# 40. WESTERN NORTH ATLANTIC: SEDIMENTARY EVOLUTION AND ASPECTS OF TECTONIC HISTORY[1]

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## ABSTRACT

Drilling results from DSDP Leg 43 and earlier legs in the western North Atlantic are synthesized to examine the sedimentary history and aspects of the tectonic evolution of the basin. In the Late Jurassic, reddish calcareous claystones and limestones were deposited in a pelagic environment near and above the CCD. Limestone deposition persisted throughout the basin until the end of Neocomian time when a sharp rise of the CCD to less than 2800 meters occurred. This event was coeval with the beginning of "black-clay" deposition and formation of euxinic basin conditions at least below the CCD. The stagnant deep basin probably was created by the formation of circum-Atlantic deep-circulation barriers coincident in time with tectonic readjustments in the North Atlantic and the initial opening of the South Atlantic. At the same time, rudist-coral reefs flourished along the continental margin as far north as Nova Scotia. Although the CCD remained shallow until the Maestrichtian, the deep basin became oxygenated in the late Cenomanian/Turonian, probably because of the introduction of deep and bottom water from the South Atlantic, and multicolored clays were deposited at true pelagic rates of 1-3 m/m.y. Little terrigenous detritus entered the deep basin because of the Cretaceous transgression and because of (now largely extinct) shelf-edge barrier reefs. Volcanic detritus from the then active New England Seamounts and from other background volcanism comprises a locally significant component of the multicolored clays. A sharp depression of the CCD to more than 5400 meters occurred in the late Maestrichtian to earliest Paleocene. This event resulted in basin-wide chalk deposition and may relate to the end-Cretaceous extinctions. Terrigenous detritus began to enter the basin in the early Paleocene, and locally deposited black clays indicate that deep-basin circulation was again sluggish. Highly siliceous sediments were deposited during the latest Paleocene to late Oligocene in response to developing abyssal circulation, upwelling, and enrichment of nutrients in surface water by volcanism. The latter played an especially significant role in late-early to early-middle Eocene rapid deposition of biogenic silica and development of "Horizon A" cherts. Siliceous turbidites flooded the basin during this interval, but the Bermuda Rise was isolated from the influx when it was uplifted during the middle to late Eocene. During the Oligocene to earliest Miocene, the abyssal circulation intensified because of climatic cooling and tectonically controlled introduction of bottom water from the Norwegian-Greenland seas; several hundred meters of sediment were eroded along the existing continental rise, forming a major unconformity. A sharp change to depositional conditions occurred in the early to middle Miocene, and rapid Neogene hemipelagic sedimentation controlled by moderately strong bottom currents is responsible for formation of the present continental rise and outer-ridge systems. The CCD has gradually deepened from about 4 km to more than 5 km during the Tertiary, but little carbonate is present in sediments other than on the Mid-Atlantic Ridge because of dilution by current-transported terrigenous debris.

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## INTRODUCTION

Leg 43 of the Deep Sea Drilling Project was designed to study three primary problems of the history of the western North Atlantic. First, drill sites were selected in areas where we anticipated that we could core previously unrecovered sections of the sedimentary record in order to document more completely the sedimentary evolution of the basin. These sites also were located specifically in areas where the basin's major seismic reflectors could be cored, dated, and correlated with sediment types. Determining whether these seismic markers were of the same age and indicated similar lithofacies throughout the basin was necessary to test the validity of reflector mapping as a tool for interpreting paleosedimentation patterns.

Secondly, a major objective of Leg 43 was to sample basalt from the prominent *J*-Anomaly basement ridge between the New England Seamounts and the Grand Banks. A very strong magnetic anomaly, locally greater than 1000 gammas, is associated with this feature, and geochemical and magnetic studies of the basalt would help establish whether unusual basement composition was responsible for the anomaly. The anomaly does not result from a geomagnetic field effect because a similar anomaly is not associated with Pacific crust of comparable age.

Finally, two sites were drilled along the New England Seamounts to determine the age and nature of volcanism that constructed this linear seamount chain. If volcanism had occurred as the North American plate moved north-westward over a stationary mantle hot-spot, progressively more youthful volcanism toward the southeast should be observed along the seamount chain. Conversely, volcanism which was essentially synchronous along the chain could indicate widespread activity along a tectonic (fault?) boundary within the plate.

The first two of these subjects are covered in this report. A companion paper (Vogt and Tucholke, this volume) explores the evolution of the New England Seamount Chain.

Six holes at six sites (382 to 387) were drilled on Leg 43 to meet the objectives outlined (Figures 1 and 2). Site

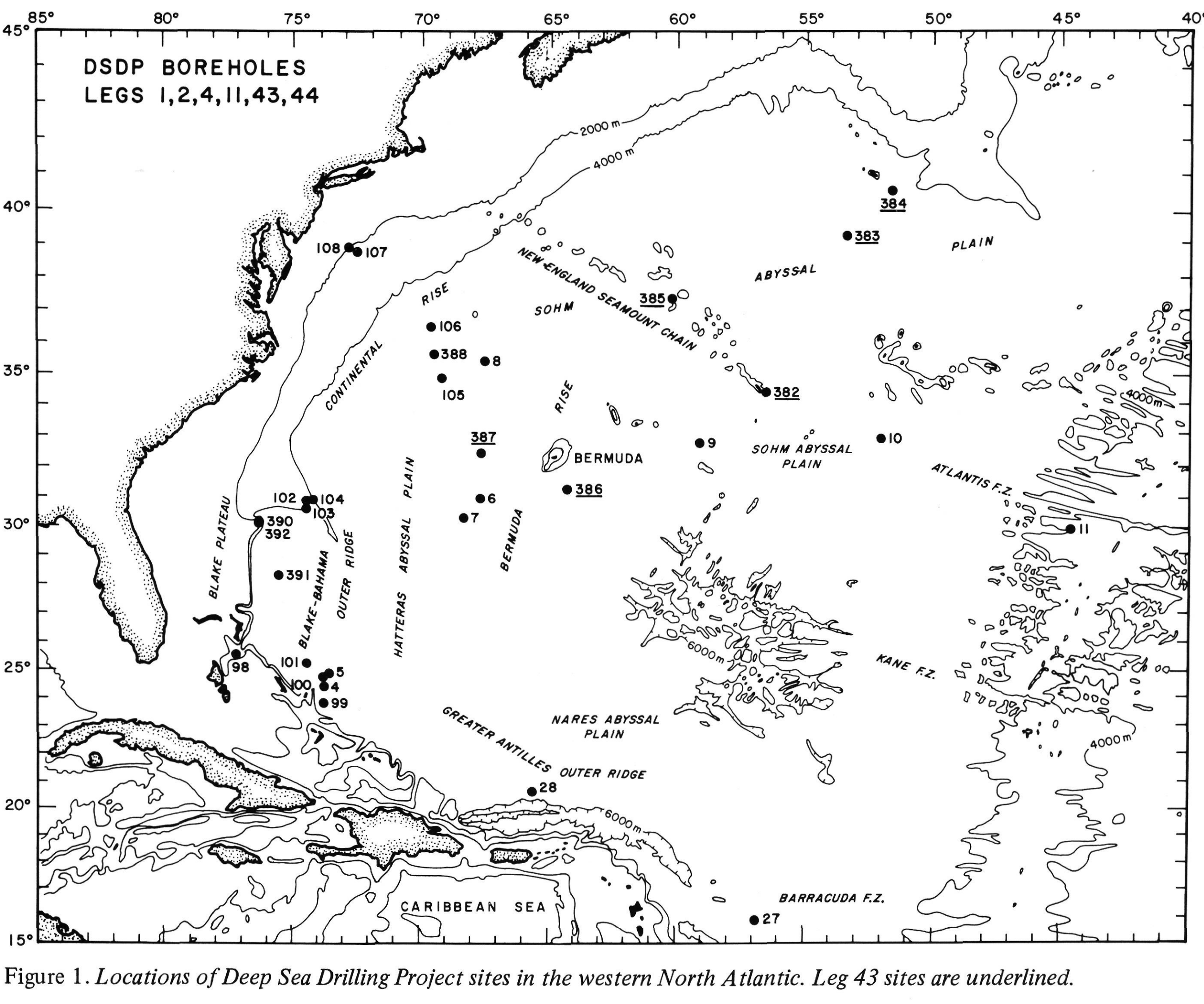


Figure 1. *Locations of Deep Sea Drilling Project sites in the western North Atlantic. Leg 43 sites are underlined.*

383, drilled into the Sohm Abyssal Plain above the *J*-Anomaly Ridge, was the only site which failed to meet our minimum objectives; the site was abandoned after only 120.3 meters of penetration when caving Pleistocene sands threatened to anchor the drill string.

In the following pages we summarize major aspects of the depositional and tectonic history of the North American Basin, based on Leg 43 results and integrated with data from DSDP Legs 1, 2, 4, and 11 in the western North Atlantic. The reader will find it useful to refer to the color foldouts at the back of this volume which summarize lithology versus age for western North Atlantic drill sites. Summaries of sediment compositions in this chapter are based on appropriate results reported in this volume and in Ewing, Worzel, et al. (1969); Petersen, Edgar, et al. (1970); and Hollister, Ewing, et al. (1972).

In our discussion we will refer to a summary diagram showing the history of the calcite compensation depth (CCD) in the western North Atlantic (Figure 3). Details on the construction of this figure are given in Appendix A. The CCD marks the level in the water column at which carbonate input from surface waters is balanced by dissolution in the undersaturated deep water. For the purpose of constructing Figure 3, we have followed the usage of van Andel (1975), placing the boundary between calcareous and non-calcareous pelagic sediments at 20 per cent $CaCO_3$; a boundary of 2 per cent $CaCO_3$ was used for hemipelagic deposits with major dilution by terrigenous sediment.

## UPPER JURASSIC LIMESTONES

Leg 43 sites did not penetrate Jurassic sediments, but a brief discussion of earlier results from Leg 11 is in order here. Figure 3 shows a compilation of lithofacies variations along sea-floor age-depth curves of western North Atlantic sites. The Oxfordian/Kimmeridgian reddish limestones generally contain more clay and less carbonate than overlying limestones; in fact the two Oxfordian cores above basalt at Site 105 have 20 per cent or less carbonate, indicating that the CCD was shallower than 3 km at this site. At Site 100, however, Oxfordian greenish gray limestones have higher carbonate content, and they suggest a deeper CCD near the Bahama Banks. Alternatively, it also is possible that Site 100 and Holes 4, 5, and 99A nearby have suffered anomalous subsidence because of their location near the Bahama Banks. If so, these sites accumulated carbonate above the CCD level depicted in Figure 3. Because the biostratigraphic control on ages at these sites is still imprecise, it is difficult to obtain any detail on Late Jurassic fluctuations of the CCD. However, in view of the apparently elevated Oxfordian CCD at Site 105, it would not be surprising to find that some pre-Oxfordian sediments in the North Atlantic are clay-rich and carbonate-poor. The subsequently deposited, carbonate-rich Tithonian limestones demonstrate depression of the CCD that continued into the Early Cretaceous.

## LOWER CRETACEOUS (NEOCOMIAN) LIMESTONES

### Description

Site 387 was the only Leg 43 drill site that penetrated Neocomian sediments; however, the Site 387 results are especially important because these sediments record Neocomian ridge-flank deposition, in contrast to previously drilled "basin-margin" holes (4, 5, 99A, 100, 101, 105).

At Site 387 the top of the limestones correlates with Horizon $\beta$ (Site 387 Report; Tucholke, this volume). Similar correlations occur at Sites 5, 101, and 105 where Horizon $\beta$ can be identified unambiguously (Ewing and Hollister, 1972). Horizon $\beta$ in the vicinity of Site 387 tends to drape over the irregular acoustic basement, and the seismic interval between the horizon and basement does not exhibit clear acoustic lamination. This is in marked contrast to the strong acoustic lamination in the $\beta$ to basement interval near Sites 5 and 101, but it is similar to the pre-$\beta$ acoustic section at Site 105.

The Neocomian limestones cored below Horizon $\beta$ generally have similar compositions and textures in all the western North Atlantic drillsites. At Site 387 there are two predominant limestone facies; one is a light colored, hard limestone which has high carbonate content (>90%) and is extensively bioturbated. This facies dominates the central part (Cores 40-47) of the $\beta$ to basement limestone sequence. At the top and bottom of the sequence (Cores 38-39 and 48-49) a second limestone facies is more common and consists of softer, darker, laminated limestones with lower carbonate content (<50-90%). Although these facies tend to be dominant in the intervals as described, the two facies are interbedded throughout the section. Overall core recovery in the limestones was low (<25%) at Site 387, but it generally was comparable to that encountered in the limestones at other sites. The noncarbonate fraction of the sediment at Site 387 contains mostly clay minerals, primarily illite and minor montmorillonite (Koch and Rothe, this volume). Traces of clinoptilolite and 1-10% quartz (probably authigenic) are present. Compared to Site 387, Neocomian limestones at other sites have an average composition with less illite (Holes 99A, 100, 101, 105) and more montmorillonite (Site 105); part of this difference probably results from different sampling densities in clayey versus limey interbeds at the various sites. The micritic calcite at all sites probably is derived from nannoplankton which are often well preserved in the clayey interbeds but are totally recrystallized in the hard limestones. Radiolarians, often pyritized, occur locally, and foraminifers are rare or absent. Organic carbon contents of Site 387 limestones tend to be high (up to 4.8%) in comparison to other sites (mostly <1%), but this may be due to the Site 387 shipboard sampling being somewhat biased toward darker, marly interbeds in the limestones. Quartzose chert is present throughout the $\beta$-to-basement sequence at Site 387, and it also is common in the limestones at Holes 99A and 100;

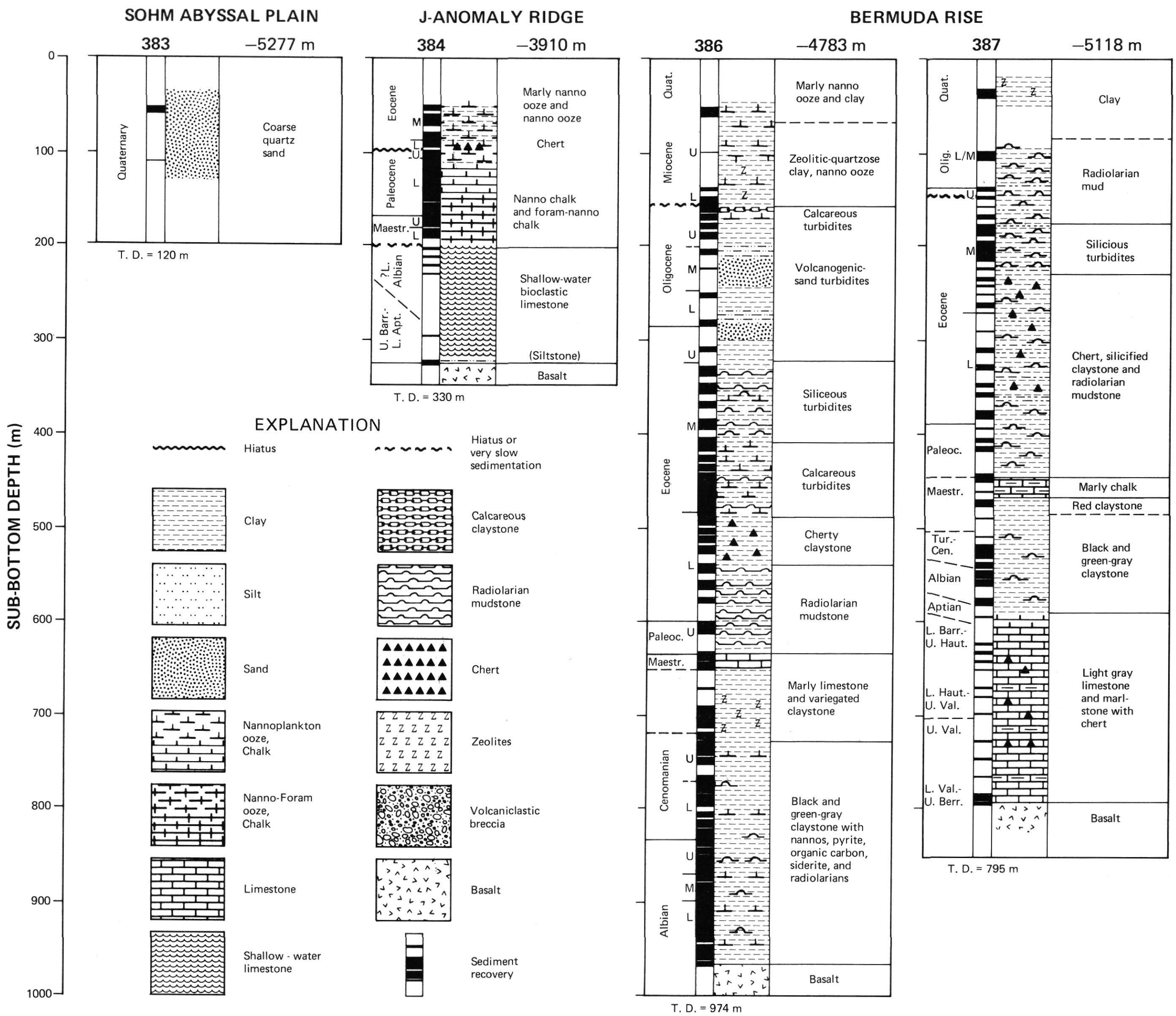


Figure 2. *Lithologic summary of Leg 43 drill sites.*

in contrast, limestones at Sites 101 and 105 rarely contain chert. Rates of accumulation for the Neocomian limestones at Site 387 are about 11 m/m.y. (2.4 kg/cm²-m.y. at an average dry bulk density of 2.16 g/cm³), which is slightly higher than the rate for the corresponding lithofacies at Holes 5, 99A, and 105 (Table 1).

## Depositional Environment

The depositional environment during the Neocomian probably was much the same throughout the southern part of the North American Basin. The CCD was deeper than 4 km (Figure 3), but the carbonate lysocline for foraminifers probably was shallower, thus accounting for the paucity of foraminifers in the limestones. Pelagic deposition on a quiescent, oxygenated sea floor persisted during the Berriasian, but beginning in the Valanginian there is evidence for increasingly common brief episodes of poorly oxygenated bottom water and consequent deposition of dark, carbon-rich, finely laminated, and burrow-free clayey interbeds in the limestones. At Site 387, cyclic variations in deposition are observed, consisting of (from the base): (a) dark, laminated, marly beds, (b) gray laminated limestones, and (c) light bioturbated limestones. One possible interpretation is that each cycle represents a progressive increase in deep-water oxygenation and surface productivity of nannoplankton. The sluggish circulation suggested for the deep water by the darker, carbonaceous interbeds may also have prevailed in surface waters, with reduced upwelling of nutrient-rich water, reduced productivity, and consequent deposition of marly rather than limey sediments. Alternatively, cycles of deep-water sluggish circulation and of surface water productivity may have acted independently, with the former cycle determining the preservation of any calcareous organisms reaching the sea floor.

There is little doubt that radiolarians were the main source of chert-forming silica at Holes 99A, 100, and

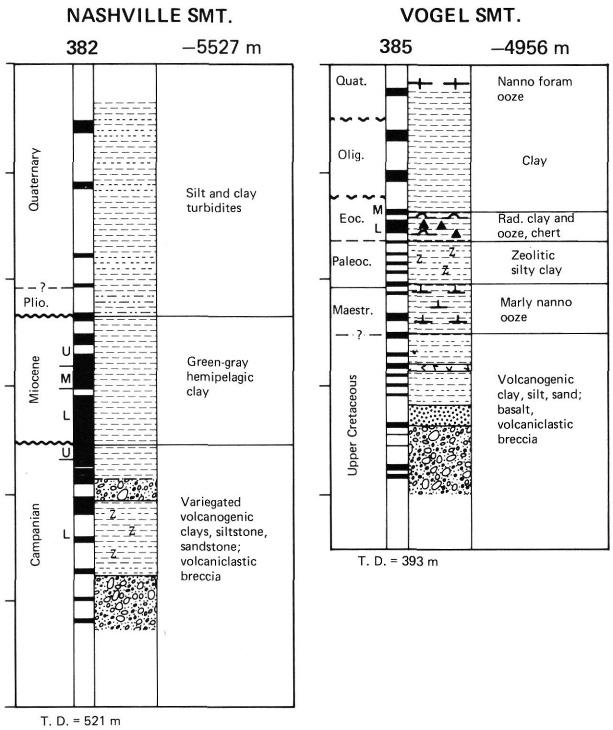


Figure 2. *(Continued).*

387, and that siliceous organisms were more common in the original sediments at these sites than is now observed. This observation, considered with the scarcity of radiolarians nearer the continental margin (Site 105), suggests that siliceous productivity may have been restricted to a zone extending from the Bahama Banks into the central basin of the Atlantic. Such a productivity belt could be expected along the path of presumed circumglobal flow through the Tethys, Atlantic-Caribbean, and Pacific in the Early Cretaceous, when the easterlies wind belt extended perhaps 10° farther north than it does today (Berggren and Hollister, 1974; Luyendyk et al., 1972).

