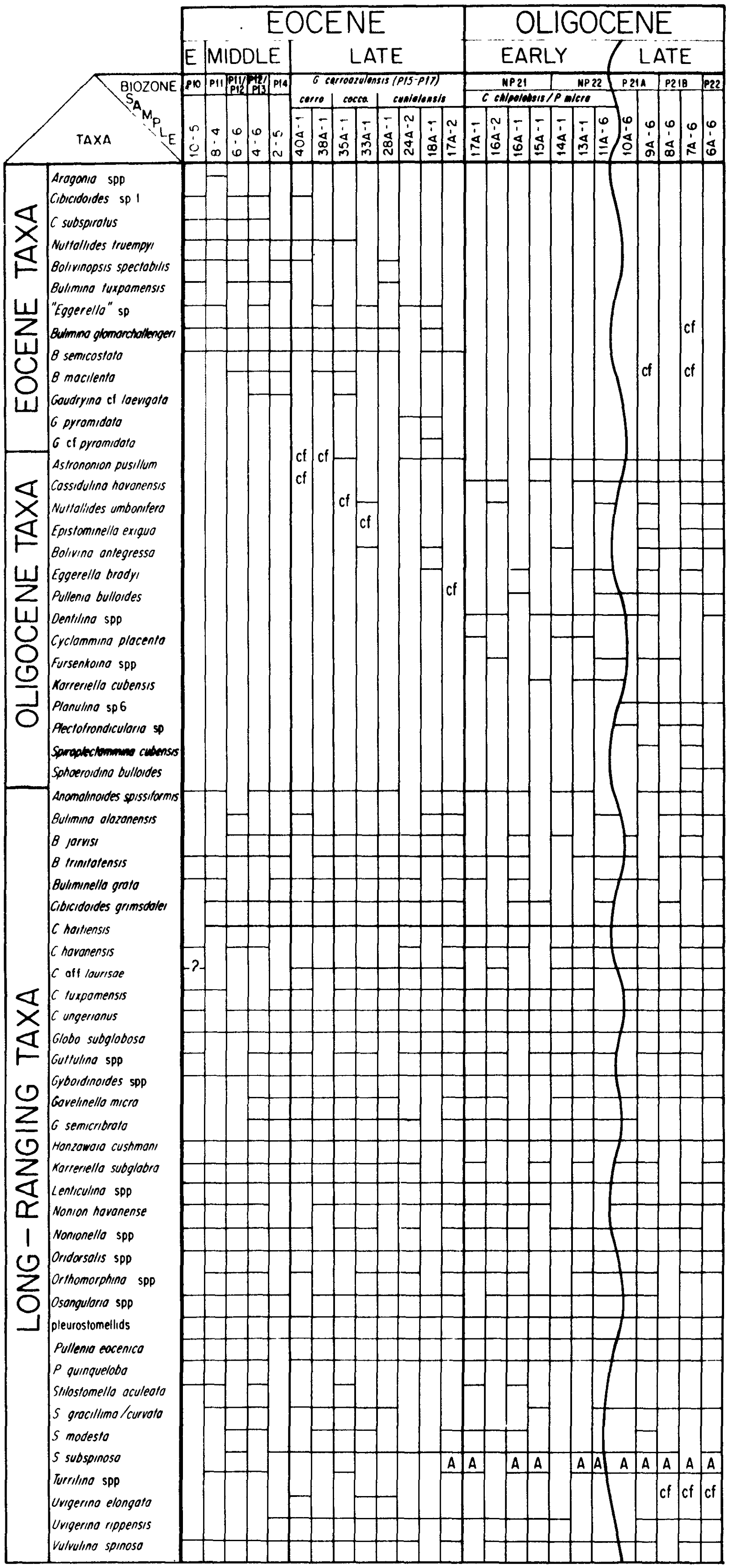
Distibution of Nuttallides truempyi in the South Atlantic, Gulf of Mexico, and Caribbean Sea. (Data after Tjalsma and Lohmann, 1982). Error bars on percentages indicate 80 °/o confidence interval.

Nuttallides truempyi assemblage was diachronous, occurring near the early/middle Eocene boundary in "shallow" sites (< 2km) such as Site 357. However, close inspection of their data shows that Site 357 is the only site in which the replacement occurs at this time (Fig. 11). In both shallower (~ 1.4 km paleodepth, Conslope borehole) and slightly deeper boreholes ~ 2.0-2.5km paleodepth, Sites 94, 363, and 19), the change occurs near the middle/late Eocene boundary (Fig. 11). The diachrony of the change at Site 357 relative to other locations may be due to uplift of that site in the Eocene (D.A. Johnson, personal communication), although this would neccessitate uplift above the upper depth limit of Nuttallides truempyi (~ 500-600m; Berggren and Aubert, 1974). Within the biostratigraphic error bars assigned here and by Tjalsma and Lohmann (1982) and excluding Site 357, the replacement of the N. truempyi assemblage is synchronous in the North Atlantic, Gulf of Mexico, Caribbean, and South Atlantic.

In the deeper locations (> 2km) in the North Atlantic (Site 549, Site 401, Barbados), there is a strong suggestion that abundance of Nuttallides truempyi displays two acmes in the Eocene. Peaks in abundance occur in the early middle and late middle Eocene, and decreases in abundance occur in the middle Eocene and late Eocene (Figs. 10,11). Although the replacement of the N. truempyi assemblage in the earliest late Eocene may be a useful biostratigraphic event, care must be taken to differentiate this decrease in abundance from the earlier decrease.

No major change in faunal abundance takes place across the Eocene/ Oligocene boundary, although the decrease in abundance of the buliminids pre-dates this boundary by < 0.5 million years. This agrees with the suggestion of Corliss (1979a, 1981), Tjalsma and Lohmann (1982), Tjalsma (1982), and Miller (in press) that the Eocene/Oligocene boundary was not a

Figure 12



Range chart, Site 549. A = Turrilina alsatica.

"catastrophe" for benthic foraminifers.

It is apparent from Figure 12 that a large number of first and last appearances occurred in the late Eocene to earliest Oligocene. *Nuttallides umbonifera*, *Epistominella* cf. *exigua*, *Astrononion pusillum* (Plate 5) appear in the late Eocene; these species constitute important faunal elements of the deep Oligocene assemblage at Site 119 (Miller, in press), Site 400A, and Site 550 (see below), and their first appearances in the late Eocene may be useful datum levels. Rarer taxa also appear in the late Eocene of Site 549: *Bolivina antegressa*, *Eggerella bradyi*, and *Uvigerina elongata*. *Cyclammina placenta*, *Karreriella cubensis*, *Dentilina* spp., *Fursenkoina* spp., and probably *Cassidulina havanesis* appear just above the level of the Eocene/Oligocene boundary. Some of these early Oligocene first appearances are local: 1) *Karreriella cubensis* appears in Zone P11; and 2) *C. havanesis* appears in Zone P11 (Tjalsma and Lohmann, 1982).

Several last appearances occur in the late Eocene. *Nuttallides truempyi*, *Cibicidoides* sp. 1, *Bolivinopsis spectabilis*, *Gaudryina laevigata*, *G pyramidata*, *G.* cf. *pyramidata*, "*Eggerella*" sp., and *Bulimina macilenta* have their last appearances in the late Eocene and may be useful datum levels. *Bulimina tuxpamensis* and B. *semicostata* have their last appearances in the late Eocene of Site 549, but elsewhere range into the Oligocene (Tjalsma and Lohmann, 1982).

Other important first appearances occur in the early to middle Eocene of Site 549: 1) *Gavelinella semicribrata* s.s. appears consistently in Zone P12 in the Bay of Biscay (Miller, in press) and elsewhere (Tjalsma and Lohmann, 1982)(Plate 7); 2) *G. micra* appears in Zone P12 at Site 401 (Miller, in press); however, elsewhere in the Atlantic the first appearance pre-dates P12 (Tjalsma and Lohmann, 1982); and 3) *Bulimina jarvisi* appears in Zone P11/P12 in Site 549, in Zone P14 in Site 401 (Miller, in press), and in Zone P15 in the South Atlantic (Tjalsma and Lohmann, 1982). The apparent diachrony of the appearance of this latter taxon, in part, may be due to differing taxonomic concepts; for example, Schnitker (1979) noted the appearance of *B. jarvisi* in Zone P6 in Site 401. Oligocene first appearances (all Zone P21) include *Plectofrondicuilaria* sp. and *Sphaeroidina bulloides*.

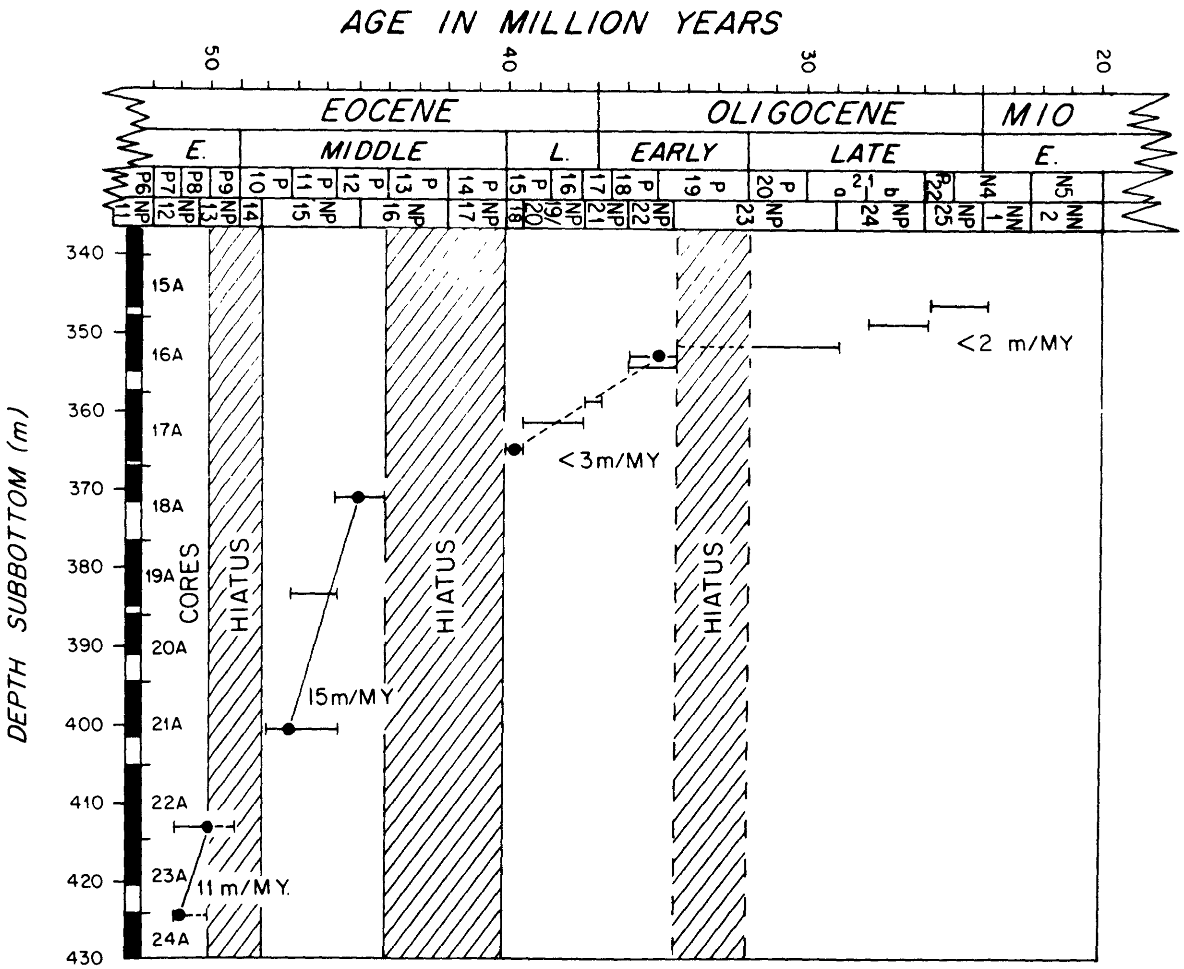
Biostratigraphy, Site 548

The Eocene through Oligocene section at Site 548 was rotary drilled. The biostratigraphic age assignments for both nannofossils (M.-P. Aubry, personal communication) and planktonic foraminifers are given in Table 4 and Figure 13. The entire middle Eocene through Oligocene section is only ~ 60m thick. Major hiatuses occur between the early Eocene and the middle Eocene (between samples 21-4, 55 and 22-6, 94), the middle Eocene and the late Eocene (between samples 17-5, 121 and 18-3, 98), and the early Oligocene and late Oligocene (between 16-3, 66 and 16-4, 69)(Table 4). A minor hiatus may occur just above the Eocene/Oligocene boundary, for the early Oligocene portion of Zone NP21 is missing, and the sedimentation rates calculated for this section are low (< 3 m/my). In addition, the sedimentation rates computed for the late Oligocene are also very low (~ 1.0 m/my), suggesting that other hiatuses may occur in this thin (~ 10m) interval.

The punctuated nature of the Eocene to Oligocene stratigraphic record recovered at Site 548 makes it less desirable for benthic foraminiferal studies. In addition, reworking appears to be more of a problem at Site 548 that at Site 549, for middle Eocene acarininids are found in the overlying upper Eocene and Oligocene sections.

Figure 13

SITE 548



Zonal age assignment and estimated average sedimentation rate Site 548. Error bars indicate the length of the biozone(s) to which the sample have been assigned (Table 4). Solid dots indicate ties for interpolating sedimentation rate and age estimates.

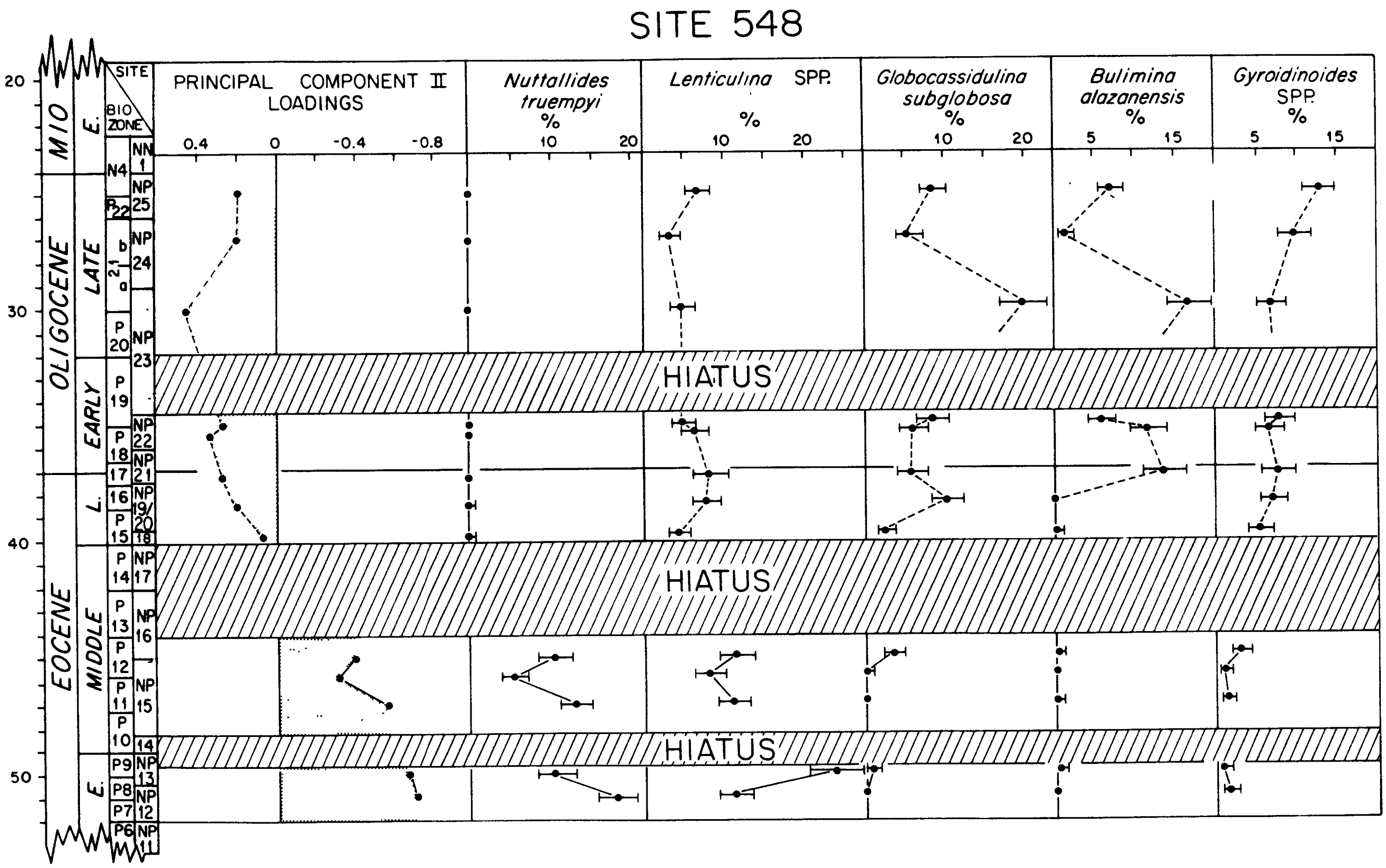
Benthic Foraminiferal Assemblages, Site 548

As in Site 549, the major Eocene to Oligocene source of faunal variation in Site 548 is the replacement of the *Nuttallides truempyi* assemblage. We made a Q-mode principal component analysis of 13 samples from Site 548. Principal component 2 explains 38.2 $^{o}/o$ of the faunal variation (for the 48 taxa used by Tjalsma and Lohmann, 1982 and Miller, 1982; Table 2) in the Eocene to Oligocene section of Site 548. The early and middle Eocene is dominated by *N. truempyi* and *Lenticulina* spp. (Figs. 10,14). The *N. truempyi*-*Lenticulina* assemblage is represented by high negative loadings on principal component 2 (Fig. 14). High positive loadings on principal component 2 represent the late Eocene through Oligocene assemblage dominated by *Gyroidinoides* spp., *Bulimina alazanensis*, and *Globocassidulina subglobosa* (Fig. 14). Although the incomplete record at Site 548 prevents firmly dating the transition from the *N. truempyi*-*Lenticulina* assemblage to the *Gyroidinoides*-*B. alazanensis*-*G. subglobosa* assemblage, loadings on principal component 2 show an increase in the late Eocene (Fig. 14). This suggests that the major faunal change at Site 548 begins just above the middle/late Eocene boundary, and is therefore coincident with the faunal replacement of the *N. truempyi*-dominated assemblage at Site 549 (Fig. 10).

