46. GENERAL SYNTHESIS OF CENTRAL AND SOUTH ATLANTIC DRILLING RESULTS, LEG 39, DEEP SEA DRILLING PROJECT

Peter R. Supko,[1] Scripps Institution of Oceanography, La Jolla, California
and
Katharina Perch-Nielsen, Eidg. Techniche Hochschule, Zürich, Switzerland

INTRODUCTION

Leg 39 drill sites were selected to address several partially interrelated geologic problems. The Vema Fracture Zone (Site 353) was drilled in an attempt to recover igneous rocks of crustal Layer 3 in an area where these deeper rocks might be exposed at drillable depths as a result of fracture zone tectonics (Bonatti, 1971). The crest portion of the Ceará Rise (Site 354) was drilled to detail its sedimentation history, to obtain as complete a low-latitude biostratigraphic section as possible, and to determine the nature of basement under this physiographic and structural high in the Guiana Basin. Sites 355 and 358 were drilled to date oceanic magnetic anomalies and to detail the sedimentation histories of deepwater locations in the Brazil and Argentine basins. The São Paulo Plateau site (356) was selected to sample sediments of ages equal to and older than the presumed Aptian salts which form an important diapiric structural province on the plateau and on the opposing Angolan borderland; to determine the nature of several subbottom reflectors of regional extent; and to attempt to sample basement. Site 357 was chosen in the vicinity of the largely unsuccessful Leg 3 Site 22, in order to obtain a very complete sedimentary succession on the little understood Rio Grande Rise, and to attempt to sample basement. Site 359, on the crest of a high in the seamount province of the Walvis Ridge, was an unplanned site of opportunity, chosen to use drilling time remaining after being blown off Site 358.

Detailed observations are presented in the Site Report chapters of this volume; discussions appear in these same chapters and in other chapters written by both shipboard scientists and shore laboratory collaborators. This chapter is not an attempt to assess the state of overall knowledge of the geological history of the South Atlantic as reflected by drilling results; such an attempt will be more appropriate when the critical results of Legs 40 and 41 are published. The following discussions center on those aspects of marine geology affected by Leg 39 results, such as the nature and occurrence of hiatuses, sedimentary successions at oceanic and marginal sites, nature of oceanic rises, nature of seismic reflectors, and the dating of oceanic magnetic anomalies. The *Glomar Challenger* track and Leg 39 drill sites are shown in Figure 1, which also shows the location of pertinent sites drilled on Legs 3, 36, 40, and 41.

[1]Present address: National Research Institute for Oceanology, P.O. Box 320, Stellenbosch, South Africa.

HIATUSES

That the sediment record of ocean basins is not continuous at all locations was recognized from the earliest DSDP results, and it soon became apparent that regional temporal and spatial trends existed (e.g., Moore, 1972). The regional importance of Atlantic hiatuses was noted by Pimm and Hayes (1972), whose compilation of Leg 14 and earlier data indicated that much of the Tertiary section was missing at many sites. They further noted that the hiatuses represented longer time spans in the western portion of the basin, implying current control by eroding westward-intensified thermohaline currents. Kennett et al. (1972, 1974) interpret regional mid-Tertiary hiatuses in the southwest Pacific and Southern oceans as resulting from initiation of the circumpolar current. Davies et al. (1975) summarize the data on Indian Ocean hiatuses. Rona (1973) notes that several major hiatuses have global significance and suggests they result from tectono-eustatic changes in sea level. In a very important paper, Moore et al. (in press) synthesize the data on all hiatuses reported by DSDP through Leg 31, and discuss their spatial and temporal distributions in terms of possible causes. Fischer and Arthur (in press) plot hiatuses as a function of geologic time for five major ocean basins. Figure 2 is a modification of their figure illustrating the ubiquity of hiatuses at certain times; data are shown as percentage of cored intervals of any given age, in which gaps representing more than 1.5 m.y. have been recognized, plotted at 2.5-m.y. intervals. Data for Legs 1-30 are from Initial Reports, and data for Legs 31-41 are from summaries published in Geotimes. On a world ocean basis, hiatuses were more common in the Turonian, late Maestrichtian-Paleocene, late Paleocene-early Eocene, at the Eocene/Oligocene boundary, and into the Oligocene. Better shown in the plot of Moore et al. (in press) is a mid- to late Miocene hiatus, occurring in all basins but most pronounced in the South Atlantic.

Hiatuses encountered on Leg 39 are shown in Figure 3, which also contains information from several Leg 3 sites (Maxwell, Von Herzen, et al., 1970), Leg 40 sites (Bolli, Ryan, et al., 1975), and Leg 36 sites on the Falkland Plateau and Malvinas Outer Basin (Barker, Dalziel, et al., 1974). Solid bars indicate hiatuses proved by missing biostratigraphic zones in continuously cored sequence; the wiggly lines indicate hiatuses inferred in undatable sediment sequences or

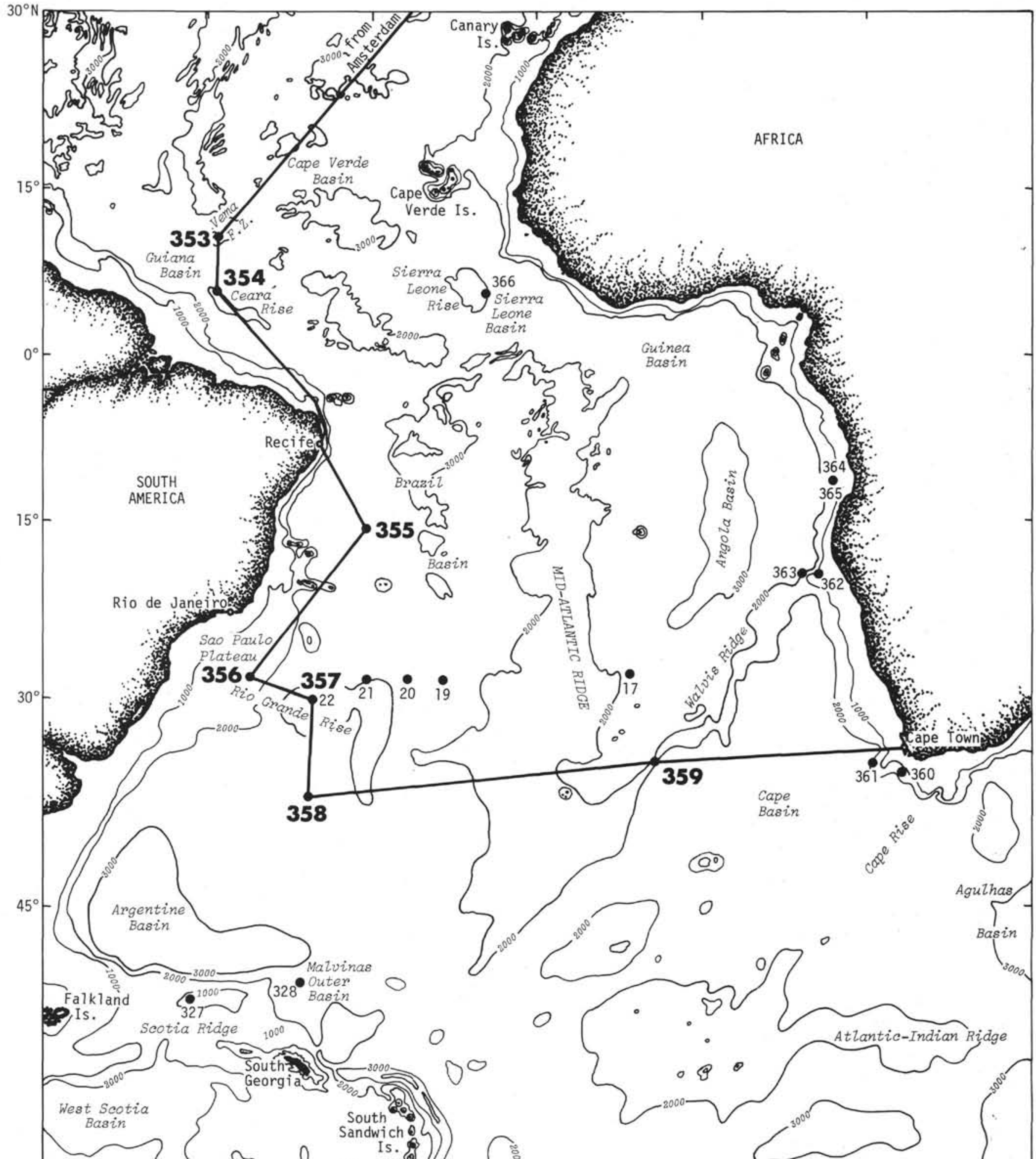


Figure 1. *Base map of locations of selected DSDP sites drilled in the South Atlantic on Legs 3, 36, 39, 40, and 41.*

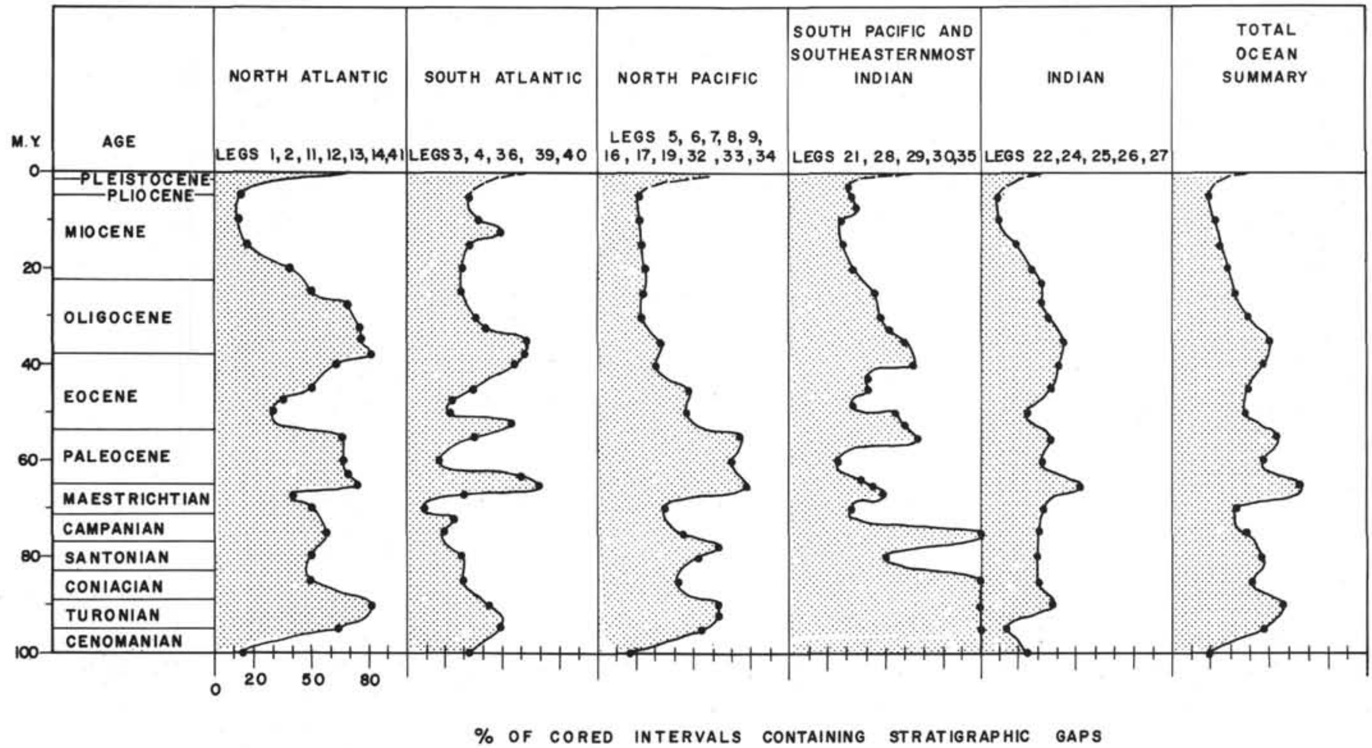


Figure 2. *Worldwide stratigraphic hiatuses based on DSDP cores. Values plotted (at2.5-m.y. intervals) are percentages of cored intervals in which gaps of 1.5 m.y. or more have been recognized (modified from Fisher and Arthur, in press, with kind permission).*

cored or uncored intervals where the rates of deposition would be much lower than the average rates typical of the expected facies[2]. Hachures designate periods of increased dissolution in the southeastern Atlantic (Bolli, Ryan, et al., 1975). Horizontal lines indicate the oldest sediment recovered at each site, and hachuring under the line indicates that basalt basement was cored.

Cretaceous hiatuses occur centered in the Turonian and Campanian. Lower Tertiary hiatuses occur at the Maestrichtian/Paleocene boundary and the Paleocene/Eocene boundary. The Eocene/Oligocene boundary is marked by hiatus at a number of sites, and the period of non-deposition or erosion continued through the Oligocene at a number of them. The mid- to late Miocene was a time of major hiatus in the South Atlantic.

The Cenomanian, Turonian, and (?) Coniacian section is missing at Site 363 in the Frio Ridge section of the Walvis Ridge, and the Cenomanian and lower Turonian are missing at the São Paulo Plateau (Site 356). The numerous erosional contacts in the uppermost limestones of late Albian age at Frio Ridge and the mid-Turonian conglomerates at the São Paulo Plateau site indicate that erosion caused this partially coeval hiatus at the two sites; the sites are on and adjacent to topographic highs, probably parts of the same east-west-trending structural feature which formed a barrier to circulation between the northern and southern parts of the South Atlantic until early Senonian time. A mid-Albian to Santonian hiatus occurs on the east Falkland Plateau (as well as at Site 249 in the Mozambique Basin), and may represent bottom-current erosion at a time when the eastern part of the Falkland Plateau and the southern tip of Africa had separated enough to allow a major circulation pattern to develop.

Slower than normal accumulation occurred in the Campanian on the São Paulo Plateau and the Rio Grande Rise. Sediments at both sites show similar trends from semi-restricted and periodic reducing conditions and terrigenous input in the Santonian, to progressively more oxidizing and open marine conditions through the Santonian and into the Campanian. On the basis of these similarities, we believe the two features had a similar subsidence history. If we then assume that the east-west structural portion of the Rio Grande Rise subsided as a unit block and followed a subsidence curve similar to that applicable to basaltic ocean crust (Sclater et al., 1971), we can use certain indicators of depth (discussed in a later section) to define a subsidence history for the rise. According to this subsidence history, the main plateau of the rise, at a present depth of 1800 meters, subsided below sea level in the Campanian. We suggest that the Campanian hiatuses at Sites 356 and 357 are associated with extreme dissolution of foraminifers, and are local features resulting from increased current activity associated with the submergence or from a CCD effect. Davies et al. (1975) cite numerous Upper Cretaceous

[2]Note that in the text the word "hiatus" is used in a general sense to refer to both cases; the reader should refer to Figure 3 for specific cases.