Increasingly frequent episodes of bottom-water stagnation near the end of the Neocomian are indicated by the abundance of dark marly interbeds near the top of the Horizon $\beta$ to basement limestone sequence. This process culminated during the Barremian in a probably sharp rise of the CCD (Figure 3) and resultant deposition of black clays. Horizon $\beta$ marks this change in lithofacies, the reflector arising from the impedance contrast at the top of the shallowest high-velocity carbonates underlying the black clays. Possible causes for the change in depositional regime are discussed in the following section.

## BARREMIAN TO CENOMANIAN BLACK CLAYS

### Description

Sites 386 and 387 both penetrated Cretaceous black clays. The 240-meter black-clay section at Site 386 is exceptionally thick because the site was drilled within a fracture valley that accumulated pelagic sediment rapidly shed from adjacent bathymetric highs. The only other locations with such thick sections are Site 101 (193-296 m) and Site 391 (about 275 m; Benson, Sheridan et al., 1976) near the continental margin; at Site 101 the top of the black clays is truncated by an erosional unconformity (Horizon $A^U$, see Tucholke, this volume), so the original thickness must have been even greater. Other sites (5, 105, and 387) all have a black-clay thickness of about 100 meters.

Site 386 is also unusual in that the black clays are deposited directly on lower Albian basaltic basement. Although this fracture-valley basement was below the mean ridge-crest elevation during Albian time, it shows that the black clay depositional field extended to depths at least as shallow as 3200 meters (Figure 3). Some of the adjacent peaks which shed calcareous debris into the valley were above the CCD and probably above the level of anoxic bottom water. The shallowest of these peaks, a few kilometers south of Site 386, rises 1800 meters above the fracture valley basement and had a ridge-crest paleodepth approximately 1400 meters below sealevel.

TABLE 1
Thickness and Approximate Accumulation Rates[a] of Some Dominant Lithofacies in the North American Basin

| Site | 101 | 100 | 5 | 105 | 99A | 8 | 387 | 7 | 6 | 385 | 386 | 9 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Hemipelagic clay | >203m (20m/m.y.) | – | – | >193m (19m/m.y.) | – | >177m (~13m/m.y.) | 41-98m (?2m/m.y.) | – | 161-190m (?4m/m.y.) | >108 (4-5m/m.y.) | 148m (9m/m.y.) | – |
| Eocene siliceous turbidites | – | – | – | – | – | >56m (?) | >210m (~20m/m.y.) | ⩾10m | >66m (?15-20m/m.y.) | – | 285-314 m[b] (~27m/m.y.) | – |
| Maestrichtian chalks/ limestone | – | – | – | – | – | – | 25.6m | – | – | 45m | 2.5m | – |
| Multicolored clays | – | – | – | 49-97m (??2m/m.y.) | – | – | 9.5-19.1m (<1m/m.y.) | >18m | – | – | 90-109m (~3m/m.y.) | >75m (~3m/m.y.) |
| Black clay | 293-340m (~12m/m.y.) | – | 100-105m (~3m/m.y.) | ~110m (~5m/m.y.) | – | – | 95-114m (4-5m/m.y.) | – | – | – | 240m (~16m/m.y.) | – |
| White and gray limestone | >92m (?18m/m.y.) | >9m (?) | >93m (~7m/m.y.) | ~139m (~10m/m.y.) | 158-181m (~7m/m.y.) | – | 199-208m (10-11m/m.y.) | – | – | – | – | – |
| Red clayey limestone | – | 39-64m (~8m/m.y.) | – | 75m (~10m/m.y.) | >13m | – | – | – | – | – | – | – |
| Homogeneous green-gray limestone | – | 41m (~8m/m.y.) | – | – | – | – | – | – | – | – | – | – |

[a]In parentheses.
[b]Includes calcareous turbidites.

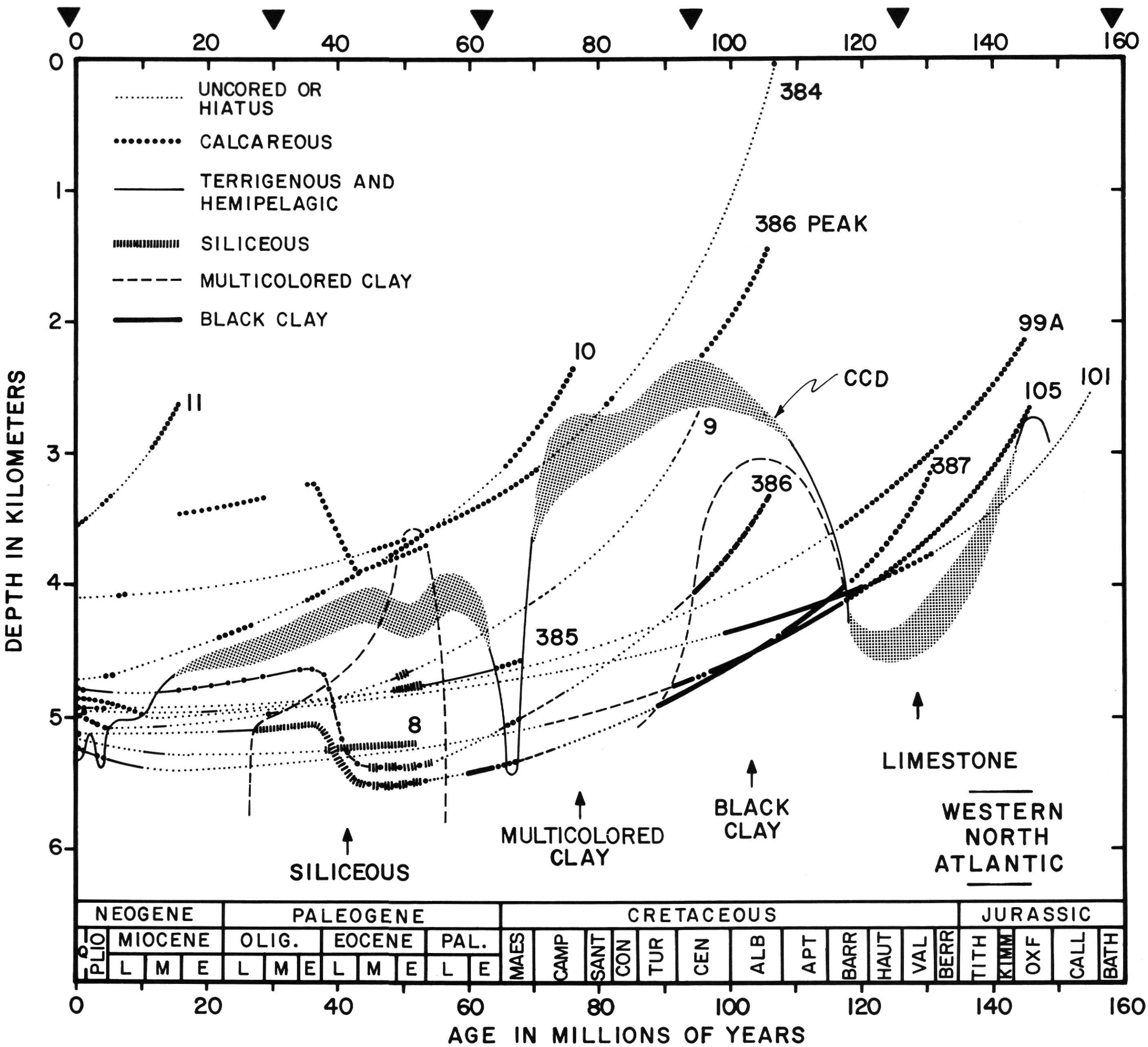


Figure 3. *Tracks of sea-floor depth versus time for western North Atlantic drill sites showing variations in dominant lithofacies along the tracks. The thin solid line shows fluctuations in the calcite compensation depth (shaded where poorly controlled). Depositional fields in time and space for highly siliceous sediments, multicolored clays, and black clays are indicated (outlined by light dashed lines). Derivation of the age/depth tracks is given in the Appendix. Filled triangles at top show oligotaxic (cool) episodes of Fischer and Arthur (1977) which alternate with polytaxic (warm) intervals. Time scales after Berggren (1972) and van Hinte (1976a, b).*

Black clays in the western North Atlantic drill sites all have broad compositional similarities, although there are local compositional peculiarities as well. All sites (101, 105, 386, 387) contain an average of 50 per cent or more clay minerals consisting primarily of montmorillonite and illite. At Sites 101 and 105 illite tends to dominate over montmorillonite, but this trend is reversed at Site 387, and at Site 386 there is a downhole change from dominantly montmorillonitic to dominantly illitic clays (Koch and Rothe, this volume; Zemmels et al., 1972). The downhole enrichment of illite at Site 386 may result from conversion of smectite (Eberl and Hower, 1976). Site 101 also consistently contains several per cent of chlorite and kaolinite, but these minerals are rare or absent at other sites. Feldspars in quantities of 2 to 4 per cent are found at all sites, with somewhat higher percentages at Site 101. Site 105 is unusual in containing abundant clinoptilolite, probably as an alteration product of volcanic ash (Zemmels et al., 1972), whereas this zeolite is rare at other sites. Finely disseminated pyrite appears to produce the black coloration in sediments at all these sites and large pyrite concretions are locally common.

Abundance of biogenic components in the black clays is highly variable between sites, although none of the sites contain foraminifers in other than trace quantities. In a reconstruction of sea-floor paleodepths (Figure 3), Site 386 had the shallowest sea floor (about 3200 m) during the period of black clay deposition. Both radiolarians and nannoplankton are abundant at this site, probably because of the combined effects of local pelagic turbidites from adjacent peaks and of the shallow sea floor (although still below the CCD). Site 387, which had the deepest sea floor (up to 4900 m), accumulated almost no calcareous nannoplankton, but radiolarians are common and are disseminated throughout the sediment; this contrasts with Site 386 where radiolarians are mostly concentrated in radiolarian-rich sand layers (McCave, this volume). At Site 101 radiolarians and nannoplankton are rare, and at Site 105 radiolarians are rare and nannoplankton common. Thus, while the calcareous microfossil distribution shows a general relationship to the depth of the depositional surface, the siliceous microfossils do not. In fact, a zone of siliceous productivity associated with a circum-global current across the central basin may have persisted from the time of deposition of the underlying limestones.

The abundance of other mineral components appears to have a direct relationship to the distribution of the biogenic components. Quartz averages 10 to 20 per cent at Sites 101 and 105 where radiolarians are uncommon, but it is two to three times as abundant in the radiolarian-bearing clays at Sites 386 and 387. Diagenetic transformation of the opaline silica probably accounts for a substantial portion of this quartz (see Riech and von Rad, this volume). Carbonates such as siderite and rhodochrosite also tend to be more abundant in the black clays at Sites 105 and 386 where calcareous biogenic material is found; these minerals are uncommon in the mostly non-calcareous section at Site 387.

The "black-clay" facies is characterized by dark greenish gray and black colors, often alternating in a cyclic manner. At Sites 386 and 387, the truly black sediments account for less than half the section; the proportion may be higher at Sites 101 and 105, although these sections were sampled more sparsely and the quantity of black sediments is more difficult to document. The black clays are finely laminated and very rarely exhibit burrow mottling, while the interbedded gray clays vary from laminated to, more commonly, strongly burrow-mottled. Contacts between the two colors range from sharp to gradational or mottled.

Organic carbon content of the black clays at all sites varies widely, usually ranging from less than 1 per cent up to 4 per cent (Figure 4). The organic matter is of two types: amorphous kerogen probably derived from marine plankton, and hydrogen-poor, vitrinitic organic matter of terrigenous origin (Kendrick, this volume). Curiously, although terrigenous vitrinitic matter is found at both Sites 386 and 387, sporomorphs are completely lacking at Site 386 (D. Habib, personal communication, 1976); the cause is presumed to be differential sorting of the organic material during its airborne and waterborne transport to the sites. At the three sites which have recovered the upper Cenomanian "top" of the black clay facies (Sites 105, 386, and 387) a prominent peak of 12 to 15 per cent organic carbon is observed, and most of this organic matter appears to be derived from marine plankton. It is also important to note that variations in organic carbon content are relatively independent of sediment color and type of sedimentary structures (Kendrick, this volume). One exception is that the bioturbated, green-gray sediments have uniformly low organic carbon contents. Thus, sediment blackness is not an index to the abundance of organic matter.

Typical accumulation rates for the black clay facies are about 3-5 m/m.y. (Sites 5, 105, 387); rates are much higher at Site 386 (~16 m/m.y.) because of the site location in a fracture valley, and at Site 101 (~12 m/m.y.), which is near the continental margin and which according to seismic profiles has an unusually thick black-clay section.

### Depositional Environment

Assuming that representative material was obtained in drill cores, we can compare the *non-carbonate* mass accumulation rates of the Neocomian limestones with the *non-carbonate* rates of the overlying black clays at Sites 105 and 387, where normal sequences of both facies are well documented. The rates in the two facies are essentially identical, about 0.65 kg/cm$^2$-m.y. at Site 387 and about 0.7 kg/cm$^2$-m.y. at Site 105; the calculated rates are based on average dry bulk densities of 1.52g/cm$^3$ for black claystones and 2.16 g/cm$^3$ for limestones (see Site 387 Report). In addition, organic carbon content and organic carbon mass accumulation rates, on a carbonate-free basis, actually *decrease* after deposition of the limestones and during deposition of the black clays (Figure 4). It could be argued that the high organic carbon contents in the carbonate-free component of the limestones are an artifact of short intervals of very low input of inorganic debris superimposed on a background of more or less steady carbon input. However, the same argument would probably apply to the black clays, because the similarity of non-carbonate accumulation rates for the limestones and black clay suggests that the carbonate-free fraction of both facies accumulated by similar mechanisms. In addition, because the black clays were deposited in an anoxic to poorly oxygenated environment (see discussion below), the preservation of organic carbon should be enhanced in the black clays in comparison with the limestones. Thus the decrease in organic carbon input to the sediment during deposition of the black clays probably is real rather than an artifact of inorganic accumulation rates or of preservation. The question also arises whether the higher carbon input to the limestones could be primarily "terregenous" carbon transported to the deep basin while sea level was lower in the Early Cretaceous. At least at Site 387, however, similar C/N

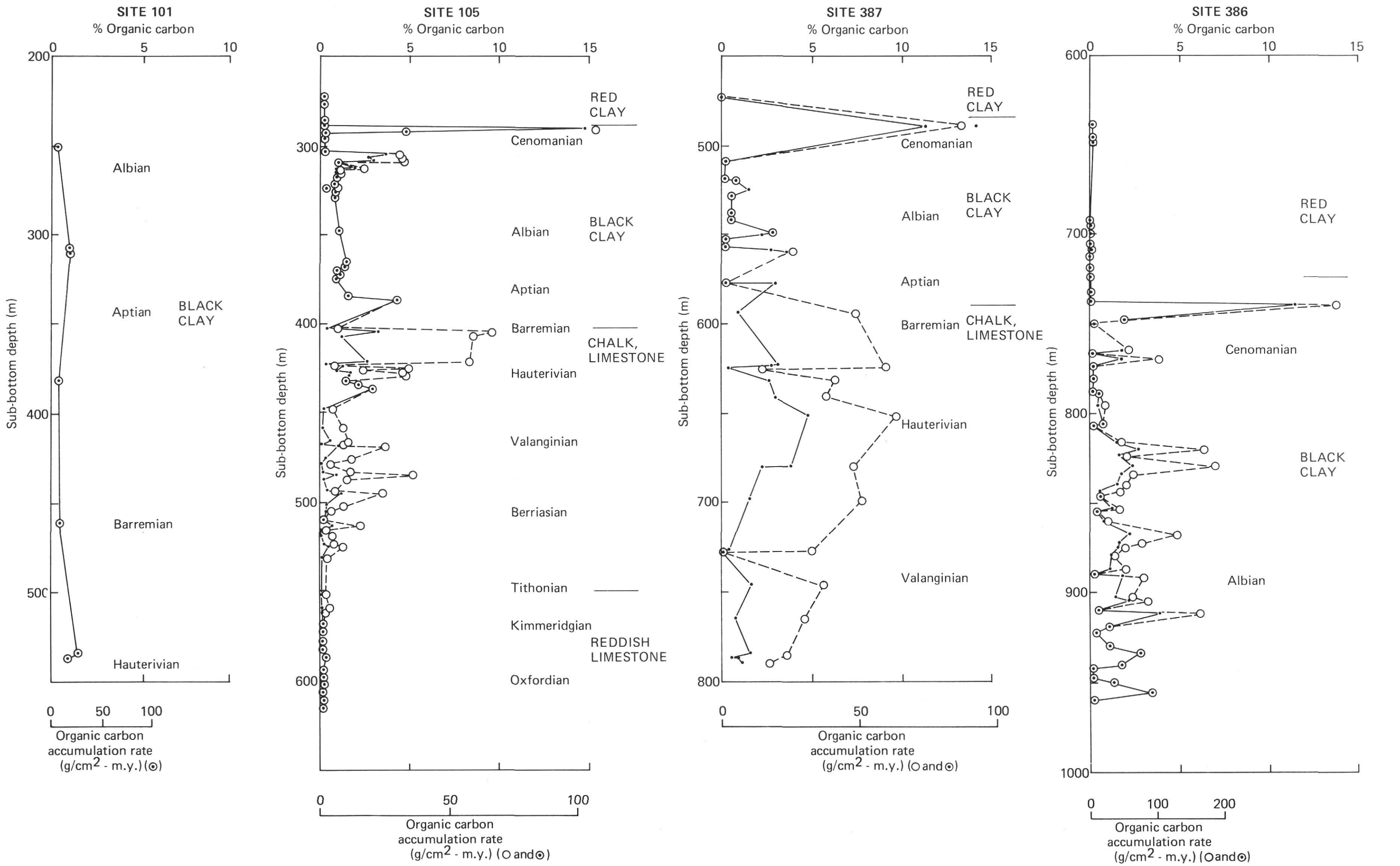


Figure 4. *Organic carbon content of Mesozoic sediments at western North Atlantic drill sites. Carbon as per cent of total sediment is shown by dots connected by solid lines; carbon as per cent of carbonate-free sediment is shown by open circles. Approximate carbon accumulation rates are indicated by the lower scale and calculated from the following: black-clay dry bulk density 1.52 g/cc (Sites 101, 105, 387) and 1.66 g/cc (Site 386); limestone dry bulk density 2.16 g/cc; average carbonate content in black clays 2 per cent at Site 387 and 8 per cent at Site 105; average carbonate content of limestones 72 per cent at Site 387 and 66 per cent at Site 105; linear accumulation rates from Table 1. These parameters were derived from the physical properties, carbon-carbonate, and age data reported in this volume and in Hollister, Ewing, et al. (1972).*

ratios in the limestones and black clays suggest compositionally similar carbon in both facies. Aside from the changes in input of carbon and carbonate, sediment sources for the limestones and black clays appear to be similar, particularly at Sites 101 and 105 where the composition of the clay mineral fraction is nearly identical for both facies.

The sea-floor environment within which the black clays were deposited can be assessed on the basis of textural and compositional parameters of the sediments. The presence of black, laminated, and unburrowed sediments indicates that the sea floor was sporadically anoxic; interbedded dark green-gray clays which are burrowed probably represent an environment sufficiently oxygenated to support benthic life but not so well oxygenated that reducing conditions within the sediment were totally dissipated. The lack of direct correlation between sediment blackness and organic carbon content demonstrates that the reducing conditions were not produced solely by organic carbon influx. Rather, it suggests that two independent cycles may have been operative, one an oxygenation cycle and the other a cycle of carbon influx (Figure 5). Note that in this model we can assume that the cycle of carbon influx represents long-period productivity patterns ("long blooms," McCave, this volume), and the state of preservation of biogenic debris will vary, depending upon the geochemical environment resulting from the interaction of the carbon and oxygen cycles in the bottom water.

An important question is whether the reducing cycles (black interbeds) were produced beneath anoxic bottom water or resulted from intrasediment reducing conditions beneath oxygenated bottom water. The lack of burrow mottling in most black sediments indicates anoxic bottom water, although rare black beds which are burrowed may have been produced by only intrasediment reduction of sulfate to form pyrite. Kendrick (this volume) presents evidence that more sulfur is partitioned into organic matter and pyrite within the black clay facies than can be accounted for purely by reduction of sulfate present in the interstitial water. The excess sulfur therefore is thought to be derived from reduction by anaerobic bacteria at the seafloor of sulfate contained in the bottom water. This observation strengthens our conclusion that the bottom water was anoxic.