Biostratigraphy and Benthic Foraminifera, Site 550

The middle Eocene through Oligocene section (all rotary drilled) recovered at the deep Site 550 (4432m water depth) occurs between ~ 306 and 320m subbottom (Fig. 15). Examination of this section reveals it to be very disturbed. Planktonic foraminiferal assemblages show evidence of

Figure 14



Benthic foraminiferal assemblage composition, Site 548. Principal component analysis was performed on 13 samples from this site. Error bars indicate 80 o/o confidence interval.

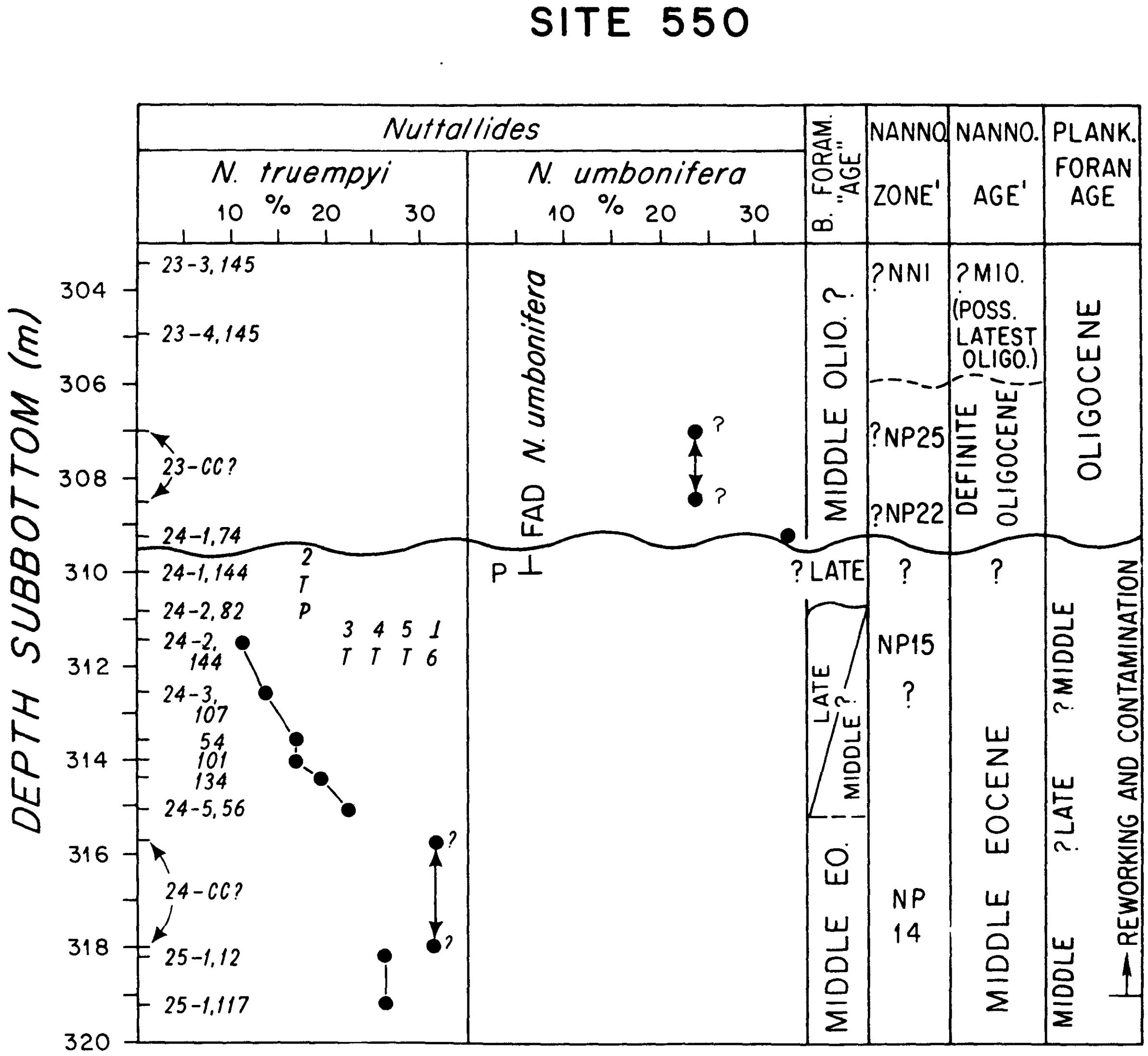
contamination by Plio-Pleistocene and Cretaceous forms. In addition, planktonic foraminifers, nannoplankton, and benthic foraminifers (Plate 7, fig. 6-12) all show evidence of strong dissolution.

An unconformity probably occurs at ~ 309.5m subbottom. The hiatus separates the Oligocene from the Eocene; biostratigraphic resolution of the age of sediments above and below the disconformity is lacking, although it appears to separate middle Oligocene from upper Eocene strata (see below; Figs. 15,16). Another hiatus, separating the upper Eocene from middle Eocene strata may occur somewhere in the interval between 311 and 316m subbottom.

Plantkonic age assignments for the Eocene to Oligocene section at Site 550 are confused. Middle Eocene planktonic foraminifers occur in samples 25-1, 117 through 24-5, 56, while sample 24-4, 134 contains abundant *Globorotalia opima nana* indicative of post-middle Eocene age. However, the overlying samples (24-4, 101 through 24-1, 144) contain middle Eocene planktonic foraminifers (*Acarinina* cf. *aspensis*, *A*. *densa*, *A*. cf. *pentacamerata*, *A*. cf. *broedermani*). Based upon nannofossils, the entire interval between samples 25-1, 117 and 24-2, 144 is assignable to the early middle Eocene (M.-P. Aubry, personal communication)(Fig. 15).

Benthic foraminifers may be used in sorting out the confused stratigraphy of Site 550 (Fig. 15). *Nuttallides truempyi* dominates the benthic foraminiferal assemblages between 311 and 320m subbottom; a decrease in abundance in *N*. *truempyi* occurs between ~ 316m and 311m subbottom. This decrease in abundance can be interpreted in at least three ways: 1) it may represent the transition from the *N*. *truempyi*-dominated assemblage to the *Oridorsalis*-*Gyroidinoides*-*G*. *subglobosa*-*C*. *ungerianus* assemblage that occurs near the middle/late Eocene boundary in Site 549; 2) it may represent the decrease in abundance of *N*. *truempyi* that occurs in the early middle Eocene in

Figure 15



Stratigraphy of Site 550. [1] Data after M.-P. Aubry, personal communication; [2] LAD Abyssammina spp.; [3] LAD Clinapertina spp.; [4] LAD Gaudryina pyramidata; [5] LAD Alabamina dissonata; [6] FAD Gavelinella semicribrata. All benthic foraminiferal datums after Tjalsma and Lohmann (1982).

Site 549; or 3) it may represent a mixing interval, due to bioturbation or core disturbance. Several last appearances of benthic foraminiferal taxa that occur elsewhere near the middle/late Eocene boundary (Fig. 15) occur in sample 24-2, 144, suggesting that this decrease in abundance represents the middle/late Eocene boundary. However, the nannofossil evidence suggests that this decrease in abundance represents the early middle Eocene decrease, and that the last appearances of benthic foraminiferal taxa are truncated at this site. The poor record at this site precludes further resolution.

The overlap of Nuttallides umbonifera (FAD late Eocene) and Abyssammina (LAD late Eocene; Tjalsma and Lohmann, 1982) suggests that sample 24-1, 144 is late Eocene. However, considering the disturbance noted here, this remains tentative.

Nuttallides umbonifera occurs in high (> 20 $^{0}/_{0}$) abundance in samples 24-1, 74 and 23-CC. An acme of this species was noted at the nearby Site 119 and Site 400A in the middle Oligocene (Miller, in press), suggesting that these samples are middle Oligocene. This is consistent with the nannofossil and planktonic foraminiferal evidence.

The Eocene/Oligocene Boundary and Benthic Foraminifers

Our data from Site 549 definitively show the nature and timing of the major Eocene to Oligocene benthic foraminiferal assemblage changes at intermediate depths (~ 2km) in the North Atlantic. In both Sites 549 and 548, the major faunal change is the replacement of the Nuttallides truempyi assemblage just above the middle/late Eocene boundary; this event can be shown to have occurred throughout the North Atlantic, South Atlantic, Caribbean, and Gulf of Mexico at this time. Other faunal changes noted in Site 549 include

1) a series of extinctions and local last appearances of taxa in the late Eocene; 2) a series of first appearances (some of which are local) in the late Eocene to earliest Oligocene; and 3) the replacement of a buliminid assemblage just below the Eocene/Oligocene boundary.

The most dramatic faunal change occurs not at the shallow and intermediate depths but at the deeper (> 3km paleodepth) sites (119, 400A, 550)(Miller, in press). Dominant Nuttallides truempyi, Clinapertina spp., and Abyssammina spp. (together with Alabamina dissonata, Aragonia spp., and other elements of the endemic deep-Eocene assemblages) become extinct between the middle Eocene and Oligocene; they are replaced by abundant wide- and long-ranging taxa (Miller, in press).

The debate over "terminal Eocene" benthic foraminiferal changes is somewhat a semantic problem. No "crisis" (Kennett and Shackleton, 1976) or "peak in faunal turnover" (Schnitker, 1979) is associated with the level of the Eocene/Oligocene boundary; however, major benthic foraminiferal changes occurred between the middle Eocene and early Oligocene (Corliss, 1979, 1981; Tjalsma, 1982; Tjalsma and Lohmann, 1982; Miller, in press; this paper). The data presented here suggest that faunal changes occurred at intermediate depths throughout the late Eocene to earliest Oligocene. Certainly the material studied (depending upon recovery, preservation, geographic location, paleobathymetry) and the approach in analysis (viz. size-fraction studied, studying ranges versus abundances, R-mode versus Q-mode principal component, factor, or cluster analyses, plotting against time versus depth) can affect the perceptions and conclusions drawn as to the timing of the faunal change. Nevertheless, Corliss (1979; 1981), Tjalsma (1982), Tjalsma and Lohmann (1982), Miller (in press) and this study are in agreement that a series of faunal changes occurred in benthic foraminifers between the middle Eocene and early Oligocene, although the timing of the changes in the deepest sites (> 3km paleodepths) is unconstrained in all of these studies.

The relationship of the benthic foraminiferal isotopic and assemblage changes is well defined in Site 549. The faunal changes begin ~ 1 million years prior to the 1 ^{o}/oo oxygen isotopic enrichment, which occurs between ~ 38.5 and 36.5 Ma (late Eocene to earliest Oligocene). However, the decrease in abundance of Nuttallides truempyi (Fig. 7,8,10) occurs at the level of the first major increase in $\delta^{13}C$ (~ 0.5 ^{o}/oo; ~ 38-40Ma; Figs. 3,6). Since the $\delta^{13}C$ records of Site 292 (Keigwin, 1980) and Site 549 do not covary, the carbon isotopic record at Site 549 probably does not reflect global changes in the late Eocene. We interpret the ^{13}C enrichment as reflecting changing water-mass characteristics, from old, corrosive bottom water to younger, less corrosive bottom water in the early late Eocene.

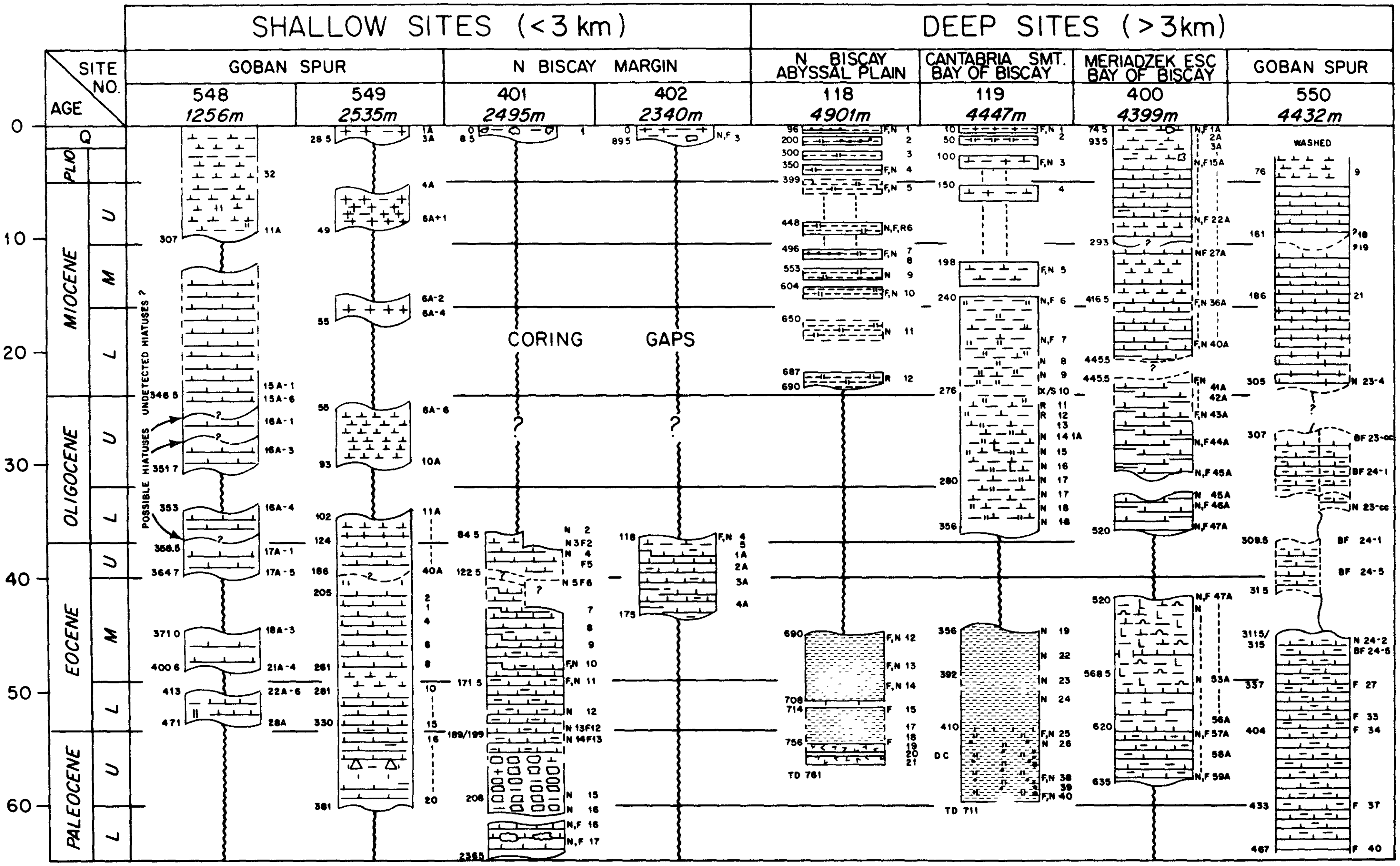
Benthic foraminiferal $\delta^{13}C$ values subsequently decrease in the late Eocene, while $\delta^{18}O$ values increase ~ 0.3 ^{o}/oo. These isotopic fluctuations occur coevally with fluctuations in abundance (Fig. 8,9) and with many first and last appearances of taxa (Fig. 12), some of which are local. Following the sharp increase in both $\delta^{18}O$ and $\delta^{13}C$ just above the Eocene/Oligocene boundary, the benthic foraminiferal assemblages are dominated by taxa that may be interpreted as tolerant of environmental changes (e.g. the bathymetrically wide-ranging and stratigraphically long-ranging Globocassidulina subglobosa, Gyroidinoides spp., Cibicidoides ungerianus, Oridorsalis umbonatus, Pullenia eocenica, Stilostomella subspinosa; Fig. 9). In this we concur with Corliss (1981) who suggested that Late Paleogene benthic foraminifers, in general, may have had wide environmental tolerances. This is further illustrated by the many taxa that range throughout the Eocene to Oligocene section (Fig. 12) despite changes in abyssal circulation inferred from the isotopic, litho-, and seismic stratigraphic records. For this reason it may be that the greatest isotopic changes, the earliest Oligocene ^{18}O and ^{13}C enrichments, is not associated with the greatest faunal abundance change, viz. the replacement of the _Nuttallides truempyi_ assemblage.

With the resolution provided by Site 549, we can compare the changes in faunal isotopic and assemblage composition with the circulation model of Miller and Tucholke (in press). The benthic foraminiferal and isotopic evidence point toward the development of abyssal circulation over a several million year period (~ 40-36.5 Ma). The change from sluggish to vigorous, intense bottom water circulation can be dated from the seismic record in the Rockall region as occurring in the late Eocene to earliest Oligocene (Miller and Tucholke, in press). In general, the circulation change associated with reflector R4 is synchronous with the late Eocene to earliest Oligocene isotopic and assemblage changes. Roberts (1975) and Miller and Tucholke (in press) suggested that reflector R4 is latest Eocene to earliest Oligocene (~ 36-38 Ma) and that the circulation event associated with this horizon was a geologically rapid event, although uncertainties in the biostratigraphy and the correlation of reflector R4 into the various boreholes prevent absolutely establishing this. If this age assignment is correct, reflector R4 and the associated circulation change correlates with the major ^{18}O enrichments, but post-dates the initial faunal changes.

Our suggestion that the faunal and isotopic changes in the North Atlantic may be attributed to the influx of cold, northern sources of bottom water may be debated in view of the isotopic and stratigraphic records from other basins. The faunal and isotopic changes discussed above have been noted from several different ocean basins over a wide range of paleodepths. However, the intepretation that the North Atlantic had an Oligocene northern source of bottom water is derived from the seismic stratigraphic record (Miller and Tucholke, in press), and does not rely upon faunal and isotopic data. Reflector R4, which is associated with the intiation of more vigorous bottom water circulation, is found in bathyal (ie. the Hatton-Rockall Basin) through abyssal locations, indicating influence of northern sources of bottom water over a wide range of depths in the North Atlantic. Southern bottom water influences on the northeast North Atlantic were probably minimal due to the restriction of flow from the south by the Azores-Biscay Rise (Fig. 1), Azores-Gibraltar Rise, and Madeira-Tore Rise (Miller and Tucholke, in press). Still, evidence of widespread unconformities in the Southern and Pacific Oceans (Moore et al., 1978) supports the idea that these basins had a southern (Antarctic) source of Oligocene bottom water, as suggested by Kennett and Shackleton (1976). We speculate that both northern sources and southern sources of cold bottom water began near the end of the Eocene in response to climatic cooling; this is evinced in a synchronous increase in benthic foraminiferal $\delta^{18}O$.