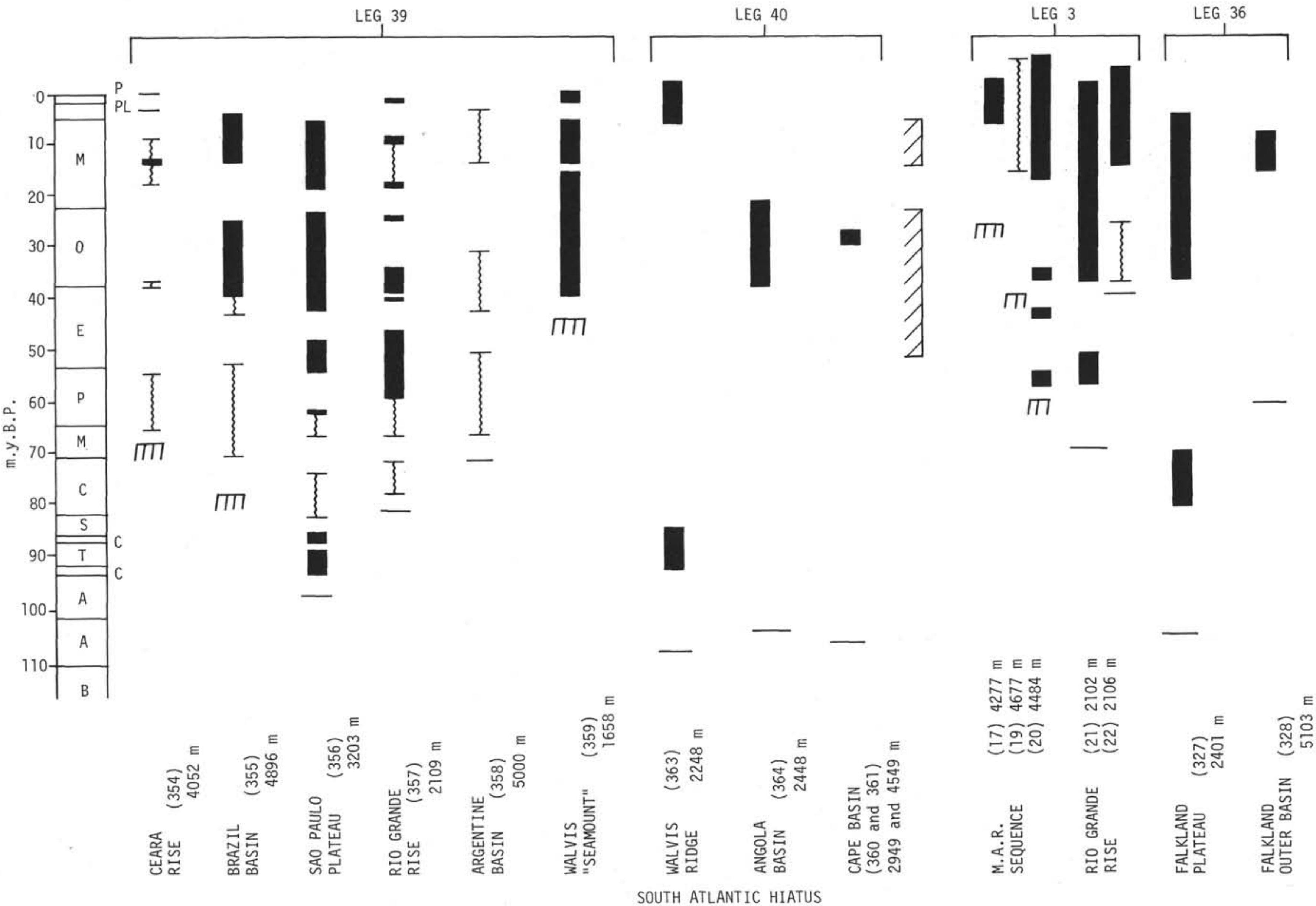


Figure 3. *Hiatuses at Leg 39 sites and other selected South Atlantic sites. Solid black lines indicate actual hiatuses, wavy lines represent inferred hiatuses, and/or abnormally low rates of sedimentation in sections not continuously cored. Short horizontal lines designate the oldest sediments reached; hachures below indicate basement was reached. Hachured bars to the right of the Leg 40 sites indicate times of general dissolution in the southeast Atlantic.*

unconformities, and Douglas et al. (1973) note an unconformity ending in the Campanian. However, sampling of the Upper Cretaceous oceanic record is not sufficient to attempt to attach regional or global significance to these scattered observations.

Times of slow sedimentation were found at the Cretaceous/Tertiary boundary at the five Leg 39 sites extending into the Cretaceous. Paleocene sediment accumulation rates were high on the São Paulo Plateau, but very low or nil at all other sites (Figure 4). The Paleocene/Eocene boundary is represented by hiatuses at all sites except 354, where sedimentation resumed in the late Paleocene, but the boundary was not recovered. Paleocene hiatuses are present at the two Leg 3 sites (20 and 21) which penetrated the Paleocene.

The latest Maestrichtian through earliest Eocene was a time of worldwide maxima in hiatuses (Figure 2; Moore et al., in press); these authors find that on a worldwide basis there is a secondary maximum of hiatuses at the Paleocene/Eocene boundary. Leg 39 drilling has confirmed that the Paleocene/Eocene boundary is represented by a hiatus at least in the western part of the South Atlantic basin. The hiatus is present at shallow sites (São Paulo Plateau and Rio Grande Rise), as well as deep sites (Brazil Basin, Argentine Basin), a characteristic also noted for the lower Tertiary hiatus in the Indian Ocean (Davies et al., 1975).

We have little knowledge of paleoceanographic conditions that may have caused the hiatuses in the Late Cretaceous and early Tertiary. By the end of the Cretaceous and beginning of the Tertiary, the South Atlantic was a fairly wide basin (Ladd et al., 1974); the Rio Grande Rise had subsided as a topographic barrier

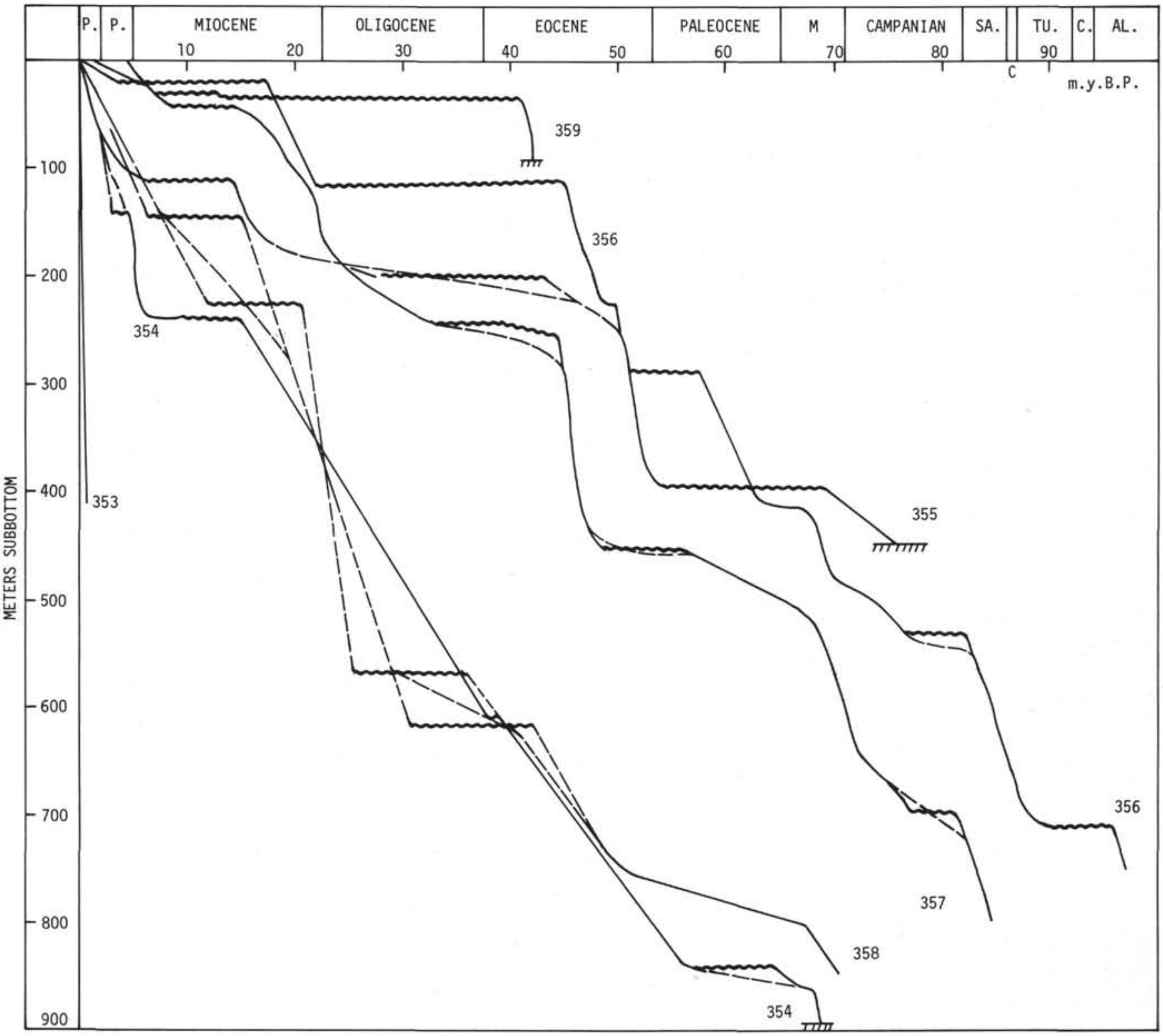


Figure 4. *Sedimentation rates at Leg 39 drilling sites. For specific details of any site, refer to the respective site report chapter.*

to surface circulation, and surface water mass exchange had begun between the North and South Atlantic (Reyment and Tait, 1972), producing a basin susceptible to the development of meridional flow (McCoy and Zimmerman, this volume). Paleotemperature data (Saito and van Donk, 1974; Savin et al., 1975) show a drop in bottom temperature at the end of the Cretaceous and beginning of the Tertiary, probably indicating the arrival of cooler waters from high southern latitudes, which in turn may have resulted from a change in heat distribution patterns caused by global tectonic events. Water would sink and move north as a bottom current, causing hiatus formation by erosion or dissolution. Strong bottom currents may have continued through the Paleocene as a result of global regression (Rona, 1973), and may have been slightly intensified at the end of the Paleocene and the beginning of the Eocene, in consequence of circulation changes associated with the separation of Australia and Antarctica (Weissel and Hayes, 1971). Current control is implied in the fact that the South Atlantic lower Tertiary hiatus is well developed only at the western basin sites drilled on Leg 39. At Site 20, the hiatus is restricted to the lower Paleocene, and deposition was continuous in the eastern South Atlantic Basin, although sedimentation rates were low across the Cretaceous-Tertiary boundary (Bolli, Ryan, et al., 1975). Such would be expected if the eroding (and dissolving) currents were intensified in the western part of the basin, as is the case with the present-day Antarctic Bottom Water (Wüst, 1937).

Worsley (1974) has proposed a terminal Cretaceous event which entailed worldwide shoaling of the carbonate compensation level to very shallow depths (approaching the photic zone), which in turn caused a time of non-deposition of pelagic carbonate facies and thus a hiatus at the Cretaceous/Tertiary boundary. Such a theory is not in line with the continuously deposited carbonate across the boundary at Site 356 and in the southeast Atlantic (Bolli, Ryan et al., 1975) and, especially, with the complete fossil suite of calcareous nannofossils and foraminifers across the boundary at Site 384 in the western North Atlantic (Tucholke, Vogt, et al., 1975). This section accumulated at abyssal depths, and has been deep ever since.

The lower Tertiary hiatus ends abruptly in the uppermost Paleocene at the Ceará Rise and in the lowermost Eocene at all other sites.

The middle Eocene was a time of high sedimentation rates at all Leg 39 sites (Figure 4). It was a time of a slight warming trend superimposed on the gradually deteriorating global Tertiary climate (Savin, et al., 1975; Boersma and Shackleton, this volume), and was generally a time of high productivity of siliceous organisms (Ramsay, 1973). The Antarctic continent was probably largely ice-free during the Eocene; the slight amount of ice-rafted erratics reported by Margolis and Kennett (1970) probably represents only very local glacial conditions in west Antarctica (Kennett and Schackleton, 1976).

The hiatus at the Eocene/Oligocene boundary is very well developed in the South Atlantic; it is present or inferred at all western Atlantic sites and the two Leg 3 sites on the Rio Grande Rise, but carbonate sedimentation continued across the Eocene/Oligocene boundary at Leg 3 Sites 19 and 20. The hiatus continues through most or all of the Oligocene in the Brazil and Angola basins, on the Falkland Plateau, São Paulo Plateau, and Walvis "seamount." In the Argentine Basin, a hiatus may only be inferred across the Eocene/Oligocene boundary and in the lower Oligocene; a high sediment accumulation resumed (Figure 4) at the beginning of the late Oligocene. Generally lower accumulation rates are reported for the Eocene at Leg 36 sites, and a hiatus representing much of the Oligocene appears likely at Site 329 in the eastern Falkland Plateau (Barker, Dalziel, et al., 1974). It is possible that much of the sediment missing here was transported north into the Argentine Basin. Only a short upper Oligocene hiatus is present in the Cape Basin (Sites 360 and 361), although most of the Eocene and Oligocene (and mid- to late Miocene) were times of carbonate dissolution in the eastern South Atlantic basin (Melguen et al., 1975).

The hiatus across the Eocene/Oligocene boundary was of global importance (see Figure 2; also Moore et al., in press). Its onset was coincident with a sharp drop of bottom water temperatures by about 4°C, which may have occurred in less than 100,000 years, i.e., geologically "instantaneously"; it probably reflects the first important formation of sea ice around the Antarctic continent, alghough the continent itself probably remained ice-free until the Miocene (Kennett and Shackleton, 1976). The sharp temperature drop in the late Eocene is indicated for the South Atlantic by isotopic paleotemperature studies from Site 357, Rio Grande Rise (Boersma and Shackleton, this volume). Sinking and equatorward flow of the cold bottom water initiated a thermohaline circulation similar to that of today, and represented the beginning of the oceanic "psychrosphere" (Benson, 1975). The Oligocene was a time of major worldwide regression, and Rona (1973) points out that periods of regression appear to be periods of maximum development of strong bottom currents capable of erosion, as indicated by maxima in hiatus abundance.

It had earlier been noted that the central Pacific and South Atlantic were both regions of rapid pelagic carbonate accumulation in the Oligoocene (Davies et al., 1975), but this no longer holds as a generality for the South Atlantic after Leg 39 drilling, although Sites 17, 19, and 20 of the Mid-Atlantic Ridge sequence were accumulating calcareous ooze in the Oligocene, as was the Ceará Rise. We have calculated the depth of the deposition area for Sites 17, 19, 20, 355, 358, and 354 for the mid-Oligocene (30 m.y.B.P.) according to the method and assumptions outlined by Berger (1972); the data are in Table 1. On the basis of similar calculations, Berger and von Rad (1972) show the Oligocene CCD deepening to 4000 meters; Sites 355 and 358 were significantly deeper than the Leg 3 sites in the mid-Oligocene, and thus did not accumulate carbonate.

The Oligocene hiatus ended abruptly in the latest Oligocene-earliest Miocene in the Brazil Basin, São Paulo Plateau, and Angola Basin. Kennett et al. (1974)

TABLE 1
Calculated Depths of Deposition Areas at Mid Oligocene (30 m.y.B.P.)[a] for Several South Atlantic DSDP Sites

Site	Present Water Depth	Basement Age (m.y.B.P.)	Post-30 m.y. Sediment Cover (m)[b]	Depositional Depth (m) at 30 m.y.B.P.	Facies
17	4300	36[c]	90	3000	Calcareous ooze
19	4677	53[c]	40	3720	Calcareous ooze
20	4512	70[c]	30	3890	Calcareous ooze
355	4896	80[d]	200	4400	Hiatus (zeolitic clay)
356	3175	?	100	?[g]	Hiatus
357	2086	?	200	?[g]	Chalk
358	5000	75[e]	380	4565	Siliceous mud
354	4052	75[f]	400	3650	Chalk

[a]Depth at time of deposition ("B" depth of Berger, 1972).
[b]Estimated to 10-20 meters.
[c]Estimated from magnetic anomalies.
[d]K/Ar and Paleontology
[e]Estimated sedimentation rates and paleontology.
[f]Paleontology.
[g]Less than present depths of Oligocene sediments.

indicate that deep circulation was initiated between the South Tasman Rise and Antarctica around 25 m.y.B.P. This is roughly coincident with the opening of the Drake Passage (Barker, Dalziel, et al., in press). Davies et al. (1975) have suggested that the Oligocene hiatus may have ended as a result of reduced thermohaline circulation; i.e., the Circum-Antarctic Current may have deflected to the east heavy, cold water masses, formed in the Weddell Sea, which otherwise would have flowed northward into the Atlantic Basin.

The early Miocene was a time of general sediment accumulation, except for Sites 19 and 20 of the Mid-Atlantic Ridge sequence of Leg 3, deposition sites far from land and in deep (~4000 m) water where non-deposition or erosion was occurring below the CCD. Site 22 on the Rio Grande Rise contains a lower Miocene hiatus, but the nearby Site 357 was mainly accumulating sediment; this indicates the sometimes local nature of some gaps in areas of topographic extremes.

The early Miocene was characterized by a brief warming trend (Savin et al., 1975; Boersma and Shackleton, this volume). It was also the beginning of formation of substantial quantities of North Atlantic Deep Water by subsidence of the Faeroe Island-Greenland Ridge to a critical sill depth at about 20 m.y.B.P. (Vogt, 1972). If this water mass penetrated to high southern latitudes, it could have caused an upwelling system of nutrient-rich water similar to that of the present, with attendant high productivity, especially in the thick calcareous sequences at Sites 356 and 357, and biosiliceous sequences at Site 358.

The middle to late Miocene was a time of general hiatus in the southwest Atlantic (Figures 3, 4), and of increased dissolution in the southeast Atlantic—as indicated by the poor preservation of calcareous skeletons (Melguen et al., 1975). The cored hiatus on the Ceará Rise shows evidence of increased calcium carbonate dissolution, as well as current-controlled deposition at an extremely low average rate of only 0.004 m/m.y.