Dean et al. (1978) argued on the basis of DSDP Leg 41 data that the anoxic bottom water might be only a thin layer created by a rapid influx of organic carbon. However, the lack of correlation between anoxic (black sediment) episodes and organic carbon content in western Atlantic sites shows that anoxic bottom water probably was produced independently of the carbon cycle. Furthermore, the decrease in mass accumulation rates of organic carbon at the onset of black-clay deposition (Figure 4) appears to contradict this hypothesis. Other workers (Fischer and Arthur, 1977; Thiede and van Andel, 1977) have invoked the concept of an expanding oxygen-minimum layer to account for the anoxic basin conditions. This concept, perhaps valid in the Late Cretaceous South Atlantic (Figure 6), also may be reasonable for thin carbonaceous facies drilled on elevated areas such as the Shatsky Rise in the Pacific (Ryan and Cita, 1977). However, we find no evidence for deep-basin oxygenated water lying beneath a zone of anoxic water in the western North Atlantic (Figure 3), and the entire basin below 3200 meters appears to have suffered episodic anoxic conditions during deposition of the black clays.

Fischer and Arthur (1977) have presented evidence for long-term (32-m.y. period) variations in diversity of pelagic biota and character of oceanic circulation. Their "oligotaxic" episodes (−30 m.y., −62 m.y., −94 m.y., etc.) are times of lowered marine temperatures, increased latitudinal and vertical temperature gradients, absence of anaerobic marine deposition, and lowered diversity in pelagic organisms. Intervening "polytaxic" episodes have maximum diversity, warmer and more uniform oceanic temperatures, and widespread marine anaerobic conditions. Fischer and Arthur (1977) have correlated the widespread Aptian-Albian black clays (Figure 6) with a polytaxic episode of global proportions. Because of the warmer, more uniform temperatures during this polytaxic phase, we anticipate reduced upwelling and reduced productivity. This effect may be demonstrated in the organic carbon accumulation rates at sites in the western North Atlantic (Figure 4). At these sites, carbon accumulation is minimal during the Aptian-Albian (110 m.y.) and the Late Jurassic (142 m.y.) polytaxic episodes and is greatest during the intervening Early Cretaceous oligotaxic phase which had presumably cooler climates, stronger temperature gradients, and increased upwelling.

In itself, the polytaxic cycle concept described above fails to explain the highly variable stratigraphic extent of the black clays in various ocean basins (Figure 6), and it is probable that tectonic events which have controlled circulation patterns in the basins are primarily responsible for the stratigraphic distribution of black clays.

In the instance of the North Atlantic, the dramatic change from the limestone to the black clay facies near the end of the Neocomian can be explained by the construction of barriers to intermediate and deep circulation in the Tethys and/or in the Caribbean (Figure 7). Unfortunately, little is known yet about potential Cretaceous barriers in these two areas, and specific reconstructions are impossible. However, these areas provided the only passages for renewal of Atlantic deep water in the Early to Middle Cretaceous, and constriction of one or both of the passages could result in the production of stagnant bottom water. Coincidentally, there is an apparent spreading-rate change in the North Atlantic at the end of the Neocomian, with spreading half rates averaging about 0.5 cm/yr during anomalies *M*10-*M*4, compared to half rates of 1.0 to 1.5 cm/yr during the remainder of the *M*-series (Schouten et al., 1976). This slowdown is thought to have been accompanied by compression in the Caribbean (K. Klitgord, personal communication, 1977) which could have caused uplift of circulation barriers. However, there is still enough uncertainty in the magnetic time scale for the *M*-series anomalies that this slowdown interpretation

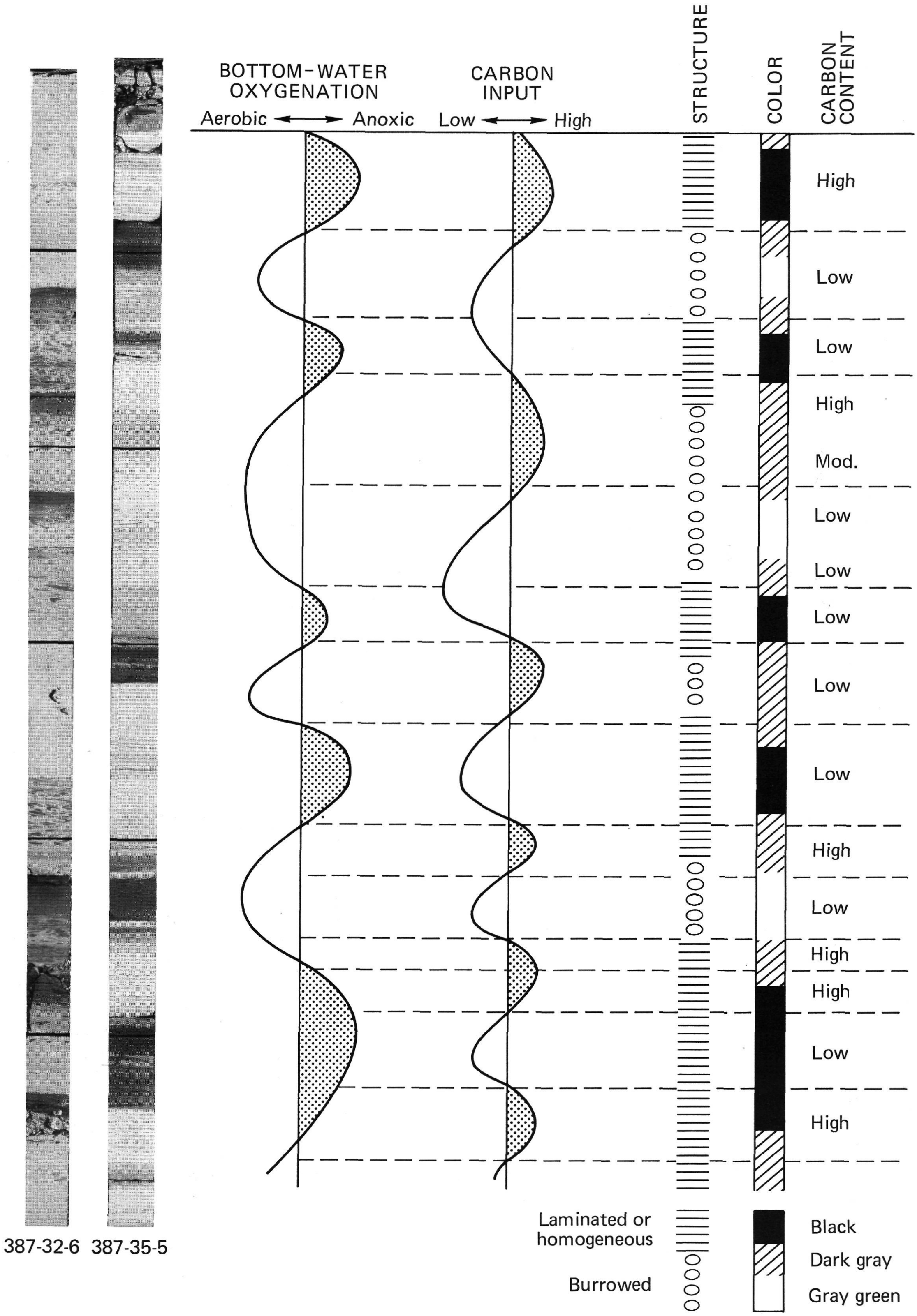


Figure 5. *From left – photographs of sections of two cores from the black-clay facies at Site 387 showing black layers and sedimentary structures, schematic representation of independent cycles of carbon input and bottom-water oxygenation (center), and the resulting sedimentary record (right). Note the assumption that organic carbon is* preserved *only beneath weakly oxygenated to anoxic bottom water.*

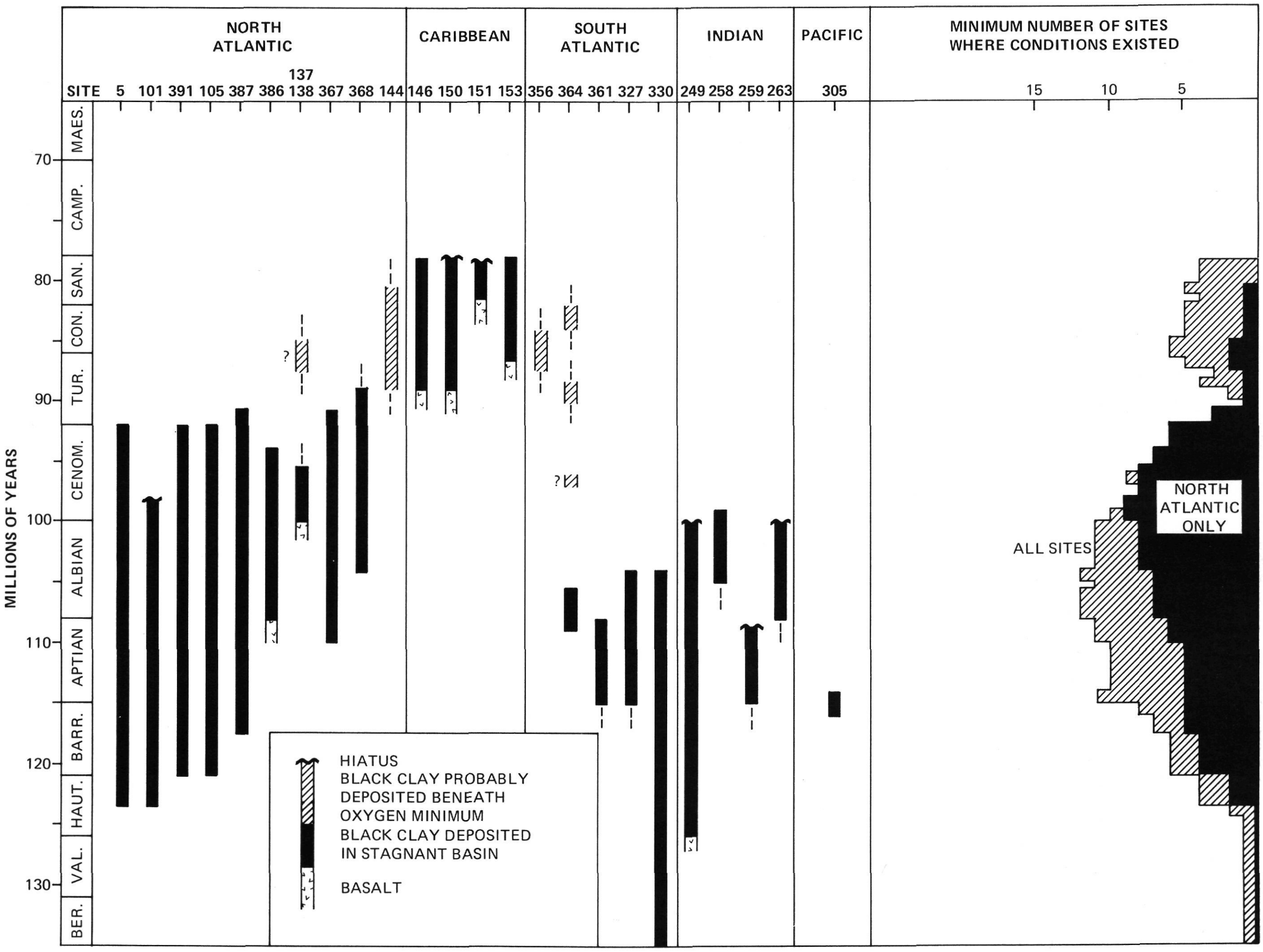


Figure 6. *Summary of drill sites which have recovered "black-clay" facies. Black bars – deposition in anoxic basin. Shaded bars – deposition probably in oxygen-minimum zone at intermediate depths (question marks where uncertain) (Thiede and van Andel, 1977). Dashes indicate uncertain limits on extent of black clays.*

can be debated. For example, Vogt and Einwich (this volume, their figure 11) suggest constant spreading rates throughout the *M*-series.

Another possible deep circulation barrier is the *J*-Anomaly Ridge which formed as an anomalously shallow segment of the Mid-Atlantic Ridge between *M*-2 and *M*-0 (Figures 8-10) (Vogt et al., 1971). However, the strong relief at the northern end of the ridge (profile 1, Figure 9) is greatly subdued south of about 40°N, and it is questionable whether this feature had any major effect in producing deep-basin stagnation.

The widespread distribution of Tethyan biota throughout the Middle Cretaceous seaway (Berggren and Hollister, 1974) indicates that the barriers discussed above did not block circum-global surface water exchange between the Atlantic, Pacific, and Indian oceans. In fact, upwelling and productivity associated with "equatorial" flow across the Atlantic could help explain the increased rates of carbon accumulation away from the western Atlantic margin and near the center of the basin at Site 386 (Figure 4). This flow is presumed to be a weaker version of the earlier Neocomian circulation that may have been responsible for central-basin productivity of siliceous organisms. The intensity of the circum-global flow may have been the controlling factor in the oxygenation of the deep water, the degree of upwelling, and the surface water productivity; however, these factors need not have varied in direct relationship to one another. For example, moderate flow could have produced upwelling that resulted in production and deposition of carbon-rich sediments in an anoxic deep basin, while more intense flow could have caused deposition of organic-rich sediments on an oxygenated or mildly anoxic sea floor. These kinds of variations may help explain why sediment blackness does not always correlate with the abundance of organic matter (Figure 5).

Termination of black clay deposition in the North Atlantic is represented by a geologically rapid transition to accumulation of red (multicolored) clays near the end of the Cenomanian. The rejuvenated deep circulation could have been caused by any one of several potential new bottom-water sources. Exchange of deep water with the South Atlantic is one of the most likely causes.

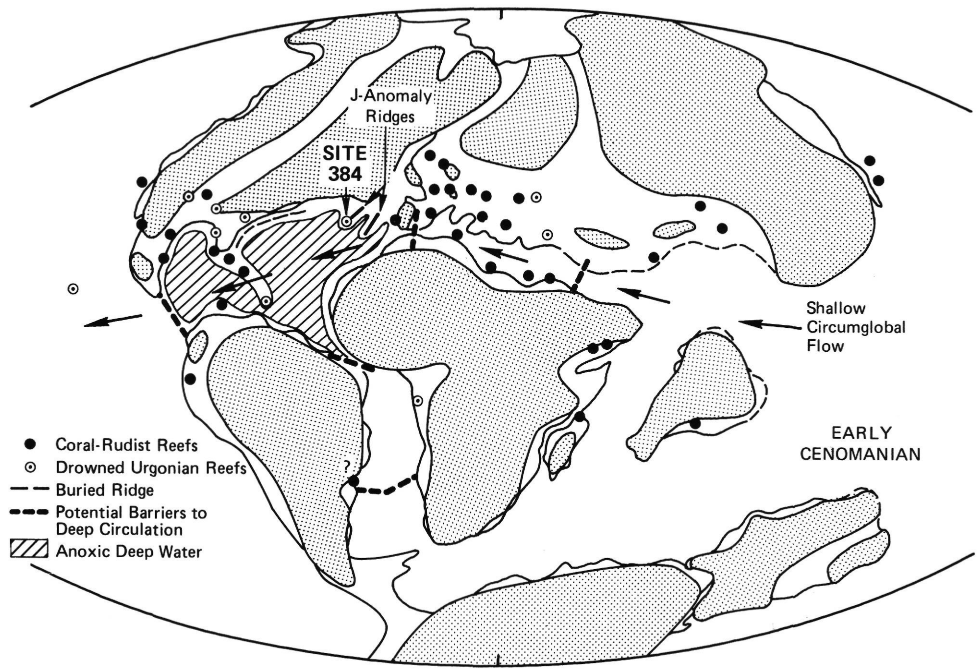


Figure 7. *Early Cenomanian reconstruction showing distribution of coral-rudist reefs and extent of deep-water anaerobic conditions. Modified from Douglas et al. (1973).*

South Atlantic boreholes indicate that black clays were deposited in an anoxic deep basin there during the Early Cretaceous but that anoxic-basin conditions ceased in the mid-Albian (Figures 6 and 7). The Falkland Plateau and South Africa separated about this time (Rabinowitz and LaBrecque, in preparation), probably allowing exchange of deep water between the South Atlantic and the Southern Ocean. However, the North and South Atlantic were still separated by the juncture between the north Brazilian margin and the opposing African coast (Figure 7). Oddly enough, it appears that the Rio Grande Rise/Walvis Ridge system did not prevent oxygenation of the deep basin in the northern South Atlantic. Surface circulation between the North and South Atlantic was established sometime between late Albian and early Turonian (Reyment, 1969; Reyment and Tait, 1972; Kennedy and Cooper, 1975). Compression between Africa and South America along the north Brazilian margin changed to extension at the end of the Aptian (Rabinowitz and LaBrecque, in preparation), and moderately deep passages allowing exchange of bottom waters between the North and South Atlantic probably developed by the end of the Cenomanian. It is noteworthy that black-clay deposition recurred at Sites 356, 363, and 364 in the South Atlantic and at Site 144 in the central Atlantic during the Turonian-Santonian (Figure 6). Thiede and van Andel (1977) have suggested that these sediments were deposited beneath waters with a true oxygen minimum shallower than about 2500 meters and that the deeper sea floor was oxygenated.

Other possible sources for oxygenated deep water in the North Atlantic at the end of the Cenomanian could have been provided if Tethyan and/or Caribbean circulation barriers were breached. The presence of Turonian-Santonian black clays in the Caribbean (Figure 6) seems to preclude deep-water connections to the Pacific, although these black clays could have been deposited beneath an oxygen minimum like that in the Late Cretaceous central and South Atlantic. In the northern North Atlantic, rifting was initiated between Labrador and the Iberian Peninsula in the Early Cretaceous (Jansa and Wade, 1975), but it is unclear when, if at all, this rifting opened a passage to a northern source for Cretaceous bottom-water renewal.

The late Cenomanian peak in organic carbon accumulation (Figure 4) may be a consequence of the improved circulation in the North Atlantic. Upwelling of nutrient-rich deep water could have sustained a very long and intense bloom of pelagic organisms. Carbon-isotope data from European chalk facies also substantiate this episode of high productivity (M. Arthur, personal communication, 1976; Scholle and Arthur, 1976).

## ORIGIN AND AGE OF THE *J*-ANOMALY RIDGE

The basement structural feature termed the *J*-Anomaly Ridge (Figure 9) formed in crust of anomaly age *M*2-*M*0 near the time when black clays were initially deposited in the North Atlantic. This asymmetrical ridge generally can be identified between the Grand Banks and the New England Seamounts. It is associated with the high amplitude (up to 1000$\gamma$) "*J*-Anomaly" in this region and in the conjugate area of the eastern Atlantic (Rabinowitz et al., this volume); the *M*-2 to *M*-0 magnetic anomaly complex exhibits much lower amplitudes to the south across the Bermuda Rise, and there is no prominent basement ridge in this region. A major change in relief of the ridge occurs near 40°N where the basement rises abruptly northward about 1200 meters and emerges above the Sohm Abyssal Plain

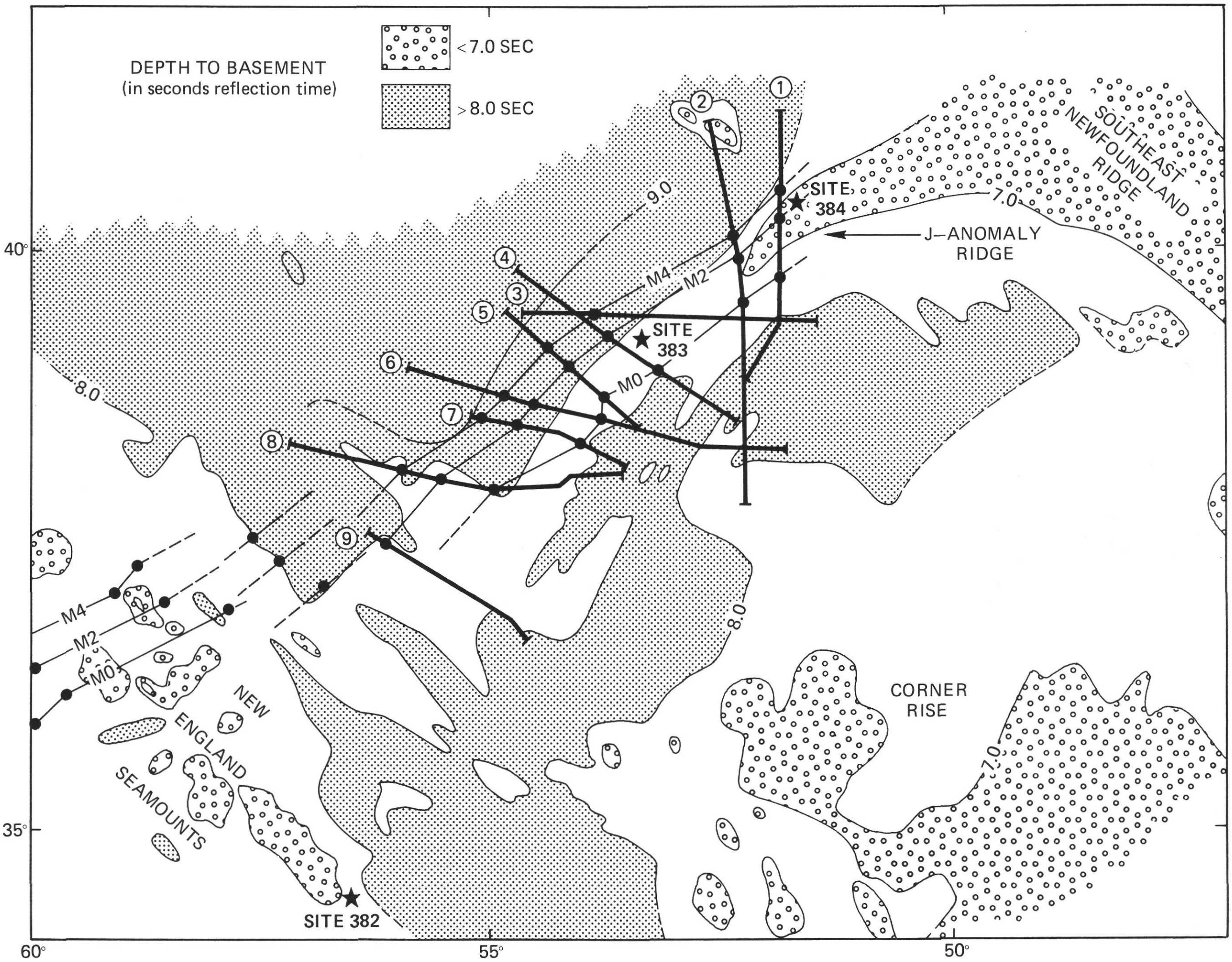


Figure 8. *Depth to basement in the vicinity of the* J-*Anomaly Ridge (in sec two-way travel time below sea level). Light lines and filled circles show the positions of Keathley anomalies* M0, M2, *and* M4, *and heavy numbered lines show position of profiles in Figure 9.*

and continental rise. South of 40°N the feature is described better as a basement step rather than a ridge.