Distribution of Hiatuses in the Bay of Biscay

A major middle Oligocene hiatus occurs in both Site 549 and Site 548. A similar hiatus may occur at Sites 401 and 402 on the Armorican margin (Fig. 17). However, coring gaps prevent constraining the timing in Site 401 and Site 402, and the hiatus there may correlate with a hiatus associated with the Eocene/Oligocene boundary in Sites 119, 400A, 550, and possibly 548. A middle Oligocene hiatus probably occurs in the deep Site 400A; although the Oligocene of the deep Site 119 is apparently continuous, the biostratigraphic control is poor enough that a minor hiatus may lie undetected (Miller, in press).


SHALLOW SITES (<3 km)
DEEP SITES (>3km)
SITE NO.
AGE
GOBAN SPUR
N BISCAY MARGIN
N BISCAY ABYSSAL PLAIN
CANTABRIA SMT. BAY OF BISCAY
MERIADZEK ESC BAY OF BISCAY
GOBAN SPUR
548 1256m
549 2535m
401 2495m
402 2340m
118 4901m
119 4447m
400 4399m
550 4432m
Q
PLIO
MIOCENE
OLIGOCENE
EOCENE
PALEOCENE
U
M
L
0
10
20
30
40
50
60
CORING GAPS
POSSIBLE HIATUSES
UNDETECTED HIATUSES ?
WASHED
TD 761
TD 711

Age versus lithofacies for Tertiary sediments recovered at DSDP sites in the Bay of Biscay. Site locations in Figure 1. Lithologic symbols follow standard DSDP format. To left of column are subbottom depth in meters (T.D. = total depth); to right are indicated core number and fossil group upon which age assignment is based. (N = nannoplankton; F = planktonic foraminifers; BF = benthic foraminifers; R = radiolaria). Heavy wavy line indicates hiatuses; dashed lines indicate inferred continuous sedimentation.

The middle Oligocene hiatus is found predominately in "shallow" sites (Sites 549, 401/402; 2-2.5km paleodepth; Site 548, ~ 1.0km paleodepth)(Fig. 17), although the unconformity can probably be traced to deep sites. This hiatus correlates with a major middle Oligocene (basal Zone P21A; 29 Ma) sea-level lowering indicated by Vail et al. (1977), but it is not clear whether the two are related. The hiatuses cannot have resulted from subaerial exposure, for this would require kilometers of uplift (or sea-level fall) and immediate subsidence (or sea- level rise). Vail et al. (1977) suggested that deposition during sea-level lowstands (e.g. the middle Oligocene) results in low-stand deposits in deep-marine basins. Thus, deposition might be expected at these intermediate depths during low stands, and not the development of hiatuses (Tucholke, 1981). Alternate possible causes of the middle Oligocene hiatus include: 1) erosion by increased canyon cutting that occurs during lowstands (Vail et al., 1977); 2) interruption of sedimentation by tectonism associated with the Pyrenian orogeny (such tectonism is widespread throughout the northeastern Atlantic, including the Bay of Biscay; Boillot et al., 1979; Montadert et al., 1979); and 3) erosion by bottom currents.

In the deep Bay of Biscay (Sites 118, 119, 400A, 550; all paleodepths > 3km), the major hiatus appears to be associated with the Eocene/Oligocene boundary (Fig. 17) and thus related to the abyssal circulation event noted by Miller and Tucholke (in press). Again, this hiatus may be partially the result of Eocene-Oligocene tectonism. However, Miller and Tucholke (in press) noted the development of a similar latest Eocene to early Oligocene hiatus in the Rockall Plateau region, correlated this hiatus with reflector R4, and argued that the hiatus resulted from a major circulation change.

The shallower, middle Oligocene event is diachronous with the deeper event. If the record at 1-2km (paleodepth) reflects eustatic sea level, then there is no causal link of eustacy with the current-controlled event. This agrees with the suggestion of Tucholke (1981) that there may be no consitent causal link between sea level change and the development of unconformities in the deep-sea. However, Olsson et al. (1980) suggested that sea-level curve of Vail et al. (1977) is miscorrelated and that the major sea-level lowering occurred somewhere in the late Eocene to early Oligocene. In view of this, we present an alternative scenario that could explain the development of Late Paleogene hiatuses. 1) A latest Eocene to earliest Oligocene hiatus developed in the deep North Atlantic in response to a bottom-water event. 2) This circulation event correlates with a major eustatic sea-level lowering which resulted in the development of a latest Eocene to early Oligocene hiatus on continental shelves (Olsson et al., 1980). 3) At intermediate depths (Sites 549, 548) deposition was relatively continuous across the Eocene/Oligocene boundary due to increased sediment input to the basin during a lowstand. 4) At intermediate depths, middle Oligocene deposition was discontinuous due to sediment starvation during a rise in sea level (Olsson et al., 1980), although this hiatus may be in part due an erosional pulse of bottom water associated with reflector R3 in the Rockall region.

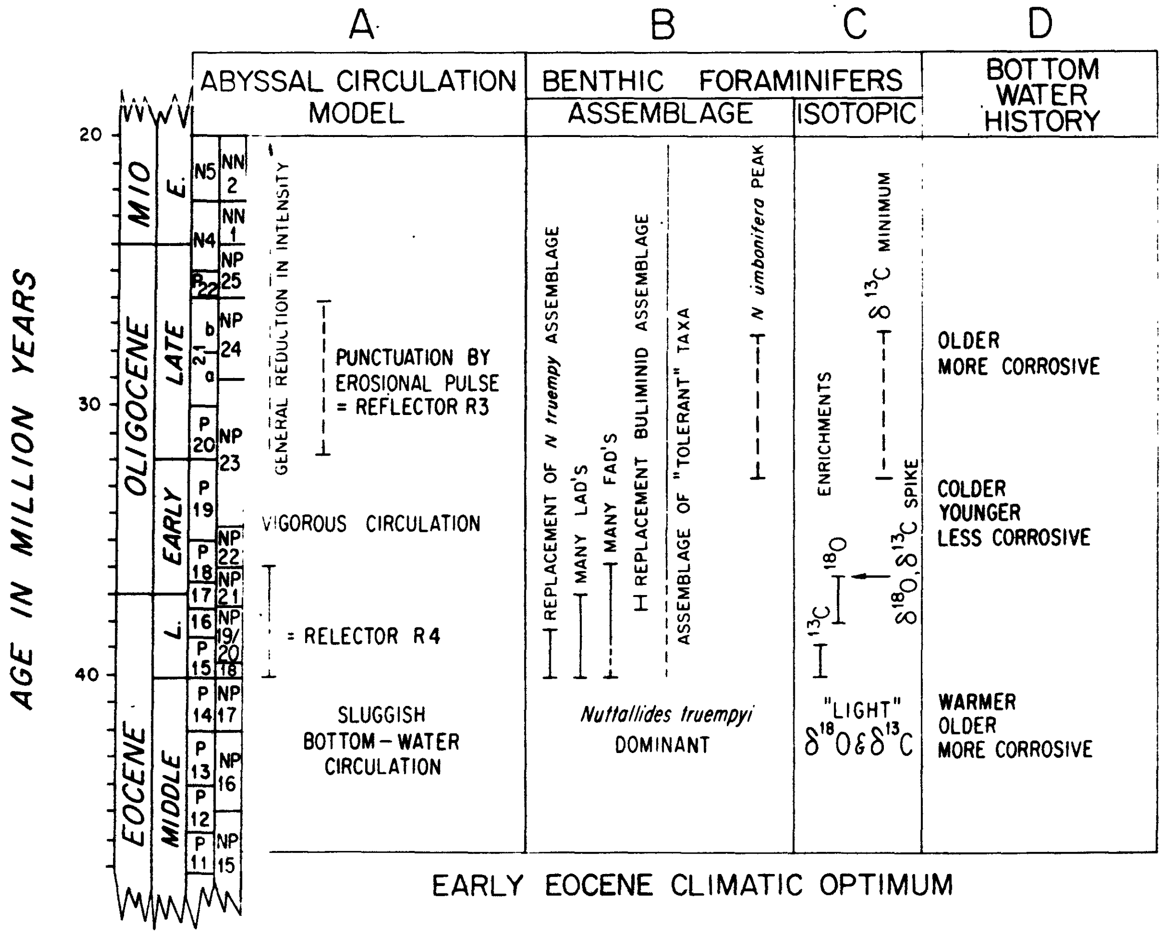
Paleoceanographic Synthesis

Site 549 is the first North Atlantic DSDP site with continuous recovery of well-dated upper Eocene through lower Oligocene sediments, apparently unaltered by diagenetic or dissolution effects. This allows: 1)a detailed comparison of the benthic foraminiferal assemblage and isotopic record; and 2) the interpretation of these records in view of the development of North Atlantic abyssal circulation. Although the middle Oligocene record was poorly recovered by Leg 80, we have used evidence for middle Oligocene abyssal circulation changes from Site 119 and Site 400A (for a full discussion of this data see Miller and Curry, 1982; Miller, in press) and from the abyssal circulation model of Miller and Tucholke (in press).

Figure 17 shows a summary of development of abyssal circulation, and demonstrates that the evolution of the Oligocene ocean from the Eocene ocean was not a rapid event as suggested by Kennett and Shackleton (1976). The major faunal and isotopic events span ~ 4 million years. Nevertheless, there is some evidence that these "gradual" (Corliss, 1979, 1981) or "sequential" (Tjalsma and Lohmann, 1982) changes were punctuated by a geologically rapid event as suggested by Kennett and Shackleton (1976). Although the overall ^{18}O enrichment occurs in a ~ 2 million year interval (assuming constant sedimentation rates, Figs.2, 3), the early Oligocene 0.7 $^{o}/oo$ ^{18}O and 0.6 $^{o}/oo$ ^{13}C enrichments occur in an interval of < 0.5 million years. This agrees well with the estimates for the duration of the $\delta^{18}O$ change previously noted by Kennett and Shackleton (1976) and Keigwin (1980).

The major change in abyssal circulation inferred from the seismic record apparently post-dates the first faunal and isotopic changes. We suggest this scenario for the data in Figure 17: 1) a series of hydrographic changes, that may represent the first influence of Arctic and/or Norwegian-Greenland Sea water on the deep North Atlantic, began by the late Eocene; 2) these changes culminated in a major temperature drop, decrease in age of bottom water, and a strong increase in intensity of circulation near the Eocene/Oligocene boundary; and 3) subsequently in the middle Oligocene, the general intensity of abyssal circulation decreased and bottom waters became slightly older.

Figure 17


A
B
C
D
ABYSSAL CIRCULATION MODEL
BENTHIC FORAMINIFERS
ASSEMBLAGE
ISOTOPIC
BOTTOM WATER HISTORY
AGE IN MILLION YEARS
20
30
40
MIO
OLIGOCENE
EOCENE
E.
LATE
EARLY
L.
MIDDLE
N5
N4
P 22
21 b
a
P 20
P 19
P 18
17
16
P 15
P 14
P 13
P 12
P 11
NN 2
NN 1
NP 25
NP 24
NP 23
NP 22
NP 21
NP 19/20
18
NP 17
NP 16
NP 15
GENERAL REDUCTION IN INTENSITY
PUNCTUATION BY EROSIONAL PULSE = REFLECTOR R3
VIGOROUS CIRCULATION
= RELECTOR R4
SLUGGISH BOTTOM-WATER CIRCULATION
REPLACEMENT OF N truempy ASSEMBLAGE
MANY LAD'S
MANY FAD'S
REPLACEMENT BULIMINID ASSEMBLAGE
ASSEMBLAGE OF "TOLERANT" TAXA
N umbonifera PEAK
Nuttallides truempyi DOMINANT
ENRICHMENTS
^{13}C
^{18}O
$\delta^{13}C$ MINIMUM
$\delta^{18}O, \delta^{13}C$ SPIKE
"LIGHT" $\delta^{18}O$ & $\delta^{13}C$
OLDER MORE CORROSIVE
COLDER YOUNGER LESS CORROSIVE
WARMER OLDER MORE CORROSIVE
EARLY EOCENE CLIMATIC OPTIMUM

Summary of paleoceanographic events. Column A indicates abyssal circulation events inferred from Miller and Tucholke's (in press) seismic straigraphic studies, column B indicates benthic foraminiferal assemblages and timing of changes, column C indicates timing of major benthic foraminiferal isotpic events, and column D indicates bottom water history inferred from the data in columns B and C. The middle Oligocene faunal and isotopic events are taken from Miller and Curry (1982) and Miller (in press); the early Eocene climatic optimum is taken from Haq (1981).

CONCLUSIONS

1. Major benthic foraminiferal changes occur in Site 549 in the late Eocene to early Oligocene:

    a. A *Nuttallides truempyi*-dominated assemblage is replaced in the early late Eocene (~ 38.5-40 Ma).

    b. The succeeding buliminid assemblage is replaced by an assemblage dominated by stratigraphically long-ranging and bathymetrically wide-ranging taxa just below the Eocene/Oligocene boundary (~ 37.5 Ma).

    c. In addition to these abundance changes, a series of last appearances occurs in the late Eocene, and a series of first appearances occurs in the late Eocene to earliest Oligocene.

2. Isotopic results from Site 549 show:

    a. A $\delta^{13}C$ increase of ~ 0.6 $^{o}/oo$ correlates with the replacement of *N. truempyi*.

    b. A major $\delta^{18}O$ increase begins at ~ 38 Ma (late Eocene), culminating in a rapid (< 0.5 my) increase in $\delta^{18}O$ just above the Eocene/Oligocene boundary (~ 36.5 Ma). A rapid major $\delta^{13}C$ correlates with this latter $\delta^{18}O$ increase.

3. Faunal results from Site 548 show that a *N. truempyi-Lenticulina* dominated assemblage is also replaced between the middle Eocene and the late Eocene by an assemblage dominated by a *Gyroidinoides-Bulimina alazanensis-Globocassidulina subglobosa* assemblage. The timing of this faunal change is more obscure due to numerous hiatuses present in this site.

4. The Leg 80 data is combined with faunal and isotopic data from Legs 12 and 48 and with an abyssal circulation history derived from seismic stratigraphic studies of the Labrador Sea and Rockall region in the following scenario:

   a. Old, warm, corrosive, and sluggishly circulating bottom water was replaced by younger, colder, and more vigorously circulating bottom water which had a northern source (Arctic and/or Norwegian-Greenland Sea). This developed throughout the late Eocene (40-37 Ma), but was punctuated by a rapid event just above the Eocene/Oligocene boundary (~ 36.5 Ma).

   b. During the middle Oligocene, bottom water circulation was reduced, and the age and corrosiveness of bottom water increased.

5. Based upon the sediment distribution in the Bay of Biscay and Rockall region, we present a scenario for the development of Late Paleogene hiatuses:

   a. A latest Eocene to earliest Oligocene hiatus developed in the deep North Atlantic in response to a bottom-water event.

   b. At intermediate depths, middle Oligocene deposition was discontinuous due to sediment starvation during a rise in sea level, although this hiatus may be in part due to an erosional pulse of bottom water associated with reflector R3 in the Rockall region.

Although the scenarios for the development of abyssal circulation and hiatuses are far from being completely developed or correct in detail, they account for the lithostratigraphic, seismic stratigraphic, biostratigraphic, and isotopic data that must be considered in paleoceanographic studies.

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BENTHIC FORAMINIFERAL CENSUS DATA -- SITE 549 continued

SAMPLE NUMBER	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25
AGE IN M.Y.	25.5	26.5	27.4	28.4	29.5	35.3	35.7	36.0	36.4	36.7	36.8	36.9	37.0	37.1	37.5	37.8	38.3	38.7	39.5	40.0	42.0	44.0	45.9	47.5	49.5
TOTAL NUMBER COUNTED	416	552	251	572	430	288	344	259	308	324	302	283	329	320	350	352	473	389	383	314	293	289	202	374	350
PERCENT UNACCOUNTED (1)	22.8	44.6	24.7	30.9	18.6	17.0	21.5	28.6	18.2	25.0	20.2	18.4	30.1	8.8	9.1	7.4	19.0	15.9	8.9	10.2	5.1	17.6	15.3	8.6	15.7
------------------------------ SPECIES INCLUDED IN TJALSMA AND LOHMANN'S (1982) CENSUS --------------------------------																									
Anomalina spissiformis	0.0	0.2	10.0	1.2	0.0	0.0	0.6	3.1	0.0	1.9	0.7	0.7	1.8	10.9	0.0	1.7	1.1	2.8	0.8	1.6	2.0	1.0	0.0	2.7	0.9
Gavelinells capitata/ semicribrata	0.7	0.9	2.4	0.9	0.5	0.3	0.3	1.2	0.0	0.0	0.3	0.7	0.6	0.0	0.0	1.1	1.7	1.0	0.5	1.0	1.7	1.7	0.0	0.0	0.6
Gavelinella micra	0.0	0.0	0.0	0.0	0.9	0.7	2.9	1.9	0.3	1.2	0.7	0.4	1.2	0.6	0.9	0.6	0.4	0.8	0.8	0.0	0.0	1.7	0.0	0.0	0.0
Nuttallides truempyi	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.5	9.9	22.3	16.4	2.8	14.4	26.7	10.3
Oridorsalis umbonatus	8.4	5.6	4.8	4.7	7.4	11.8	5.8	8.1	6.2	17.3	12.6	18.7	8.5	8.8	11.4	6.5	13.7	2.3	4.4	7.0	5.5	7.6	5.9	4.8	11.4
Osangularia spp.	0.0	0.0	0.0	0.2	0.0	0.0	2.3	1.2	1.6	0.6	1.0	0.0	0.9	3.1	44.9	0.6	3.2	2.8	1.0	0.0	0.0	5.9	5.0	2.1	7.1
Alabamina dissonata	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Hanzawaia cushmani	1.0	0.2	2.4	0.5	0.2	2.8	4.9	1.2	0.6	0.9	1.0	0.7	2.1	0.3	1.1	1.7	2.1	1.5	1.6	3.5	3.4	5.5	3.5	0.0	0.3
Globocassidulina subglobosa	7.0	5.6	1.2	11.4	15.1	10.4	7.8	1.9	1.9	8.3	5.6	1.4	7.9	20.6	1.4	10.8	5.7	2.8	0.0	1.0	4.1	0.3	0.5	9.1	3.4
Cibicidoides grimsdalei	0.0	0.0	0.4	0.0	0.0	0.3	0.0	0.0	0.3	0.6	0.0	0.0	0.0	0.0	1.1	0.6	3.4	3.1	0.0	0.0	0.3	0.7	0.0	0.0	0.0
C. haitiensis	1.7	1.4	0.4	0.3	2.8	1.0	2.3	2.3	1.9	1.9	0.7	2.8	0.6	0.3	0.9	0.9	3.0	1.0	1.6	2.5	3.8	4.2	5.0	0.8	0.0
C. havanensis	0.2	0.4	0.0	0.5	0.2	0.7	0.3	0.8	0.0	0.3	1.3	0.7	0.9	0.0	0.3	0.0	0.0	0.0	0.0	0.0	0.0	1.7	1.0	0.0	0.3
Cibicicoides ungerianus	8.4	9.6	14.7	6.8	6.3	10.4	9.3	3.9	17.9	10.5	7.6	13.4	3.6	4.1	8.0	10.8	3.6	2.3	10.2	4.1	3.8	2.1	0.5	0.0	7.7
C. tuxpamensis	3.4	0.9	0.4	1.7	0.2	0.0	0.6	0.4	0.3	2.5	2.6	1.1	4.0	1.9	1.4	0.6	3.4	2.1	0.0	0.6	6.8	0.0	0.0	2.1	1.1
C. cf. pseudoperlucidus	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
C. subspiratus	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.7	2.0	3.2	3.7
Cibicidoides laurisae	1.4	2.2	1.2	0.7	8.4	5.9	13.1	8.9	0.0	0.0	0.3	0.0	2.4	1.6	0.9	0.3	1.7	0.8	1.6	0.3	0.0	0.0	0.0	0.0	1.7
------------------------------ SPECIES NOT INCLUDED IN TJALSMA AND LOHMANN'S CENSUS --------------------------------																									
Astrononion pusillum	1.2	7.8	0.3	14.0	0.2	0.3	0.9	14.7	1.0	0.0	1.0	0.0	0.6	0.6	0.3	0.0	0.0	0.5	sp.	sp.	0.0	0.0	0.0	0.0	0.0
Cibicidoides sp. 1	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	6.9	6.4	0.0	7.1
Nuttallides umbonifera	1.0	0.7	0.0	0.2	0.0	0.3	0.0	0.0	0.0	0.0	0.7	0.0	0.0	0.0	0.0	0.0	0.8	cf.	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Epistominella exigua	0.2	2.0	0.0	1.7	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Bolivina antegressa	0.5	2.5	2.4	1.0	0.0	0.0	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0.6	0.0	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Cassidulina havanensis	2.1	9.8	8.0	1.6	1.1	1.7	2.0	0.0	0.3	0.0	0.3	0.4	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.9	0.0	0.0	0.0	0.0	0.0
Nonionella spp.	0.0	0.9	2.0	0.7	2.6	1.4	0.9	2.3	0.0	1.5	3.0	0.4	1.8	0.0	0.3	0.	0.2	0.5	0.0	1.3	0.7	0.3	0.5	1.9	0.6
Orthomorphina	1.0	1.1	0.0	1.2	0.2	1.0	1.5	0.0	0.0	0.6	3.6	2.5	2.1	0.0	0.6	0.6	1.7	0.0	0.0	0.6	0.0	1.0	0.1	0.3	0.0