The mid- to late Miocene hiatus was tied to global climatic events. The early middle Miocene was a time of sharp changes in O^{18} ratios of planktonic and benthic foraminifer tests, which Savin et al. (1975) interpret as indicating a drop of bottom temperature (or high-latitude surface temperature), caused by a rise, not a drop, in low-latitude surface temperatures (Savin et al., 1965). These authors see this as an indication of how solar energy was distributed in and by the world ocean. The Circum-Antarctic Current caused latitudinal isolation of surface waters, and ultimately led to formation of the present-day continental ice sheet of eastern Antarctica. This probably produced a strong thermohaline circulation much like that of today, with seasonal production of cold, saline Antarctic Bottom Water which flowed strongly northward in the western basin beneath a strong southward flow of less dense North Atlantic Deep Water formed in the Norwegian and Labrador seas. A current regime was produced which had potential for eroding as well as for dissolving carbonate (and silica). The CCD shoaled to about 3200 meters in the South Atlantic in the mid-Miocene (Berger and Von Rad, 1972).

The ocean's overall sediment budget in space and time will require much more study. The amount of sediment potentially able to accumulate in the ocean per unit time at any given time is subject to many variables well known and too detailed for discussion here. In general, however, one might expect that time intervals represented by hiatus at some places in basins would be represented by sedimentation in others. This is true at certain South Atlantic sites; for example, Leg 36 sites show generally reduced or missing Eocene and lower Miocene sections, but Eocene and the early Miocene were periods of sedimentation in the Argentine Basin. By contrast, some of the major hiatuses, such as the lower Tertiary and the Eocene/Oligocene boundary hiatuses, are more general, and occur in both shallow and deep water Leg 39 sites. Davies et al. (1975) noticed the same in the Indian Ocean sites.

It is accepted that some of these periods of major regional or global hiatus were also periods of increased sediment transport and buildup in certain local oceanic areas. Thus, Moore et al. (in press) cite other studies (chiefly Ruddiman, 1972) which indicate in the North Atlantic periods of rapid local sediment deposition, coincident with the major Atlantic hiatuses at the end of the Eocene and the beginning of the Oligocene, during the Oligocene, and in the mid- to late Miocene. Rona (1973) recognized the enigma of the synchronism between periods of global oceanic hiatus, such as the Paleocene and Oligocene, and periods of major marine regression—periods in which an increased sediment load should have been delivered to the oceans as result of rejuvenation of erosive forces on land and exposure of continental shelves. The net sedimentary budget to the oceans should be one of increased accumulation at times of regression. The process of rapid local or regional accumulation in the deep basins may be sufficient to account for this sediment. Perch-Nielsen et al. (1975) also suggest that these sediments might be found on prograding continental slopes.

MARGIN SITE—SÃO PAULO PLATEAU

Site 356 on the São Paulo Plateau is the only Leg 39 site that may be considered a "margin" site in terms of expected facies progression through time, although Site 357 on the northern flank on the Rio Grande Rise also has a tendency toward a marginal facies succession, as is discussed later. Schneider (1972) proposes for the sedimentary evolution of rifted continental margins a model in which he identifies main stages of basin development. The facies progression through time is from clastic sediments to evaporites to sapropels and mudstones rich in organic matter to progressively open marine facies, with occasional periods of isolation. Transition in such a sediment facies is observed in bore holes on land and from deep-sea drilling in the Red Sea and the South Atlantic. In the Red Sea, thick accumulations of evaporites (mostly halite) are known from borings on shore, and are at least locally underlain by wedges of clastic sediments. Sites 225, 227, and 228 were drilled on DSDP Leg 23 (Whitmarsh, Weser, Ross, et al., 1973) in the main trough region just landward of the axial trough, in present water depths of about 2000 meters. For present purposes, stratigraphies at the three sites may be generalized. Upper Miocene evaporites consist of halite interbedded with bedded and nodular anhydrite, with occasional layers of pyritic dolomitic claystone. The evaporites are overlain by uppermost Miocene-lowermost Pliocene claystones rich in organic material, and by dolomite and pyrite. Geochemical studies on dolomites (Supko et al., 1973) indicate a general trend from extreme hypersalinity to more normal salinity. The reduced sediments are overlain by Pliocene calcareous silts and clays, formed under normal oceanic conditions; these in turn are overlain by calcareous oozes and chalks with variable terrigenous components. Periods of isolation in the Pleistocene, associated with eustatic drops of sea level, are marked by precipitation of aragonite cements under conditions of above-normal salinity.

Sedimentation at Site 356 on the São Paulo Plateau has been similar to that proposed by the model and observed for the Red Sea. The sedimentary succession and geologic history at Site 356 is also very closely parallel to that at Site 364 in the Angola Basin (Bolli, Ryan, et al., 1975). Figure 5 shows lithofacies at Sites 356 and 364, as well as the composite stratigraphic sections from Sites 360 and 361 (Cape Basin) and Sites 362 and 363 (Frio Ridge section of the Walvis Ridge); see Figure 1 for site locations. Kumar et al. (this volume) discuss Site 356 results in the broad context of the evolution of the northern South Atlantic Basin.

Structurally, Sites 356 and 364 are in an area of the northern South Atlantic which, early in its history, was a semi-isolated small ocean basin, bounded to the south by the topographically continuous feature made up of the São Paulo Ridge and the Frio Ridge section of the Walvis Ridge, to the north by a series of equatorial marginal fracture ridges (Gorini and Bryan, 1976), and to the east and west by continents. The east-west-trending barriers to water exchange and meridional flow exerted great control over sedimentation.

Site 361 in the Cape Basin bottomed in uppermost Barremian or lowermost Aptian sandstones, siltstones, and sapropelic shales, approximately 50 meters above acoustic basement. The sediments were deposited at high rates under euxinic conditions (Bolli, Ryan, et al., 1975). Such conditions probably persisted since the period of initial separation in the Neocomian (Larson and Ladd, 1973), and may have been preceded by a pre-drift period of subaerial clastic sedimentation in an early rift valley stage. Reducing conditions ended in the Cape Basin in the late Aptian, perhaps as a result of opening of the basin to free water exchange with the separation of the Falkland-Agulhas Fracture Zone from the southern tip of Africa. Terrigenous shale of abyssal plain or distal fan facies was deposited at a rapid rate in the opening Cape Basin between the late Aptian and the Maestrichtian, at which time

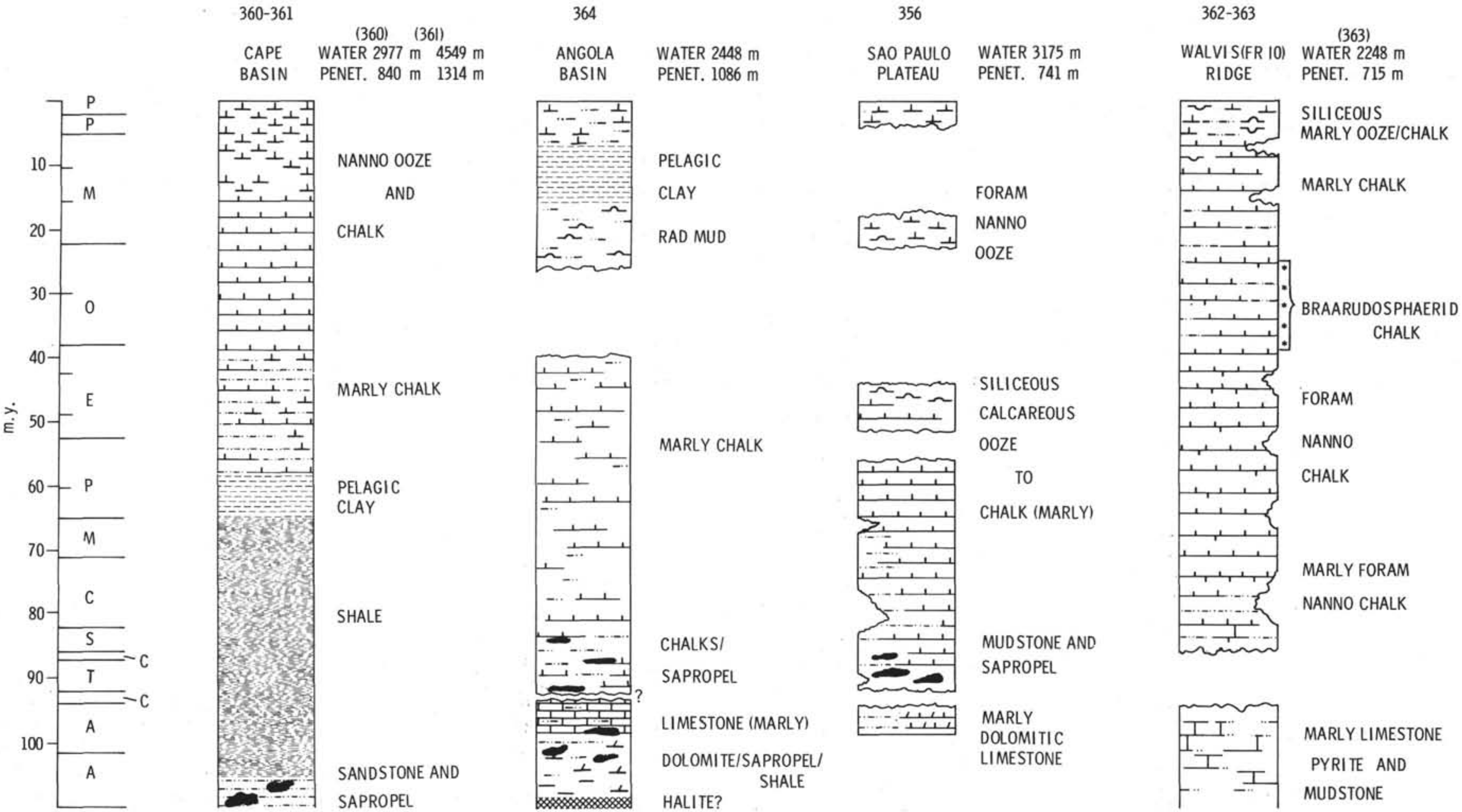


Figure 5. *Lithostratigraphy versus age for drilling sites located at the margins of the South Atlantic Basin. Hiatuses shown as breaks in sediment column; times of inferred hiatus, slow sedimentation, or times of dissolution shown by indentations of sides of column.*

terrigenous supply to the basin was sharply reduced and open sea abyssal conditions began. Paleocene pelagic clay is overlain in turn by Eocene marly chalk, Oligocene chalk, and Neogene calcareous oozes (Figure 5). Thus the Cape Basin, relatively open to the south but semi-restricted in its early stages of development, and situated in temperate latitudes under humid conditions (Bolli, Ryan, et al., 1975), did not undergo an evaporite phase of sedimentation; rather, the sediment sequence in the opening and expanding basin went from clastics to a reduced facies (sapropels), to a transitional stage of rapid terrigenous contribution (shales), to open marine conditions (pelagic clay, chalk, oozes).

The drilling sites on the São Paulo Plateau and the Angola Basin bottomed in the upper Albian and upper Aptian, respectively, so samples representing the early sedimentary history of the northern South Atlantic Basin were not taken. Both sites are presumably underlain by basaltic oceanic crust, created by normal sea-floor spreading since the initiation of drift ($\sim$127 m.y.B.P., according to Larson and Ladd, 1973). Mascle and Renard (1976) believe that a clastic layer overlies the basalt basement, but Leyden (1976) believes that evaporites directly overlie basement. Marine geophysical studies have indicated with near certainty the existence of evaporites beneath the São Paulo Plateau (see Kumar et al., this volume, and references therein), and drilling in contiguous coastal basins shows the salt to be Aptian. Although the salt was not reached at Site 364, nor its correlative facies at Site 356, high interstitial water salinities in the black muds at the base of Site 364 indicate the presence of halite only some tens of meters below the bottom of the hole.

The presumed salt is overlain by shales and sapropels in the upper Aptian-lower Albian at Site 364; the basin became increasingly open-marine during the Albian. The increasingly oceanic conditions probably occurred because of water exchange with the southern South Atlantic; Reyment and Tait (1972) show that surface circulation between the North and South Atlantic oceans was probably only initiated in the early Turonian; Premoli-Silva and Boersma (this volume) report reworked Tethyan Cenomanian fauna. It seems likely that the initiation of surface water interchange with the southern South Atlantic was associated with tectonic events. On the basis of salt boundary morphology and the greater width of the northern South Atlantic west of the present ridge, Kumar et al. (this volume) postulate a ridge jump to the east at the end of the Aptian.

The dolomitic and sapropelic mudstones of the Angola Basin site extend into the lower Albian, and are overlain by more than 250 meters of Albian limestone and marly limestone. This limestone is very similar in mineralogy and gross aspect to the upper Albian marly limestones recovered at the base of Site 356, except that the latter are dolomitic and more characteristically laminated. Minor indications of reducing conditions occur in the Albian on the Frio Ridge (Sites 362 and 363).

A period of stagnation in the Late Cretaceous is clearly shown by the occurrence of sapropelic sediments at both the Angola Basin and the São Paulo Plateau. At the Angola Basin site, the reduced sediments extend from the uppermost Albian to the Coniacian-Santonian (the Cenomanian may be missing); at the São Paulo Plateau, the upper Albian-upper Turonian are represented by a hiatus and the sapropelic calcareous mudstones, and clay-pebble conglomerates are restricted to the upper Turonian-Coniacian. The uppermost Albian to Coniacian section is missing at the Frio Ridge. Sedimentation continued in the Cape Basin throughout the Late Cretaceous with the deposition of shales under aerobic but acidic conditions (Bolli, Ryan, et al., 1975), indicating that the Late Cretaceous period of stagnation was restricted to the area north of the São Paulo Ridge-Rio Grande Rise-Walvis Ridge. Evidence of Late Cretaceous periods of stagnation has been reported from the western North Atlantic (Hollister, Ewing, et al., 1972), eastern North Atlantic (Hayes, Pimm et al., 1972), and Caribbean (Edgar, Saunders, et al., 1974). The Turonian-Coniacian period of stagnation in the northern South Atlantic may be related to the global Turonian regression; the drop in sea level made the subsiding São Paulo Ridge-Rio Grande Rise-Walvis Ridge complex a more effective barrier to water interchange than it was in the Albian, although it is difficult to imagine that such a mechanism of isolation could affect the northwest Atlantic. Kumar et al. (this volume) suggest the possibility of basin-wide oxygen-minimum conditions in the Late Cretaceous, although according to Fischer and Arthur (in press), great increase in thicknesses of the oxygen minimum layer should occur during periods of global transgression, not regression.

Sedimentation at the São Paulo Plateau, Angola Basin, and Frio Ridge became open-oceanic in the Santonian-Campanian with the deposition of marly chalks, which at Site 356 show evidence of considerable dissolution. Detrital components remained very important until the end of Cretaceous time, when the terrigenous contribution dropped abruptly, probably because of global transgression. Subsequent sedimentation was pelagic, either entirely biogenic (as at the São Paulo Plateau) or biogenic with intervals of pelagic clay (Angola Basin). Slumping was in important sedimentary process at Site 356 in the late Paleocene and Neogene. Correlation of major seismic reflecting horizons with changes in lithology at Site 356 is good (Site 356 chapter, this volume). Kumar et al. (this volume) extrapolate Site 356 lithologies (and ages) over the entire plateau. The lowest reflector at Site 356 (at 0.75 sec) represents the top of the upper Albian limestone; a well-developed reflector at 0.73 sec at Site 364 corresponds to sediments of the same lithology and age.

BASIN SITES—BRAZIL, ARGENTINE, AND MALVINAS OUTER BASIN

We define basin sites as those on ocean crust formed at divergent plate margins sufficiently after opening of the basin that the trailing edges of the continents were already far enough away not to have a serious direct effect on sedimentation. Sediments accumulating were pelagic and largely controlled by ridge processes, unlike the sequences characteristic of marginal sites discussed above. Figure 6 shows a very simple model using a sea-floor spreading from an ocean ridge, subsidence of crust as a function of age (and distance from the ridge, since time and distance are equated at a constant spreading rate), and a depth of carbonate compensation. The section at T_1 was deposited entirely when the site was above the CCD, and is pelagic carbonate ooze. As the site passed below the CCD, carbonate accumulation ceased, and only biogenic silica, authigenic minerals, and pelagic clay accumulated, as at T_2. T_3 represents a time of increased bottom circulation, during which, we propose, diagenesis increased at the sea floor, in this example remobilizing the biogenic silica and causing it to reprecipitate as cement (see following discussion on Horizon A). T_4 represents a time when terrigenous sediment was added, either because of changes in the source area or by changing deep current patterns in the expanding basin.