Drilling on the *J*-Anomaly Ridge was intended to sample the igneous basement in order to investigate the cause of the exceptionally high amplitude magnetic anomaly and to determine whether the ridge is compositionally anomalous, perhaps like elevated ocean crust around Iceland and the Azores. The unusual amplitude of the anomaly cannot be a geomagnetic field effect because similar amplitudes are not observed at the young end of the *M*-series in the Pacific. Site 383, drilled over the ridge crest just on the young side of anomaly *M*-1 (Figure 8), failed to reach basement. Site 384 was drilled near anomaly *M*-2 on the ridge crest farther north where it emerges above the Sohm Abyssal Plain. The limited profiler and magnetic data available in this area suggest that the *J*-Anomaly Ridge may not be strictly parallel to sea-floor isochrons but is formed in slightly younger crust toward the south (Figure 9). Similar time-transgressive ridges have been suggested on the Reykjanes Ridge southwest of Iceland (Vogt, 1974).

Post-cruise modeling of the magnetic source of the *J*-Anomaly also suggests that anomalous magnetization may be centered about anomaly *M*-0 (Rabinowitz et al., this volume). Thus the drilling results near anomaly *M*-2 at Site 384 in themselves do not definitively explain the origin of the *J*-Anomaly.

**Site 384 Magnetism and Petrochemistry**

One of the objectives in drilling the *J*-Anomaly Ridge was to determine if the basalt composition and magnetic properties were anomalous, such that they would explain the existence of the high-amplitude magnetic anomaly. Another objective was to compare the basalt with LIL-enriched basalt associated with other morphologically similar features which may have resulted in part from a "primary mantle plume" chemistry (Schilling, 1973; Vogt, 1974). Although the zone of intense (or thickened) magnetization is centered somewhat east of the ridge crest (Rabinowitz et al., this volume), the magnetic anomaly and the basement ridge no doubt are genetically associated, and there was good reason to ex-

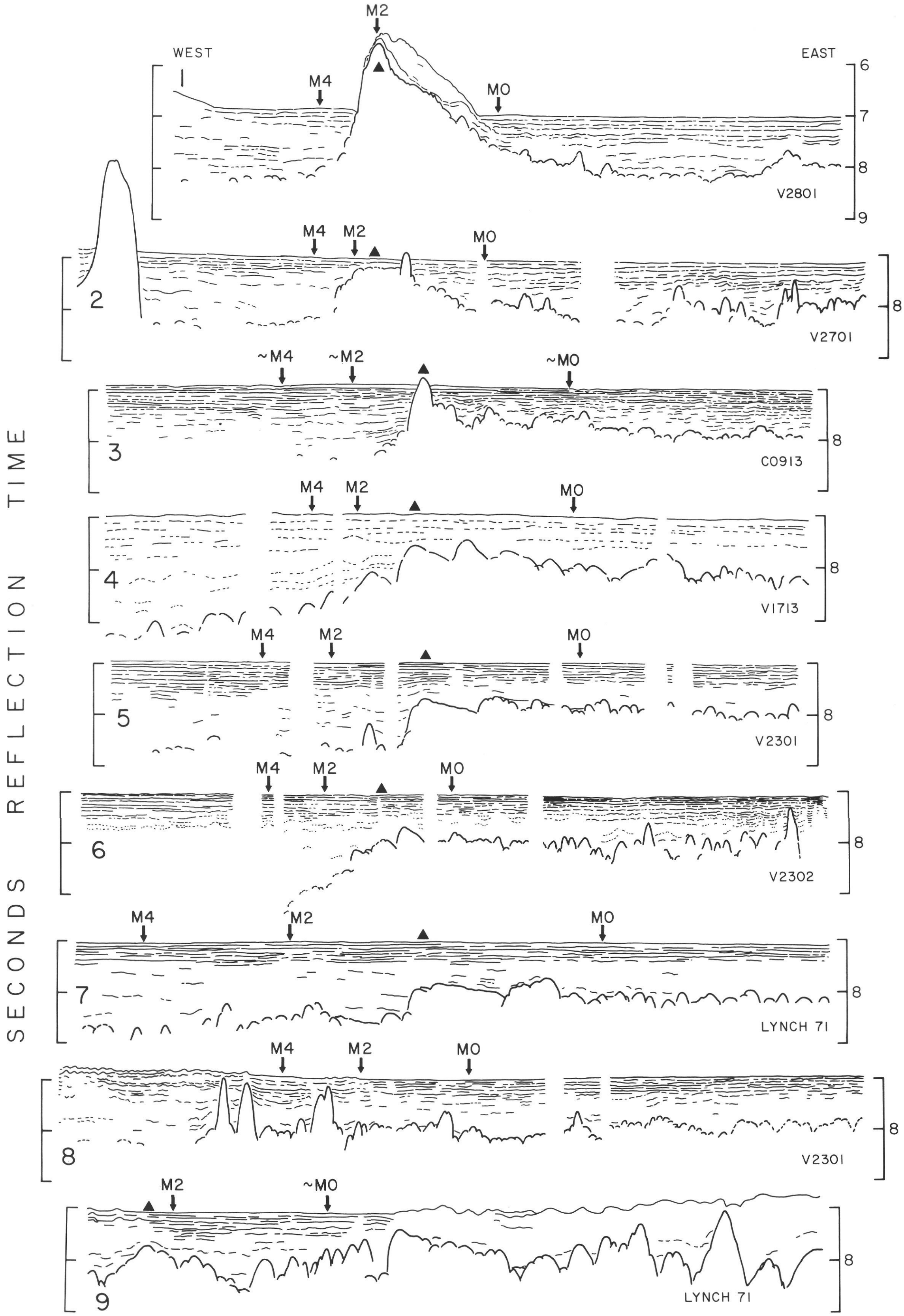


Figure 9. *Tracings of seismic profiler records across the* J-*Anomaly Ridge (locations in Figure 8), with magnetic anomaly positions illustrated. The ridge "crest" (or lip of escarpment) is marked by triangles.*

pect anomalous basalt composition at the ridge crest. However, the three basalt flows sampled proved to be rather typical ridge-crest tholeiites not significantly enriched in $FeO^T$ (6.99-8.48 wt. % for three samples) or $TiO_2$ (1.19-1.68%), (Houghton, this volume, table 3). Although all samples have experienced some alteration, the effects on $FeO^T$ and $TiO_2$ concentration probably are small.

Magnetic properties of the recovered basalts are similarly undistinguished (Petersen et al., this volume). The 10 samples measured gave means of $2.17 \pm 1.80 \times 10^{-3}$ Gauss for NRM, and $1.41 \pm 0.59 \times 10^{-3}$ Gauss/Oe for susceptibility. The relatively low Q ratio is more characteristic of subaerial than submarine basalts and thus supports other evidence for subaerial eruption. Presumably there are subaqueously erupted basalts at greater depth, having greater NRM and Q values.

Two distinct groups of stable magnetic inclination directions were obtained by Petersen et al. The upper eight samples showed steep normal directions (mean, 62.8°) close to the present dipole field inclination at the site (59.5°). Examination of the opaque mineralogy of these basalts suggests that they lost their primary remanence by low-temperature oxidation and that the present remanence is chemical in nature. The lowest two samples show a shallow reversed inclination (−36.8°) consistent with a thermo-remanence acquired during a reversed period in the Early Cretaceous (perhaps *M*-3, or 121.5 - 123 m.y.B.P., on the van Hinte [1976b] time scale).

All samples revealed some low-temperature oxidation effects, as expected for old basalts sampled away from the immediate spreading axis. More important is the evidence for earlier high-temperature deuteric oxidation which is rare in ocean floor basalts but is typical of subaerial basalt. This provides another line of evidence that the *J*-Anomaly Ridge at Site 384 was emergent early in its existence.

We conclude from major element chemistry and rock magnetic studies that if Fe-Ti basalts of high remanence, such as those reported along a section of the Galapagos spreading axis by Vogt and de Boer (1976), are responsible for the *J*-Anomaly, then they must lie east of Site 384 (Rabinowitz et al., this volume) and also possibly at greater depth at the drill site.

Besides Fe-Ti basalt or simply a thickened mass of ordinary pillow basalt, there are few possible source materials that could account for the large magnetic anomaly. One possibility is the alteration of a large mass of mantle peridotite to serpentinite, which would yield magnetite as a secondary mineral. Intensely magnetized gabbros, particularly ferrogabbros, are another possibility. For example, at Stardalur Farm in southwest Iceland there is an anomaly of 18,000 $\gamma$ amplitude at ground level and 4000 $\gamma$ amplitude at 1 km elevation, where the anomaly is 6 km wide (Kristjansson, 1970). The calculated source rock magnetization (0.05-0.06 Gauss) exceeds that commonly observed for basalts, although peculiar basalt with such high magnetization was recovered at shallow depth above the source body. Perhaps significantly, Rabinowitz et al. (this volume) calculate a similar value of magnetization for the *J*-Anomaly source. Kristjansson (1970) suggested that gabbro intrusives, representing upward protruberances in Layer 3, are responsible for the Stardalur and similar, less spectacular magnetic features elsewhere in Iceland. Supporting this is the finding that Icelandic gabbroic rocks, including olivine gabbro and pyroxinite, generally have considerably higher values of remanence and susceptibility than Tertiary lavas on Iceland (Kristjansson, 1970). These gabbros possess abundant titanomagnetite grains (2.5-3.0 volume per cent). Because the density of gabbro (2.9 g/cm³) exceeds that of basalt (2.7 g/cm³), positive gravity anomalies would be expected to correlate with magnetic features if the gabbroic intrusives occur at the expense of basaltic crust. Such gravity anomalies are observed on Iceland and would under this hypothesis be expected in the *J*-Anomaly area. However, large positive gravity anomalies are not observed across the *J*-Anomaly, and gabbroic intrusives, if present, must be of minor importance.

It is noteworthy that of the three Leg 43 drill sites that penetrated basaltic basement (384, 386, 387), only the basalts from Site 384 show significant amounts and relatively large sites of chromite crystals (Petersen et al., this volume). Chromite is not an uncommon constituent of Mid-Atlantic Ridge basalt. Between 30° and 40°N, eight out of 18 dredge hauls that recovered basalt along the spreading axis contained at least some chromite-bearing basalt (Sigurdsson and Schilling, 1976). At each site, the chromite-bearing basalts were the least differentiated (lowest FeO/(FeO + MgO). The differentiation index averaged 0.516 for the Mid-Atlantic Ridge basalts and is somewhat higher for the Site 384 basalt (0.54) (compare Sigurdsson and Schilling, 1976, with Houghton, this volume). This may suggest, but without great statistical significance, that the *J*-Anomaly basalt is somewhat more differentiated, had lost some of its chromite due to early precipitation, and came from a Cr-rich parent magma. Although it might be argued on this basis that the *J*-Anomaly is underlain by a deep-seated, dense (4.5-4.8 g/cm³), magnetized chromitite body, the lack of an associated gravity anomaly appears to preclude significant accumulation of such chromium spinel-bearing ultrabasic rocks.

In drilling the *J*-Anomaly Ridge, we wanted to test whether it falls in the same category as the Azores platform, or Iceland and the upper Reykjanes Ridge. Both of these so-called "hot-spots" exhibit paired basement ridges which are parallel or sub-parallel to the spreading axis and are similar to the *J*-Anomaly Ridge in dimensions, although they lack prominent magnetic anomalies (Vogt, 1971, 1974; Vogt et al., in preparation). Basalts recovered along and transverse to the Mid-Atlantic Ridge near these hot-spots are characteristically enriched in LIL elements (Schilling, 1973, 1975a,b, Brooks and Jakobsson, 1974) although the magnitude of the enrichment varies from one hot spot to another. Gradual transitions to typical LIL-depleted basalt chemistries are observed southward away from the Azores and Iceland hot-spots, and an abrupt change was found northward beyond the north coast of Iceland. We suspected that the *J*-Anomaly Ridge would be composed of LIL enriched basalt. However, the three sampled basalt flows are only slightly enriched in most LIL elements relative to typical Mid-Atlantic Ridge basalt, and even this slight enrichment might reflect alteration (Houghton, this volume). The relatively high $K_2O$ concentrations of Site 384 basalts undoubtedly does reflect alteration. On the other hand the ratio $(La/Sm)_{E.F.}$ was found to be surprisingly invariant across normal (depleted) segments of the mid-ocean ridges out to crustal ages of 60 m.y. (Schilling, 1975b). Thus the values of 0.56 to 0.83 computed for $(La/Sm)_{E.F.}$ from Houghton's data at Site 384 may indicate a chemically "transitional" ridge segment rather than the effects of alteration. In variability and average $(La/Sm)_{E.F.}$, the three flows sampled at Site 384 are quite similar to a typical dredge haul taken on the Reykjanes Ridge near 62°N where the LIL enrichment is transitional (Schilling, 1973) and basement ridges morphologically similar to the *J*-Anomaly Ridge are well developed (Vogt, 1971, 1974). The same may be said for $P_2O_5$ (0.13, 0.18, and 0.2 wt. %) and $TiO_2$ (1.19, 1.42, and 1.68 wt. %) determined for Site 384 basalts.

The *J*-Anomaly Ridge shoals toward the northeast and reaches its greatest width and elevation some 200 km northeast of Site 384. If a hot spot was responsible for the *J*-Anomaly Ridge and was centered 200 km northeast of Site 384, this drill site would be in a transitional setting similar to the northern Reykjanes Ridge. Future drilling could readily test this hypothesis. Another possibility, requiring deeper drilling for its verification, is that more depleted basalts erupted on top of earlier, unsampled, LIL enriched basalts at Site 384 when the "plume mantle" was used up and replaced by normal, depleted mantle. This is apparently the case in the Faeroes (Schilling and Noe-Nygaard, 1974). In any case, the apparent absence of strong LIL-enrichment at Site 384 on the crest of the *J*-Anomaly Ridge shows that the relationship between basement morphology and chemistry is not simple, and this finding reinforces DSDP results in the Greenland-Norwegian Sea (Talwani, Udintsev, et al., 1976).

### Age and Origin

Both radiometric and biostratigraphic age data were used to date the *J*-Anomaly Ridge basement at Site 384. Radiometric ages of 88 ± 5 m.y. (whole rock $^{40}Ar/^{39}Ar$) and 106 ± 4 m.y. (whole-rock K-Ar) on a sample free of amygdaloidal chlorite were obtained by Houghton et al. (this volume). The first age is not considered reliable because the curve of apparent age versus incremental temperature is irregular. The two ages are likely to be lower limits, and we infer a minimum absolute age of 106 ± 4 m.y. for the basalt.

Perkins (this volume) examined rudists from the limestones overlying the basalt. Specific identifications could not be made on the steinkerns in Core 16 (123m above basalt), but they generally resemble lower Albian forms in central Texas and northern Mexico. In Core 20 (29m above basalt) rudists indicate an age of middle Aptian through Albian, but probably not older than Albian. In core 21 immediately above basalt, Schroeder and Cherchi (this volume) identified orbitolinid fauna characteristic of the late Barremian and earliest Aptian. The available age data thus suggest a minimum basement age of about 113-118 m.y. according to the van Hinte (1976a) time scale.

Several factors need to be considered in evaluating this age. First it is uncertain how much time elapsed between cessation of volcanism and initial accumulation of limestones. Several lines of evidence suggest that the ridge crest at Site 384 was emergent during its early history (see later discussion). The time that elapsed between formation and subsidence to sea level is difficult to estimate, but reasonable erosion and subsidence rates (Vogt, 1972) suggest it is unlikely to be more than a few million years. Once the ridge crest reached sea level, shallow-water carbonate accumulation probably was quickly established.

Later igneous activity also may have buried older sediments, and the question arises whether the sampled basalt flows erupted significantly later than those below the bottom of the drill hole. This seems unlikely, for several reasons. The basalt is rather typical ridge crest tholeiite (Houghton, this volume), not the alkali basalt *normally* erupted a few tens of kilometers or more from the rift axis. Morphologically, the *J*-Anomaly Ridge and its eastern Atlantic counterpart probably formed an anomalously elevated segment of spreading axis much like the central block of the Reykjanes Ridge southwest of Iceland (Vogt, 1974). The Reykjanes Ridge has a narrow accretion axis and shows no evidence of igneous activity away from the crest. Significant "late" volcanism on the crest of the *J*-Anomaly Ridge also probably would have created equidimensional topography, e.g., conical seamounts. If the basalt cored at Site 384 had been erupted long after ridge formation (of the order 10 m.y. or more) the ridge probably would have subsided at least a kilometer, which is much too deep considering both the deuteric high-temperature oxidation of the opaque minerals in the basalt recovered (Petersen et al., this volume) and the high basalt vesicularity.

It also is important to consider the interval between the time the *J*-Anomaly Ridge began to form and the time it reached sea level. There are two likely ways that the ridge could have formed: (1) as a tectonic, fault-bounded horst, or (2) as a progressive, constructional buildup at the accretion axis, such as would result from increased output of basaltic magma. In tectonically active areas vertical motions as great as approximately 1 cm/yr are possible. At this rate the *J*-Anomaly Ridge could have been tectonically erected in 0.3 m.y. Longer tectonic uplift, say approximately 3 m.y., would violate the assumption that uplift occurred in the narrow axial accretion zone of the Mid-Atlantic Ridge, and because the *J*-Anomaly Ridge is oriented nearly parallel to magnetic lineations and has a counterpart (although less well developed) in the eastern Atlantic, formation of the ridge along the immediate spreading axis is strongly indicated.

In the "constructional" model the buildup time equals the horizontal distance from the crest to the western base of the *J*-Anomaly Ridge basement escarpment (about 10-30 km), divided by the spreading half-rate (of the order 1 cm/yr). This time interval, approximately 1-3 m.y., is the same as that more reliably estimated for the buildup time of the inner and outer basement escarpments on the Reykjanes Ridge southwest of Iceland (Vogt, 1974). Morphologically, the *J*-Anomaly Ridge closely resembles these features, and it may have had a similar origin.

On the basis of all these considerations we infer that the *J*-Anomaly Ridge began to form during Barremian time (121-115 m.y.B.P.) and had completed its growth and subsequent erosion to sea level in about 5 million years (Figure 10).