(1) Percent of total number counted not accounted for by the first 48 taxa in the table (i.e. the 48 species used by Tjalsma and Lohmann (1982)).

Sample number refers to designations given on Table 3.

TABLE 1

BENTHIC FORAMINIFERAL CENSUS DATA -- SITE 549

SAMPLE NUMBER	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	26
AGE IN M.Y.	25.5	26.5	27.4	28.4	29.5	35.3	35.7	36.0	36.4	36.7	36.8	36.9	37.0	37.1	37.5	37.8	38.3	38.7	39.5	40.0	42.0	44.0	45.9	47.5	49.5
TOTAL NUMBER COUNTED	416	552	251	572	430	288	344	259	308	324	302	283	329	320	350	352	473	389	383	314	293	289	202	374	350
PERCENT UNACCOUNTED (1)	22.8	44.6	24.7	30.9	18.6	17.0	21.5	28.6	18.2	25.0	20.2	18.4	30.1	8.8	9.1	7.4	19.0	15.9	8.9	10.2	5.1	17.6	15.3	8.6	15.7
------------------------------ SPECIES INCLUDED IN TJALSMA AND LOHMANN'S (1982) CENSUS --------------------------------																									
SIMPLE AGGLUTINANTS	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Spiroplectammina spp.	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.0	0.0	0.0	0.6	1.4	0.0	0.5	0.8	0.3
Gaudryina cf. *laevigata*	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.5	0.0	0.0	0.3	0.3	0.0	0.0	0.0
Gaudryina pyramidata	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.3	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3
Gaudryina cf. *pyramidata*	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.3	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Vulvulina spp.	0.5	0.5	0.0	1.2	0.2	0.7	1.7	1.5	0.6	0.3	0.3	0.0	1.8	0.0	1.1	4.0	4.0	0.5	0.5	0.3	1.4	0.7	1.0	1.3	1.1
Karreriella subglabra	0.5	0.0	0.0	0.5	0.2	0.0	0.3	0.0	0.6	0.0	0.0	0.4	0.0	0.0	0.3	0.6	0.6	0.5	0.5	1.9	0.7	0.3	2.5	0.5	1.4
Pleurostomella spp.	4.8	2.4	4.8	3.3	10.7	5.9	7.3	2.7	5.5	4.3	8.3	5.3	2.4	7.2	2.0	3.1	2.7	3.6	1.8	2.2	2.0	3.1	5.0	2.4	1.1
Lenticulina spp.	1.2	1.4	0.0	1.0	1.4	1.7	0.9	1.5	0.6	1.9	0.3	1.4	1.5	0.3	0.6	1.1	1.5	0.5	0.3	1.3	2.4	5.2	5.0	3.2	6.9
Stilostomella aculeata	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.6	0.0	0.4	0.0	0.0	0.0	0.0	0.0	0.3	0.0	0.0	0.0	3.5	0.0	0.8	1.4
S. gracillima	12.3	5.3	5.6	9.1	3.3	0.7	0.0	0.0	0.6	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.4	1.5	0.3	0.3	0.0	5.9	0.5	1.3	0.0
S. subspinosa	4.3	0.0	6.4	2.4	0.9	4.9	0.9	2.7	0.6	1.2	3.0	1.8	4.6	0.6	0.6	4.0	2.3	4.1	3.1	1.9	1.7	0.0	2.0	0.0	0.0
Aragonia spp.	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.8	0.0
Bulimina alazanensis	0.0	0.0	0.0	0.0	0.0	0.3	0.0	0.0	0.0	0.6	0.3	0.7	0.3	0.6	0.0	0.0	0.0	0.0	0.0	2.5	0.0	0.0	0.5	0.0	0.0
B. tuxpamensis	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.0	0.0	0.0	0.0	6.1	5.5	0.0	0.0	1.7
Buliminella grata	0.5	0.0	0.0	0.2	0.5	0.3	0.0	0.0	0.0	0.3	0.0	1.1	2.7	0.0	0.9	0.9	1.5	4.6	2.9	1.0	3.4	0.0	0.0	2.7	1.4
Bulimina jarvisi	0.0	0.5	0.0	0.0	0.7	0.0	0.0	0.8	0.0	0.3	0.0	0.0	0.3	0.3	0.3	7.4	3.2	3.9	6.3	1.6	4.8	1.4	3.0	0.0	0.0
B. macilenta	0.0	0.5	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.5	0.0	0.3	0.7	0.5	0.0	0.0
B. callahani	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
B. semicostata	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.3	0.0	4.0	1.5	2.1	6.8	2.9	1.4	4.8	4.5	6.4	3.7
B. glomarchallengeri	0.0	0.2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.6	0.0	11.4	3.6	3.9	10.4	4.1	4.4	4.5	6.4	3.2	3.4
B. trinitatensis/ impendens	0.2	0.2	0.4	0.0	2.8	0.7	0.3	0.0	0.0	0.3	6.6	1.4	1.8	2.8	1.4	2.8	3.0	7.2	4.2	0.0	3.8	1.7	0.5	7.8	3.1
Turrilina spp.	0.5	0.4	0.8	0.3	0.9	0.3	3.2	0.0	0.6	0.6	0.0	0.4	1.8	0.0	0.0	0.0	0.0	0.3	0.0	0.0	0.0	3.1	1.5	0.3	0.0
Uvigerina rippensis	0.5	0.2	3.6	4.7	0.0	2.8	0.0	0.4	26.6	1.9	5.3	4.6	1.2	2.8	3.7	3.7	0.2	2.3	14.4	5.4	3.1	0.0	0.0	0.0	0.0
Quadrimorphina sp.	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	2.8	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Abyssamina spp.	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Clinapertina spp.	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Nonion havanense	4.6	4.0	2.4	2.1	2.8	0.0	0.0	0.0	2.3	0.9	1.3	1.4	2.1	2.5	0.6	1.7	3.4	11.1	0.8	3.5	0.0	0.0	2.5	1.1	4.6
Pullenia eocenica	4.8	4.2	3.6	3.0	1.9	7.3	2.6	1.2	1.0	2.2	3.6	6.0	4.0	3.8	0.3	4.3	4.2	6.9	0.5	4.1	4.8	1.7	5.0	2.9	0.6
Pullenia quinqueloba	2.2	2.0	1.6	3.3	3.3	4.9	2.0	6.2	2.3	3.1	2.3	4.9	1.5	2.5	1.1	1.7	1.1	1.5	0.8	4.1	0.0	0.3	0.5	0.8	0.9
Gyroidinoides spp.	8.2	6.5	8.4	8.6	9.8	8.0	9.0	19.7	9.1	10.5	13.9	11.0	8.8	10.3	5.4	5.1	4.9	4.6	4.7	8.0	5.1	2.4	5.9	3.5	3.7

TABLE 1

BENTHIC FORAMINIFERAL CENSUS DATA -- SITE 548 continued

SAMPLE NUMBER	1	2	3	4	5	6	7	8	9	10	11	12	13
AGE IN M.Y.	25.0	27.0	30.0	33.8	35.0	37.2	38.1	39.0	45.0	45.8	47.0	50.0	51.0
TOTAL NUMBER COUNTED	?83	337	393	429	331	360	411	350	344	312	437	263	405
PERCENT UNACCOUNTED (1)	20.8	23.4	26.7	31.2	20.5	29.4	22.6	17.1	12.2	13.8	2.7	24.3	16.5
--------------- SPECIES INCLUDED IN TJALSMA AND LOHMANN'S (1982) CENSUS ---------------													
Anomalina spissiformis	1.2	1.2	0.0	0.7	2.1	0.0	0.0	1.4	0.0	0.0	4.3	7.2	4.7
Gavelinella capitata/ semicribrata	1.2	2.7	0.5	2.3	0.3	0.3	1.5	2.9	3.5	3.5	3.4	0.8	0.7
Gavelinella micra	0.0	0.6	0.3	0.0	0.0	0.0	0.5	1.1	0.3	0.3	2.7	0.0	1.7
Nuttallides truempyi	0.0	0.0	0.0	0.0	0.0	0.0	0.2	0.0	10.2	5.1	12.8	10.3	18.0
Oridorsalis umbonatus	5.1	9.8	2.5	3.5	2.4	6.7	4.4	5.4	9.0	4.8	10.3	1.1	2.7
Osangularia mexicana	0.0	0.6	0.0	0.0	0.0	0.3	0.0	0.9	0.6	0.0	0.9	1.1	0.7
Alabamina dissonata	0.0	0.0	0.0	0.0	0.0	0.3	0.0	0.0	0.0	2.6	0.0	0.0	0.0
Hanzawaia cushmani	0.0	0.0	0.0	0.0	0.0	0.0	0.2	2.0	0.0	1.9	6.9	0.0	0.0
Globocassidulina subglobosa	6.1	5.6	14.8	6.5	6.0	4.2	10.2	2.6	3.5	0.3	0.0	0.8	0.0
Cibicidoides grimsdalei	0.0	0.0	0.0	0.0	0.0	0.3	0.7	0.0	0.3	0.0	1.6	0.0	0.0
C. haitiensis	2.3	1.8	1.5	2.1	3.6	1.9	1.0	1.7	4.1	1.3	0.0	0.0	3.2
C. havanensis	0.0	0.0	0.0	0.2	0.0	0.3	0.5	0.0	0.6	0.3	1.4	0.8	0.0
C. ungerianus	20.8	11.0	1.8	11.0	2.7	3.6	11.2	18.0	12.5	4.5	9.2	1.5	3.0
C. tuxpamensis	0.6	2.4	6.1	6.5	7.3	2.2	5.1	1.1	0.0	1.6	0.0	0.0	0.0
C. cf. *pseudoperlucidus*	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
C. subspiratus	0.0	0.0	0.0	0.0	0.0	0.6	0.2	0.0	0.0	0.0	0.0	0.0	0.2
C. laurisae	1.3	5.3	1.5	2.8	4.5	1.7	1.9	7.1	0.9	1.6	0.2	0.0	3.7

(1) Percent of total number counted not accounted for by the first 48 taxa in the table (i.e. the 48 species used by Tjalsma and Lohmann (1982)).

Sample number refers to designations given on Table 4.

TABLE 2

BENTHIC FORAMINIFERAL CENSUS DATA -- SITE 548

SAMPLE NUMBER	1	2	3	4	5	6	7	8	9	10	11	12	13
AGE IN M.Y.	25.0	27.0	30.0	33.8	35.0	37.2	38.1	39.0	45.0	45.8	47.0	50.0	51.0
TOTAL NUMBER COUNTED	?83	337	393	429	331	360	411	350	344	312	437	263	405
PERCENT UNACCOUNTED (1)	20.8	23.4	26.7	31.2	20.5	29.4	22.6	17.1	12.2	13.8	2.7	24.3	16.5
--------------- SPECIES INCLUDED IN TJALSMA AND LOHMANN'S (1982) CENSUS ---------------													
SIMPLE AGGLUTINATES	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.1	0.0
Spiroplectammina spp.	0.3	0.3	1.0	0.0	0.0	0.0	0.0	0.0	0.3	1.3	1.1	0.0	0.2
Gaudryina cf. laevigata	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	9.3	0.0	4.6	0.0	0.2
Gaudryina pyramidata	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.4	1.1	0.0
Gaudryina cf. pyramidata	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.6	0.0	0.0
Vulvulina spp.	1.9	1.2	3.3	2.3	4.5	1.9	3.4	0.9	0.9	3.5	3.0	0.4	2.5
Karreriella subglabra	0.7	0.9	0.0	0.0	0.9	0.0	0.0	0.6	0.3	0.6	1.6	0.4	0.0
Pleurostomella spp.	0.1	0.6	1.0	1.4	2.4	0.0	0.2	0.9	2.0	2.6	0.0	0.4	2.2
Lenticulina spp.	4.8	3.3	3.6	3.7	6.3	5.8	7.8	4.3	11.3	8.0	11.0	23.6	11.1
Stilostomella aculeata	0.7	0.3	2.0	0.7	0.3	1.7	0.0	0.0	0.0	0.0	0.0	0.0	0.2
S. gracillima	3.5	3.3	3.1	1.4	2.4	2.2	0.5	3.1	0.0	0.3	3.9	2.3	4.7
S. subspinosa	0.6	0.6	0.0	0.5	0.3	0.8	1.2	1.1	2.6	8.3	0.9	0.0	0.0
Aragonia spp.	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.0	0.6	0.5	0.0	0.0
Bulimina alazanensis	5.0	1.5	12.5	4.4	11.5	9.7	0.2	0.3	0.3	0.0	0.2	0.4	0.0
B. tuxpamensis	0.0	0.0	0.0	0.0	0.0	0.0	1.0	0.0	4.4	6.1	0.7	0.4	0.0
Buliminella grata	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	5.8	0.9	0.4	0.0
Bulimina jarvisi	0.0	0.0	0.0	0.5	0.0	0.8	0.0	0.0	0.0	0.0	0.0	0.0	0.0
B. macilenta	2.5	0.6	0.5	0.9	0.3	0.0	0.2	0.0	0.6	0.3	1.1	0.4	0.0
B. callahani	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
B. semicostata	0.0	0.0	0.0	0.0	3.6	5.0	0.0	4.0	0.0	0.0	0.0	0.0	0.0
B. glomarchallengeri	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.3	0.2	16.3	6.4
B. trinitatensis/ impendens	4.8	4.7	5.3	4.7	0.0	2.8	3.9	11.7	2.0	5.1	4.8	1.5	2.0
Turrilina spp.	0.0	0.9	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.0	0.0	0.8	6.2
Uvigerina rippensis	0.7	0.0	0.0	1.2	0.9	1.7	1.2	3.4	0.0	0.3	0.2	0.0	0.0
Quadrimorphina sp.	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Abyssamina spp.	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Clinapertina spp.	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Nonion havanense	0.4	2.4	1.5	4.0	1.2	0.6	1.0	1.1	0.0	2.9	2.7	0.4	5.2
Pullenia eocenica	4.1	4.5	5.1	1.2	6.9	7.2	11.2	0.9	2.3	7.7	0.2	0.0	0.0
Pullenia quinqueloba	1.2	0.9	0.3	0.2	1.5	1.7	0.7	0.6	3.2	2.6	1.4	1.5	2.0
Gyroidinoides spp.	9.1	9.8	5.1	6.1	6.6	5.6	7.1	5.4	2.9	1.0	1.4	0.8	1.7

TABLE 2

TABLE 3

ZONAL AGE ASSIGNMENTS, SITE 549

With notes on planktonic foraminiferal zonations

<table>
<tr><th>SAMPLE</th><th>NP ZONE[1]</th><th>FORAMINIFERAL ZONE</th><th>ESTIMATED AGE (Ma)</th></tr>
<tr><td colspan="4">HOLE 549A</td></tr>
<tr><td>1. 6-6 100-105cm</td><td>NP24/NP25</td><td>*G. ciperoensis*[2-4] (=P22/N4 ref. 5-7)</td><td>25.5</td></tr>
<tr><td colspan="4">Based upon the ranges of *G. quadriloatus primordius* (FAD latest Oligocene[4]), *G. dehiscens* (FAD latest Oligocene[4]), *G. ciperoensis* (Oligocene[4]),*G. binaiensis* (range *G. ciperoensis* Zone[4]; P22-N5, ref. 7), *G. opima nana*, *G. selli* (LAD P22, ref. 7), *G. angulofficinalis* (LAD P22, ref. 7), *C. chipolensis*, *G. suteri*, *D. galavisi* (LAD P21, ref. 7), *G. mendacis* (P21-N4, ref. 7), *C. unicavus*.</td></tr>
<tr><td>2. 7-6 118-123cm</td><td>NP24/NP25</td><td>P21B ref. 5-7</td><td>26.5</td></tr>
<tr><td colspan="4">Based upon *G. opima opima* (range P21, possibly P20, ref. 4), *G. angulisuturalis* (FAD base P21, ref. 4), *G. opima nana*, *G. anguliofficinalis*, *G. ciperoensis*, *G. psuedobesa*, *C. unicavus*, *C. dissimilis*, *G. praebulloides*, *D. galavisi*. Above the LAD of *Chiloguembelina* spp. (LAD P21A). Apparently anomalous occurrence of *S.* cf. *angiporoides*, for although Blow[7] notes its range into Zone P21, Stainforth et al.[4] note its extinction at the base of the *G. ampliapertura* Zone.</td></tr>
<tr><td>3. 8-6 60-65cm</td><td>NP24/NP25</td><td>P21B ref. 5-7</td><td>27.4</td></tr>
<tr><td colspan="4">As in 7-6 with *G. ouachitaensis* (LAD P21, ref. 7) and *G. tripartita*</td></tr>
<tr><td>4. 9-6 70-75cm</td><td>NP24/NP25</td><td>P21A ref. 5-7</td><td>28.4</td></tr>
<tr><td colspan="4">As in 8-6, but with *Chiloguembelina cubensis*. The overlap of this species and *G. opima opima* is indicative of Zone P21A (ref. 5-7).</td></tr>
<tr><td>5. 10-6 143-148cm</td><td>NP23</td><td>P21A ref. 5-7 (possibly P20)</td><td>29.5</td></tr>
<tr><td colspan="4">*G. opima opima* (range P21, possibly P20, ref. 4), *Chiloguembelina cubensis*, *G. permicra* (LAD P21, possibly base P22, ref. 7), *G. postcretacea* (LAD in P21, ref. 4), *G. euaperta*, *G. gortanii*, *G. munda*, *G.* cf. *pseudovenzulana* (LAD P19/20, ref. 7), *S.* cf. *angiporoides* (LAD lower P20, ref. 4). The absence of *G. angulisuturalis* suggests that this may be Zone P20, and therefore *G. opima opima*, which is found in abundance, has its FAD in that zone.</td></tr>
</table>

HIATUS

6. 11-6 143-148cm NP22 C. chipolensis/P. micra[2-4] 35.25

Based upon the occurrence of Psuedohastigerina barbadoensis (LAD top C. chipolensis/P. micra Zone[4]), G. ampliapertura (LAD top P20, ref. 4), S. angiporoides, G. opima nana, G. increbescens (LAD P19/20, ref. 7; late Eocene to early Oligocene[4]), G. euaperta (FAD P19/20, ref. 7), C. unicavus, D. galavisi.