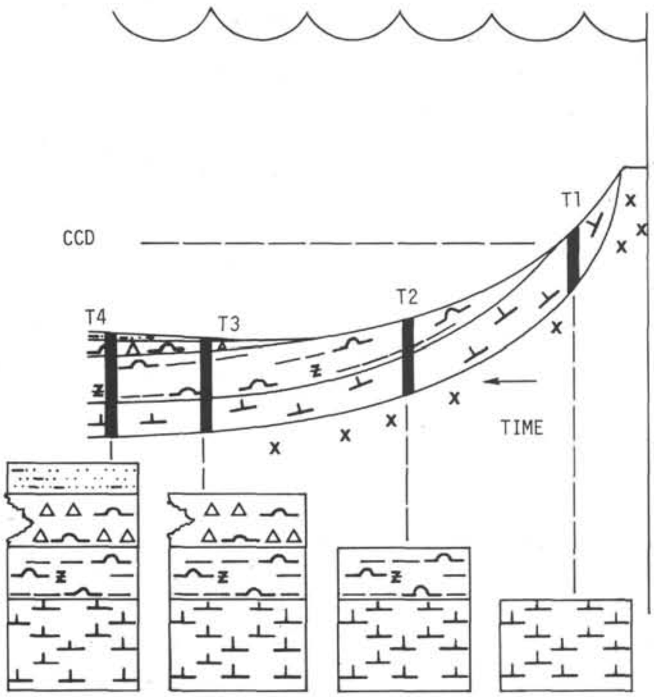


Figure 6. *Simple model of facies transition at a basinal site as it moves away from a spreading ridge. See text for discussion.*

Site 355 in the Brazil Basin and Site 358 in the Argentine Basin are typical basin sites, and their sedimentation histories are summarized here, along with Site 328, drilled in the Malvinas Outer Basin on Leg 36. There is some justification for also considering Site 354 on the Ceará Rise a basin site, but we exclude it here, largely because it has always remained above the CCD; Site 354 is discussed with "ridge sites." A more detailed account of the sedimentation history of the South Atlantic is given in McCoy and Zimmerman (this volume).

The basal sediments at Sites 355 and 358 are carbonates, and were deposited when the sites of deposition were above the CCD. The lowermost sediments sampled at Site 328 in the Malvinas Outer Basin are zeolitic muds. The absence of carbonate here may reflect local shoaling of the CCD, a result of the high southern latitude.

The lowermost 45 meters of sediment at Site 355 in the Brazil Basin consist of nannofossil ooze with a carbonate content between 40% and 80%. The ooze contains veins of sparry calcite, probably products of hydrothermal processes associated with crustal generation. Basement age at this site is estimated at 76-68 m.y., and the youngest nannofossil ooze is of early Maestrichtian age, or ~70 m.y. The ooze is overlain by 10 meters of non-fossiliferous ferruginous clay, which is in turn overlain by lower Eocene zeolitic muds. If the cessation of carbonate deposition resulted from the site passing below the CCD by normal subsidence as a function of sea-floor spreading, and if the uppermost nannofossil ooze truly represents the last carbonate deposited at the site, an early Maestrichtian CCD level of about 3000 meters is indicated. If, however, carbonate deposition continued until the middle Eocene, with the additional carbonate later removed by erosion, an early Eocene CCD of about 3900 meters is indicated (see Figure 13, Site 355 chapter, this volume). The basal sequence at Site 358 in the Argentine Basin extends from 730 meters subbottom to the bottom of the hole at 842 meters, and probably to the calculated basement depth of 897 meters; it consists of marly chalk (50%-70% $CaCO_3$) alternating with mudstone which contains little or no carbonate. Basement age is about 72 m.y. The last up-hole occurrence of carbonate is in the middle Eocene, at 730 meters subbottom. If the site underwent only normal subsidence, the middle Eocene CCD in the Argentine Basin at Site 358 was about 3900 meters, slightly deeper than the 3700 meters determined for the equatorial Atlantic by Berger and von Rad (1972). Sedimentological and paleontological data

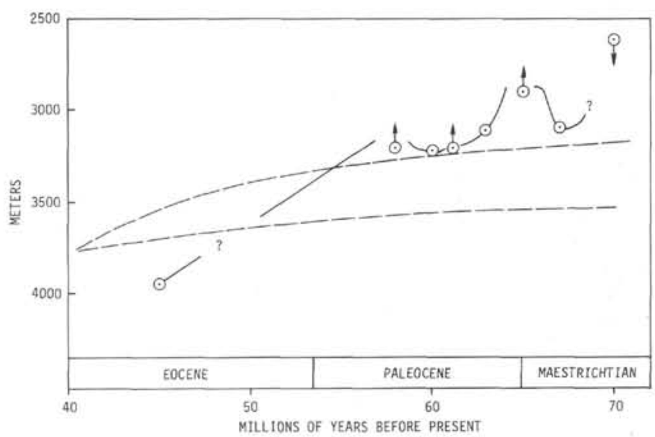


Figure 7. *Depth variation of the CCD at Site 358 during the Late Cretaceous and early Paleogene. Data points are after Boersma (this volume). Dashed lines indicate the range of the CCD for the Atlantic as determined by Berger and von Rad (1972).*

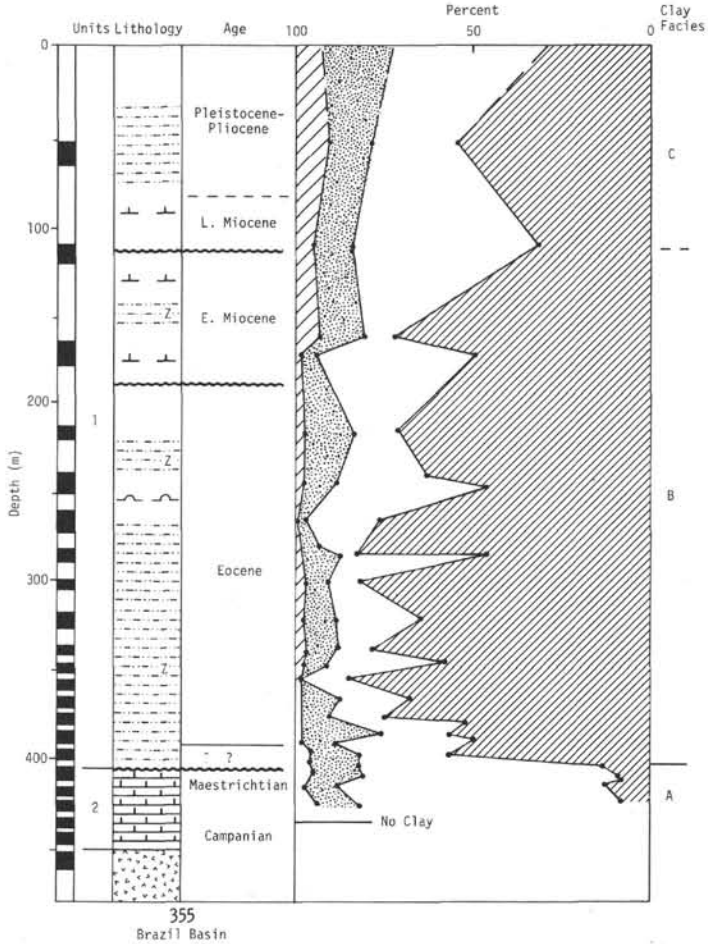
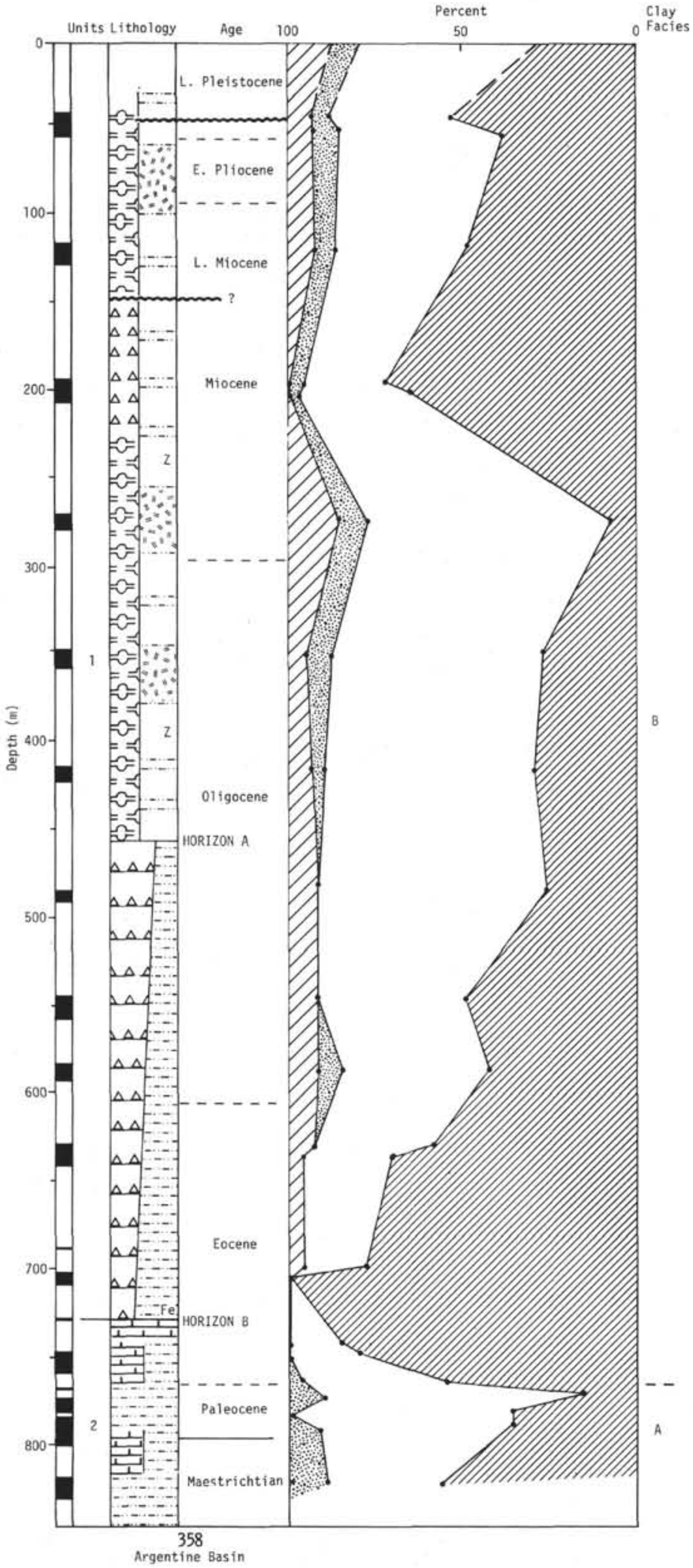


Figure 8. *Sediment facies as a function of depth and age at basinal sites in the western South Atlantic Ocean. Site 328 data from Barker, Dalziel, et al. (in press) Clay minerology data for Sites 355 and 358 from Zimmerman (this volume).*

indicate that Site 358 was near the CCD for a long time. Interlayered mudstones and chalks indicate alternating deposition of calcareous and of non-calcareous sediment. Boersma (this volume) uses faunal dissolution criteria as indicators of the CCD (or lysocline) at various times from the late Maestrichtian to the middle Eocene, assuming normal crustal subsidence of the site. Her data points (corrected for isostatic sediment load) are plotted in Figure 7, and assume a CCD about 300 meters below the lysocline. The data indicate a rise in the CCD at the end of the Cretaceous and beginning of the Tertiary, and then a general increase in the CCD to about 3900 meters in the middle Eocene.

In the Brazil Basin, the ooze sequence is overlain by mudstones and zeolitic mudstones (Figure 8). The lower section of this Tertiary sequence comprises zeolitic mudstones with numerous thin laminae of quartz-zeolite sands and silts. The laminae are interpreted as distal turbidite facies, on the basis of displaced biogenic material and surface morphologies of the quartz grains (Krinsley and McCoy, this volume). Accumulation rates of these zeolitic muds in the early Eocene were very high (Figure 9). The lower to middle Eocene also contains radiolarian mud and dolomite mud sub-units. The section above the Eocene-Oligocene hiatus is predominantly silty clay with local silt and sand laminae; zeolites are less common. Carbonate content is negligible in the Tertiary section, except for bioclastic turbidites deposited below CCD in the Miocene and Plio-Pleistocene and preserved by rapid burial. There is a sharp change in clay mineralogy (Figure 8) across the Upper Cretaceous/Paleocene unconformity. Below the unconformity, the basal section contains predominantly illite, and above the unconformity there is a distinct increase in montmorillonite. Zimmerman (this volume) and McCoy and Zimmerman (this volume) suggest that the Eocene was a period of general vulcanism in the South Atlantic with the main locus of activity along the Rio Grande Rise-Walvis Ridge trend, but also extending into tropical latitudes, as indicated by an abundance of montmorillonite in the Eocene at the Ceará Rise. The illitic clays below the Upper Cretaceous-Tertiary hiatus correspond to the "continental" facies of Heath (1969); the montmorillonite clays of the Eocene constitute the



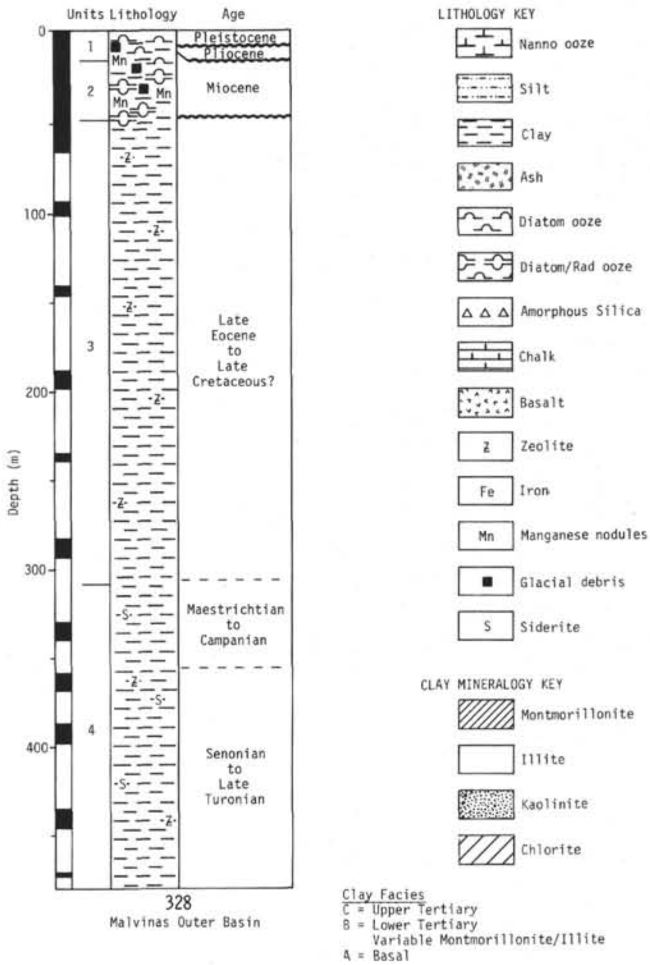


Figure 8. *(Continued).*

"oceanic" facies. There is some indication of a return to the illitic "continental" facies above the mid- to upper Miocene hiatus in the Brazil Basin, as is the case at the Ceará Rise in the section of post middle Miocene age, but wide core spacing in the Brazil Basin makes this less apparent.

The section of post middle Eocene age in the Argentine Basin consists of biosiliceous mudstones. Again, the montmorillonite to illite ratio increased sharply in the Eocene; montmorillonite constitutes virtually 100% of the clay fraction in the sediment of early Eocene age (Figure 8). Montmorillonite decreased at the expense of illite until the early Miocene, then increased from the middle Miocene to the present. Silica occurs as tests of radiolarians and frustules of diatoms, except in the middle Miocene and the Eocene to middle Oligocene, where the biogenic silica has been diagenetically altered to an amorphous silica cement (Figure 8). Volcanic ash is present only in those sections containing unaltered biogenic silica; presumably the diagenetic process which has altered the biogenic silica also affected the ash. We speculate that the diagenetic alteration may be a result of increased bottom current activity at the end of the Eocene and beginning of the Oligocene and in the middle Miocene (see discussion on Horizon A).