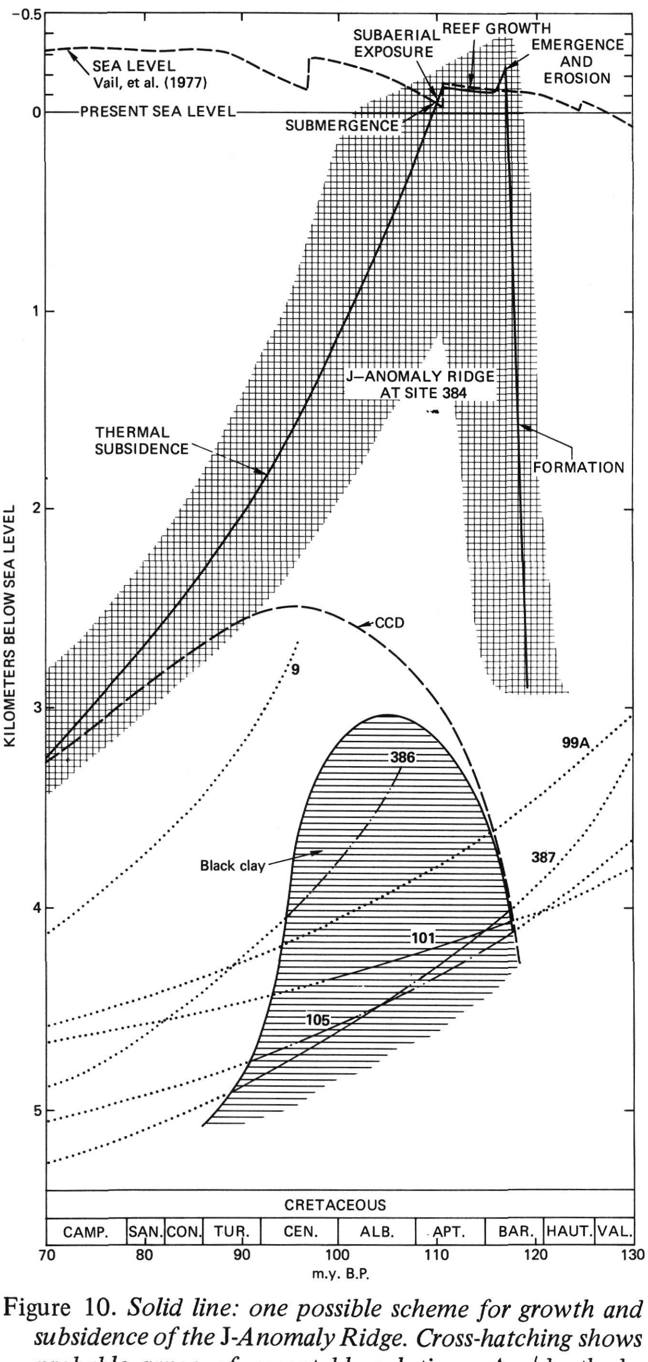


Figure 10. *Solid line: one possible scheme for growth and subsidence of the* J-*Anomaly Ridge. Cross-hatching shows probable range of acceptable solutions. Age/depth depositional field of western Atlantic black clays (horizontal shading), determined from other drill sites, possibly relates to formation and subsidence of circulation barriers such as the* J-*Anomaly Ridge.*

If the determined age is correct within a few million years, we can draw the following conclusions. The age confirms our initial working hypothesis that the *J*-Anomaly Ridge at Site 384 was formed near the time of anomaly *M*-2, according to the time scale of Larson and Hilde (1975). Indeed, the Site 384 basement age, late Barremian to early Aptian, helps confirm the Larson-Hilde time scale at the youngest end of the Keathley (*M*-series) reversal sequence, where no basement ages were previously available (Vogt and Einwich, this volume). These results from Site 384 also contradict the suggestion of Barrett and Keen (1976) that the *J*-Anomaly Ridge formed during the middle of the *M*-series, or about 125-130 m.y.B.P. on the van Hinte (1976a) time scale. However, we reiterate that the Site 384 basement age is a minimum age, and we had to infer that the true age is only a few million years greater.

It is also noteworthy that the *J*-Anomaly Ridge was created at the transition from a long period of relatively slow spreading to another long period of faster spreading (Vogt and Einwich, this volume) or shortly after a spreading half-rate increase from about 0.5 cm/yr (*M*10-*M*4) to 1.0-1.5 cm/yr (post *M*-4) (Schouten et al., 1976; K. Klitgord, personal communication, 1977). If this coincidence is not fortuitous, a causal relationship between the increased spreading rates and formation of the ridge is suggested.

**Emergence and Subsidence**

A surprising result at Site 384 was the discovery that at least the crest of the *J*-Anomaly Ridge had been at and even above sea level early in its existence. Independent indicators from the basalt and the paleontology, sedimentology, and geochemistry of the overlying bioclastic limestone provide this evidence. Of these clues, some merely indicate shallow water, others shallow water and/or emergence, and a few argue unequivocably for emergence.

The basalt recovered at Site 384 exhibits high vesicularity (20-30%) (Site 384 Report), suggesting it was erupted at depths of less than 500 meters (Moore and Schilling, 1973) and perhaps even above sea level. Subaerial emplacement is more strongly indicated by the high-temperature oxidation experienced by the titanomagnetites (Petersen et al., this volume). Such oxidation is common in subaerial basalts but rare in ocean floor basalts.

A diverse, typically shallow-water fossil assemblage of pelecypods, gastropods, echinoderms, forams, bryozoans, solitary corals, and possibly red algae was recovered from the bioclastic limestone formation. Specifically, a shallow-water "shelf-bank" environment of the order of 10 meters depth and Aptian-Albian in age is indicated by the rudists (Perkins, this volume). Admixture of line mud suggests a relatively quiet backreef environment, at least temporarily. Among observed diagenetic effects, moldic porosity from dissolution of aragonite skeletons clearly indicates exposure to meteoric water (Rothe, this volume). This idea is reinforced by relatively high $\delta^{18}O$ values which reflect diagenetic alteration of carbonates from aragonite and high-Mg calcite to low-Mg calcite. The alteration was caused by percolating meteoric water (Rothe and Hoefs, this volume).

The evidence thus suggests that the ridge crest was initially above sea level (Figure 10). Subsidence to sea level probably was accomplished by subaerial erosion as well as by tectonic and thermal subsidence. A shallow-water "shelf-bank" environment was provided and colonized by diverse fauna and flora. Because the bioclastic limestone is 123 meters thick we assume that biological accretion temporarily kept pace with subsidence. Core recovery was so poor and so disturbed by drilling that little can be deduced about the detailed bio- and lithostratigraphy and evolution of the limestone facies. Evidence for exposure to meteoric water might be ascribed to temporary tectonic uplift or eustatic sea level lowering that was sufficiently vigorous to counteract the effect of general subsidence of the ridge and thus subaerially expose the limestones (Figure 10). Exceptional storms may also have constructed subaerial banks, and the role of ebb tides in exposing the shallow sea floor to rainwater or the presence of a sub-sea level, freshwater lens extending from subaerial outcrops might have to be considered. Also, on the basis of the small amount of core recovered, we cannot assume that the entire 123-meter-thick formation was emergent at the same time.

On the tectonic grounds it would appear unlikely that the 123-meter-thick bioclastic formation represents an extended period of time. At "thermal" subsidence rates of normal ocean crust, the ridge crest would have sunk 120 meters in less than 2 million years. Isostatic adjustment of such a massive ridge formed at the spreading axis would be much faster, but most of this adjustment probably occurred within the period of ridge formation. For example, on Reykjanes Peninsula in Iceland, subsidence rates have been of the order 0.5 to 1 cm/yr for at least the last $10^5$ years (Tryggvason, 1974). Even a volcanic edifice constructed far from the spreading axis may subside rapidly; deep drilling on São Miguel in the Azores revealed at least 0.1 cm/yr, possibly 1.0 cm/yr subsidence rates for the last few hundred thousand years (Aumento and Sullivan, 1974).

If rapid tectonic/thermal subsidence actually occurred, it might explain why the bank community perished. On the other hand, similar rudistid reef or bank communities also became extinct in Albian times on widely scattered equatorial Pacific guyots (Heezen et al., 1973), and it is suggested that the Albian-Cenomanian transgression drowned these reefs (Douglas et al., 1973). Recent detailed studies of global eustatic sea level cycles (Vail et al., 1977) and attempts to calibrate Cretaceous sea level (Pitman, in press; Sleep 1976) generally support this interpretation. These studies suggest that, following a late Aptian regression, sea level rose on the order of 250 meters by early Cenomanian time. On the *J*-Anomaly Ridge, extinction of the reef-bank community may have resulted from subaerial exposure during the Aptian regression (possibly accentuated by renewed tectonic uplift) before the Cenomanian sea-level rise (Figure 10). The Cenomanian sea-level rise suggested by Vail et al. (1977) is about 18 cm/1000 years, much faster than the 1 cm/1000 year rate expected as a response to increasing spreading rates along the mid-ocean ridges (Hays and Pitman, 1973; Pitman, in press). Thus, if this rate is realistic, other as yet undiscovered processes must be responsible.

The subsidence history of the *J*-Anomaly Ridge after its final exposure above sea level, and of the adjacent ocean floor, is summarized in Figure 11 and Table 2. We have analyzed this subsidence in two ways: (1) the ridge and basement parcels east and west of it are assumed locally compensated for sediment loading, and each parcel subsides along the empirical North Atlantic age/depth curve; (2) the ridge is assumed "welded" to the crust on either side, and all three crustal parcels subside to their present depth from an initial depth determined by the observation that the crest was at sea level about 110 m.y.B.P.

Using the empirical age/depth curve for the Atlantic (see Figure A1), and correcting for isostatic loading by sediments, we can backtrack the basement east and west of the *J*-Anomaly Ridge to initial depths of around 2.7 to 3.0 km, which is only a few hundred meters deeper than the average present crestal depth of the Mid-Atlantic Ridge (Figure 11). This slight excess subsidence (Column 9, Table 2) could be explained in part as an effect of flexure of the lithosphere by the load of the ~2 km high *J*-Anomaly basement ridge (see fig. 2 of Watts et al., 1975). The crest of the *J*-Anomaly Ridge, however, does not backtrack to the zero depth demanded by the Site 384 borehole data. The calculated initial depth is about 1 km too deep (Figure 11). Therefore, if the adjacent basement was initially at typical Mid-Atlantic Ridge crestal depths, the *J*-Anomaly Ridge must have subsided differentially by 1 km with respect to the adjacent basement. A few hundred meters of this depth descrepancy can be explained as downward flexing of the ridge by the presence of adjacent thick sediments of the surrounding abyssal plain and continental rise. Possibly, much of the extra subsidence was accomplished tectonically, by normal faulting of the adjacent, initially thin plate as the basement ridge was formed on it. Unfortunately, available seismic reflection records do not resolve deep structure well enough that we can search for the implied marginal faults. Another possible explanation is the downward flexing of the lithosphere by the load of the basement ridge. A 2-km-high basaltic ridge imposed on very young lithosphere no doubt would depress the basement level considerably more than the required amount (Watts et al., 1975). However, crustal adjustment in response to an imposed load occurs over time periods short enough (~$10^4$ years) to keep pace with the construction of most volcanic loads. Thus, it is difficult to account for 1000 meters of differential subsidence of the ridge as a late-stage phenomenon using this mechanism.

In the above discussion the *J*-Anomaly Ridge was allowed to subside differentially with respect to the adjacent basement, which we have supposed to follow the empirical North Atlantic curve. A second approach is to

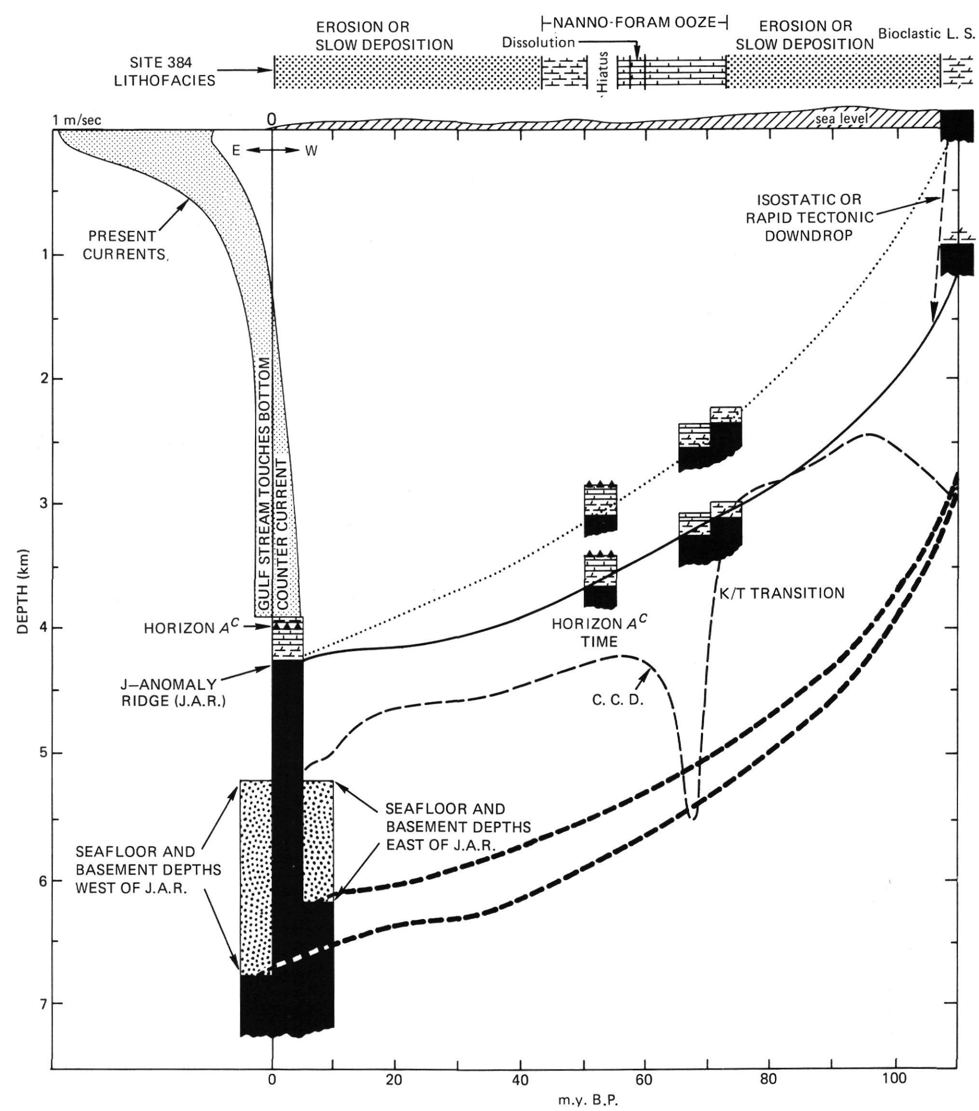


Figure 11. *Subsidence history of the J-Anomaly Ridge. Heavy solid and heavy dashed curves are basement depth as a function of age for the ridge crest and points 20 km east and west. Curves computed from the present time backwards using empirical Atlantic age/depth relation (Figure A1), with corrections for isostatic loading by sediment (assuming local compensation). One-half of sediment load is attributed to abyssal plain turbidites deposited within the last 10 m.y. – this explains downward bend at tail of subsidence curve. Because crest of ridge does not "backtrack" to sea level, initial rapid tectonic and/or isostatic downdrop of 1 km is postulated. Alternative model, uniform but relatively fast subsidence of ridge (and adjacent crusts), is shown by a dotted line. Western North Atlantic CCD (dashed) is plotted from Figure 3. Approximate present present current speeds estimated from Neumann (1968), Grant (1968), and Boisvert (1967), and sea-level fluctuations are from Vine (1973).*

assume that the ridge remained "welded" to the adjacent basement. In this model, not only the ridge but also the surrounding crust must have subsided by 4.1 to 4.2 km (Column 4, Table 1), and no particular age/depth relationship need be assumed. An initial depth of 2.5-2.9 km then is implied for the basement west of the ridge, which is typical of the present-day Mid-Atlantic Ridge crest. The initial basement depth east of the ridge would have been somewhat shallower, about 2.0 to 2.4 km (Column 5, Table 2). The shallower basement on younger crust east of the ridge, as well as the asymmetrical profile of the ridge itself suggests that the mantle disturbance responsible for the feature affected the spreading axis abruptly but then dissipated more gradually. Features with a similarly asymmetrical morphology occur on the Reykjanes Ridge flanks southwest of Iceland, and south of the Charlie Gibbs fracture zone (Vogt, 1971). A model explaining their main features in terms of plume-generated flow under the accretion axis was presented by Vogt (1974), suggesting

TABLE 2
**Details of Total *J*-Anomaly Ridge (J.A.R.) Subsidence Near Site 384 With Respect to Present Sea Level (in km)**

| Locality | (1) Present Seafloor Depth | (2) Sediment (km) | (3) Corrected Depth (km) | (4) Implied Subsidence | (5) Implied Initial Depth (km) |
|---|---|---|---|---|---|
| ~20 km west of J.A.R. crest. | 5.2 | 1.5-1.8 | 6.2-6.5 | (4.1-4.2) | (2.5-2.9) |
| J.A.R. crest. | 4.0 | 0.5 | 4.3-4.4 | 4.1-4.2 | 0 |
| ~20 km east of J.A.R. crest. | 5.2 | 1.0-1.3 | 5.9-6.1 | (4.1-4.2) | (2.0-2.4) |
| | (6) Theoretical Depth (km) | (7) Avg. Atl. Observed Depth (km) | (8) Predicted Total Subsidence (km) | (9) Excess Subsidence | (10) Present Depth Anomaly (km) |
| ~20 km west of J.A.R. crest. | 5.9 | 5.8 | 3.7-4.0 | (–0.1 to –0.5) | –0.3 to –0.7 |
| J.A.R. crest. | 5.85 | 5.75 | 3.35-3.55 | –0.55 to –0.85 | +1.35 to +1.45 |
| ~20 km east of J.A.R. crest. | 5.8 | 5.7 | 3.5-3.7 | (–0.4 to –0.7) | –0.1 to –0.4 |

Note: Columns (1) and (2) self-explanatory: Average values from seismic reflection profiles; (3) basement depth adjusted for isostatic loading by sediment, assuming local compensation; (4) total subsidence from sea level, based on Site 384 results. Values in parentheses apply if crust adjacent to J.A.R. is assumed "welded" to it; (5) present basement depth minus column (4); (6) average Atlantic basement depth appropriate to crust of this age, computed by model of Parsons and Sclater (1977); (7) empirical equivalent to (6) (Parsons and Sclater, 1977); (8) total subsidence computed from age/depth curves (6 or 7) and corrected for isostatic loading by sediment; (9) column (4) minus (8); (10) column (3) minus (6) or (7).

that the crust formed just subsequent to such a ridge is likely to be anomalously shallow and the crust formed just prior to it more normal in depth. This comparison lends credence to the "welded-crust" model for *J*-Anomaly Ridge subsidence, but an anomalous, excessive amount of *total* subsidence of the ridge area is still required to explain the present basement depth. At present we cannot pinpoint the mechanism producing such additional subsidence, and the actual total-subsidence curve (thermal subsidence plus loading by sediment [Parsons and Sclater, 1977], plus anomalous regional subsidence) can only be estimated (Figure 11).

## EARLY CRETACEOUS RUDIST REEFS

Site 384 on the crest of the *J*-Anomaly Ridge recovered cores from a 123-meter-thick bioclastic limestone sequence. Components include abundant rudists, other pelecypods, gastropods, foraminifers, ostracodes, bryozoans, corals, and red algae. Unfortunately, the generally poor preservation and core recovery make it difficult to document the exact environment of deposition. The monopleurid rudists in Core 16 probably reflect a very shallow restricted marine environment, and the caprinid rudists in Core 20 a shallow-water "shelf bank" environment. However, neither is necessarily associated with barrier or shelf-edge reef assemblages (Perkins, this volume). Rothe (this volume) points out on the basis of sedimentological criteria that the carbonates could represent a quiet back-reef environment, and this interpretation seems to be supported by the morphology of the entire carbonate bank in the *Challenger* profile across Site 384 (see Figure 6 in Site 384 Report). On this basis, we interpret the carbonate bank capping the *J*-Anomaly Ridge as having both true reef (not cored) and back-reef environments.

Coral-rudist reefs flourished in near-equatorial marginal seas and on oceanic islands during the Early Cretaceous (Douglas et al., 1973) at the same time that the deep North Atlantic Basin was substantially anaerobic (Figure 7). Shallow-water limestones of reefal affinity have been dredged at several locations along the Blake Escarpment (Heezen and Sheridan, 1966; Sheridan et al., 1969, 1971), and Meyerhoff and Hatten (1974) have traced this Cretaceous marginal reef from the Blake Escarpment through the Bahamas and Cuba to the West Florida Escarpment.

Seismic reflection and refraction studies of the U.S. Atlantic continental shelf since the 1950's also have suggested the presence of a ridge structure beneath the edge of the continental shelf and slope (Figure 7). Evidence for this ridge recently has been summarized by Emery et al. (1970), Mayhew (1974), and Mattick et al. (1974). These authors suggest that the ridge is a buried reef complex, essentially a northward extension of the Blake Escarpment reef complex. Recovery of apparent reefal material at Site 384 fills in a broad gap between Florida and Iberia in known occurrence of coral-rudist reefs, and it provides strong additional evidence that the buried ridge complex beneath the outer shelf and continental slope is of reef origin.

If an extensive Early Cretaceous barrier reef system was present along the North American continental margin, it could have major implications for the sedimentary history of the deep North American Basin. The combined effects of this shelf-edge barrier, rising Cretaceous sea level, and rapid margin subsidence (Whitten, 1976) should have prevented much terrigenous debris from entering the deep sea. Presumably, terrigenous debris reaching the continental shelf was transported only in the shelf bottom-water, thus allowing the reef to flourish in relatively clear water. If this assessment is valid, these controls and the reef system in particular provide an explanation for the uniform non-carbonate accumulation rates of the Neocomian limestones compared with black clays, and for the consistent thickness ($\sim$100m) of the Cretaceous black clays both near and distant from the continental margin (Sites 105 and 387, respectively). It is interesting to note that Sheridan et al. (1969) have suggested that Great Abaco Canyon near the southern end of the Blake Escarpment was a major breach in the Early Cretaceous barrier reef system. This may help explain the presence of unusually thick black-clay sections at Site 101 (Hollister, Ewing, et al., 1972) and at Site 391 (Benson, Sheridan, et al., 1976). Mapping of black-clay thickness from seismic profiler records also suggests that these sediments thicken abruptly beneath the central and upper continental rise (Tucholke and Mountain, 1977). Presumably these thickened wedges contain terrigenous sediment that also reached the deep sea through local breaches in the reef system, but they may contain substantial shallow-water carbonate debris.