7. 13-1 143-148cm NP22 C. chipolensis/P. micra[2-4] 35.6

As in 11-6 with G. psuedovenezulana, G. increbescens, G. postcretacea, G. tripartita, G. senilis.

8. 14-1 116-121cm NP21 C. chipolensis/P. micra[2-4] 36.0

G. tapuriensis (range P18-P19/20, ref. 7), G. opima nana, C. unicavus, D. galavisi, G. euaperta, G. permicra

The occurrence of G. euaperta in this sample together with nannoplankton indicative of Zone NP21 after the extinction of characteristic Eocene planktonic foraminifers (=Zone P18) disagrees with the range given by Blow (FAD P19/20, ref. 7)

9. 15-1 116-121cm NP21 C. chipolensis/P. micra[2-4] 36.4

10. 16-1 138-141cm NP21 C. chipolensis/P. micra[2-4] 36.7

11. 16-2 24-27cm NP21 C. chipolensis/P. micra[2-4] 36.7

12. 17-1 24-27cm NP21 C. chipolensis/P. micra[2-4] 36.9

Samples 11-6 to 17-1 contain low diversity, predominantely simple globigeriniform planktonic foraminifers which are not age diagnostic. This interval is assigned to the C. chipolensis/P. micra Zone based upon the presence of Pseudohastigerina after the extinction of the G. cerroazulensis group and Hantkenina spp.

--------------------EOCENE/OLIGOCENE BOUNDARY--------------------

Hole 549A

13. 17-2 43-45cm NP21 G. cerroazulensis cunialensis[8] 37.0

Based upon the presence of G. cerroazulensis cunialensis (FAD base of G. cunalensis Zone[8]; ~ upper P16-lower P17), G. cerroazulensis cocoaensis, G. c. cocoaensis transitional to cunialensis and G. cerroazulensis cerroazulensis, Cribrohantkenina, Hantkenina alabamensis, G. opima nana, G. ampliapertura, S. angiporoides, P. micra s.s.

14. 18-1 103-106cm NP21 *G. cerroazulensis cunialensis*[8] 37.1

As in 17-2.

15. 24-2 126-130cm NP21 *G. cerroazulensis cunialensis*[8] 37.5

As in 17-2 with *G. mexicana*, *G. tropicalis*, *G. postcretacea* (FAD *G. cerroazulensis* s.l. Zone[4])

16. 28-1 143-146cm NP19/NP20 *G. cerroazulensis cunialensis*[8] 37.8

Toumarkine and Bolli[10] noted that in the Possagno section, Italy *G. ampliapertura* had its first appearence just above the FAD of *G. cerroazulensis cunialensis*. The absence of *G. ampliapertura* together with the presence of the nominate *G. cerroazulensis cunialensis* in sample 28-1 suggests that this sample may be assigned to the lower portion of the zone.

17. 33-1 143-146cm NP19/NP20 *G. cerroazulensis cocoaensis*[8] 38.3
(~ upper P15-P16 ref. 5-7)

18. 35-1 26-28cm NP18* *G. cerroazulensis cocoaensis*[8] 38.7
(~ upper P15-P16 ref. 5-7)

Samples 33-1 and 35-1 are assigned to the *G. cerroazulensis cocoaensis* Zone based upon the first appearance of the nominate taxa in 35-1 (ref. 8; ~ upper *G semiinvoluta* to lower *G. cerroazulensi* s. l. Zone of ref. 2-4). The underlying two samples are thus placed in the basal late Eocene *G. cerroazulensis cerroazulensis* Zone[8]

19. 38-1 49-53cm NP18 *G. cerroazulensis cerroazulensis*[8] 39.5
(~P15 ref. 5-7)

Based upon the presence of *G. cerroazulensis cerroazulensis*, *G. cerroazulensis pomeroli*, *G. opima nana*, *S. angiporoides*, and *G. increbescens* prior to the appearence of *G. cerroazulensis cocoaensis*.

20. 40-1 94-97cm NP17/NP18 *G. cerroazulensis cerroazulensis*[8] 40.0
(~P15 ref. 5-7)

As in 38-1 with *A. spinuloinflata* (ranges into Zone P15, ref. 7)

Hole 549

21. 2-5 126-129cm -- P14 ref. 5-7 (possibly earliest P15) 42.0

Based upon the presence of *G. cerroazulensis pomeroli* (= centralis of Blow[5,7])(FAD upper P12, ref. 8), *M. spinulosa* (P10-P14, ref. 4), *T. rohri* (LAD top P14, ref. 2-5; LAD base P15, ref. 7), *G. opima nana*, *S. angiporoides* (both FAD P15, ref. 4). This is a very similar assemblage to that noted in 548A: 17-5 with the exception of the absence of *G. semiinvoluta* which is used by Blow[7] to differentiate Zone P14 from P15. The overlap of *G. opima nana* and *S. angiporoides* with *M. spinulosa* and *T.*

rohri is not considered to be the result of reworking or contamination, and therefore contradicts the ranges for the former taxa given by Stainforth[7], but agrees with Blow's ranges.

22. 4-6 145-148cm NP15/16 P12 (P12-P13) ref. 5-7 44.0

Based upon the presence of G. frontosa (LAD P12 ref. 4,8), M. spinulosa coronata (P10-P13, ref. 7), G. cf. cerroazulensis possagnoensis (upper P11 to lower P12, ref. 8), T. topilensis (P12-lower P14, ref. 7), M. spinulosa, A. densa, D. galavisi, S. linaperta, P. micra, P. cf. wilcoxensis, G. index.

23. 6-6 148-150cm NP15/16 upper P11-lower P12 (P11-P12) ref. 5-7 45.9

Based upon the presence of G. frontosa (P9-P12, ref. 4; LAD P11, ref. 4,7), G. index (FAD late P11, ref. 4), G. higginsi (P9-P11/early P12, ref. 7), T. rohri, A. bullbrooki (= A. densa), S. crociaperta, D. galavisi, G. linaperta.

24. 8-4 145-147cm NP15/16 P11 ref. 5-7 47.5**

Based upon the presence of G. broedermanii broedermanii (P6b-P11, ref. 4; to lower P12 ref.), G. broedermanii anapetes (upper P10-P11, ref. 7), P. micra (FAD P11, ref. 4,9), A. bullbrooki (= A. densa)(FAD P9, ref. 5,7; range P10 to ~P11, ref. 4), M. spinulosa, P. wilcoxensis.

25. 10-5 57-61cm NP13 P9 ref. 5-7 49.5

Based upon the presence of G. caucasica (P8-P9, ref. 4), A. bullbrooki (= A. densa)(FAD P9, ref. 5,7; range P10-~P11, ref. 4), D. galavisi (FAD P9, ref. 5,7), G. renzi (P8-P14, ref. 4), G. frontosa (FAD P9, ref. 4), G. cf. broedermanii, P. sharkriverensis (FAD early middle Eocene[9]).

REFERENCES:

1. Data provided by M.-P. Aubry, personal communication.

Zonation of 2. Bolli 1957a-b; 3. Bolli 1966; and 4. Stainforth et al., 1975. 5. Blow, 1969; 6. Hardenbol and Berggren, 1978; 7. Blow, 1979. 8. Zonation of Toumarkine and Bolli, 1970. 9. Cordey, et al., 1970; 10. Toumarkine and Bolli, 1975

* Age discrepancies between planktonic foraminiferal and nannofossil age assignments. ** Age discrepancies between zonal age assignment and estimated age in Ma result from interpolation between tie points (assuming constant sedimentation rates between points underlined above; Fig. 2).

TABLE 4

ZONAL AGE ASSIGNMENTS, SITE 548

SAMPLE	NP ZONE[1]	FORAMINIFERAL ZONE	ESTIMATED AGE (Ma)
1. 15-6 52-56cm	NP24/25	*G. ciperoensis*[2-4]	25.0
2. 16-1 99-102cm	NP24	P21B ref. 5-7	27.0
3. 16-3 66-69cm	NP23	μP20 ref. 5-7	30.0

Note: the very low sedimentation rates computed for the above section (less than ~ 1 m/m.y.) suggest that it is discontinuous.

-------------------- MAJOR HIATUS --------------------

SAMPLE	NP ZONE[1]	FORAMINIFERAL ZONE	ESTIMATED AGE (Ma)
4. 16-4 66-69cm	NP22 (possibly NP23)	*C. chipolensis*/*P. micra*[2-4]	35.0
5. 16-5 23-26cm	NP22	*C. chipolensis*/*P. micra*[2-4]	35.4

possible minor hiatus

SAMPLE	NP ZONE[1]	FORAMINIFERAL ZONE	ESTIMATED AGE (Ma)
6. 17-1 99-102cm	NP21	*G. cerroazulensis* s.l.[2-4]	37.2
7. 17-3 102-105cm	NP19/20	*G. cerroazulensis* s.l.[2-4]	38.4
8. 17-5 121-124cm	NP18	earliest P15 ref. 7	39.8

-------------------- MAJOR HIATUS --------------------

SAMPLE	NP ZONE[1]	FORAMINIFERAL ZONE	ESTIMATED AGE (Ma)
9. 18-3 98-101cm	NP15/16	P12 ref. 5-7	45.0
10. 19-5 88-91cm	NP15/16	P11 ref. 5-7	45.8
11. 21-4 55-58cm	NP15/16	P10-11 ref. 5-7	47.0

-------------------- MAJOR HIATUS --------------------

SAMPLE	NP ZONE[1]	FORAMINIFERAL ZONE	ESTIMATED AGE (Ma)
12. 22-6 94-97cm	NP12	P8-9 ref. 5-7	50.0
13. 24-1 38-41cm	NP12	P8	51.0

REFERENCES:

1. Data provided by M.-P. Aubry, personal communication.

Zonation of: 2. Bolli 1957a-b; 3. Bolli 1966; and 4. Stainforth et al., 1975. 5. Blow, 1969; 6. Hardenbol and Berggren, 1978; 7. Blow, 1979. 8. Zonation of Toumarkine and Bolli, 1970.

TABLE 5

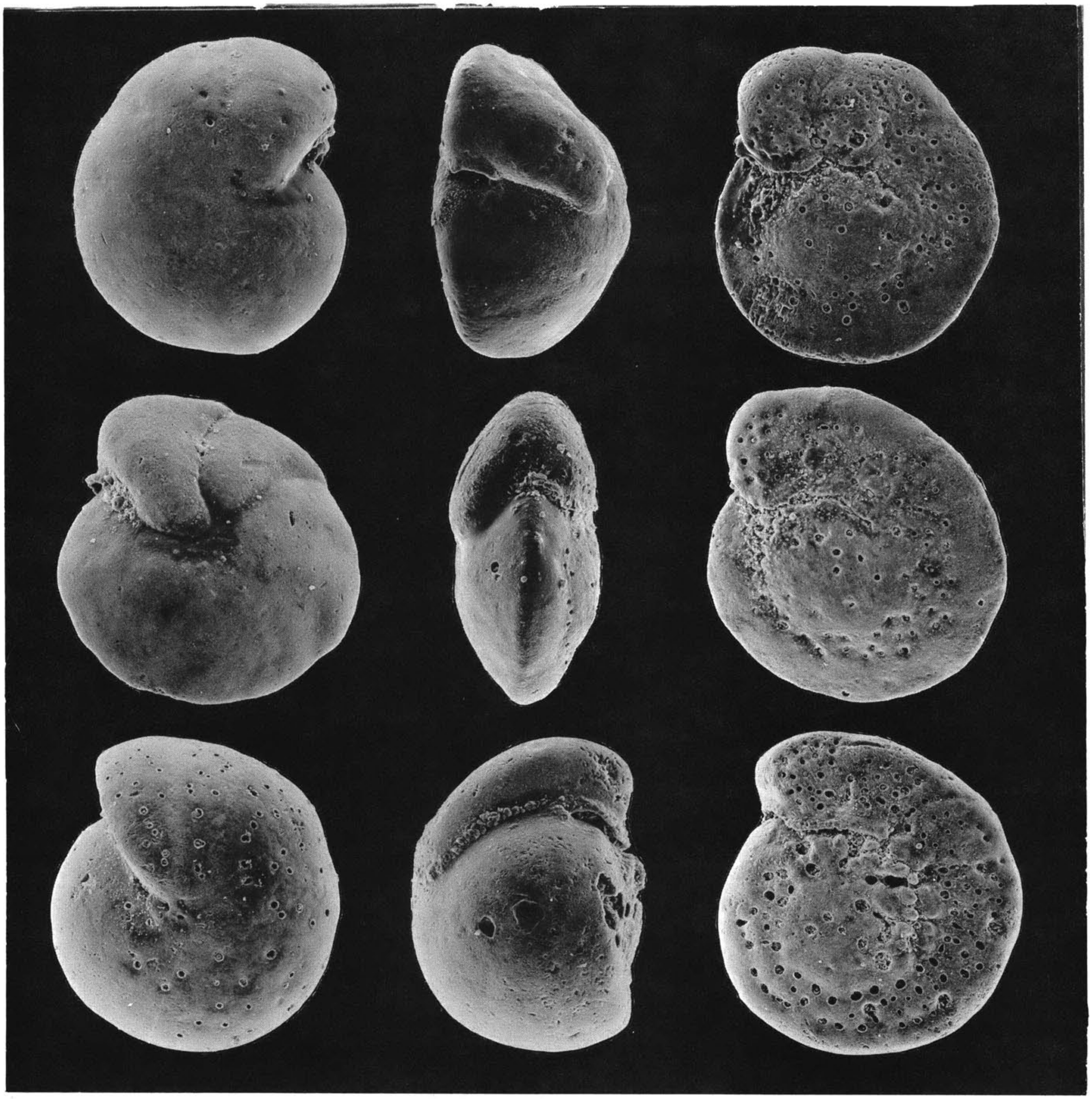
BENTHIC FORAMINIFERAL OXYGEN AND CARBON ISOTOPIC DATA -- SITE 549
All analyses performed on monogeneric *Cibicidoides* spp.

	SAMPLE	DEPTH	$\delta^{18}O$	$\delta^{13}C$	NANNO. ZONE	FORAM. ZONE
HOLE 549A						
1.	6-6	100-105cm	1.17 1.47	0.83 0.87	NP24/NP25	P22
2.	7-6	118-123cm	1.17	0.52	NP24/NP25	P21B
3.	8-6	60-65cm	1.38	0.76	NP24/NP25	P21B
4.	9-6	70-75cm	1.41	0.73	NP24/NP25	P21A
5.	10-6	143-148cm	1.45	1.00	NP23	P21A (P20)
				HIATUS		
6.	11-6	143-148cm	1.22 1.18	1.15 1.04	NP22	*C. chipolensis/P. micra*
7.	13-1	143-148cm	1.34	1.23	NP22	*C. chipolensis/P. micra*
8.	14-1	116-121cm	1.31	1.58	NP21	*C. chipolensis/P. micra*
9.	15-1	116-121cm	1.58	1.57	NP21	*C. chipolensis/P. micra*
10.	16-1	138-141cm	0.85	1.03	NP21	*C. chipolensis/P. micra*
11.	16-2	24-27cm	0.96	1.18	NP21	*C. chipolensis/P. micra*
12.	17-1	24-27cm	0.94	1.21	NP21	*C. chipolensis/P. micra*
				--------------------EOCENE/OLIGOCENE BOUNDARY--------------------		
13.	17-2	43-45cm	0.80	0.99	NP21	*G. c. cunialensis*
14.	18-1	103-106cm	0.97	1.17	NP21	*G. c. cunialensis*
15.	24-2	126-130cm	0.77 0.93	1.12 1.15	NP21	*G. c. cunialensis*
16.	28-1	143-146cm	0.65	1.00	NP19/NP20	*G. c.cunialensis*
17.	33-1	143-146cm	0.43	1.30	NP19/NP20	*G. c. cocoaensis*

TABLE 5 (cont.)

BENTHIC FORAMINIFERAL OXYGEN AND CARBON ISOTOPIC DATA -- SITE 549
All anlyses performed on monogeneric Cibicidoides spp.