In the Malvinas Outer Basin, accumulation of zeolitic clay and claystone was rapid and continuous, except for a possible short hiatus at the end of the

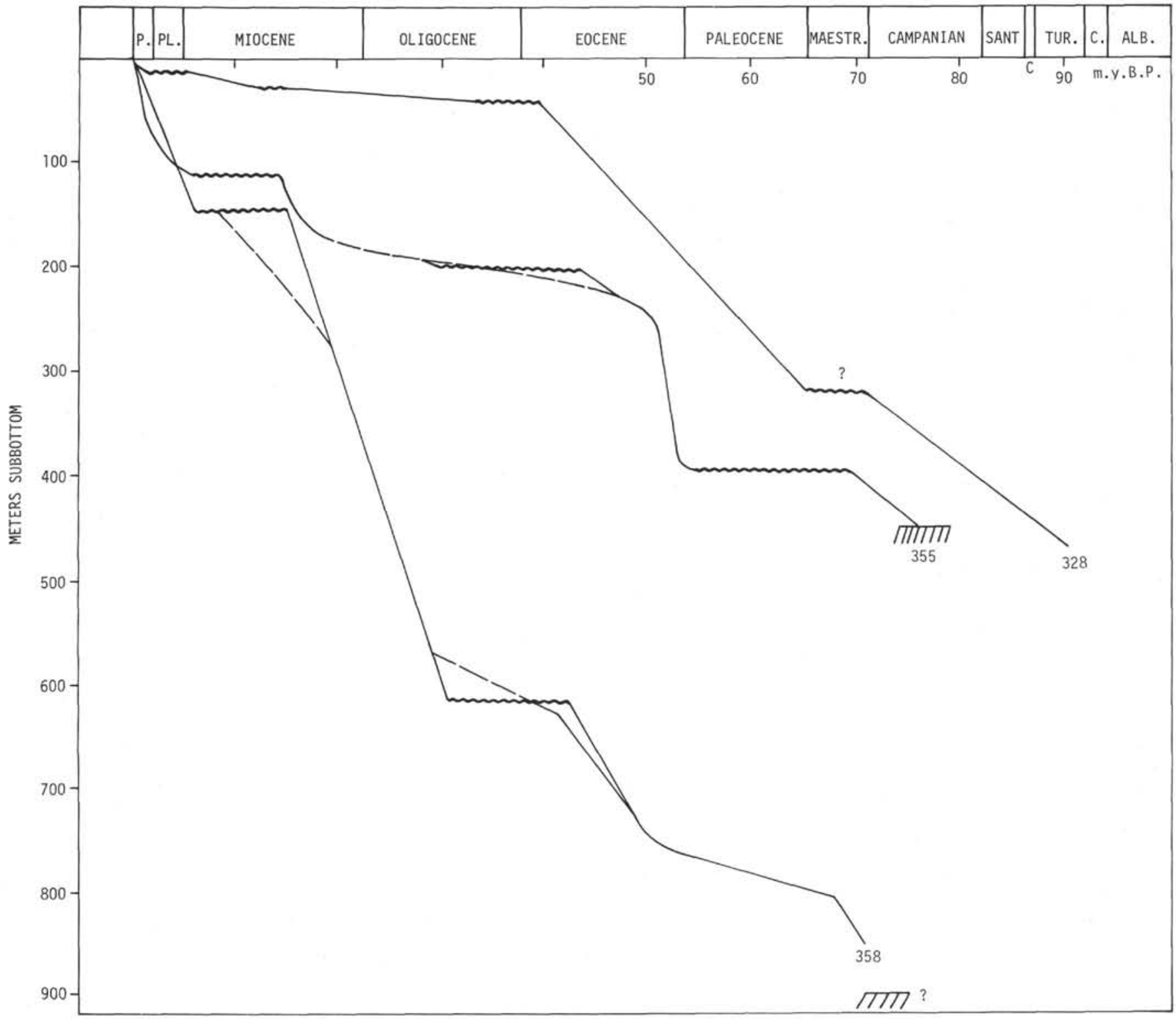


Figure 9. *Sediment accumulation rates for basinal sites in the western South Atlantic Ocean.*

Cretaceous and beginning of the Tertiary. Siliceous microorganisms are scarce, which indicates low productivity during this time. The probable source of the clay was the Andean Cordillera and West Antarctic Cordillera (Barker, Dalziel, et al., in press). The sedimentation pattern changed abruptly after a late Eocene-early Oligocene hiatus. The Oligocene and lower Miocene sediments are biogenic siliceous clays; diatoms in the lower part of the section show effects of dissolution. Sediments above a middle Miocene hiatus are diatomaceous oozes with abundant manganese nodules, sand, and large ice-rafted clasts.

Accumulation rates at Sites 328, 358, and 355 are shown in Figure 9. Hiatuses are inferred across the Cretaceous/Tertiary boundary in the Malvinas Outer Basin and Brazil Basin; accumulation of siliceous ooze was slow in the Argentine Basin in the early Tertiary, and increased in the early and middle Eocene. A hiatus across the Eocene/Oligocene boundary is inferred at all three sites, as is a hiatus in the middle Miocene. The late Oligocene-early Miocene was a time of very slow sediment accumulation in the Malvinas Outer Basin, but of very rapid accumulation in the Argentine Basin. A change in current pattern associated with initiation of the thermohaline circulation probably resulted in transport of sediment from the Malvinas Outer Basin to the Argentine Basin. The very high rate of accumulation in the Argentine Basin in the early Miocene also resulted from greatly enhanced biogenic productivity brought on by the development then of the Antarctic Convergence.

DATING OF MARINE MAGNETIC ANOMALIES

Sites 355 and 358 are on prominent magnetic anomalies at the older end of the Cenozoic-Late Cretaceous magnetic anomaly sequence originally described by Heirtzler et al. (1968) and later modified by Larson and Pitman (1972). Before Leg 39, only 11 drilling sites (10, 20, 100, 105, 137, 138, 166, 239, 245, 303, and 304) provided reasonably well dated basement contacts in locations of anomaly 30 age or older. Sites 137, 138, 100, 105, 166, 245, 303, and 304 reached basement older than that represented by the Heirtzler scale. Site 239 was drilled on anomaly 31 and Site 10 on anomaly 32; these results are discussed below, but there are problems associated with the interpretation of each. The oldest reliable date on the Heirtzler anomaly scale before Leg 39 was Site 20, drilled on anomaly 30 and dated by microfossils at 65-67 m.y. (Maxwell, Von Herzen, et al., 1970). Site 355 is between anomalies 33 and 34, and Site 358 is to the west (older side) of anomaly 32. Site locations are shown in Figure 10, along with magnetic anomaly patterns in the vicinity, as interpreted from aeromagnetic data and data collected on cruises of research vessels *Conrad* and *Vema* of Lamont-Doherty Geological Observatory (R. Larson, personal communication).

The stratigraphically lowest core obtained at Site 358 (situated to the west of anomaly 32) is a ferruginous marly chalk, dated by planktonic foraminifers as belonging to the lower Maestrichtian *Globotruncana tricarinata* Zone. The bottom of the lowermost core catcher sample was assigned by coccoliths to the *Tetralithus trifidus* Zone of late Campanian to early Maestrichtian age, for present purposes taken as 70 m.y. Basement was not reached, but was calculated to lie at 897 meters subbottom, 70 meters below the lowest core. The accumulation rate of the ferruginous marly chalk and mudstone sequence was approximately 17 meters/m.y. (see Site 358 chapter). If accumulation of the lowermost 70 meters of sediment was constant at the same rate, and if acoustic basement represents true oceanic basement, then the basement age should be about 4 m.y. older, or 74 m.y. The age predicted for this anomaly by Heirtzler et al. (1968) and Larson and Pitman (1972) is 76 m.y. Sclater et al. (1974) argue that ages on the Heirtzler scale prior to anomaly 5 (10 m.y.B.P.) should be shortened by about 8%, or that anomaly 32 should be dated at 70.7 m.y.B.P. The Site 358 calculated age of 74 m.y. for anomaly 32 is between the two values, and does not help us determine which of the two time scales (Heirtzler versus "shortened") is more nearly correct. Site 10 (Leg 2) is also supposed to be on anomaly 32, although the exact identification of the anomaly is in doubt (Sclater et al., 1974; Larson and Pitman, 1975—more recent work by Cande and Kristoffersen, in preparation, indicate Site 10 is nearly on anomaly 33). Van Hinte (1976) calculates basement at Site 10 to be 72.8 m.y. old (he gives a range of 72-75 m.y. for the basement age); Larson and Pitman plot the basement age at Site 10 as 76 to 80 m.y., based upon an 82 m.y. age for the base of the Campanian (the age accepted for Leg 39 use). If Site 10 represents anomaly 32, the Site 358 results indicate the 72 to 75 m.y.B.P. date of van Hinte (1976) may be better than the older date of Larson and Pitman (1975). The paleontological data at Site 239, drilled just seaward of anomaly 31 in the southwest Indian Ocean, are unclear. Simpson, Schlich, et al. (1974) date the basement as "pre-late Campanian" (which by their scale yields a basement older than 71 m.y.), and Larson and Pitman (1975) take this as corroboration of the age of the older end of the original Heirtzler scale, as modified (Larson and Pitman, 1972). According to Thierstein (personal communication, 1976), the oldest sediment above basement belongs to the *Micula mura* Zone of latest Maestrichtian age, or 66 m.y. Van Hinte (1976) reports that the basal sediments indicate an age range of latest Campanian to latest Maestrichtian, 65 to 72 m.y. by his time scale.

Site 355 lies to the west (older side) of anomaly 33 in the Brazil Basin, within the reversely magnetized zone between anomalies 33 and 34. The basal sediment is nannofossil ooze, which directly overlies the tholeiitic basalt interpreted as oceanic basement (Site 355 report; Fodor et al., this volume). Coccoliths indicate that the sediment immediately above basement belongs to the *Broinsonia parca* Zone of early Campanian age (Perch-Nielsen, this volume). The range of absolute age is 76 to 81 m.y. The Larson and Pitman (1972) modification of the time scale of Heirtzler et al. (1968) predicts an age of 81.5 m.y. for anomaly 33; the McKenzie and Sclater (1971) modification of the later Late Cretaceous portion of the Heirtzler time scale gives an age of 82 m.y. If the "shortened" time scale of Sclater et al. (1974) is accepted, the respective ages are 75.8 and 76.2 m.y. Again, because of the length of the coccolith zones represented and the question which absolute time scale is most appropriate for the Late Cretaceous, the Site 355 basement age for anomaly 33 cannot be used to decide definitively whether the "shortened" age scale of Sclater et al. (1974) is more appropriate for the Late Cretaceous part of the geomagnetic reversal sequence.

A fresh sample of the tholeiitic basalt basement rock at Site 355 was dated by the K/Ar method (McKee and Fodor, this volume). The radiometric age is 78.1 ±9 m.y. Although the mean value falls almost in the middle of the indicated age range, the large analytical uncertainty does not allow assignment to a specific stage.

Alvarez et al. (in press) have analyzed the magnetic stratigraphy of an Upper Cretaceous-Paleocene pelagic limestone sequence in the Umbrian Appennines at Gubbio, Italy. The section is well-dated by foraminifers. The magnetic reversal sequence for the Gubbio Section is shown in Figure 11, and is correlated with planktonic foraminifer zones (Premoli-Silva, in press) and with the geomagnetic reversal pattern for the South Atlantic as determined from Lamont-Doherty magnetic profiles C1102 (Ladd, 1974) and V3101 (Larson, personal communication). The length of the South Atlantic marine geomagnetic pattern is adjusted so that the reversals at anomalies 29 and 34 coincide with those of the Gubbio section; absolute age boundaries for Late Cretaceous stages are listed according to Bukry (1974) and van Hinte (1976). Site 355 and 358 locations are indicated on the marine magnetic profile as taken from the reference profiles (Figure 10).

The geomagnetic reversal pattern measured at Gubbio is identical with that determined from marine magnetic anomalies in the South Atlantic, except that

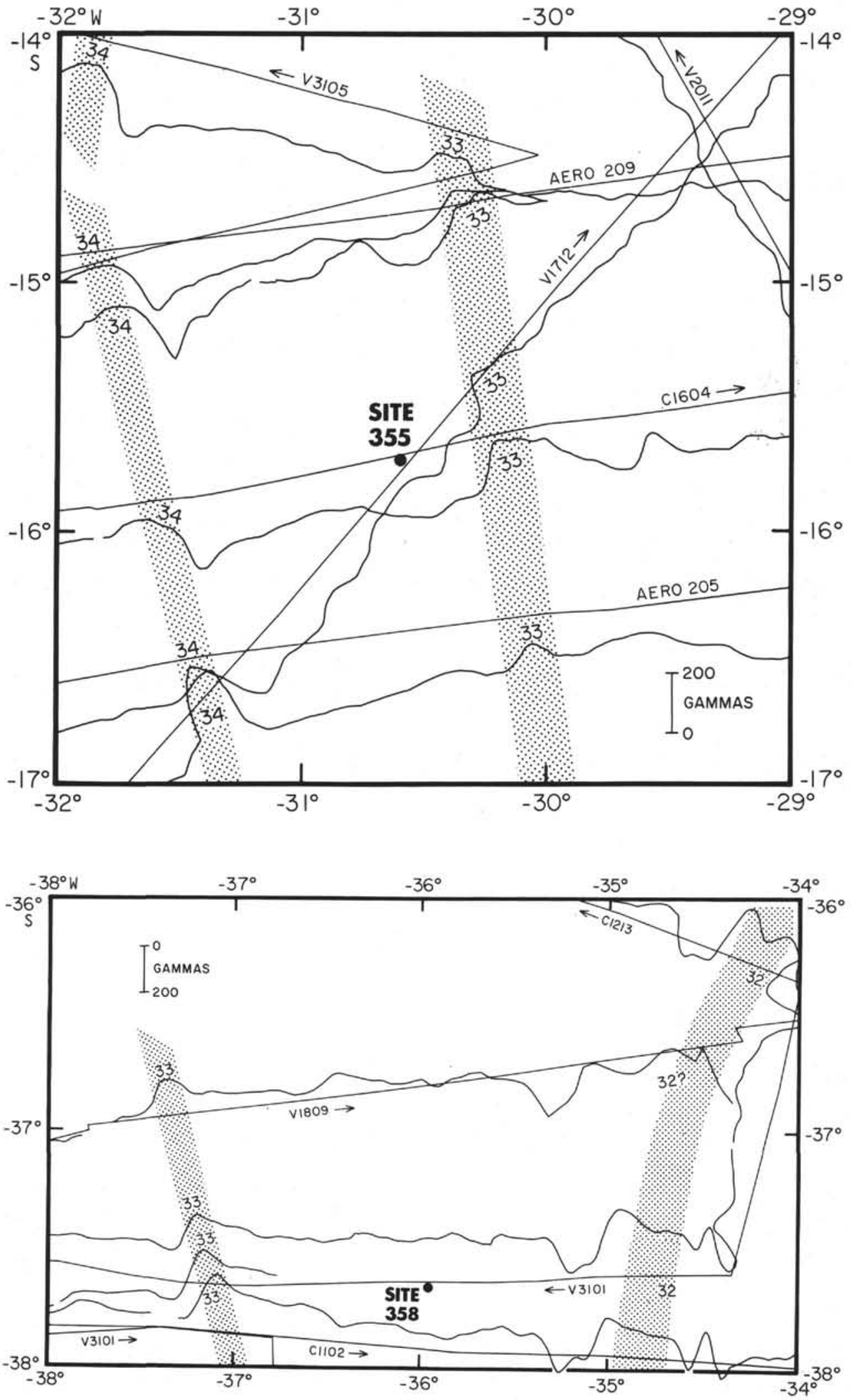


Figure 10. *(A) Location of Site 355 in the Brazil Basin in relation to magnetic anomalies 33 and 34. (B) Site 358 in the Argentine Basin in relation to magnetic anomalies 32 and 33. Magnetics were plotted and anomalies were identified by Roger Larson (personal communication).*

the Gubbio section does not show the short period of normal magnetization, within the reversed interval immediately before anomaly 32, which is seen in the South Atlantic and other ocean basins. Possible reasons for this are discussed by Alvarez et al. (in press). The Gubbio study is important in tying the marine geomagnetic reversal pattern to the magnetic stratigraphy of a biostratigraphically well-dated pelagic sequence on land. It confirms the position of the Cretaceous/Tertiary boundary between anomalies 29 and 30, as suggested by Sclater et al. (1974), and proves the existence of the reversed intervals between anomalies 32 and 33 and between 33 and the Cretaceous Long Normal interval (Larson and Pitman, 1972).