By way of contrast with most western North Atlantic sites, black clays drilled in the eastern North Atlantic (Ryan et al., 1976; Montadert, Roberts, et al., 1976; Lancelot, Seibold, et al., 1975) and in the South Atlantic (Bolli, Ryan, et al., 1975) include abundant terrigenous organic debris and generally thicker sections and higher rates of accumulation. Barrier-reef control of terrigenous influx to the deep basin along these margins must have been minor or nonexistent.

## TURONIAN-MAESTRICHTIAN MULTICOLORED CLAYS

### Description

Multicolored clays (commonly hues of red) were cored at Sites 386 and 387 on Leg 43 and also were recovered in variable quantities at earlier drill sites (7, 9, 28, 105). Only the clays at Sites 386 and 387 contained microfossils with which to date the sediments, but the clays at the other sites are inferred to be of comparable age. This is reasonable in terms of the accumulation rates that are then required (see color foldouts).

Although the thickness of the multicolored clays varies throughout the basin (Table 1), it probably only locally exceeds 100 meters. Accumulation rates are low, ranging from $<1$ m/m.y. at Site 387 to perhaps as much as 3 m/m.y. at Sites 386 and 9. The latter rates are probably higher than average due to abundant volcanogenic components (see below).

The multicolored clays are devoid of carbonate, organic carbon, and siliceous microfossils at almost all sites where they have been sampled. However, disordered cristobalite ranges in abundance from a few per cent (Sites 7, 387) to a dominant fraction (Site 9) and it may be derived from recrystallization of siliceous tests. From these compositional data, it is clear that the multicolored clays were deposited well below the CCD and that bottom water in the deeper parts of the basin probably also was undersaturated with respect to silica (Figure 3). In this respect, it is noteworthy in Figure 3 that the site having the shallowest sea floor in the red-clay depositional field (Site 9) also has the greatest abundance of disordered cristobalite.

Compared to the underlying black clays, the multicolored clays contain a larger clay-mineral fraction (60-80%), which is dominated by illite, with lesser montmorillonite, kaolinite, and minor chlorite. Quartz and feldspars are only locally abundant, and they average about 20 per cent and 3-5 per cent, respectively. Sites 386 and 9 are unusual in that they contain abundant clinoptilolite and locally common phillipsite, whereas these zeolites are less common in multicolored clays from other sites.

Lancelot et al. (1972) have noted that the multicolored claystones are mineralogically similar to the hot-brine deposits in the Red Sea. However they differ chemically from these deposits and from the basal metalliferous sediments formed near actively spreading ocean ridges (Murdmaa et al.; and Arthur, this volume). Chemical composition of the multicolored clays is most similar to modern abyssal "red" clays in the Pacific, but they have somewhat lower trace-element content, lower manganese, and higher iron.

### Depositional Environment

As already noted, the multicolored clays accumulated at true pelagic clay rates below a shallow CCD. The lack of major terrigenous input may be explained by two factors. First, the Late Cretaceous transgression (Sloss, 1963; Ronov, 1968; Vail et al., 1977) would have restricted deposition of continental debris mostly to the widespread continental shelves and marginal seas. In addition, the shelf-edge barrier reefs, although largely extinct, would continue to act for an unspecified time as dams preventing seaward dispersal of terrigenous debris (Emery et al., 1970).

The shallow CCD might have continued from the Early Cretaceous in response to the extensive transgression and presumably equable, maritime climate (polytaxic episode of Fischer and Arthur, 1978). In this situation, wind stress and upwelling in the open ocean would be reduced, resulting in lowered surface productivity and undersaturation of the deep ocean with respect to calcite and silica. However, it is clear that the bottom water was oxygenated, probably because of deep-water connections with the South Atlantic. In addition, the termination of black-clay deposition in the Caribbean by Campanian time (Figure 6; Edgar, Saunders, et al., 1973) suggests that deep flow connections were established there with the North Atlantic and/or Pacific oceans.

In view of the very low accumulation rates for the multicolored clays in the western North Atlantic, it is not surprising that volcanogenic and zeolitic components constitute a significant fraction of the total sedimentary record. In environments of more rapid deposition, these components would be greatly diluted. However, there is some evidence that volcanism at greater than "background levels" contributed debris to the multicolored clays. Although questionable, there may have been Cenomanian volcanism at or near Bermuda (Reynolds and Aumento, 1974), which contributed volcanogenic debris to Site 386. Similarly, Leg 43 drilling at Sites 382 and 385 established that Late Cretaceous volcanism occurred along the eastern end of the New England Seamount Chain, possibly accounting for volcanogenic components in the multicolored clays at Site 9 and in the carbonates at Site 10. If related to seamount volcanism, the increased occurrence of this volcanogenic debris only in the more easterly drill sites (386, 9, 10) could indicate that volcanism along the western part of the New England Seamount Chain had ceased by Late Cretaceous time. This possibility is supported by radiometric dating of seamount basalts, which suggests Early Cretaceous volcanism along the western seamounts of the chain (Duncan and Houghton, 1977). This problem is further discussed by Vogt and Tucholke (this volume).

## LATE MAESTRICHTIAN CARBONATES

Aside from Site 10, which recovered a totally calcareous Campanian-Maestrichtian section immediately above basaltic basement, the Leg 43 Sites 384, 385, 386, and 387 were the first boreholes in the western North Atlantic to recover upper Maestrichtian sediments. The upper Maestrichtian sediments at these sites are all nannoplankton chalks and limestones with minor foraminifers (see Site Reports). They clearly demonstrate a sharp depression of the CCD to more than 5400 meters in the late Maestrichtian, followed by a rise to 4-4.5 km in early Paleocene time (Figure 3). This CCD depression is exactly opposite to the global shoaling of the CCD suggested by Worsley (1974).

The Maestrichtian carbonate facies correlates with Horizon *A** at Sites 386 and 387 (see Site Reports). Although the carbonate layer is relatively thin, it can be traced seismically as Horizon *A** through a substantial portion of the basin west of Bermuda (Tucholke, this volume). This provides further evidence that the CCD depression was at least a basinwide event.

The mechanism of this sudden CCD depression is not clear, although we can make some preliminary speculations. The deep-water saturation with carbonate implied by the CCD lowering probably resulted from increased upwelling and productivity in the surface water. This could have resulted from increasingly "continental" climate caused by latest Cretaceous sea-level regression (Ronov, 1968; Douglas et al., 1973) and from cooler oceanic temperatures (Dorman, 1968; Ramsay, 1974; Fischer and Arthur, 1977). However, unless the magnitude of the regression and cooling are greater than presently believed, it is difficult to believe that they solely account for such a dramatic fluctuation in the carbonate compensation depth.

## THE CRETACEOUS-TERTIARY BOUNDARY

### Continuity of the Site 384 Record

Of major interest at Site 384 was the recovery of an apparently continuous fossiliferous sequence of foraminifer-nannofossil ooze extending from the middle Maestrichtian into the late Paleocene, and, above a 2 to 4 m.y. hiatus, into the middle Eocene.

In many parts of the world the Cretaceous/Tertiary boundary is marked by hiatuses or non-fossiliferous sections, and as a result little has been learned about the cause of the end-Cretaceous extinctions. Thus the results from Site 384 merit careful analysis. If the diverse taxa that became extinct did not do so simultaneously (within stratigraphic resolution), then a "continuous section" is one in which Cretaceous taxa one by one disappear upward, to be replaced by Tertiary forms. Possibly transitional Cretaceous/Tertiary sections include the Braggs section (continental shelf) in southern Alabama (Worsley, 1974), and the Shatsky Rise (DSDP Site 47.2; Fischer, Heezen, et al., 1971) in an open ocean environment. The length of the "transition" at those sites was estimated at $10^5$-$10^6$ years and about $0.5 \times 10^6$ years, respectively. Such transitional sections have not been found elsewhere, even in the Gubbio section of Italy where the lowermost Danian "*Globigerina*" *eugubina* Zone (P.1a) lies directly above upper Maestrichtian sediment, with only a 3-cm-thick clay layer between (Cita and Premoli Silva, 1974; Arthur and Fischer, 1977). A. G. Fischer has estimated that all extinctions at Gubbio occurred in the span of only $10^4$-$10^5$ years (Cita and Premoli Silva, 1974). This is near the limits of stratigraphic resolution in carbonate ooze deposited at rates of 5 to 20 m/m.y.

There are few data on the question of how long it takes a particular species to become extinct. The correlation of several Plio-Pleistocene radiolarian extinctions with geomagnetic polarity reversals suggests that at least two species with regional distribution can become simultaneously extinct within a few thousand years at most. One radiolarian of global distribution has been shown to become extinct synchronously in all areas within a few thousand years or less (Hays and Shackleton, 1976). If the end-Cretaceous marine extinctions were synchronous and took only $10^3$-$10^4$ years or less, there is little hope of resolving the actual transition interval at Site 384.

At Site 384, the lowest standard foraminiferal zone (P.1a) is present directly above the upper Maestrichtian. The nannofossil Zone NP 1 (*Markalius astroporus*) is not observed. However, this zone has been defined as the interval between the extinction of Cretaceous taxa and the first appearance of *Cruciplacolithus tenuis*, a very small, fragile, and solution-susceptible species. Because of the excellent nannofossil preservation at Site 384 and because of probable dissolution effects in the NP1 interval at other localities, Thierstein and Okada (this volume) have argued that Zone NP 1 as defined probably is only an artifact of preservation, and the "absence" of the zone at Site 384 has no bearing on the continuity of the Cretaceous-Tertiary boundary at the site.

There is a transition in the nannofossil record at Site 384 from dominantly Cretaceous forms to dominantly Tertiary forms over the interval 167.92 meters to 167.20 meters sub-bottom depth. Theirstein and Okada (this volume) have shown that this can be explained by "instantaneous" extinction of Cretaceous nannoflora near 167.74 meters, with subsequent partial benthic mixing creating a "transitional" boundary.

The apparent abruptness of the extinctions at Site 384 could have three explanations.

1) The extinctions and associated geological changes were globally abrupt ($10^4$-$10^5$ years or less) and there is no hiatus at Site 384. If so, other explanations are needed for the supposedly transitional sections in Alabama (Worsley, 1974) and on the Shatsky Rise (Bukry et al., 1971). It is entirely possible that these "transitional" sections do not represent transitional extinctions but rather are a result of lateral sediment transport and reworking. Douglas (1971), for example, has pointed out that erosion and sediment reworking of upper Maestrichtian sediment at Site 47.2 probably occurred in earliest Paleocene time.

2) The extinctions were spread over $10^5$-$10^6$ years (Worsley, 1974; Bukry et al., 1971) and the "transitional" (extinction) interval is unrepresented at Site 384, either because productivity and sedimentation rate fell essentially to zero, or because a very brief episode of sea-floor erosion has removed the transitional section. A hiatus would be required by the biostratigraphy if the time-equivalents of lowest Tertiary fauna present in other parts of the world are missing at Site 384. In fact, the lowest Tertiary faunal assemblage at the site does not typify the *"G." eugubina* Zone as described by Cita and Premoli Silva (1974). Because environmental differences may have existed between Site 384 and the Tethys, however, the Site 384 fauna in themselves do not demand a hiatus.

The *"G." eugubina* Zone (P.1a), characterized by a peculiar, dwarfed fauna, has been identified locally in the Appenines of Italy, in the USSR, North Africa, the Caribbean (DSDP Sites 38, 146, 152, and 153), and at Site 199 in the Caroline Abyssal Plain in the western Pacific (Premoli Silva and Bolli, 1973; Krasheninnikov and Hoskins, 1973). This distribution, although limited, suggests that it is a true biostratigraphic zone (Cita and Premoli Silva, 1974), but some paleontologists explain it as a locally environmentally induced, impoverished facies (see Site 384 Report).

The "*G.*" *eugubina* Zone occurs between the Maestrichtian and typically Paleocene foraminifer assemblages and is generally (but not universally) assigned to the lowest Tertiary. The length of time represented by the zone can be little more than 0.5 to 1 m.y. and is probably much less. At Site 199 in the Pacific, rare Maestrichtian foraminifers are found in the *"G." eugubina* Zone; if they are *in situ* the zone could be locally transitional between the Cretaceous and Tertiary (Krasheninnikov and Hoskins, 1973). At Site 199, NP 1 (*Markalius astroporus*) is not observed, and the nannofossils found in the "*G.*" *eugubina* sediment define Zone NP 2 (*Cruciplacolithus tenuis*) (Hekel, 1973). A similar relationship exists at Site 384 where the "*G.*" *eugubina* Zone correlates with nannofossil Zone NP 2. However, as noted earlier, the apparent "absence" of NP 1 may be only a problem of fossil preservation. At Site 47.2 on the Shatsky Rise neither the "*G.*" *eugubina* Zone nor NP 1 was identified, but a 4.5-meter-thick interval of mixed Cretaceous and Paleocene foraminifers lies between the lower Paleocene *Globigerina taurica* Zone and the upper Maestrichtian (Douglas, 1971). The mixed foraminifer transition zone, which represents roughly half a million years at prevailing sedimentation rates, correlates with the nannofossil *Tetralithus murus* Zone of the Maestrichtian without Tertiary nannofossil admixtures. On the basis of this observation, Bukry et al. (1971) suggested that "Tertiary planktonic foraminiferal species appeared before the calcareous nannoplankton species generally used to indicate the earliest Tertiary." They therefore are interpretating the transition zone as *gradual* disappearance of Cretaceous foraminifers, not as reworking. If this were true on a global scale, then a hiatus could be required at Site 384 to explain the correlative nannoplankton and foraminiferal extinction levels in the core.

3) The extinctions were spread over as much as $10^5$-$10^6$ years and there is no hiatus. In this case, the warmth-loving Maestrichtian assemblages may have abruptly disappeared only locally because of the appearance of new water masses over the site. Patterns of extinction could have been more gradual in Alabama (Worsley, 1974) and other remaining warm-water strongholds, but Cretaceous forms never reappeared at Site 384. We stress that a hiatus is not required to explain the biostratigraphic abruptness of the boundary; a swift appearance of new ecological/oceanographic conditions caused by a shift in circulation is not unlikely, especially at Site 384 which lies close to the boundary between the warm Gulf Stream and much cooler water masses to the north (Boisvert, 1967). Although the temperature drop across the northern edge of the Gulf Stream is surely more pronounced today (10 °C; Grant, 1968), the probable existence of an early Tertiary Gulf Stream (Berggren and Hollister, 1974) carries with it the implication of a water mass and thermal boundary near Site 384.

At present, it is not possible to choose among these alternatives. Because of uncertainties in biostratigraphic zonation for the earliest Paleocene (e.g., "NP 1") and uncertainties in ages of zonal boundaries (compare Berggren's [1972] time-scale to the time scale at the front of this volume), calculating sedimentation rates across the K/T boundary as a means of determining the presence or absence of a hiatus has little meaning. The paleomagnetic data at Site 384 are not of high quality, and the correct identification of magnetic intervals is uncertain (Larson and Opdyke, this volume). Thus, attempting to determine the presence or absence of a hiatus at Site 384 on the basis of the paleomagnetic data has dubious validity.

There appears to be no evidence in lithologic data for a hiatus at Site 384. The Cretaceous/Tertiary boundary falls within a zone of smoothly decreasing carbonate content, and there is no background, solution horizon, or other major lithologic break. Only a subtle upward change to slightly darker yellow chalk occurs at the contact. However, small subrounded "clasts" of the darker yellow chalk do occur in the top of the lighter, Cretaceous chalks. These contain early Danian nannofossils and foraminifers, and they indicate either some coring disturbance at the boundary or Paleocene reworking of sea-floor sediment into subsurface burrows by benthic organisms.

It is clear that the detailed biostratigraphy of the Cretaceous-Tertiary boundary is complex and still is poorly known. It is possible that the extinction levels are diachronous for different fossils (Worsley, 1974), that is, "fossil A" might become extinct *before* "fossil B" at the environment sampled by one drillsite but *after* "fossil B" at another drillsite. In light of presently available data, there is no solid evidence that the Cretaceous/Tertiary boundary is not continuous at Site 384.

**The End Cretaceous Extinctions**

A colorful variety of explanations have been advanced for the massive end-Cretaceous extinctions of marine and terrestrial taxa, and a recent review is given by Cita and Premoli Silva (1974). The data from Site 384 have significant bearing on the problem, even though they do not provide a definitive answer.

The magnetic polarities of the Site 384 Maestrichtian-early Paleocene calcareous oozes are in general agreement with recent reversal chronologies such as that of Tarling and Mitchell (1976). However, at Site 384 the Cretaceous/Tertiary boundary appears to occur at the beginning of anomaly 29 rather than in the reversed interval between anomalies 29 and 30 as observed at the Gubbio section in Italy (Larson and Opdyke, this volume). This discrepancy may be due to the lower quality of paleomagnetic data at Site 384. Thus, there is no strong evidence that the extinctions occurred because of a geomagnetic field reversal or even at the beginning of a period of frequent reversals, and it appears unlikely that the geomagnetic field played a role in the extinctions as suggested by Hays (1971).

A sharp rise of the calcite compensation depth (CCD) to sea level was proposed by Worsley (1974) as a cause for the extinctions. However, the lower Danian/upper Maestrichtian nannofossil-foraminifer oozes at Site 384 show no sign of dissolution even though water depth at that time was at least 2.3-3.1 km (Figure 11). Furthermore, calcareous oozes deposited in the late Maestrichtian at Sites 386 and 387 and in the late Maestrichtian to early Danian at Site 385 indicate that the CCD was strongly depressed, not elevated, near the K/T boundary (Figure 3).

In regard to extra-terrestrial explanations (e.g., supernovae, Russell and Tucker, 1971), Worsley (1974) has pointed out that such cataclysms should be geologically instantaneous and beyond stratigraphic resolution and they should occur without foreshadowing in the geologic record. Although some authors have claimed to see a transition zone of measurable length as noted earlier, any such transition is beyond the limits of resolution at Site 384, or else it occurred during a brief hiatus possibly present. Thus a cataclysmic, extraterrestrial cause for the extinctions cannot be excluded on biostratigraphic grounds at Site 384. However, the extinction level occurs early *within* a period of decreasing $CaCO_3$, increasing clay, and increasing montmorillonite/illite ratios (Koch and Rothe, this volume). Thus, if a "cosmic" hypothesis is invoked, it does not appear to have been the cause of the observed geological changes. Furthermore, "catastrophe" and "cosmic" hypotheses are weakened by the recent finding that terrestrial extinctions occurred about 0.5-1.5 m.y. after marine extinctions (Butler et al., 1977).

A global, eustatic lowering of sea level is yet another theory put forth for the extinctions (Hays and Pitman, 1973). The primary effect at Site 384 and generally worldwide would be to bring sources of terrestrial detritus closer to the deep ocean. North and west of Site 384, areas of Cretaceous clay-bearing shelf sediments would be exposed, thus perhaps accounting for the increasing accumulation of clay at the site. Known, nonglacial mechanisms of sea-level change are limited to rates less than 1 cm/1000 yr (Pitman, in press), and it is puzzling how such a gradual sea-level lowering could have triggered a geologically instantaneous event. If one appeals to a much sharper sea-level drop occurring at the end of the Maestrichtian, it leaves unexplained why similar extinctions did not occur during numerous other sea-level fluctuations of comparable magnitude and duration during the Mesozoic and Cenozoic (Vail et al., 1977). Furthermore, although Danian shallow-water seas were reduced in area compared to Maestrichtian seas, they were nevertheless still extensive (Cita and Premoli Silva, 1974; Worsley, 1974).

Climatic deterioration has been one of the most popular theories for the end-Cretaceous extinctions. Such deterioration might result from sea-level regression and increased in epeirogenic uplift and volcanism (Vogt, 1972a). However, oxygen isotope ($\delta^{18}O$) analysis of the calcareous foraminifers at Site 384 shows that water temperature at the surface and at the sea floor on the crest of the *J*-Anomaly Ridge (paleodepths 2.3-3.1 km; Figure 11) and elsewhere in the North Atlantic actually rose significantly across the K/T boundary (Boersma et al., this volume). Boersma et al. also have re-evaluated other $\delta^{18}O$ data for the K/T boundary in the South Atlantic (Saito and Van Donk, 1974) and in the Pacific (Douglas and Savin, 1971, 1975), and on the basis of this re-evaluation and their extensive Atlantic data, they conclude that global warming probably occurred very near the boundary. Thus available evidence does not support the hypothesis of climatic deterioration.