SAMPLE	DEPTH	$\delta^{18}O$	$\delta^{13}C$	NANNO. ZONE	FORAM. ZONE
Hole 549A					
18. 35-1	26-28cm	0.55	1.39	NP18	G. c. cocoaensis
19. 38-1	49-53cm	0.72	0.88	NP18	G. c. cerroazulensis
20. 40-1	94-97cm	0.60	0.96	NP17 (NP18)	G. c. cerroazulensis
Hole 549					
21. 2-5	126-129cm	0.50	0.72	--	P14
22. 4-6	145-148cm	insufficient material			
23. 6-6	148-150cm	insufficient material			
24. 8-4	145-147cm	-0.16	0.84	NP15	P11
25. 10-5	57-61cm	-0.99	0.90	NP13	P9



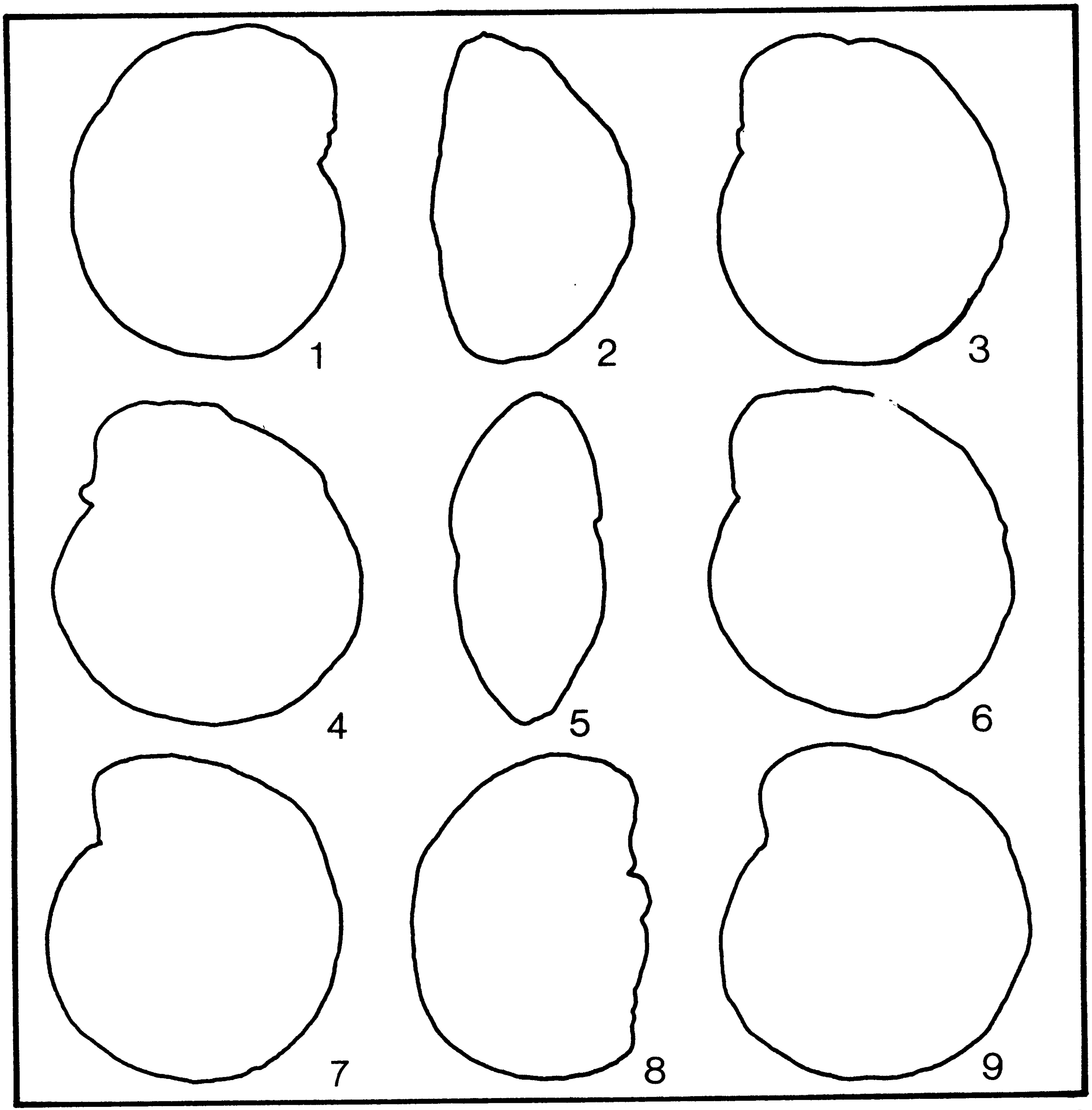
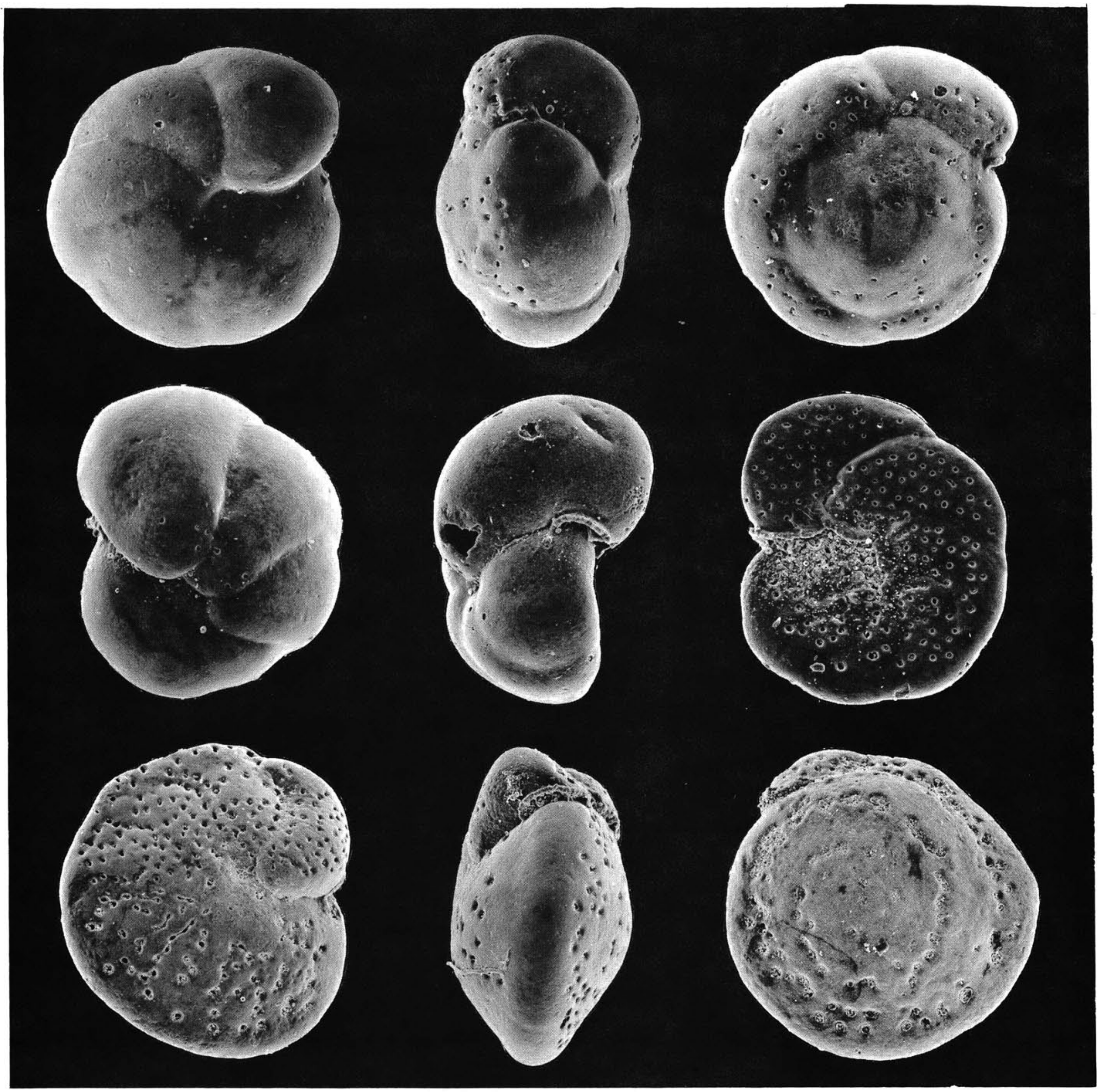


Plate 1. *Cibicidoides* spp., Site 549. Figs. 1-3. *Cibicidoides tuxpamensis*. 1. Sample 549A 17-2, 43-35cm, x120; 2. Sample 549A 13-1, 143-148cm, x 120; 3. Sample 24-2, 126-130cm, x160. Figs. 4-6. *Cibicidoides ungerianus*, Sample 549A 10-6, 143-148cm. 4. x205; 5. x180; 6. x175. Figs. 7-9. *Cibicidoides grimsdalei*. 7. Sample 549A 35-1, 26-28cm, x145; 8. Sample 549A 33-1, 143-146cm, x210; 9. Sample 549A 33-1, 143-146cm, x95.



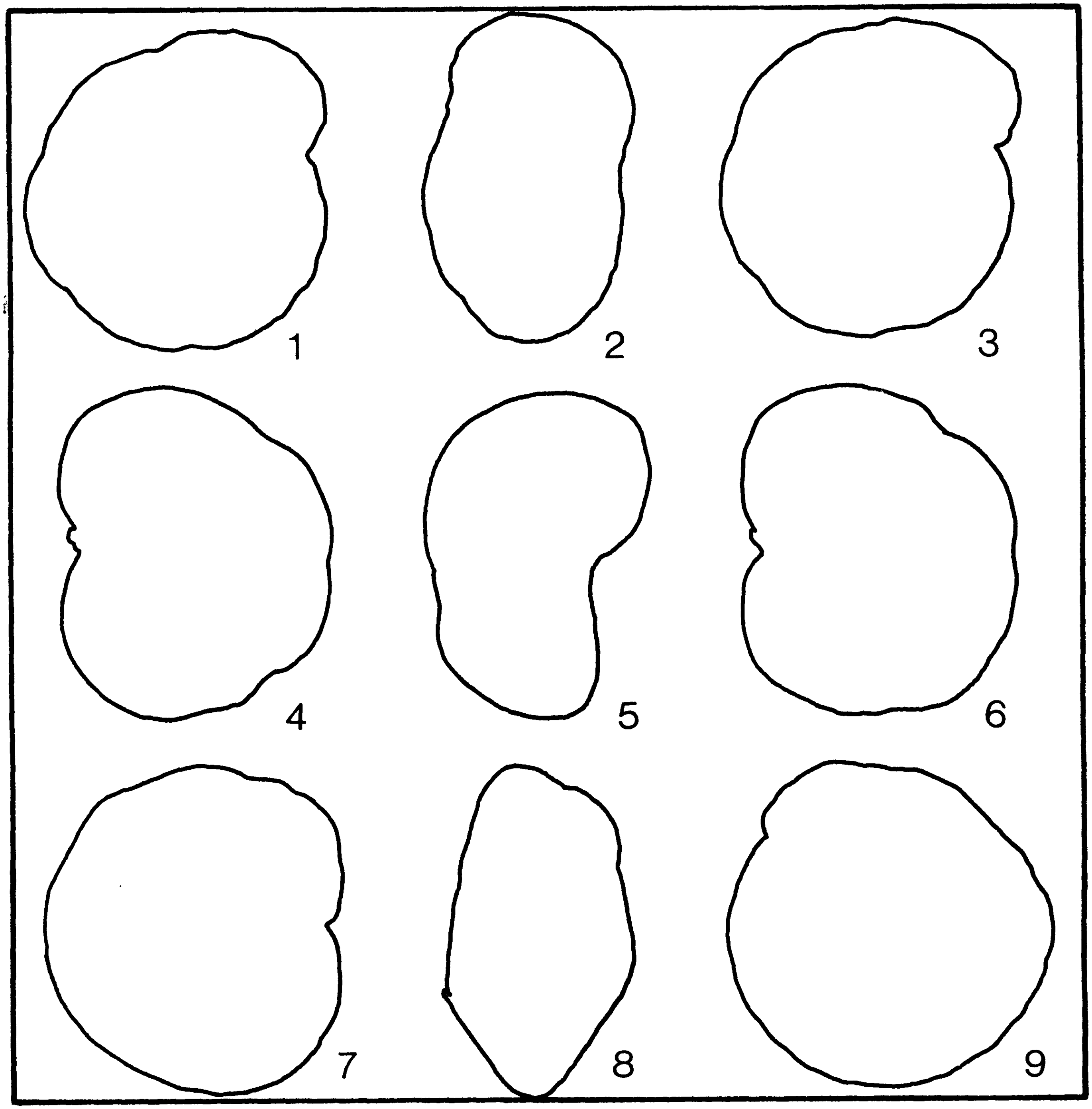
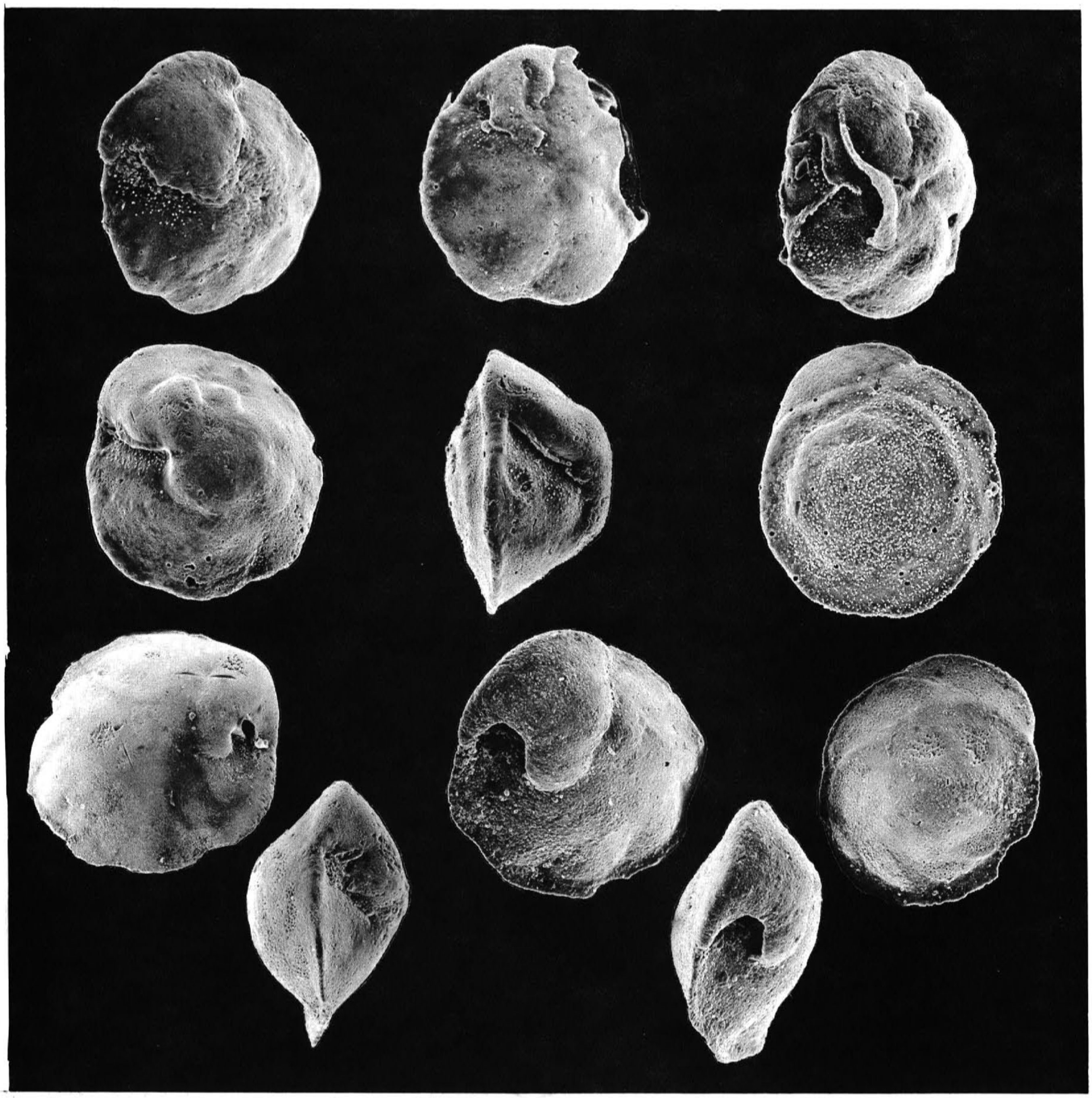


Plate 2. *Cibicidoides* spp., Site 549. Figs 1-3. *Cibicidoides haitiensis*, Sample 549A 14-1, 116-121cm. 1. x175; 2. x 200; 3. x185. Figs. 4-6. *Cibicidoides laurisae*. 4. Sample 549A 7-6, 118-123cm, x200; 5. Sample 549A 7-6, 118-123cm, x190; 6. Sample 13-1, 143-148cm, x210. Figs. 7-9. *Cibicidoides havanesis*. 7. Sample 549A 13-1, 143-148cm, x70; 8. Sample 549A 14-1, 116-121cm, x90; 9. 549A 14-1, 116-121cm, x80.