Using the correlation between the South Atlantic anomaly pattern and the reversal sequence at Gubbio, the biostratigraphic dating of the Gubbio section, and the absolute age scales of Bukry (1974) and van Hinte (1976), we can indirectly date Site 358 basement at 71.5 to 73 m.y.B.P. and Site 355 basement at 76.5 to 78 m.y.B.P. (dotted lines in Figure 11). These figures agree with the Leg 39 ages determined by coccolith dating, and are taken as the best estimates of basement age at the two drilling sites.

ARGENTINE BASIN SEISMIC REFLECTORS

Ewing et al. (1964) first described an areally widespread seismic reflector in the Argentine Basin, and termed this reflector Horizon A. In a later very detailed study of the sedimentary structure of the Argentine continental margin, Ewing and Lonardi (1971) extended the known area of Horizon A in the basin and further described a deeper reflector (Horizon B) which had earlier been reported by Ewing and Ewing (1965). Horizon A is characterized by its relative flatness and the fact that it generally parallels sea-floor topography rather than morphologic variations of acoustic basement. It marks a sharp change of reflection character in the sediment column, with many highly stratified reflectors above Horizon A and a generally transparent sequence below. The deeper reflection, Horizon B, is level, and appears to fill depressions in the basement. It is very patchily distributed and difficult to correlate between profiles. Figure 12, a simplified version of Figure 1a of Ewing and Lonardi (1971), shows the relationship of the major reflecting horizons in the Argentine Basin north of 41°S. The wavy nature of the sea floor, reflector F, and the layers under F in the right of the figure, are intended to indicate the presence in the central basin of "giant ripples," large (~5 km wavelength) sedimentary structures which extend from the sea floor down to the level of Horizon A (Ewing et al., 1971). Figure 13 shows typical air-gun reflection profiles collected in the Argentine Basin by Lamont-Doherty cruise *Conrad* 11. Location of the profiles is indicated in Figure 14. Horizon A is easily recognized on the profiles, except in the areas east of the dotted lines in Figure 14, which mark the eastward limit of recognition of a well-defined Horizon A (see profiles). Horizon B is more difficult to recognize, but may be inferred just above basement in

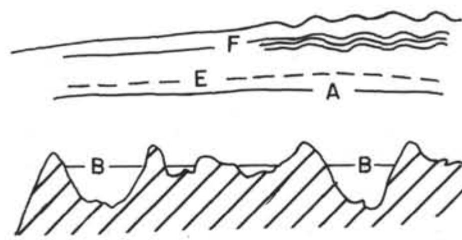


Figure 12. *Generalized structural section of the Argentine Basin showing the relationships between Horizons A, B, and F. Simplified from Figure 1a of Ewing and Lonardi (1971, p. 131).*

profile BC between about 1600 and 2300 hours. Figure 15 is an enlargement of a Lamont-Doherty *Vema* 3101 profile, shot on a heading of 270° through the location later chosen as Site 358. The line drawing is our interpretation of the depths and attitudes of Horizons A and B on this profile. Horizon B is interpreted as having some topographic expression. If this interpretation is correct, Horizon B may here be the top of a sediment blanketing basement rather than filling basement depressions, as was suggested by Ewing and Lonardi (1971).

The tops of Horizons A and B at Site 358 were picked at 0.58 and 0.88 sec, respectively, using the *Glomar Challenger* approach profile (see Site 358 report). Another reflector was seen at 0.20 sec on the approach profile, although it is not observed directly at Site 358. Basement is at 1.04 sec. Using sediment sound velocities measured aboard ship, the depth to the top of Horizon A was calculated to be 460 meters, and the depth to the top of Horizon B was calculated to be 730 meters. The uppermost reflector at 0.20 sec is determined at 155 m depth, and may correspond to Reflector F of Ewing and Lonardi (1971).

Correlation of reflectors with the stratigraphic section is shown in Figure 16. The top of the lower reflective zone (Horizon B) correlates well with the contact between the youngest carbonate sediments (middle Eocene marly chalks of Core 11) and the overlying siliceous mudstones. The Horizon A reflector at 0.58 sec correlates with a change from the upper Oligocene diatom-radiolarian mudstone in Core 6 to the siliceous mudstones of the same age in Core 7. The calculated depth to the reflector of 460 meters falls between the two cores, and a sharp reduction in drilling rate was observed at 450 meters subbottom. The chief difference in the facies above and below the top of the Horizon A reflector lies in the nature of the silica component. Above the top of the reflecting horizon, the silica occurs as relatively fresh biogenic opal in the form of radiolarian tests and diatom frustules (15%-25%). Volcanic glass shards (10%-20%) are also present. In the siliceous mudstones below, silica is present as amorphous masses in the silt to fine sand range, although smear-slide examination occasionally reveals evidence that biogenic debris served as a source for the seemingly inorganic amorphous silica cement (see Figure 8, Site 358 report). Thus, at Site 358, at least,

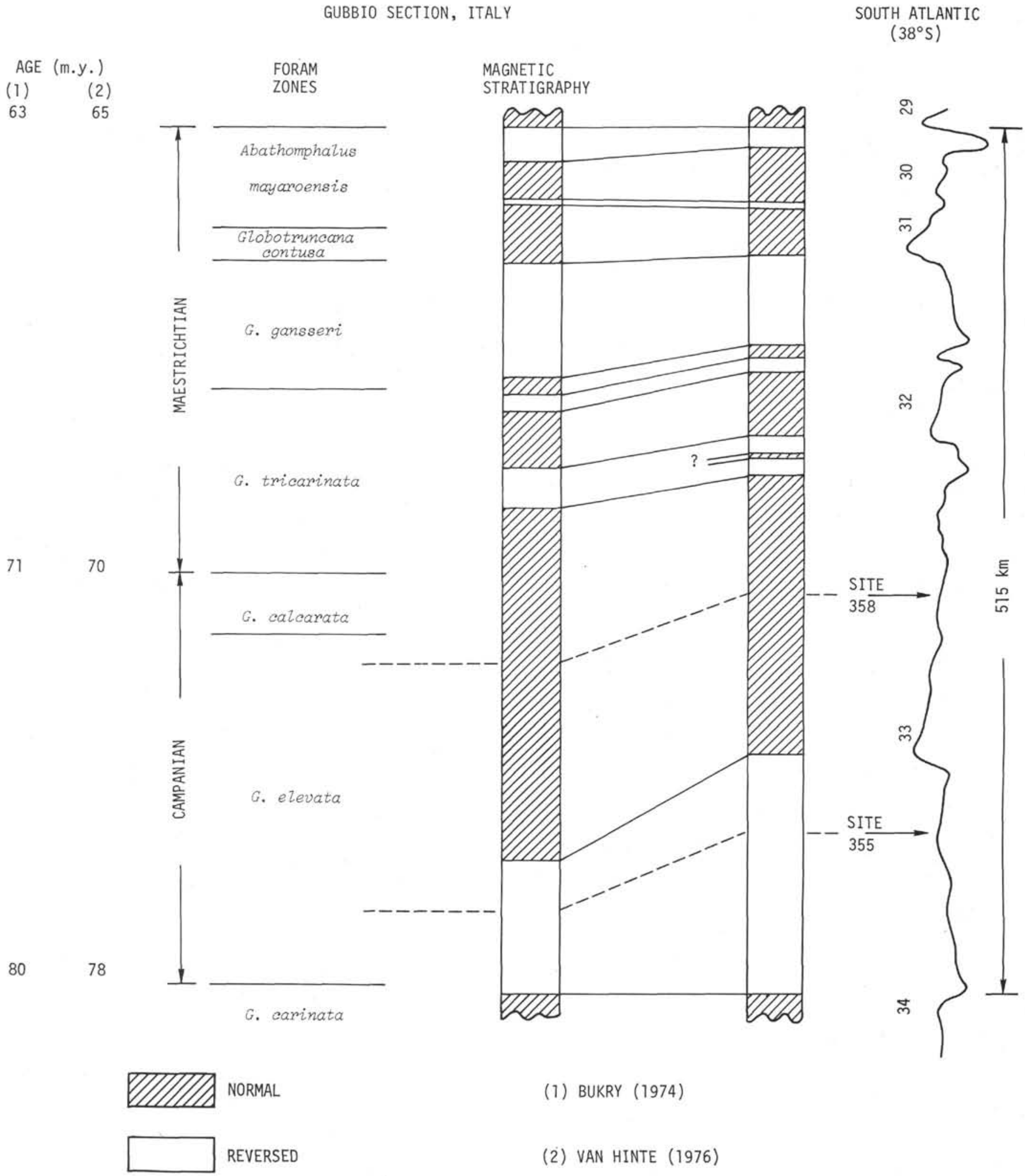


Figure 11. *Biostratigraphy and magnetic stratigraphy of the Late Cretaceous Scaglia Rossa type section at Gubbio, Italy, and the geomagnetic reversal sequence in the South Atlantic based on L-DGO magnetic profiles C1102 and V3101. The oceanic profile has been reduced to bring the beginning and end of the reversal sequence into position with the same points in the Gubbio section. Absolute ages are according to Bukry (1974) and van Hinte (1976). Site 355 and 358 locations are picked from their positions on the anomaly profiles. Adapted from Alvarez et al. (in press) with kind permission.*

Horizon A represents a diagenetic boundary in a siliceous mudstone sequence of late Oligocene age. Similarly, Horizon F represents the top of a sequence of unconsolidated siliceous muds, of middle Miocene age, in which amorphous silica and clay predominate. This facies, represented by Core 3 and probably covering the section between 155 and 220 (?) meters subbottom, is underlain and overlain by radiolarian and diatom mud in which volcanic ash is common.

Leg 36 investigators (Barker, Dalziel, et al., 1974) cored a section in the Malvinas Outer Basin (Site 328) which contained a diffuse mid-section reflector at 0.45 to 0.50 sec; this reflector was tentatively correlated with Horizon A in the Argentine Basin (a deeper horizon, which may be analogous to Horizon B in the Argentine Basin, was seen at 0.68 sec at Site 328, but drilling there was terminated before this depth was reached). Although Horizon A, because of the intervening scarp of the Falkland Fracture Zone, cannot be directly traced from the Argentine Basin to the Malvinas Outer Basin, Ewing and Lonardi (1971) suggest that the prominent mid-depth horizon of the two basins may be correlated on the basis of their character. The depth of the mid-section reflector at Site 328 puts it within sedimentary Unit 4, a variegated claystone of Maestrichtian to (?) early Eocene age which lies between 310 and 471 meters subbottom. If this reflector is representative of the horizon as seen over the basin as a whole, Horizon A in the Malvinas Outer Basin is probably the result of variations in diagenesis and lithification in an otherwise uniform sediment facies of Late Cretaceous to early Tertiary age (Barker, Dalziel, et al., in press).

Ewing and Lonardi (1971) equate Horizon A in the Argentine Basin with a similar reflector, also called Horizon A, in the North Atlantic. Horizon A in the North Atlantic was originally identified, on the basis of piston core samples, as Upper Cretaceous turbidites which form a fossil abyssal plain, (Ewing et al., 1966). Early deep-sea drilling results (Ewing, Worzel, et al., 1968; Peterson, Edgar, et al., 1969) indicated, however, that Horizon A in the North Atlantic was caused by a widespread chert layer of early to middle Eocene age (Berggren, 1969), and that Ewing et al. (1966) actually described a turbidite unit, beneath Horizon A, in an area where reflection profiles were poor and the turbidite and the overlying reflecting horizon could not be resolved. Both biogenic silica (radiolarians, diatoms, silicoflagellates) and silica deriving from alteration of volcanic glass have been suggested as sources of the silica which makes up the chert. Berggren and Phillips (1971), Dietz and Holden (1970), and Jones et al. (1970) claim that continental separations in high northern latitudes resulted in cold deep-water masses first entering the North Atlantic in the Eocene, with attendant upwelling of nutrient-rich waters and enhanced production of organisms with siliceous tests. Gibson and Towe (1971), citing evidence for abundant vulcanism in the Eocene, favor inorganic precipitation of chert by diagenetic alterantion of volcanic ash. Although release of silica by alteration of the ash could also have aided productivity, this was, they believed, only a secondary factor. More recent studies (e.g., Weaver and Wise, 1974) clearly indicate a biogenic origin for the Eocene Atlantic cherts; a paleocirculation study by Ramsay (1971) shows widespread distribution of Eocene biogenic silica. The Eocene North Atlantic cherts were almost certainly derived from alteration of siliceous biogenic material, but the importance of vulcanism in contributing silica to the ocean remains questionable.

The nature of Horizon A in the North Atlantic was found to be more complex when drilling at Sites 101 and 105 showed the reflector to correspond to a long mid-Cretaceous to mid-Tertiary hiatus (Hollister, Ewing, et al., 1972). Berggren and Hollister (1974) maintain that the Horizon A hiatus and the Horizon A chert are both consequences of the same process, the initiation of a deep-water flow regime in the North Atlantic, with high productivity of biogenic silica resulting from upwelling, and with erosion or nondeposition in the western basin resulting from westward intensification of the bottom currents.

Eocene chert occurs at Site 13 on the Sierra Leone Rise and at Site 21 on the Rio Grande Rise (Maxwell, von Herzen, et al., 1970), and it was thought that Horizon A in the South Atlantic would also largely represent Eocene chert. Although we encountered Eocene chert at Site 357, the finding that at Site 358 Horizon A marks a diagenetic change in upper Oligocene siliceous mudstones indicates the reflector is diachronous, at least from one major ocean basin to another.

Horizon A in the North Atlantic represents a relatively short period of silica deposition, but siliceous sediments have accumulated continuously in the Argentine Basin since the middle Eocene. The difference between the facies representing Horizon A and that immediately above is the presence of silica as amorphous masses rather than as recognizable biogenic debris. The same is true of the shallower (F) reflector. Two factors deserve consideration: the presence or absence of other sedimentary components and the timing of the event relative to inferred environmental events of basinal or global scope.

At Site 358, volcanic ash is common in the upper Miocene to Pleistocene radiolarian-diatom muds above 152 meters, and in the upper Oligocene to lower Miocene radiolarian-diatom mudstones between about 220 and 460 meters subbottom (see Figure 16). Volcanic ash is not recognized in the layers where the former biogenic silica has been converted to amorphous silica, i.e., the layers whose tops correspond to reflectors F and A. It may be that the volcanic ash acts in some way to inhibit silica diagenesis. Riedel (1959) first noted that volcanic material may help to preserve siliceous skeletons from dissolution after burial. He suggested that liberation of silica by weathering of the ash might be the cause; this is probably not so. Heath (1974) cites other studies which indicate that volcanic ash dissolves much more slowly than biogenic silica, and suggests alternative mechanisms whereby the presence of ash may inhibit dissolution of the tests. There clearly seems to be a

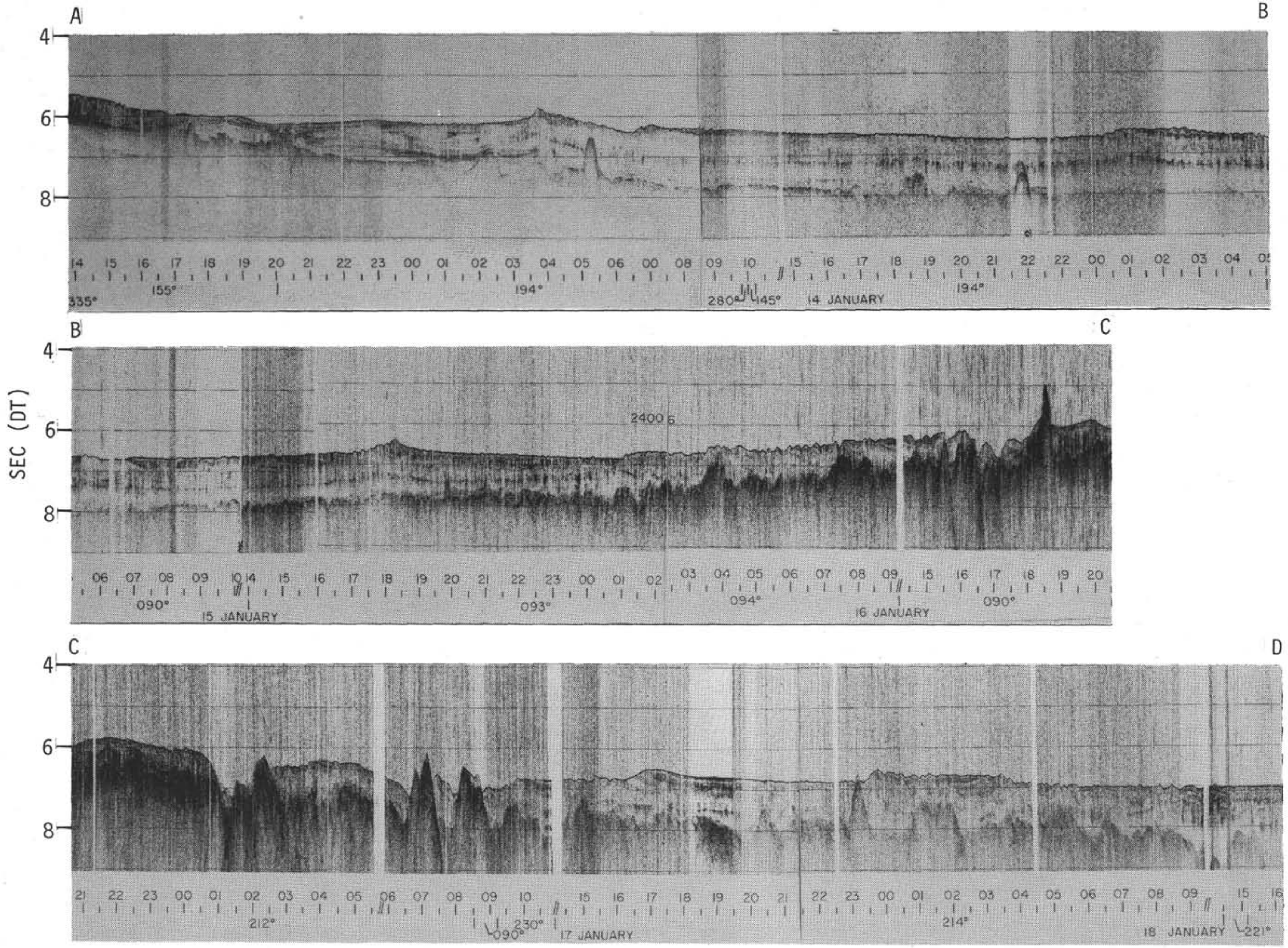


Figure 13. *Lamont-Doherty Geological Observatory seismic reflection profiles obtained in the vicinity of Site 358 (Cruise* Robert D. Conrad *11).*