The lithology at Site 384 offers some support to the hypothesis that a global increase in volcanic activity occurred at the time of the Cretaceous/Tertiary transition (e.g., Vogt, 1972a). According to this view the extinctions were caused in some way by (a) global cooling resulting from volcanic ash emission, (b) toxic trace-element contamination of the atmosphere and oceans, (c) other geochemical and climatological effects, for example $SO_2$ and $CO_2$ emission, or (d) some combination of the above. At Site 384, increasing montmorillonite and clinoptilolite concentrations are observed across the Cretaceous/Tertiary boundary (Koch and Rothe, this volume), and these components commonly are produced by the devitrification of volcanic ash (Okada and

Tomita, 1973). However, because no ash layers were observed in the Site 384 calcareous oozes, any volcanic activity could not have been nearby. Results from Sites 382 and 385 suggest that the New England Seamounts were inactive by Maestrichtian time. The age of the Corner Seamounts 500 km southeast of Site 384 is unknown, but they could have formed during the early Tertiary or Late Cretaceous (MacGregor and Krause, 1972). Volcanism in the Brito-Arctic ("Thulean") province 2500 to 3500 km northeast of Site 384 apparently did not begin until late early Danian, and it reached its peak in the late Paleocene to early Eocene (Noe-Nygaard, 1974). The nearest major volcanic province was the Caribbean, 2500 km southwest of Site 384, where deep-sea drilling results suggest increased late Maestrichtian-early Paleocene volcanism (Donnelly, 1973).

Because it was distant from centers of major volcanic activity 65 million years ago, Site 384 could be an ideal "integrator" of volcanic activity, receiving volcanic dust initially ejected to stratospheric heights by major eruptions. However, to be sure that the mineralogical changes at Site 384 are not local, it is necessary to find the Site 384 pattern repeated at other sites far from local volcanic or tectonic influences. This is not an easy task because the Cretaceous/Tertiary transition is generally marked by hiatuses, and core recovery in deep-sea drill holes often is incomplete. At two Pacific drill-sites (47.2, 199) that might be compared with Site 384 in terms of continuity of the sedimentary record and as "global integrators," there is relatively abundant montmorillonite and clinoptilolite in the Maestrichtian and lower Paleocene sections (Rex et al., 1971; Okada and Tomita, 1973). It is not yet known whether the sediments at these three sites record significantly increased volcanism at the end of the Cretaceous as compared with some other, arbitrary time interval, and it will require careful global documentation of trends of volcanism to resolve the question.

Reduced nutrient supply from the nearly leveled Late Cretaceous continents has been invoked in several extinction models (e.g., Bramlette, 1965; Tappan, 1968). Although we cannot ascribe the nutrient depletion to decreased continental erosion, the theme of nutrient depletion in surface water is an attractive possibility in view of the sharp depression of the CCD to more than 5400 meters in the middle to late Maestrichtian and the subsequent rise early in the Paleocene (Figure 3). The CCD depression indicates that most if not all of the western North Atlantic sea floor was above the CCD and that most carbonate and nutrients secreted in the calcareous tests failed to be recycled into the water column. This could have led successively to decreased nutrient supply in the surface water, failure of productivity (and extinctions), earliest Paleocene shoaling of the CCD in response to the decreased supply of carbonate from the surface water, and stabilization of the CCD at some intermediate depth as nutrients again were recycled from the deep water. For this argument to be valid, of course, we should observe marked CCD depression on a global scale, not just in the western North Atlantic. There is some evidence for a similar CCD depression in the Indian and Pacific oceans (Thierstein and Okada, this volume), but unfortunately almost all DSDP drillsites have hiatuses at the boundary and it is difficult to document the level of the CCD. The hiatuses in themselves do not argue against CCD depression because, like the western North Atlantic, most of the unconformities probably were eroded when increased abyssal circulation developed during the Tertiary.

## THE OBSCURE PALEOCENE

Paleocene sediments have been recovered in the deep basin of the western North Atlantic at only two sites (385, 387), although shallower water carbonate oozes are found at Site 384 on the *J*-Anomaly Ridge and at Site 390 on the Blake Nose (Benson, Sheridan, et al., 1976). At other sites the Paleocene is either absent because of erosion by bottom currents or was not recovered because of widely spaced coring intervals (see color foldouts).

The deep-basin sediments at Site 385 and 387 are hemipelagic clays which accumulated at rates of 5 to 6 m/m.y. The clays at Site 387 have black and green-gray colors like those of the Cretaceous black clays and they are locally carbon-rich (up to 1.3%). In contrast, the uncored Paleocene at the deep-basin Site 386 could have accumulated sediment at an average rate no greater than about 2 m/m.y., which suggests it may consist of pelagic clay similar to the underlying Upper Cretaceous multicolored clays. Alternatively, a hiatus may be present, although the Paleocene hiatuses seem to be restricted to the western margin of the basin. Sites with Paleocene sea-floor depths less than 4 km (Sites 10, 384) probably accumulated pelagic carbonates at rates of 6 to 8 m/m.y. (Figure 3).

The increased Paleocene sedimentation rates at Site 385 and 387, as compared with the underlying multicolored clays, may reflect influx of terrigenous sediment from North America. By Late Cretaceous to early Paleocene time, continental debris probably topped the shelf-reef barriers along the margin, and sea-level regression allowed sediment dispersal directly to the deep basin, thus developing an early continental rise. The Bermuda Rise had not yet formed in the Paleocene (see discussion below), but little terrigenous sediment probably reached what is now the central Bermuda Rise, as is suggested by the low sedimentation rates at Site 386.

Reconstruction of sea-floor depths (Figure 3) indicates that the sea floor at Site 387 was the deepest of any of the DSDP sites in the Paleocene. The presence of carbon-enriched black clays in this interval therefore suggests that sluggish circulation and anoxic conditions recurred in the deepest part of the basin during the Paleocene.

## EOCENE SILICEOUS SEDIMENTATION

Eocene sediments have been recovered at 14 sites in the western North Atlantic (see color foldout), and they invariably contain common to dominant quantities of siliceous organisms (radiolarians, diatoms, spicules) and/or porcelanitic chert. Carbonate content is highly variable, depending upon the Eocene sea-floor depth at the site and the mode of deposition (see discussion below). The non-biogenic fraction of the Eocene sediments contains quartz and some feldspar, but it is dominated by clay minerals, of which montmorillonite and illite are the most common.

The middle and lower Eocene sections on the western and central Bermuda Rise (Sites 6, 7, 8, 387, 386) and off the Antilles (Sites 27, 28) contain numerous graded beds, mostly of silt and finer size sediment (Ewing, Worzel, et al., 1969; Petersen, Edgar, et al., 1970; Bader et al., 1970; McCave, this volume), and they accumulated at very rapid rates (Table 1). These beds clearly are turbidites, and they were deposited from turbidity currents presumably originating along the continental margin of the eastern United States and along the Antilles, prior to regional uplift that formed the present Bermuda Rise (Ewing et al., 1969). The abundance of biogenic debris and the fine grain size in most of these turbidites indicate that they are distal deposits composed primarily of pelagic detritus entrained by turbidity currents along their path. The turbidity currents appear to have reached as far as the central Bermuda Rise (Site 386), 1000 km from the continental margin.

In modern analogs, turbidity currents probably originating on the Amazon Cone along the South American continental margin have reached comparable distances and entered the Vema fracture zone in the Mid-Atlantic Ridge (Bader et al., 1970), and sediments originating on the North American continental margin have been transported at least 1600 km and deposited as distal, clayey turbidites in fracture zones near the eastern end of the Nares Abyssal Plain (Figure 1). The top of the turbidite facies on the Bermuda Rise correlates with the uppermost reflector (Horizon $A^T$) of an acoustically laminated sequence (Tucholke, this volume), and it marks the time at which uplift formed the Bermuda Rise and isolated it from turbidity currents. The upper Eocene siliceous sediments above Horizon $A^T$ are oozes and lack graded bedding (notably Site 387).

Horizon $A^T$ at Site 386 on the central Bermuda Rise dates approximately from the middle of the middle Eocene, and this age presumably marks the time of initial uplift forming the rise. At the same time, the main phase of construction of the Bermuda Pedestal probably occurred, and volcaniclastic turbidites eroded at wave base from the pedestal were deposited at Site 386 beginning at the base of the upper Eocene (see Site 386 Report). There is some evidence for younger ages of Horizon $A^T$ toward the west (see color foldouts), which would be expected in offlapping turbidites as the Bermuda Rise was uplifted, but uncertainties in the biostratigraphic ages are large enough that this interpretation can be questioned (Tucholke, this volume).

In Eocene sediments deposited below paleodepths of about 4 km, high carbonate contents are restricted to turbidites (Figure 3), and the non-turbidite upper Eocene sediments at Site 387 have very low carbonate contents. Thus all these sediments were probably deposited below the contemporary CCD. An unusually carbonate-rich sequence of turbidites also comprises the upper lower to lower middle Eocene section at Site 386. It is underlain and overlain by dominantly siliceous turbidite sequences. A large proportion of all three turbidite sequences probably is sediment displaced from numerous adjacent basement highs, and therefore the middle, calcareous sequence may reflect the late early to early middle Eocene depression of the CCD below the average crestal level of these peaks (Figure 3).

Cherts, mostly porcelanitic, are invariably developed in all upper-lower to lower-middle Eocene sediments of the North American Basin, and they correlate with the seismic Horizon $A^C$ (Tucholke, this volume). The chertification appears to be independent of both the depth of the paleodepositional surface and the depositional environment (see Figures 2, 3, and color foldouts in cover pocket). However, one common association is that most of the cherts have interbedded calcareous sediments, emplaced either by turbidity currents or by pelagic deposition above the CCD; notable exceptions are found at Sites 8, 9, and 385. The carbonate fraction of most deep-basin turbidites probably originated along the shallow-water continental margin where dredge samples and shelf and coastal-plain boreholes show a richly calcareous facies (Gibson et al., 1968; Hathaway et al., 1976). It is tempting to attribute the formation of the cherts to their general association with calcareous sediments because the carbonates may have provided the increased permeability requisite to migration of silica-rich pore water (Lancelot, 1973). Also the silica solubility could have been enhanced in the higher $p$H, carbonate environment (Ramsay, 1974). However, these arguments fail to explain the localization of chert horizons in sections such as that at Site 384, where siliceous organisms are present throughout the post-lower-Paleocene calcareous section. Initial preservation of exceptionally large quantities of opaline tests at the Horizon $A^C$ level, possibly because of silica-rich bottom waters produced by subaerial and/or submarine volcanism (Gibson and Towe, 1971; Mattson and Pessagno, 1971; Herman, 1972), may be required. Intense volcanic activity in the northeast Atlantic (Brito-Arctic or Thulean province) and the Caribbean could have been the source of this silica (Figure 12).

The cause of the dramatic change from deposition of mostly non-siliceous sediments to deposition of siliceous biogenic sediments in the Eocene is uncertain. The intermediate level of the CCD certainly enhanced the relative abundance of siliceous debris by reducing carbonate accumulation, but it is also likely that the deep water was generally more silica rich, thus allowing preservation of a larger proportion of siliceous organisms. A causative (or correlative) factor may have been the introduction of cooler bottom water from the newly forming Norwegian-Greenland Sea, which stimulated upwelling of nutrient-rich water and enhanced productivity (Berggren and Hollister, 1974). This cooler water originated in an area of active late Paleocene and early Eocene volcanism and hence of high silica supply (Figure 12). Gibson and Towe (1971) and

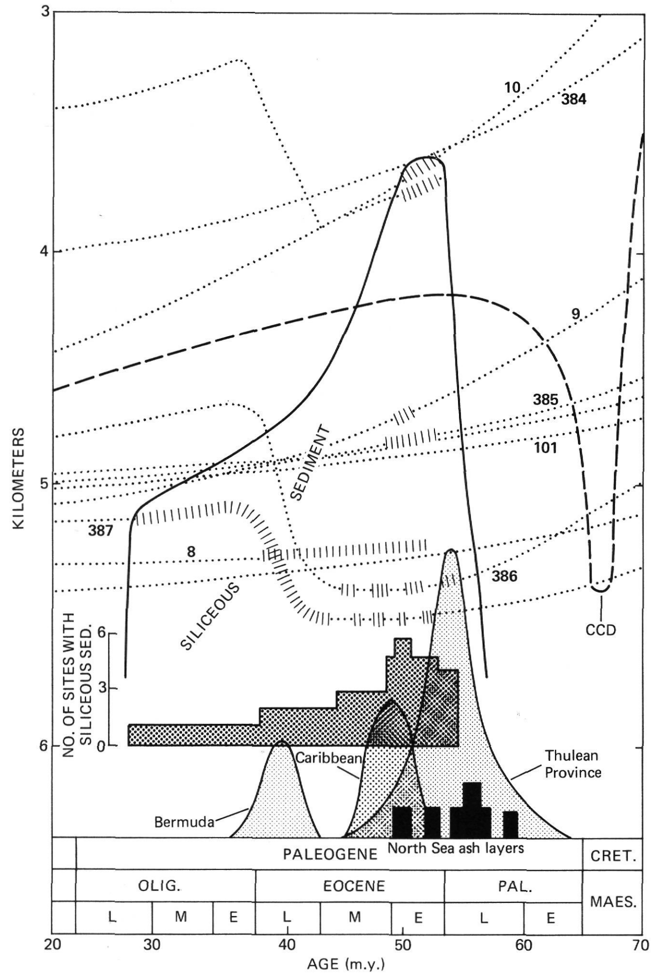


Figure 12. *Age-depth curves of western North Atlantic drill sites showing depositional field of siliceous sediments (from Figure 3), and summary of circum-North Atlantic volcanic activity (schematic and compiled from various sources, except North Sea ash layers from Jaque and Thorwenin, 1975).*

Mattson and Pessagno (1971) suggested that silica and nutrient phosphorous supplied to oceanic surface waters by Eocene volcanism was a primary stimulus to siliceous productivity. In addition, Berggren and Hollister (1974) have summarized evidence that a strong westward-flowing equatorial current system was developed through the Tethys, North Atlantic, and across the Isthmus of Panama into the Pacific, accounting for the circum-global "equatorial" production and deposition of Eocene siliceous sediments.

## EVENTS IN THE DEEP CIRCULATION

One of the most significant features in the sedimentary record of the western North Atlantic is a major erosional unconformity separating Miocene hemipelagic clays from underlying sediments as old as Early Cretaceous along the deep continental margin. The unconformity, termed Horizon $A^U$ (Tucholke, this volume) cuts into the Upper Cretaceous multicolored clays at Sites 105 and 391, into the Cenomanian black clays at Site 101, and down to Horizon $\beta$ in the outcrop area northeast of San Salvador and beneath the Blake Outer Ridge (Tucholke, unpublished profiler data). Thus, at least several hundred meters of sediment were removed along substantial portions of the continental margin. The precise timing and cause of this erosional event still are not well known, but with recent geological and geophysical data, we can limit the possibilities.

Horizon $A^U$ presumably was eroded by an intensified precursor to the present Western Boundary Undercurrent. This horizon truncates both Horizon $A^T$ (late Eocene) and Horizon $A^C$ (early-middle Eocene) just west of Site 8 (Tucholke, this volume) and the main phase of erosion therefore must have occurred no earlier than late Eocene. Subsequently, a geologically rapid change to depositional conditions occurred near the end of the Paleogene, but the exact timing is difficult to document. Borehole results at Sites 101, 105, 106, and 388 indicate that a depositional regime influenced by bottom currents was established by middle Miocene time along most of the continental rise (see color foldouts). However, deposition may have occurred slightly earlier beneath the Blake Outer Ridge, possibly because of bottom-current interaction with the Gulf Stream (Bryan, 1970). For example, at Site 104 rough extrapolation of the 190 m/m.y. middle-Miocene sedimentation rates (Hollister, Ewing, et al., 1972) downward to Horizon $A^U$ suggests that lower Miocene sediment overlies the unconformity. Thus, the erosion forming the observed unconformity beneath the continental rise probably occurred at some time between the late Eocene and early to middle Miocene.

The source of the bottom water generating the early Western Boundary Undercurrent probably was in the subpolar regions of the northern Atlantic Ocean (i.e., the Labrador and Norwegian-Greenland seas; Jones et al., 1970; Vogt, 1972b; and Berggren and Hollister, 1974, among others). General climatic cooling which began late in the Paleocene and persisted through the Tertiary (Schwarzbach, 1961; Boersma et al., this volume) allowed formation of cooler, denser bottom water in the high latitudes, although such water was not as cold or dense as the modern Norwegian Sea overflow or Labrador Sea water. Potential bottom-water influx from the Pacific was blocked by the Antilles (and perhaps Panamanian barriers) in the Caribbean, and there was no significant bottom-water contribution from the Tethys (Berggren and Hollister, 1974). Antarctic Bottom Water (AABW) probably did not flow past the Rio Grande Rise into the North Atlantic in significant quantities until the Miocene, although the early North Atlantic Deep Water (NADW) may have flowed south into the South Atlantic as early as Eocene time (McCoy and Zimmerman, 1977).

In Figure 13 we have summarized known events that may have affected the flow of the mid-Tertiary Western Boundary Undercurrent in the North American Basin. A basic question is whether the intensification of the Western Boundary Undercurrent was a gradual or a geologically rapid process. Because most of the Paleogene record is missing along the continental

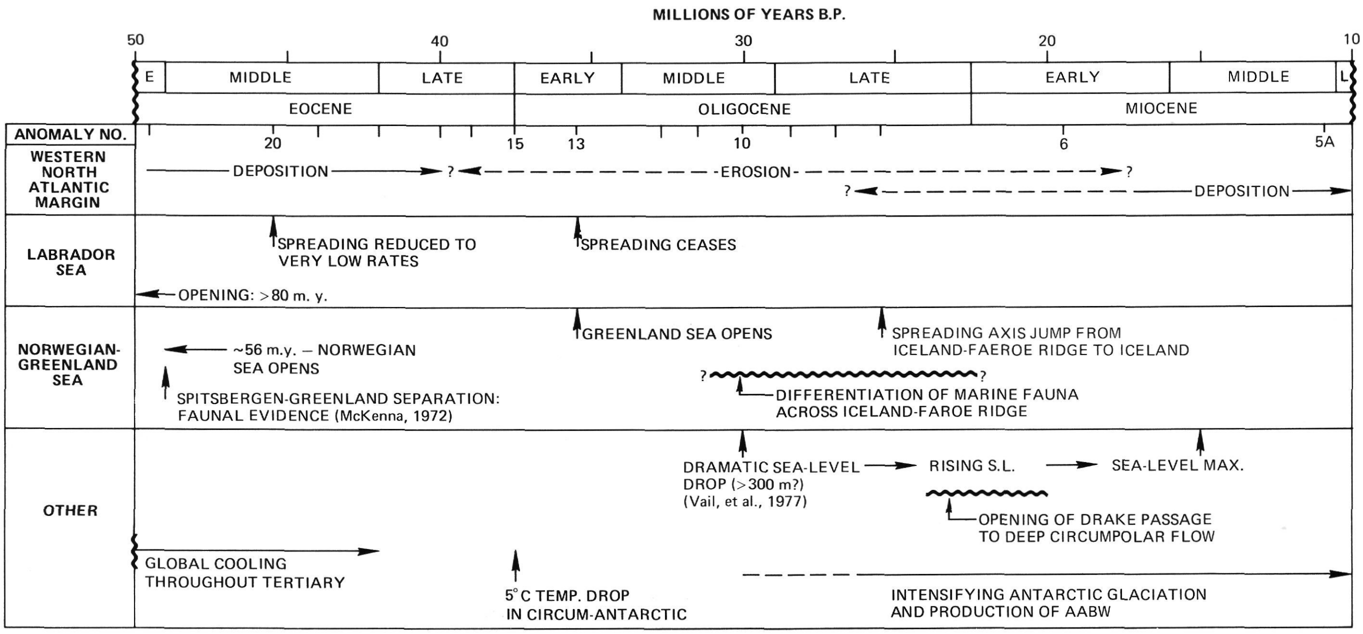


Figure 13. *Summary of Western Boundary Undercurrent effects on continental margin and mid-Tertiary events that may have affected the bottom circulation. Data from Pitman and Herron (1974) and Kristoffersen and Talwani (1977) for Labrador Sea; Talwani and Eldholm (1977) and Talwani, Udintsev, et al. (1976) for Norwegian-Greenland Sea; other sources in text. Magnetic polarity time scale after LaBrecque et al. (1977).*

margin, we unfortunately cannot determine the development of the bottom-water flow from the sediments. Global cooling which began late in the Paleocene (Boersma et al., this volume) was a gradual process, and the rate of formation of bottom water in the Labrador and Norwegian Seas, which began opening prior to 80 m.y. and about 56 m.y., respectively, probably was related directly to this cooling trend.

There are several Paleogene "events" which also may have affected the circulation. Margolis (1976) has noted a sharp temperature drop (~5 °C) near the Eocene-Oligocene boundary in the circum-Antarctic region, and if similar cooling was present in the Northern Hemisphere, it could have caused strong intensification of the early Western Boundary Undercurrent.