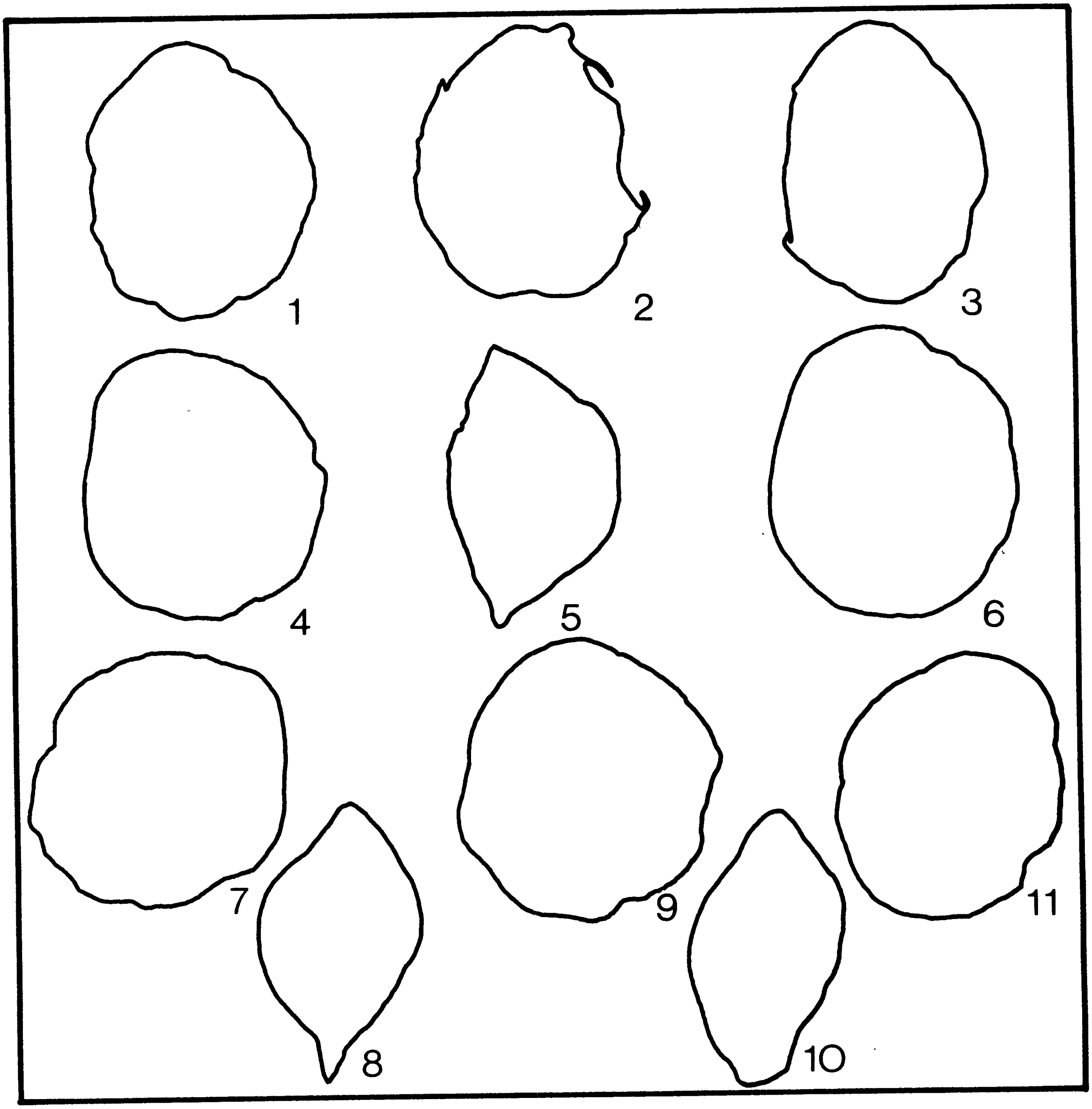
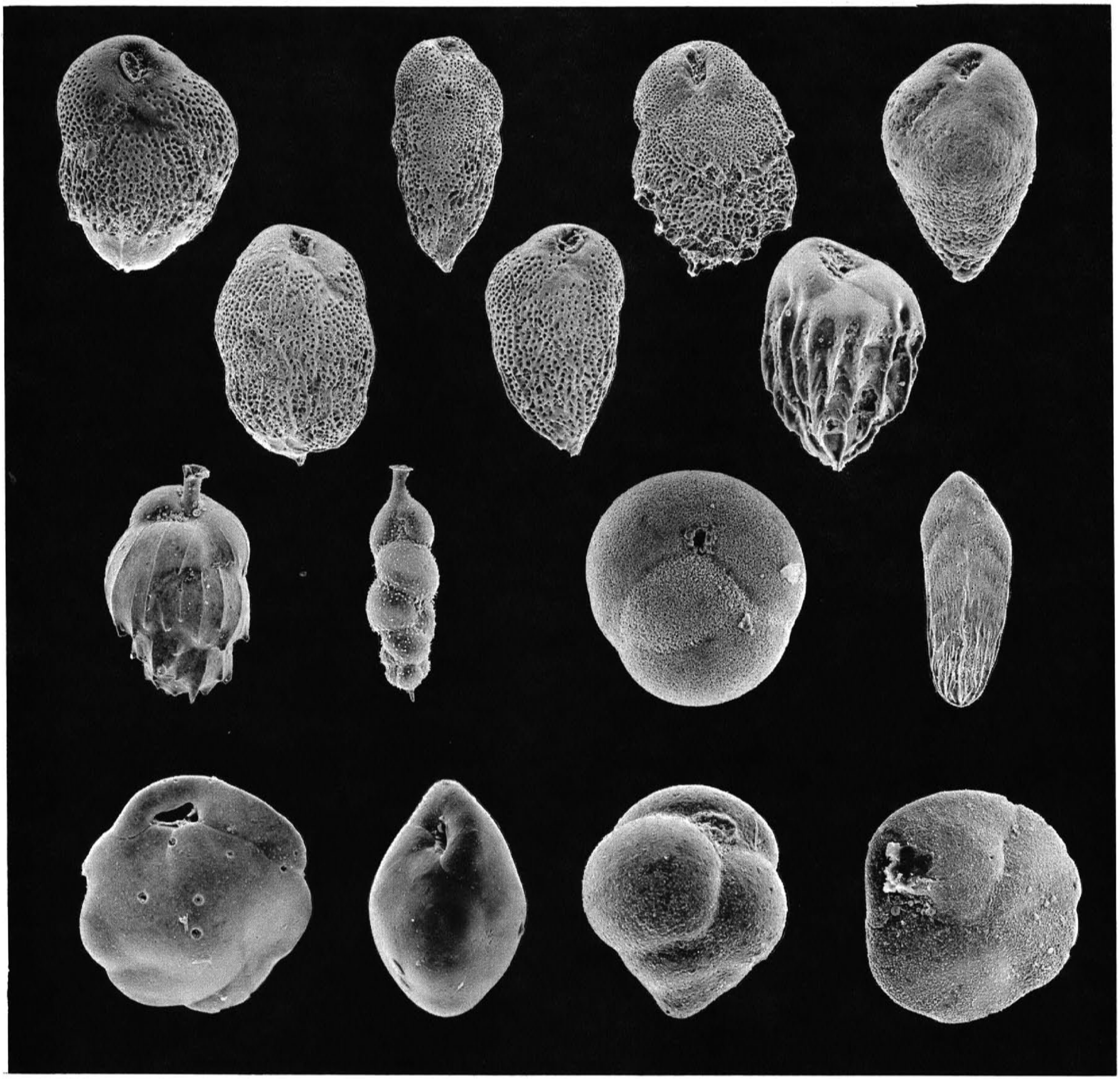


Plate 3. *Nuttallides* and *Osangularia* spp., Site 549. Figs. 1-2. *Nuttallides umbonifera*. 1. Sample 549A 7-6, 118-123cm, x210; 2. Sample 549A 33-1, 143-146cm, x195. Fig. 3. *Nuttallides* cf. *umbonifera*, Sample 549A 35-1, 26-28cm, x210. Figs. 4-6. *Nuttallides truempyi*. 4. Sample 549A 35-1, 26-28cm, x125; 5, 6. Sample 549A 40-1 94-97cm, x150. Figs. 7,8. *Osangularia mexicana*, Sample 549A 35-1, 26-28cm. 7. x65; 8. x115. Figs. 9-11. *Osangularia* sp. 2, Sample 549A 24-2, 126-130cm. 9. x150; 10. x 155; 11. x135.



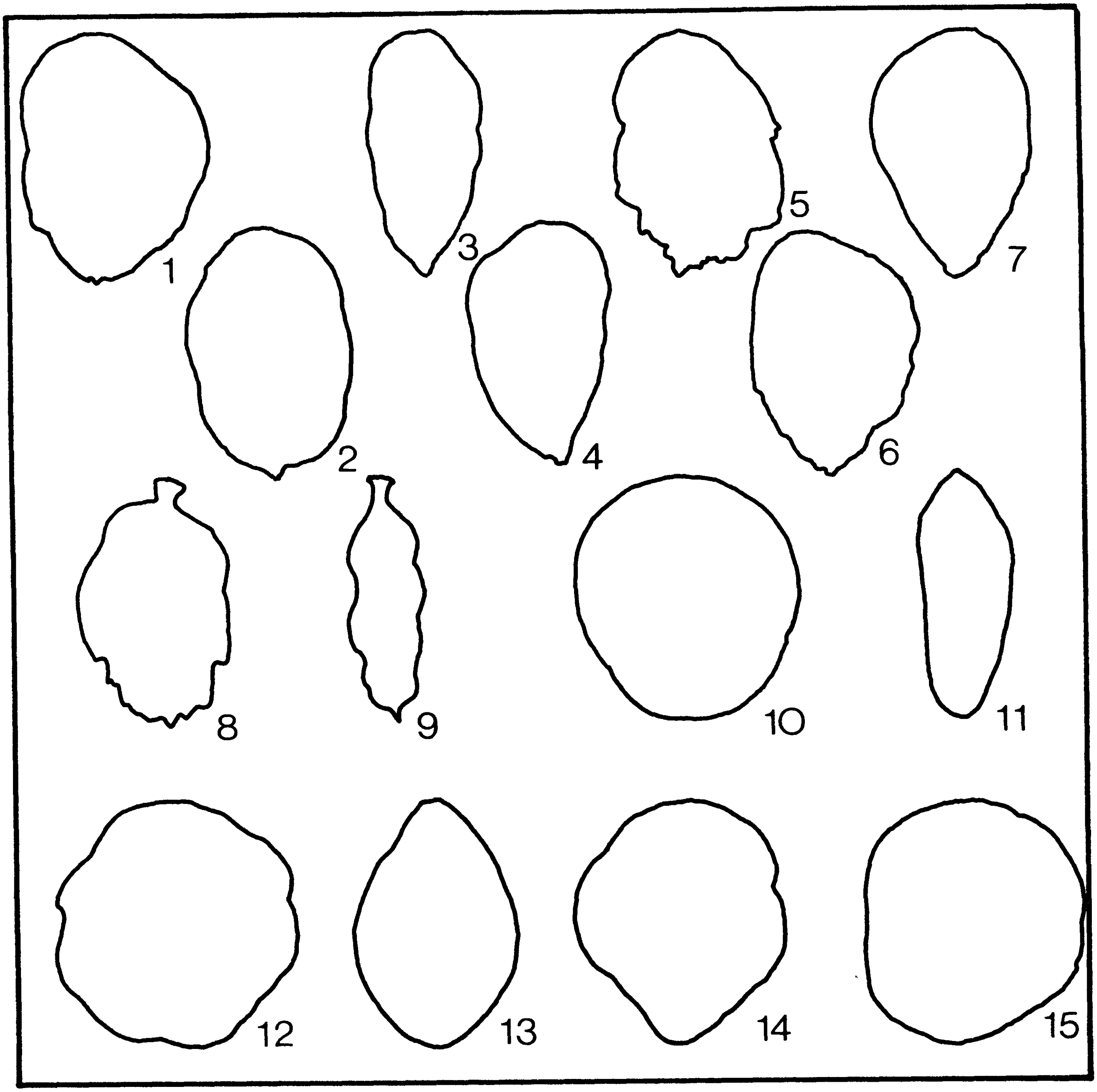
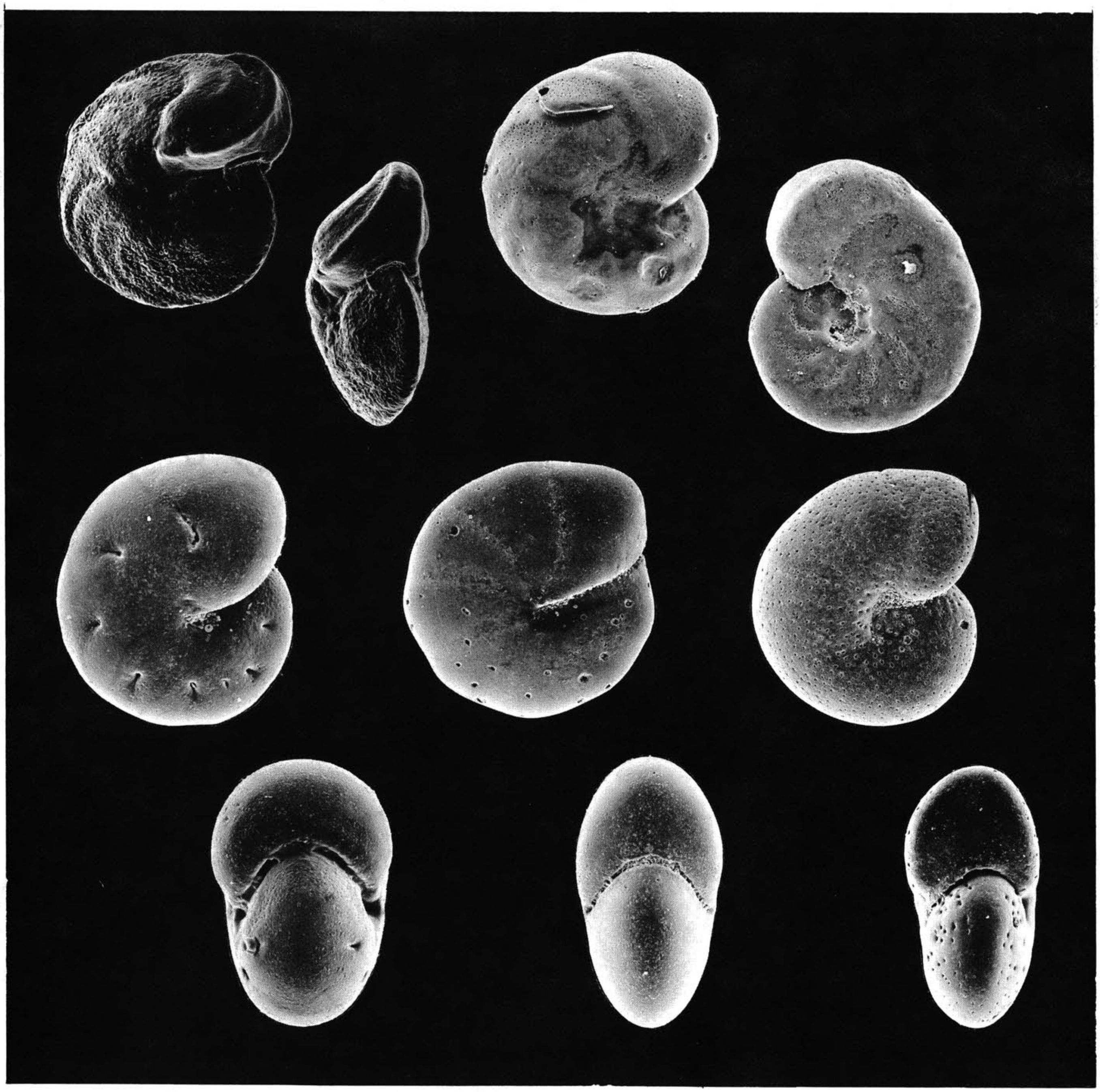


Plate 4. Buliminids and late Oligocene taxa, Site 549. Fig. 1. *Bulimina glomarchallengeri*, Sample 549A 28-1, 143-146cm, x110. Fig. 2 *B. cf. semicostata*, Sample 549A 28-1, 143-146cm, x90. Fig. 3. B. *jarvisi*, Sample 549A 28-1, 143-146cm, x50. Fig. 4. *B. semicostata*, Sample 549A 28-1, 143-146, x90. Fig. 5. *B. trinitatensis*, Sample 549A 33-1, 143-146cm, x100. Fig. 6. *B. alazanensis*, Sample 549A 16-2, 24-27cm, x165. Fig. 7. *B. tuxpamensis*, Sample 549 2-5, 126-129cm, x105. Figs. 8,9. *Uvigerina* spp., Sample 549A 6-6, 100-105cm, x105. Fig. 10. *Sphaeroidina bulloides*, Sample 549A 6-6, 100-105cm, x190. Fig. 11. *Bolivina antegressa*, Sample 549A 7-6, 118-123cm, x75. Figs. 12,13. *Cassidulina havanesis*, Sample 549A 7-6, 118-123cm, x190. Fig. 14. *Eggerella bradyi*, Sample 549A 7-6, 118-123cm, x165. Fig. 15. *Epistominella exigua*, Sample 549A 7-6, 118-123cm, x165.