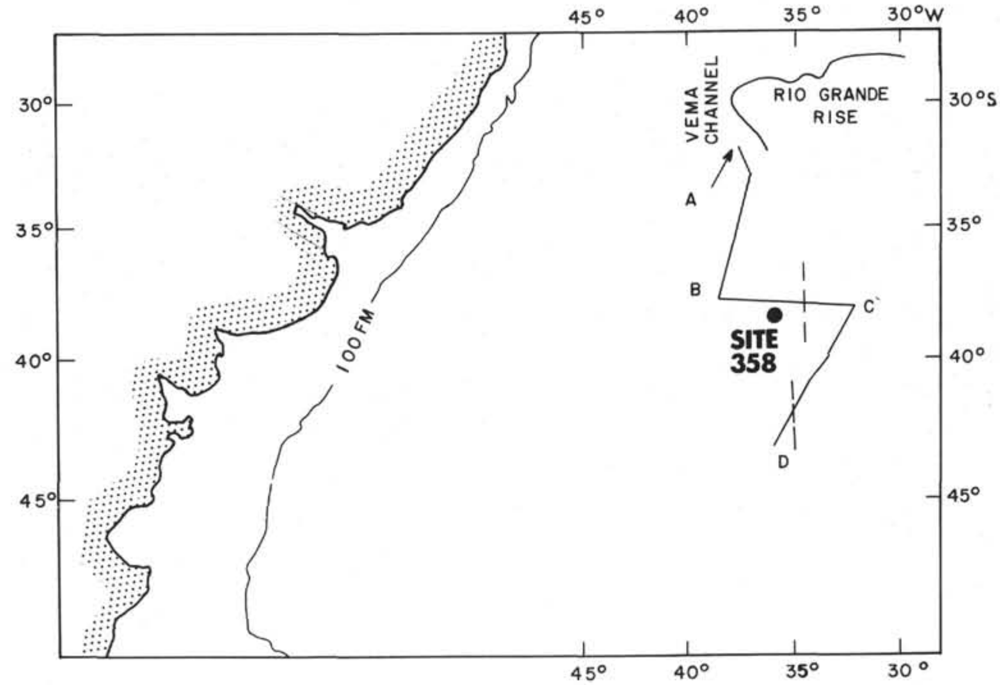


Figure 14. *Base map showing the location of profiles in Figure 13. Dashed vertical lines indicate the eastward limit of recognition of Horizon A.*

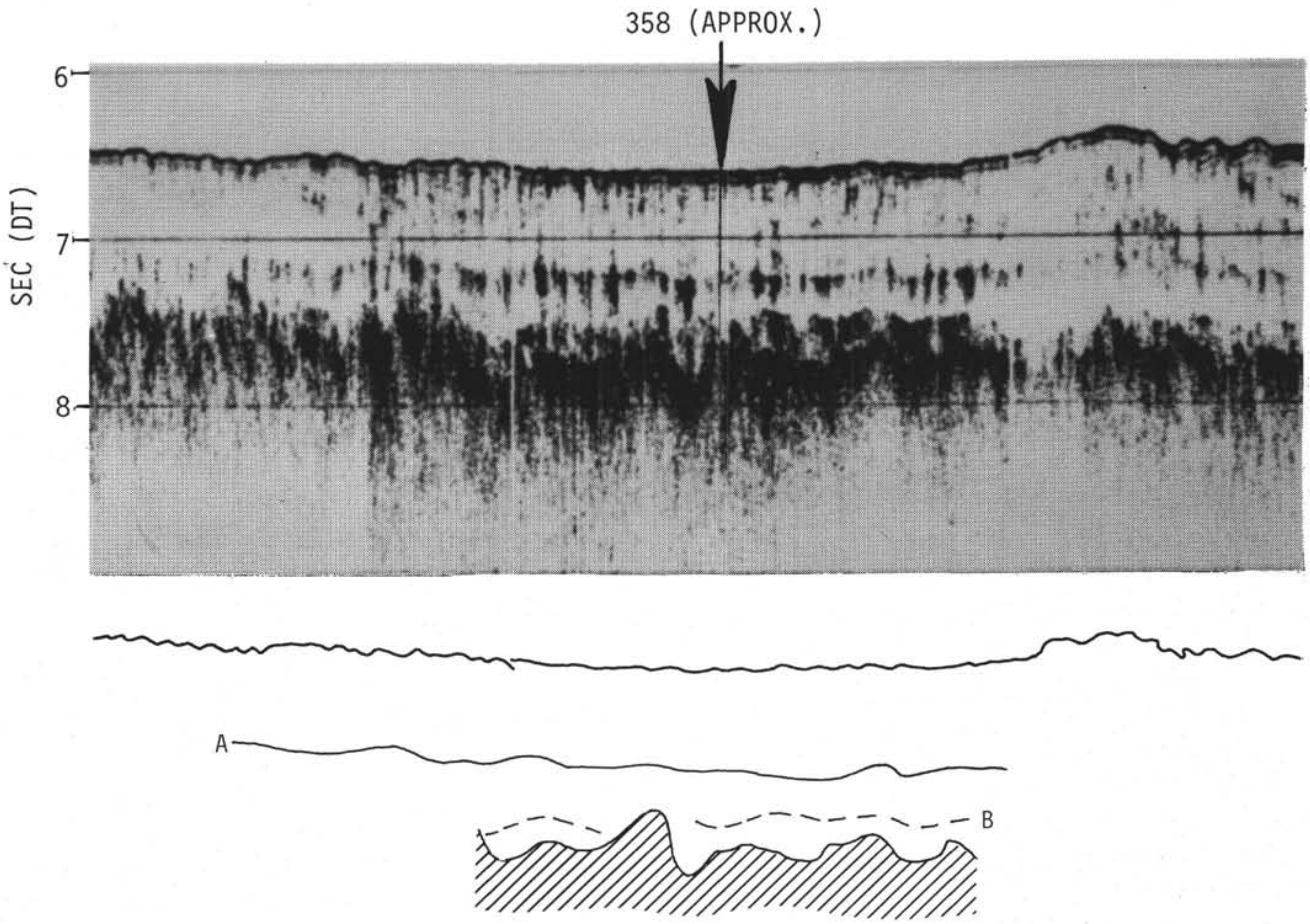


Figure 15. *Lamont-Doherty* Vema *3101 profile used for site selection with our interpretation of Horizons A and B. Note that Horizon B appears to be less of a depression-filling feature than indicated schematically in Figure 12.*

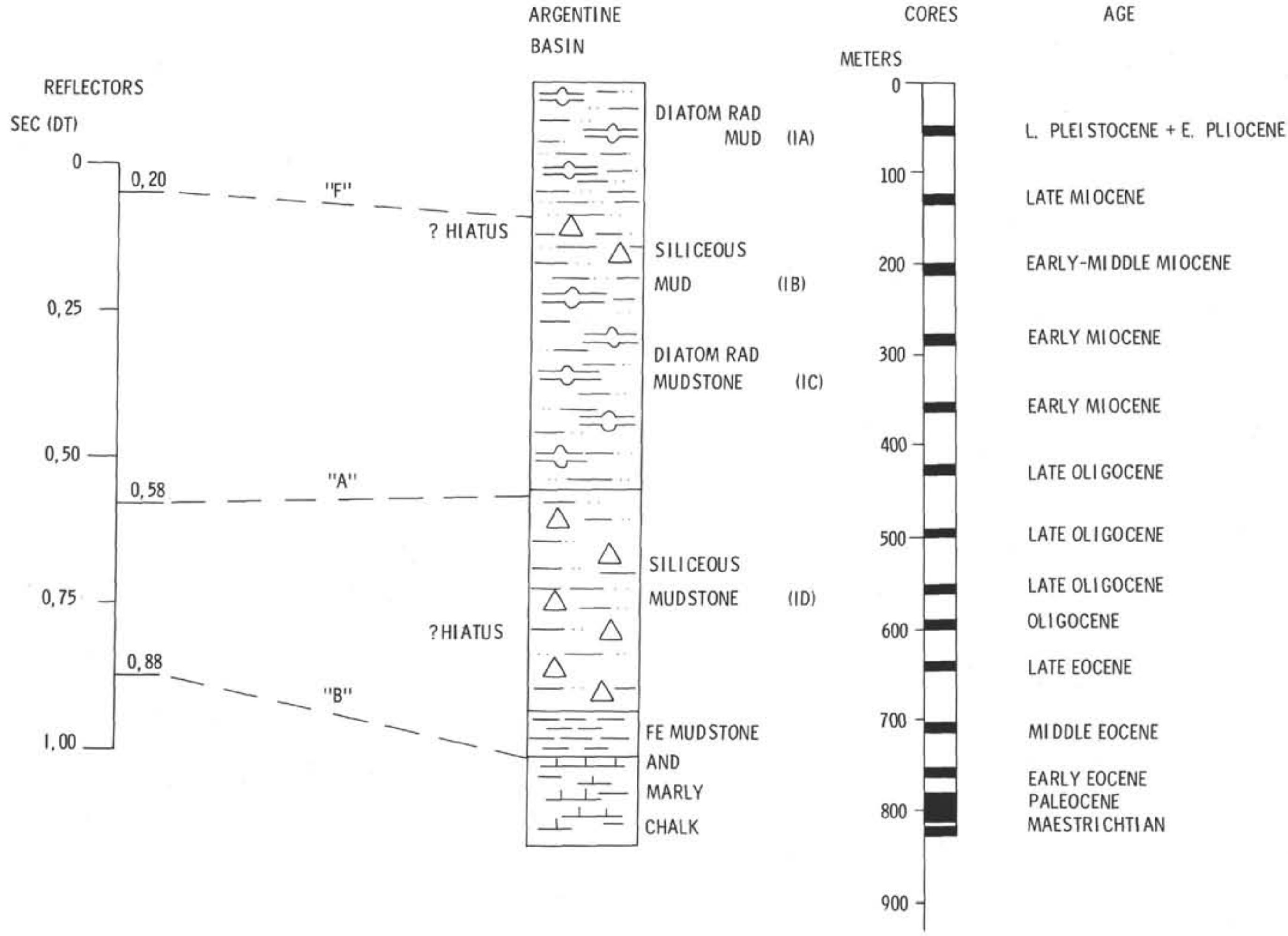


Figure 16. *Correlation of Horizons "A", "B", and "F" with sedimentary facies cored at Site 358. 1A-1D denote lithologic subunits.*

cause-and-effect relationship between presence of ash and high degree of preservation of siliceous skeletons, but the mechanism is not yet fully understood.

A second possibility is that the diagenetic change in the biosiliceous sediments may be an effect of sedimentation rate. Riedel (1959) points out that much of the dissolution of siliceous skeletons occurs while they are close to the sediment-water interface, i.e., within a few centimeters of the sediment surface, and that dissolution is thus favored by slow rates of accumulation. Lerman (1975) shows that dissolution, migration, and re-precipitation within the sediment column are minor. We observe that the siliceous mudstones of Unit 1D (the top of which corresponds to Horizon A) and Unit 1B (the top of which corresponds to Horizon F; see Figure 16), were both deposited at times of relatively slow sediment accumulation (see Figure 4).

We can infer something about processes if we try to understand the sedimentation rates in terms of sediment supply and the role of currents; we present two differing hypotheses here, both put forward by Leg 39 scientists. Both must take into account that Unit 1D, of late Eocene to late Oligocene age, correlated with the beginning of deep thermohaline circulation [associated with a sharp drop in bottom temperatures of 4°C at the Eocene/Oligocene boundary (Kennett and Shackleton, 1976; Boersma and Shackleton, this volume), caused by sea ice formation around Antarctica (Kennett and Shackleton, 1976)]. Similarly, the middle Miocene period of slow accumulation correlates in time with a transgression in the South Atlantic (McCoy and Zimmerman, this volume, Figure 2) and with formation of extensive Antarctic ice shelves (Hayes, Frakes, et al., 1975), which was presumably responsible for the formation of high salinity, low temperature Antarctic Bottom Water (AABW) similar to that of today.

Zimmerman (personal communication) believes that the slow sediment accumulation rate in the late Eocene and Oligocene resulted partly from a low rate of introduction of sediment, caused by low bottom-current velocities. He believes the deep thermohaline current (pre-Antarctic Bottom Water, or pAABW) initiated at the end of the Eocene and beginning of the Oligocene began as a weak current and intensified throughout the remainder of the Tertiary, achieving maximum development during the Quaternary. The middle Miocene period of slow sediment accumulation was the result of a transgression in the South Atlantic. An attractive aspect of this theory is that it explains the occurrence of the "giant ripples" described by Ewing et al. (1971). These features, described earlier, are evident in subbottom reflections down to, but not beyond, Horizon A. Increasing bottom current velocity over time could explain their restriction to the upper part of the section.

We suggest a second possiblity, that the periods of slow sediment accumulation may be very much the results of currents. In this view, the late Eocene-late Oligocene at Site 358 may have been a time of slow accumulation caused by initiation of pAABW and the resultant partial winnowing and bypassing. This is in line with observations that the Eocene/Oligocene boundary and much of the Oligocene are represented by hiatuses, presumably current-controlled, in many places in the oceans (see section on hiatuses). The siliceous mudstones of Unit 1D accumulated at low rates under conditions of active bottom-current influence. Note that the effect of the current itself would tend to enhance diagenesis in the same way as a simple slow sedimentation mechanism; the amount of water to which a sedimentary particle would be exposed per unit time would be increased. Rapid accumulation of the biogenic mudstones of Unit 1C occurred from late Oligocene to early Miocene. This could be a supply effect, or could be consequent to a drop in bottom-current velocity, perhaps because of formation of a significant deep latitudinal transport in the Circum-Antarctic Current (see discussion in section on hiatuses).

In Unit 1B, siliceous skeletons have again been diagenetically altered by re-precipitation to fine grains of silica. The single core from this unit is of early to middle Miocene age. The middle Miocene at Site 358, as at other South Atlantic sites, was a period of hiatus or slow accumulation (Figures 3, 4). Presumably this was associated with strong bottom currents of cold, saline Antarctic Bottom Water which formed when the Antarctic ice shelves formed in the middle Miocene (Hayes, Frakes, et al., 1975). This facies therefore also would represent early diagenetic conversion of siliceous skeletons to fine grains of silica, by solution and re-precipitation, at a time of slow sedimentation associated with maximum bottom-current activity.

We realize that this hypothesis, in which slow sediment accumulation rates result almost entirely from current control, may require periods of faster bottom currents in the late Eocene to late Oligocene and in the middle Miocene. The hypothesis is also hard to reconcile with the down-section termination of the "giant ripples" at Horizon A, although this difficulty may be an artifact of the acoustic records presently available. We do not, however, see these problems as serious, in light of our poor knowledge of circulation factors and associated phenomena with regard to fine-grained sediments, and we consider the slow sediment accumulation-current control theory a valid one.