In addition, at anomaly 13 time (about 35 m.y.B.P.), the Greenland Sea began to open between Greenland and Spitsbergen (Talwani and Eldholm, 1977), providing an even higher latitude source for cool bottom water than was available in the Norwegian Sea (McKenna [1972] has provided evidence from terrestrial fauna that the Greenland-Spitsbergen separation may have occurred even earlier, about middle Eocene time). However, it is unclear whether much of the bottom water formed in either the Norwegian or Greenland seas prior to the middle Oligocene was able to pass over the Greenland-Iceland-Faeroe Ridge. Distinct differences in marine plankton at Sites 336 and 352 on either side of the Iceland-Faeroe Ridge indicate that the ridge blocked significant exchange of water with the North Atlantic, and perhaps was even emergent, at least during the late-middle and late Oligocene and probably earlier (Vogt, 1972b; Talwani, Udintsev, et al., 1976) (Figure 13). This ridge presumably was constructed by hot-spot volcanism along the spreading-ridge axis between Iceland and the Faeroes (Talwani, Udintsev, et al., 1976) and its development as a circulation barrier either may have resulted from, or been accentuated by, a sharp sea level drop as great as 400 meters late in the middle Oligocene (Vail et al., 1977). The increased continentality caused by the sea-level lowering would be expected to result in still cooler climates (Donn and Shaw, 1977), although there is no such evidence in Margolis' (1976) $\delta^{18}O$ data. A westward jump in the Norwegian Sea spreading axis from the Iceland-Faeroe Ridge to Iceland probably occurred about 25 m.y.B.P. (Talwani and Eldholm, 1977; Figure 13), and subsequent subsidence of the ridge coupled with rising sea-level "permanently" established the Norwegian Sea overflow (temporary reductions in sill depth recurred during Plio-Pleistocene eustatic sea level lowerings).

In view of the evolution of the Greenland-Iceland-Faeroe Ridge, it appears possible that the unconformity along the western North Atlantic continental rise was eroded by: (1) bottom water derived from the Labrador Sea mostly during Oligocene time, (2) combined Norwegian Sea overflow and Labrador Sea water during the earliest Miocene, or (3) a combination of the above.

The first possibility is attractive in that a substantial interval of geologic time (i.e., the Oligocene) was available for bottom currents to erode the large volume of sediment that was removed from the continental rise, and erosion rates need not have been high. There is evidence that bottom currents were geologically significant during the Oligocene. For example, there was a significant shift in the locus of most rapid deposition from the southern to northern Bermuda Rise about this time (Ewing et al., 1970); at Site 9 on the northern rise, sedimentation rates more than doubled between middle Eocene and middle Miocene time (see color foldouts). Also at Sites 382 and 385 along the New England Seamounts, unconformities capped by lower Miocene sediments may have been eroded by abyssal currents. However, if Oligocene erosion by bottom currents derived from Labrador Sea water accounts for Horizon $A^U$, it is puzzling that similar erosion is not observed in the present basin where bottom-water sources, rates of production, and velocities presumably are greater.

One method of evaluating the cause of the erosion is to examine events that would modify an energetic abyssal circulation and produce the dramatic change to rapid deposition that is observed in the latest Paleogene or early Neogene. One such event is the late Oligocene jump in spreading axis from the Iceland-Faeroe Ridge to Iceland (Figure 13). Although this jump could have blocked an existing western passage for Norwegian-Greenland Sea water entering the Atlantic, the existence of such a passage seems unlikely because of the middle Oligocene (and earlier?) faunal differentiation across the Iceland-Faeroe Ridge. There are no other known high-latitude events in the Northern Hemisphere that might have diminished the intensity of abyssal circulation, but two events in the southern hemisphere merit consideration. First is the intensified glaciation of Antartica beginning in the late Oligocene/early Miocene (Hayes, Frakes, et al., 1975; Hollister, Craddock, et al., 1976). The increasing production of early AABW and its northward flow into the Atlantic Basins presumably would have caused the early NADW flux into the South Atlantic to decrease and its flow in the North American Basin to become less intense. Secondly, the Drake Passage probably blocked all but shallow circumpolar flow until the end of the Oligocene (Tucholke et al., 1976; Barker and Burrell, 1976). Until this time, large volumes of early NADW may have been drawn through the North and South Atlantic and into the base of the circumpolar flow east of the Drake Passage. The deep breaching of the Drake Passage at the end of the Oligocene would have replaced this strong draw on bottom water from the Atlantic with continuous circumpolar deep flow, complementing the effect of increased northward flow of the early AABW.

In addition to these considerations purely of bottom circulation, there may have been increased input of terrigenous detritus to the North American Basin because of the sea-level lowering late in the middle Oligocene (Figure 13). The combined effect of this increased sediment supply and the possibly decreasing competence of bottom currents to transport and erode sediments near the end of the Oligocene could account for the change to depositional conditions in the basin.

If the erosion that created the Horizon $A^U$ unconformity is attributed to bottom water originating in the Norwegian-Greenland Seas, it must have occurred during the latest Oligocene or early Miocene after subsidence of the Greenland-Iceland-Faeroe Ridge below sea level. Furthermore, because a depositional regime became established by the middle Miocene or even in the early Miocene, the time available for deep erosion of the continental rise is at most a few million years, and rates of erosion could have approached or exceeded 100 m/m.y. Although such rapid erosion could be considered geologically "catastrophic," it is an attractive possibility because of the timing of the event and the nature of the bottom-water source. What then caused the dramatic change to a depositional regime in the early to middle Miocene? The events at high southern latitudes discussed earlier could have had an effect, but it is uncertain that they occurred late enough to have been a significant factor. Another possibility is that the very strong thermohaline gradients created by the introduction of cool, dense bottom water and responsible for the strong abyssal circulation gradually were dissipated, and an equilibrium was reached producing moderate abyssal circulation similar to that in the present basin.

In order to clarify the timing (and thus the bottom-water sources) involved in the erosion of Horizon $A^U$ along the continental rise, it will be necessary to continuously core across the unconformity in an area peripheral to the main erosional zone. In such an area the hiatus would be much shorter and would more closely pinpoint the erosional event. DSDP Site 8 is in an appropriate location, but unfortunately the cored intervals are much too widely spaced for the site to be of any value in studying the unconformity.

Where was all the sediment that was eroded from the continental rise ultimately redeposited? As noted earlier, the northern Bermuda Rise could be one locus of deposition, possibly receiving sediment derived from the Western Boundary Undercurrent near the Antilles and carried in north-flowing bottom currents along the western flank of the Mid-Atlantic Ridge. Some of the sediment was deposited to form the western sector of the Greater Antilles Outer Ridge north of Puerto Rico (Tucholke and Ewing, 1974), and much sediment may have been deposited in thinner drifts throughout the basin. A significant part of the sediment probably was transported in the South Atlantic or beyond.

There is a clear change in depositional style between Neogene sediments and older sediments along the deep-water continental margin, reflecting the influence of abyssal currents. The Neogene sediments commonly show effects of current-controlled deposition (moats, differential accumulation, migrating sediment waves) in marked contrast to older sediments which were ponded, draped (pelagic), or deposited in submarine fans. The strong late Paleogene to early Neogene erosion also markedly altered the morphology of the contemporaneous continental rise, and the present continental rise is in essence a Neogene feature that has developed under the influence of abyssal currents. Accumulation rates have ranged from 20 to 200 m/m.y. (Sites 104, 105, 106; Hollister, Ewing, et al., 1972). In addition, depositional outer ridges were created or rapidly expanded during the Neogene. Examples include the Blake-Bahama Outer Ridge (Ewing and Hollister, 1972) and the Caicos and Greater Antilles Outer Ridges (Tucholke and Ewing, 1974).

In the remainder of the basin similar changes in depositional style are less dramatic and may somewhat predate the Neogene. Throughout the entire basin, siliceous sediments like those deposited during the Eocene/Oligocene are only locally observed in the Neogene, and the siliceous debris probably has been diluted by rapid accumulation of hemipelagic sediments controlled by bottom currents. Similar dilution of carbonate is observed, even though the CCD has been relatively deep during the Neogene (4.5-5.5 km; Figure 3).

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## APPENDIX A
## CONSTRUCTION OF SEA-FLOOR PALEODEPTH CURVES

The empirical curve of subsidence for Atlantic Ocean crust, corrected for sediment overburden, is shown in Figure A1, based on tabular data in Columns 7 and 8 of Table A1. Crustal depths are modified somewhat from Sclater and Detrick (1973) and also include Leg 43 drillsites. Topographic corrections are made for the difference in elevation between site basement depth and regional basement depth at the site as determined from seismic profiles; corrections are made to the nearest 50 meters and corrections less than 50 meters are ignored. In addition, corrections were made at Sites 386 and 387 to remove the effect of the uplift of the Bermuda Rise (Table A1), values were determined from the difference in depth of Horizon $A^T$ at the two sites and under the central Hatteras Abyssal Plain in the middle Eocene, based on the assumptions that the turbidites comprising Horizon $A^T$ were horizontal at that time, that Horizon $A^T$ is isochronous, and that the Hatteras Abyssal Plain has had a normal subsidence history. In the Atlantic age/depth curve, the mean depth of the ridge crest is assumed to be 2700 meters.

The crustal subsidence curve in Figure A1 and the data in Columns 5 and 9 of Table A1 were used to determine the sea-floor paleodepths Figure 3. Paleodepths were determined from $D = Z - S\,(\rho_m - \rho_s)/(\rho_m - \rho_w)$, where D = sea-floor depth, Z = empirical crustal depth, $\varrho m$ = density of mantle (3.3 g/cm³ assumed), $\rho_s$ = density of sediments (1.8 g/cm³ assumed), $\rho_w$ = density of water (1.03 g/cm³), and S = thickness of sediments; local (Airy) compensation of the sediment load is assumed. Confidence limits on paleodepths are difficult to assess, but are probably about ± 150 meters for sites on normal sea floor (van Andel et al., 1975) and somewhat greater for sites with anomalous tectonic history.

The lithofacies indicated on the age/depth curves in Figure 3 are taken from the color foldouts in the back of this volume. The boundary between calcareous and non-calcareous sediments was set at 20 per cent $CaCO_3$ except in hemipelagic sediments where terrigenous dilution was taken into account and the boundary was set at 2 per cent $CaCO_3$ (van Andel, 1975).

Many of the western North Atlantic drill sites were not used in Figure 3 because of sparse core recovery or because of their uncertain crustal age, basement depth or tectonic history. Basement ages for sites used in Figure 3 which did not reach basement are estimates based on the position of the site in the *M*-series anomalies and on the van Hinte (1976a, b) time scales. The revisions of this time scale suggested by Leg 43 data are not large (Vogt and Einwich, this volume). Site data are given in Table A1. Comments on several of the sites used and not used in Figure 3 are given below.

**Sites Illustrated**

Site 8 — Coring was sparse at this site and the paleodepth curve is shown only to the depth cored (lower Eocene).

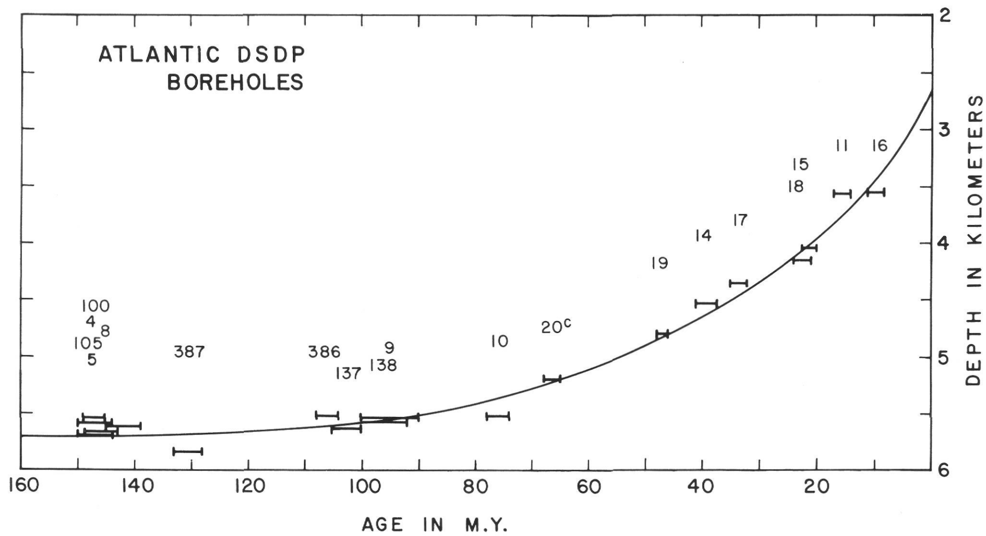


Figure A1. *Empirical age-depth curve for the Atlantic, based on data in Table A1.*

**TABLE A1**
**Derivation of Empirical Curves of Subsidence for North Atlantic Ocean Crust**

| 1<br>Site | 2<br>Sea Floor Depth (m) | 3<br>Sediment Thickness (m) | 4<br>Sediment Correction (m) | 5<br>Sediment-Free Basement Depth (m) | 6<br>Topographic Correction (m)[a] | 7<br>Mean Corrected Basement Depth (m) | 8<br>Age (m.y.) | 9<br>Age Assumed (m.y.) |
|---|---|---|---|---|---|---|---|---|
| 4[b] | -5320 | 400[c] | -264 | -5584 | 0 | -5584 | Near anomaly M-25[d] | 147 |
| 5[b] | -5354 | 500[c] | -330 | -5684 | 0 | -5684 | Near anomaly M-25[d] | 147 |
| 8 | -5169 | 660[c] | -436 | -5605 | 0 | -5605 | Near anomaly M-23[d] | 142 |
| 9 | -4981 | 834 | -551 | -5532 | 0 | -5532 | Late Cretaceous | 95 |
| 10 | -4712 | 451 | -298 | -5010 | -500 | -5510 | Early Campanian (74-78) | 76 |
| 11 | -3571 | 280 | -185 | -3756 | +200 | -3556 | Early to Middle Miocene (14-17) | 16 |
| 14 | -4346 | 107 | -71 | -4417 | -100 | -4517 | Late Eocene (37.5-41) | – |
| 15 | -3938 | 141 | -93 | -4031 | 0 | -4031 | Early Miocene (20-22.5) | – |
| 16 | -3526 | 175 | -116 | -3642 | +100 | -3542 | Late Miocene (8-11) | – |
| 17 | -4277 | 93 | -61 | -4338 | 0 | -4338 | Early to Middle Oligocene (32-35) | – |
| 18 | -4022 | 178 | -118 | -4140 | 0 | -4140 | Late Oligocene-Early Miocene (21-24) | – |
| 19 | -4685 | 141 | -93 | -4778 | 0 | -4778 | Middle Eocene (46-48) | – |
| 20C | -4484 | 72 | -48 | -4532 | -650 | -5182 | Late Maestrichtian (65-68) | – |
| 99A | -4914 | 350[c] | -231 | -5145 | – | – | Anomaly M-24[d] | 145 |
| 100[b] | -5325 | 314 | -208 | -5533 | 0 | -5533 | Early Oxfordian (145-149) | 148 |
| 101 | -4868 | 1050[c] | -694 | -5562 | – | – | Jurassic Quiet Zone[d] | 155 |
| 105 | -5251 | 623 | -412 | -5663 | 0 | -5663 | Oxfordian (143-149) | 146 |
| 137 | -5361 | 397 | -262 | -5623 | 0 | -5623 | Late Albian (100-105) | – |
| 138 | -5288 | 437 | -289 | -5577 | 0 | -5577 | Cenomanian (92-100) | – |
| 384 | -3910 | 325 | -215 | -4125 | – | – | Late Barremian (115-118) | 107[e] |
| 385 | -4956 | 1350[c] | -892 | -5848 | – | – | Between M-11 and M-12[d] | 128 |
| 386 | -4783 | 964 | -637 | -5420 | +600 | -5520[f] | Early Albian (104-108) | 107 |
| 387 | -5118 | 792 | -524 | -5642 | +200 | -5842[f] | Late Berriasian to E. Valanginian (128-133) | 131 |

[a]Topographic correction is made where seismic profiles show sites were drilled on basement features with basement depth markedly different from regional basement depth. Correction equals regional crustal depth corrected for sediment overburden minus sediment-free basement depth (nearest 50 m).
[b]Site not shown in Figure 3. Sea-floor paleodepth curve very similar to Site 105.
[c]Calculated from seismic data; basement not reached.
[d]Basement not reached; assumed age based on position in anomaly sequence (van Hinte, 1976a, b, time scale).
[e]Petrologic data at Site 384 indicate basement was near sea level until early Albian.
[f]Includes correction of –700 meters at Site 386 and –400 meters at Site 387 to remove effect of late Eocene uplift of Bermuda Rise.

Site 9 — There is substantial uncertainty in the age of basement at this site. However, a shallow Late Cretaceous CCD is required by core data even with departures of $\pm 10$ m.y. from the assumed basement age of 95 m.y.B.P.

Site 10 — High carbonate content in all cores recovered at this site constraints the minimum depth of the latest Cretaceous to Recent CCD. Abundant siliceous tests in lower Eocene sediments also provide a limit on the siliceous depositional field.

Site 384 — The basement subsidence curve for this site deviates from that depicted in Figure A1 because of the *J*-Anomaly Ridge obviously has experienced an unusual subsidence history (see text). The youngest reefal material at Site 384 (Aptian-Albian) has been subaerially weathered, indicating that basement was near sea level at least until the beginning of the Albian. Eustatic sea-level changes are ignored in this interpretation.

Site 385 — The paleodepth curve for this site on the deep flank of Vogel Seamount is derived from the empirical curve in Figure A1, but it is extended back only through the late Maestrichtian calcareous zone. If crustal subsidence had been along a steeper gradient (because of local elevation due to seamount volcanism) then the Maestrichtian sea floor would have been shallower and the CCD drop and rise even sharper than that depicted in Figure 3. True sediment thickness above original basement at Site 385 is uncertain, but departures of $\pm 50$ per cent in thickness from the assumed 1350 meters have a negligible effect on the curve depicted. A $\pm$ 10-m.y. age variation causes variations of about $\pm$ 150 meters in the Maestrichtian part of the curve.

Site 386 — The seafloor paleodepth curve is modified to reflect a 700 meter uplift of the site during formation of the Bermuda Rise. The value of 700 meters is derived from differences in middle Eocene elevation of Horizon $A^T$ at Site 386 and under the Hatteras Abyssal Plain; Horizon $A^T$ is assumed to be isochronous (middle Eocene), a formerly horizontal turbidite depositional surface (see Tucholke, this volume), and the Hatteras Abyssal Plain is assumed to have experienced a normal subsidence history. In addition, the Bermuda Rise uplift is assumed to have occurred completely within the late Eocene. Because the nature of the (thermal?) event that formed the Bermuda Rise is uncertain, the rate of post-Eocene subsidence also is poorly known. However, a qualitative confirmation of the general character of the whole curve is demonstrated by the derived original basement depth of Site 386 (about 3300 m; Figure 3); this is about 600 meters below the average ridge-crest elevation of 2700 meters and matches the 600-meter topographic correction for the site.

The Albian-Cenomanian sediments at Site 386 contain calcareous interbeds, most if not all of which are graded (McCave, this volume). They are interpreted as turbidites originating from adjacent basement highs and deposited at Site 386 below the CCD. If some of these beds are pelagic and not displaced, the Middle Cretaceous CCD should be accordingly deeper in Figure 3.

Eocene to Miocene sediments at this site also contain carbonate interbeds. These are clearly associated with turbidites that were displaced from adjacent shallower areas (Bermuda Pedestal and adjacent peaks).

Site 386 Peak — A basement peak just 6 km south of Site 386 rises some 1800 meters above the level of basement at the site. It is assumed to have experienced the same relative subsidence history as Site 386 and to have accumulated carbonate which was displaced to form calcareous turbidites in the valley at Site 386. The inferred episodes of carbonate accumulation are shown on the Site "386-Peak" curve, and they provide additional control on the fluctuations of the CCD.

Site 387 — Like Site 386, Site 387 on the Bermuda Rise is assumed to have experienced uplift during the late Eocene (400 m uplift based on Horizon $A^T$ levels). The resultant curve implies crustal depth of about 3100 meters in the late Berriasian/early Valanginian, about 400 meters below the assumed average ridge-crest depth of 2700 meters. The 200-meter-topographic correction for Site 387 crust reduces the discrepancy to 200 meters, which is within acceptable limits. The Eocene calcareous beds at Site 387 appear to be turbidites displaced from shallower water along the continental margin.

**Sites Not Illustrated**

Sites 4, 5, and 100 — These sites have paleodepth curves very close to that of Site 105 and consequently are not shown in Figure 3. Sites 4 and 5 are unusual in that they contain Middle and Upper Cretaceous calcareous beds. Some of these beds are graded and/or contain displaced shallow-water carbonate. However, other beds are nannofossil marls which are not obviously graded and may not be displaced. Thus, local depression of the CCD near the Bahama Banks could have occurred during this period. It is also possible that these three sites (and Hole 99A) have experienced and anomalous amount of subsidence because of their position near the Bahama Banks.

Sites 6, 7, and 28 — These sites were cored very sparsely, but sediments recovered confirm the siliceous depositional field in the Eocene (Figure 3). The Eocene siliceous sediments also contain carbonate-rich turbidites probably displaced from shallower water near the continental margin.