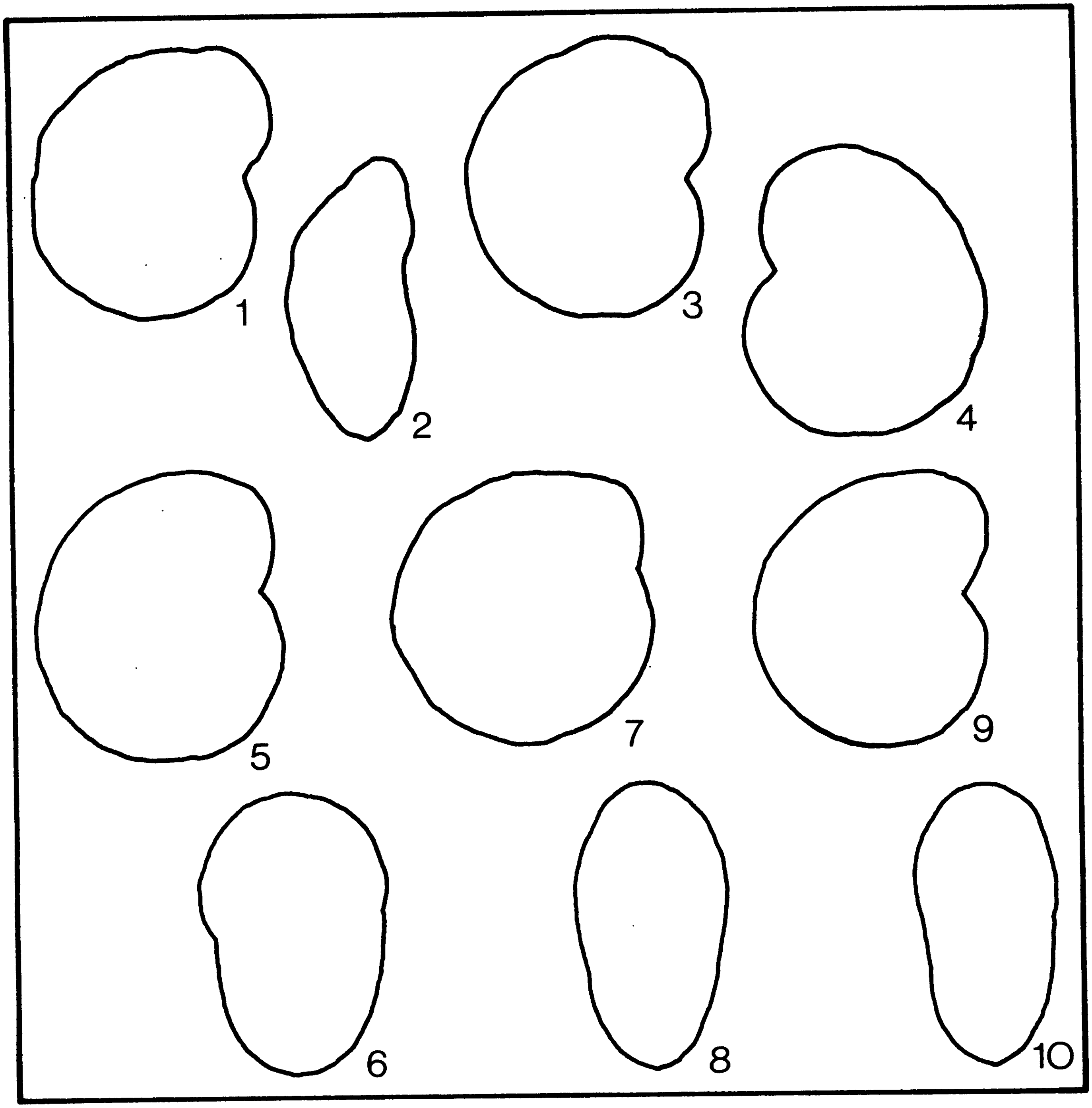
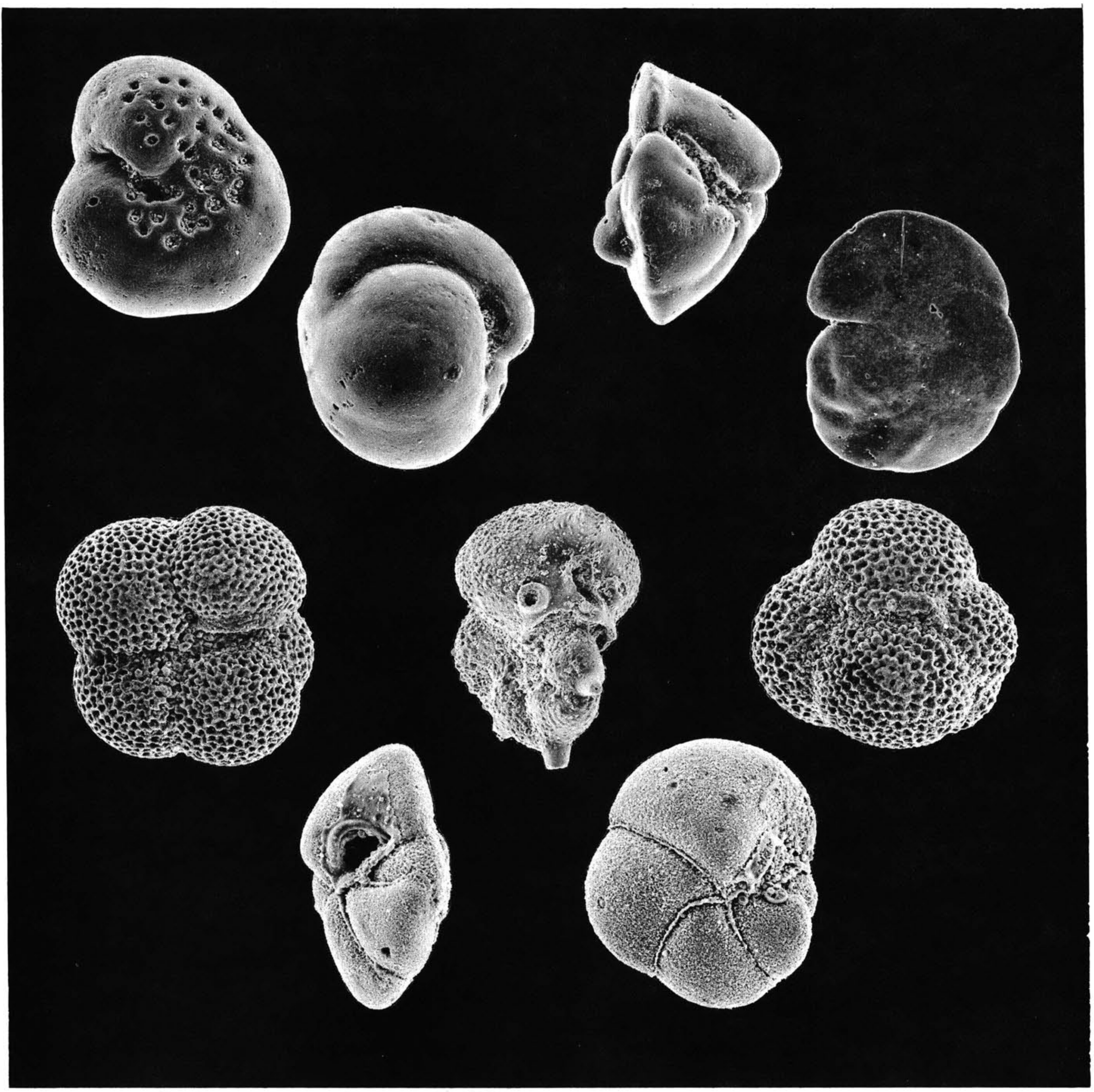


Plate 5. Oligocene taxa, Site 549. Figs. 1,2. *Planulina* sp. 6, Sample 549A 10-6, 143-148cm. 1. x70; 2. x65. Figs. 3,4. *Hanzawaia cushmani*, Sample 549A 14-1, 116-121cm. 3. x85; 4. x70. Figs. 5,6. *Astrononion pusillum*, Sample 549A 9-6, 70-75cm, x180. Figs. 7,8. *Nonion havanense*, Sample 549A 10-6 143-148cm. 7. x110; 8. x145. Figs. 9,10. *Anomalina spissiformis*, Sample 549A 16-1, 138-141cm. 9. x90; 10. x150.



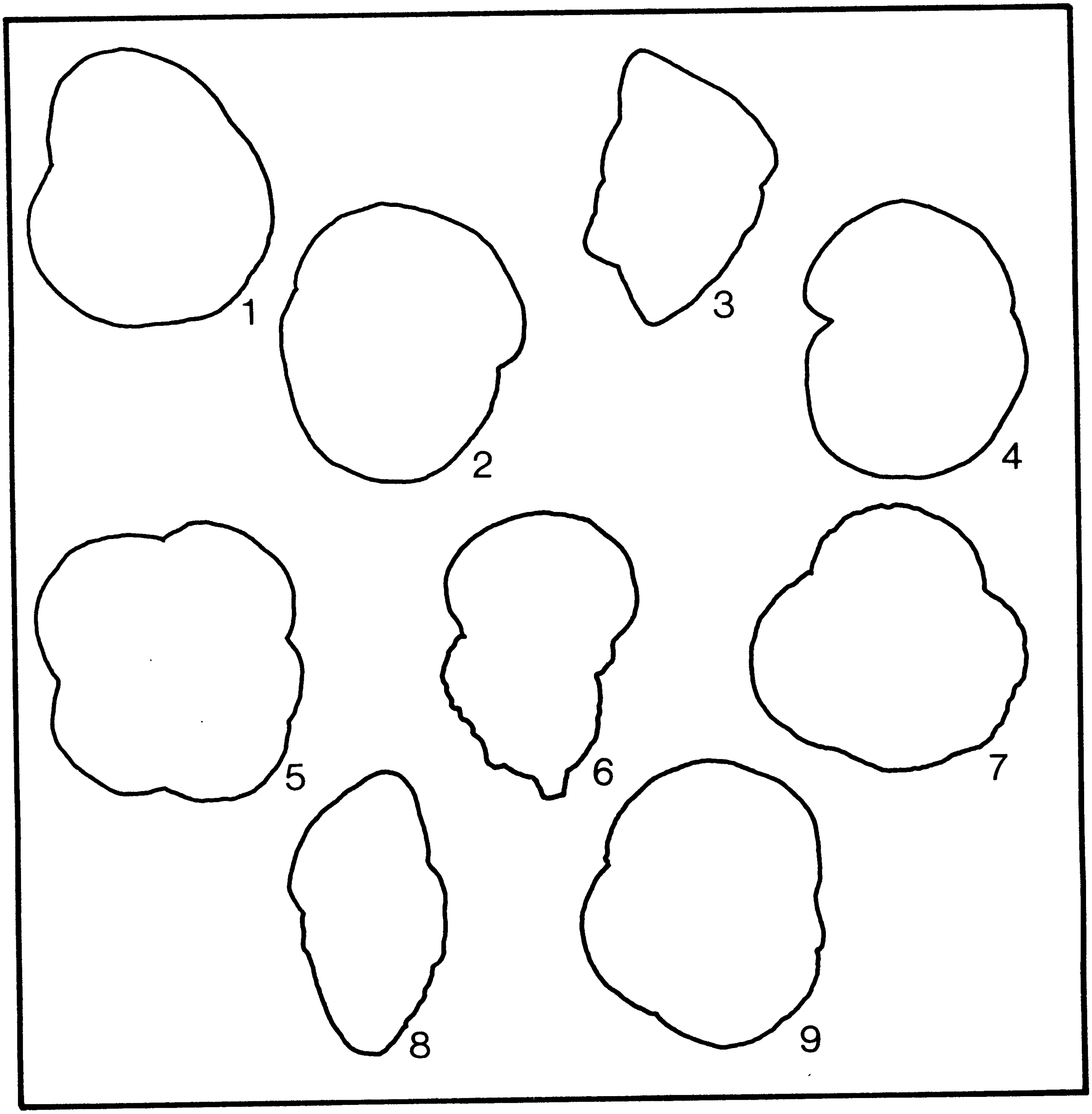
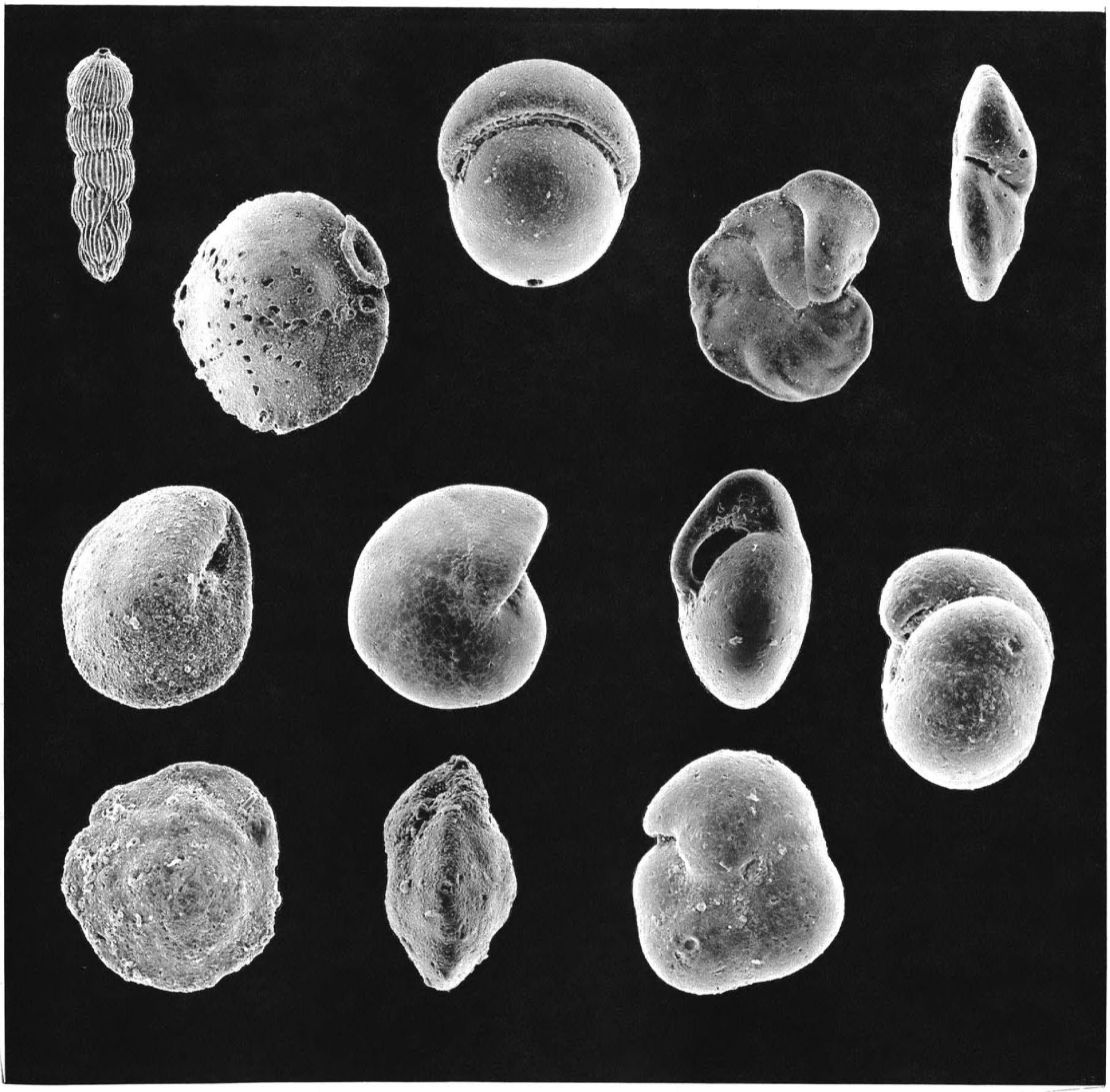


Plate 6. Miscellaneous benthic foraminifers and planktonic foraminifers, Site 549. Figs. 1,2. Gavelinella semicribrata, Sample 549A 33-1, 143-146cm. 1. x130; 2. x180. Figs. 3,4. µGyroidinoides spp., Sample 549A 11-6, 143-148cm. 3. x 150; 4. x135. Fig. 5. Globorotalia opima opima, Sample 549A 10-6, 143-148cm, x 180. Fig. 6. Cribrohantkenina sp., Sample 549A 17-2, 43-45, x90. Fig. 7. Subbotina angiporoides, Sample 549A 10-6, 143-148cm, x190. Figs. 8,9. Globorotalia cerroazulensis cunialensis, Sample 549A 28-1, 143-146cm.



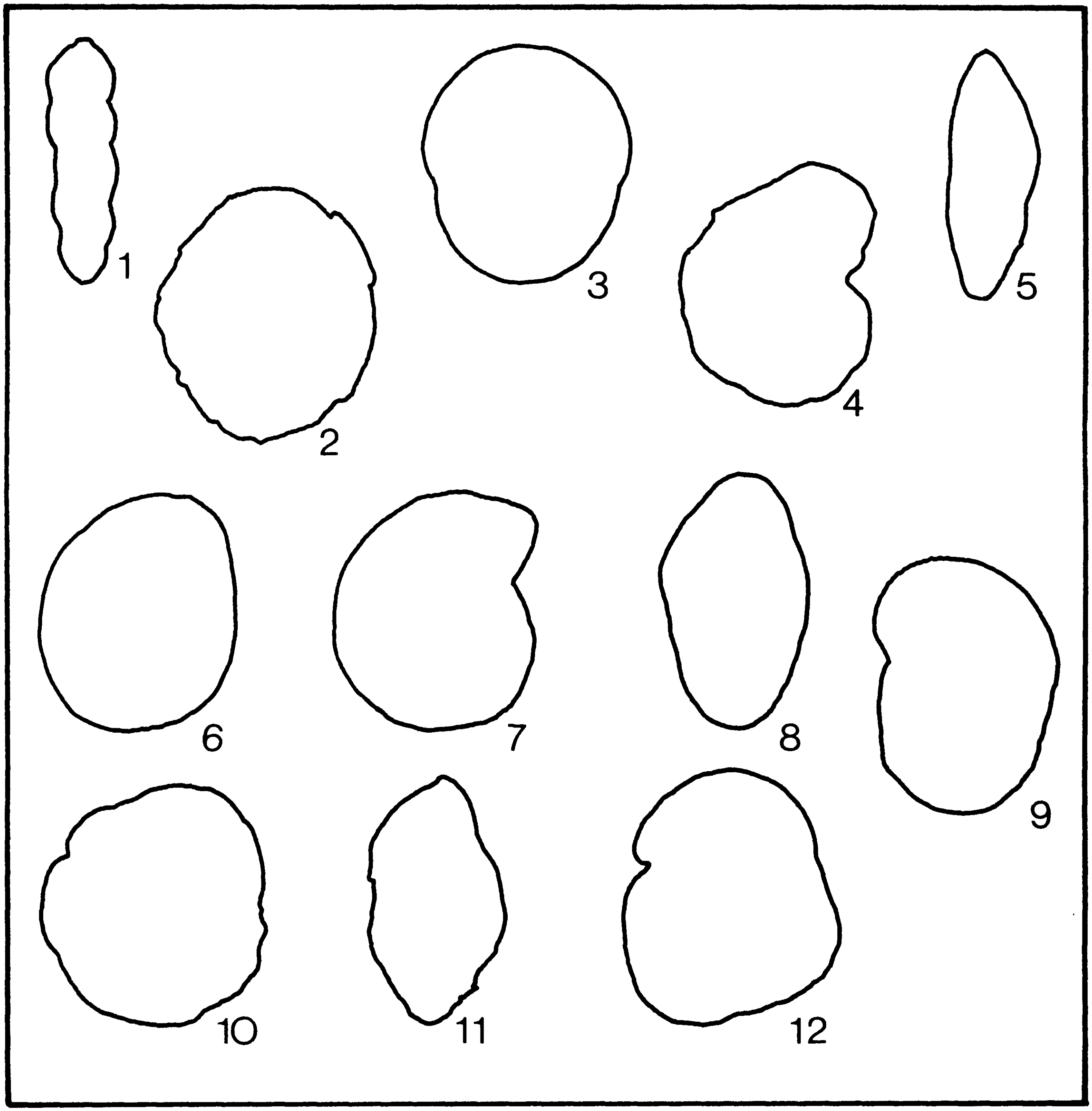


Plate 7. Top two rows: benthic foraminiferal taxa, Site 548. Bottom two rows: benthic foraminiferal taxa, Site 550. Figs. 1,2 from the Miocene, although they also occur in the late Oligocene. Fig. 1. *Rectouvigerina* sp., Sample 548, 15-2, cm, x125. Fig. 2. *Siphonina tenuicarinata*, Sample 548, 15-2, cm, x 125. Fig. 3. *Pullenia bulloides*, Sample 548 15-6, 52-56cm, x130. Figs. 4,5. *Planulina* sp. 5., Sample 548 15-6, 52-56cm. 4. x50; 5. x135. Fig. 6. *Alabamina dissonata*, Sample 550, 24-2, 144- cm, x125. Figs. 7,8. *Clinapetina inflata*, Sample 550 24-5, 56-cm. 7. x170; 8. x146. Fig. 9. *Abyssamina poagi*, Sample 550 24-CC, x195. Figs. 10,11. *Nuttallides umbonifera*, Sample 550 24-1, 74- cm. 10. x176; 11. x190. Fig. 12. *Quadrimorphina profunda*, Sample 550 24-4, 54- cm, x220.