We prefer the slow sediment accumulation hypotheses to a hypothesis requiring inhibition of solution of biogenic silica by the presence of volcanic ash. We can explain the absence of recognizable ash in Units 1B and 1D if we assume that the ash was altered to clays and amorphous silica at or close to the sediment-water interface by the same mechanism as were the siliceous tests. The alternative would be to require discrete periods of South Atlantic vulcanism in the late Oligocene-early Miocene and late Miocene to Pleistocene. Although vulcanism was probably fairly general through time, clay mineralogy studies (Zimmerman, this volume) suggest that the Eocene was a period of especially extensive vulcanism in the South Atlantic; a volcanic breccia in middle Eocene sediments was sampled on the Rio Grande Rise. The occurrence of ash, therefore, is not especially correlative with the inhibition of silica mobilization. Horizon A at Site 358 is not coeval with Horizon A as described from the North Atlantic (Hollister, Ewing, et al., 1972) or the Caribbean (Edgar, Saunders, et al., 1973). Nor is it coeval with the midsection reflecting horizon sampled at Site 328 in the Malvinas Outer Basin; this horizon represents a diagenetic change in sediments of Late Cretaceous to early Tertiary age. If this diagenetic change was associated with currents in any way, the current regime is older than that responsible for Horizon A in the northeastern Argentine Basin. If Horizon A in the Argentine Basin resulted from a paleocurrent effect, it should be fairly synchronous in the main basin; but this can be proved only by direct sampling.

OCEANIC RISES

Ceará Rise—Sierra Leone Rise

At Site 354 on the Ceará Rise an 881-meter section of predominantly pelagic sediment was penetrated; basaltic acoustic basement was reached. The sedimentary sequence (and implied oceanographic history) is very similar to that at the Sierra Leone Rise in the eastern equatorial Atlantic (Figure 1), as are the general settings of the two features and certain aspects of the seismic stratigraphy. Figure 17 shows the generalized lithofacies at the two sites, as well as interpretive drawings of the major subbottom reflectors and their correlations with lithology (Site 354, this volume; Site 366, Lancelot, Seibold, et al., 1975; Lancelot, Seibold, et al., in press).

The sedimentary section at the Ceará Rise indicates accumulation of biogenic calcareous sediments in a pelagic environment since the early Maestrichtian. The basal carbonates are enriched in iron and clay minerals. The Maestrichtian to Oligocene limestone and chalk sequence is generally marly. Siliceous components, as diatom frustules, are significant only in the early Oligocene. The middle Oligocene to middle Miocene chalks are relatively free of terrigenous material. The only significant contribution of terrigenous clastic material occurred from the middle Miocene to the present. The section is burrowed throughout, and there are no indications of periods of stagnation. Apparently, deposition of predominantly calcareous biogenic debris on the oceanic rise occurred always at depths above the level of carbonate dissolution and always in an

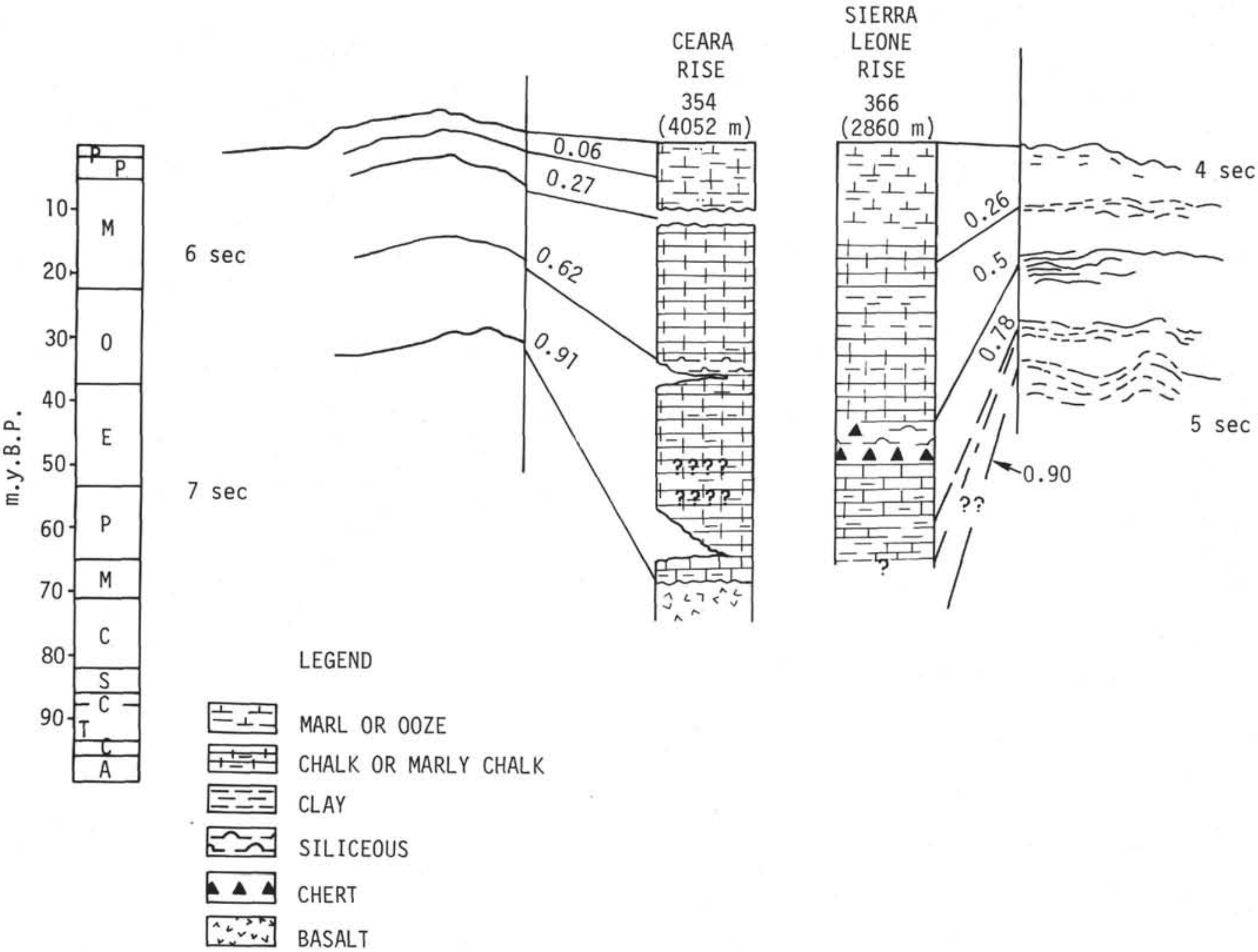


Figure 17. *Comparison of lithostratigraphy versus age and of seismic reflectors at the Ceara Rise (Site 354, Leg 39) and Sierra Leone Rise (Site 366, Leg 41).*

oxidizing environment. Periods of slow accumulation or actual hiatus are represented at the Cretaceous/Tertiary boundary and in the Paleocene, at the Eocene/Oligocene boundary; minor hiatuses occur in the middle Miocene, lower Pliocene, and parts of the Pleistocene.

At Site 366, 850.5 meters of sediment were penetrated before bottoming in Maestrichtian marlstones. The sedimentary section is very similar to that at the Ceará Rise, in that it has about the same total thickness, is composed of open pelagic calcareous ooze (and its diagenetic facies, chalk, and limestone), the basal section is richer in clay, and there is significant occurrence of silica in the mid-Tertiary. There are no hiatuses in the Sierra Leone Rise section, although periods of slow accumulation of calcareous ooze occurred in the late Eocene (perhaps corresponding to the period of slow accumulation, or hiatus, at the Ceará Rise) and in the middle to late Miocene (correspoding to the middle Miocene period of very slow accumulation at the Ceará Rise). Early study of Leg 41 cores (Lancelot, Seibold, et al., 1975) indicated the possible importance of cyclic sedimentation. Sierra Leone Rise chalks of early to middle Eocene age showed variations of clay content which were thought to represent fluctuations in rate of supply of clay, perhaps climatically controlled. The upper Eocene to lower Miocene chalks were cyclical (oozes and chalks alternating with marls and clay), but were interpreted as dissolution cycles, similar to those reported in Leg 40 drilling results (Melguen et al., 1975). The former cycles were not noted in the Ceará Rise section; fluctuations in the ratio of siliceous to calcareous material in the Oligocene Ceará Rise cores may indicate dissolution cycles, but this is hard to assess. Wide core spacing makes this site a poor one for study of such a phenomenon.

One difference in the stratigraphic sections is in the nature and age of the silica present. At the Sierra Leone Rise, the silica is in the form of porcellanite and chert, of early to middle Eocene age (Lancelot, Seibold, et al., 1975), similar to and coeval with the chert layers widely noted in the North Atlantic (Berger and von Rad, 1972, and references therein). At the Ceará Rise, lower and middle Eocene sediments were not recovered, but silica occurs in the form of relatively unaltered diatom frustules of early Oligocene age. The fact that deposition of biogenic silica did not begin until the middle Eocene at Site 358 (Argentine Basin), early Oligocene at the Ceará Rise, and the Oligocene in the Malvinas Outer Basin (Site 328), indicates that silica productivity in the western South Atlantic was caused by mechanisms different from that in the North Atlantic and the Caribbean. If the early Oligocene period of high silica productivity in the western South Atlantic was related to the oceanographic changes occurring around Antarctica at about this time (Kennett and Shackleton, 1976), and if it represents transport to the deposition areas by newly created deep northerly flowing bottom currents, then it is understandable that the Sierra Leone Rise sites might be nonsiliceous, since the Mid-Atlantic Ridge may have served as an effective barrier to the eastward flow of the bottom water. The Sierra Leone Rise siliceous sediments are similar in lithology (cherts, porcellanites) and in age to the North Atlantic cherts described earlier (Ewing, Worzel, et al., 1968; Peterson, Edgar, et al., 1969; Hollister, Ewing, et al., 1972; Hayes, Pimm, et al., 1972), and one may suspect a common origin. Several workers (Dietz and Holden, 1970; Jones et al., 1970; Berggren and Hollister, 1974) suggest that high Eocene silica productivity in the North Atlantic resulted from upwelling consequent to circulation changes associated with tectonic movements at high northern latitudes. It is clear that the diachronous deposition of silica at the Ceará Rise and Sierra Leona Rise sites must be accounted for in any proposed paleocirculation model.

Another difference between the two sections is the abundance only at the Ceará Rise of terrigenous material deposited post middle Miocene; this is explained by the beginning of Amazon River drainage in the Miocene (Damuth and Kumar, 1975).

The sequence of internal seismic reflectors at the two sites shows similarities, but also some differences. The major reflectors are traced in Figure 17, and depths in seconds of two-way travel time to the tops of major reflecting zones are given. The reflectors at both sites are subparallel to the sea floor. The shallowest identified reflector at Site 354 lies at a depth of 0.06 sec, and correlates with a lithologic unit change; it is of local significance only, and has no recognizable counterpart at Site 366. Regional reflectors at 0.27 sec at Site 354 and 0.26 sec at Site 366 correlate with a change in diagenetic facies, from ooze above to chalk below. A midsection regional reflector is well developed at both sites. At Site 354 its top is at 0.62 sec and corresponds to the first downhole occurrence of siliceous sediments and to a sharp drilling break at 450 meters subbottom. A similar reflector lies at 0.5 sec at Site 366, and corresponds to the first downhole occurrence of porcellanite and a sharp reduction in drilling rate at 480 meters. The reflector sequence with its top at 0.78 sec at Site 366 has no apparent correlative at Site 354; Leg 41 scientists had difficulty in correlating this with the stratigraphic section, although it may be related to a change in lithology at 775 meters, from limestone above to marls below (Lancelot, Seibold, et al., in press). The acoustic basement at Site 354 is at 0.91 sec, and corresponds to the basalt cored (actually, the topmost of this "group" of basement reflectors probably corresponds to the top of the red marls immediately overlying basement). The acoustic basement at 0.90 sec at Site 366 also seems to consist of a "group" of reflectors; the hole was terminated before basement was reached.

The geographic and gross morphologic similarities of the Ceará Rise and the Sierra Leone Rise led Embley et al. (1972) to suggest a common origin for these features. Perch-Nielsen, Supko, et al. (1975) and Supko et al. (1975) have discussed drilling results from Legs 39, 3, and 41 in this context. Kumar and Embley (in press) have combined Leg 39 drilling results with geophysical and sedimentological studies on the Ceará Rise, and have presented a model which accepts a common origin for the two rises within the plate tectonics framework.

The Ceará Rise and the Sierra Leone Rise lie on portions of the Atlantic and African plates, respectively, bounded by extensions of the 4°N and 8°N (Doldrums) fracture zones (Fox, 1972; Cochran, 1973). Seismic refraction studies (Houtz et al., in press) indicate that the Ceará Rise is an oceanic structure, but in the shallowest parts of the rise the acoustic basement as detected by reflection profiling corresponds to a thick (up to 2 km) layer with a seismic velocity of about 3.4 km/sec. This layer is interpreted to be composed of extrusive volcanics, perhaps interlayered with sediments, a character similar to that proposed for the acoustic basement of the Rio Grande Rise (McDowell et al., this volume). Magnetic lineations in the equatorial Atlantic are not yet well enough defined to be used for interferring crustal ages. However, the (roughly equal) distances of the rises from the present ridge crest would indicate an age of about 80 m.y. for their formation at the ridge crest, if they moved away at spreading rates interpolated from those of the North Atlantic (Pitman and Talwani, 1972) and South Atlantic (Larson and Ladd, 1973). Voluminous outpourings of basalt may have occurred at this time, in conjunction with a (proposed) readjustment of plate motions associated with a shift in the rotational pole controlling the direction of Atlantic opening (Le Pichon and Hayes, 1971; Le Pichon and Fox, 1971; Pitman and Talwani, 1972).

Site 354 results seem to support the theory that the two rises formed at the same time at the Mid-Atlantic Ridge and were broken apart and carried away from each other by sea-floor spreading processes. The diabasic nature of the basalt cored suggests it cooled relatively slowly; i.e., it may be a thick flow or sill. This is consistent with the physical properties data of Neprochnov et al. (this volume), who show that velocities and density are highest in the center of the basalt body and lower at top and bottom. A short interval of high drilling rate was noted in coring the basalt; this may indicate intercalated sediment which was not recovered (see Site 354 report, this volume). Also, the acoustic basement at both sites may contain multiple reflectors, which may indicate layering.

The oldest sediments just above basalt at Site 354 are assigned to the lower Maestrichtian *Arkhangelskiella cymbiformis* Zone, about 70 m.y.B.P. using the time scale adopted for Leg 39. This is not inconsistent with an 80 m.y. age of initiation of outpourings, if the basalts recovered represent the uppermost sill or flow in a series that may be 2 km thick.

Fossils in the lower Maestrichtian marls immediately overlying basement indicate a depositional depth of about 1000 meters (a depth of about 1000 m in the Maestrichtian was also noted for Site 366: Lancelot, Seibold, et al., in press). This is about 1600 meters shallower than the present worldwide average depth of the mid-ocean ridge system.

This only slightly exceeds the roughly 1 km elevation of the Ceará Rise basement above surrounding oceanic crust (Kumar and Embley, in press), and is very close if even a small degree of compensation is taken into account. Preliminary gravity data suggest some compensation (Embley and Hayes, 1972). Compensation is also required from subsidence theory. If the basal sediment was deposited at a depth of 1000 meters in the early Maestrichtian, subsidence at the normal rate for oceanic crust (Sclater et al., 1971) would require 2800 meters of subsidence until the present. The present depth of the basal sediments of about 4900 meters requires either an abnormally fast subsidence of the surrounding crustal plate or compensation beneath the Ceará Rise.

Rio Grande Rise

Site 357 is in 2086 meters of water on the northern slope of the Rio Grande Rise, 30 km west of Site 22, Leg 3 (Figure 1). A 797-meter sedimentary section was cored at close spacing; the lowermost sediments are marly chalks of early Santonian age, about 85 m.y. (igneous basement was not reached). The sediments are predominantly calcareous biogenic oozes, chalks, and limestone, deposited at relatively high rates (Figure 4), except for varying periods of hiatus or inferred hiatus (Figures 3 and 4). The sedimentary history at this site is described in the Site 357 chapter (this volume). This section will consider only sedimentological and paleontological evidence bearing on the origin and structural history of Rio Grande Rise. Detailed accounts of the inferred structural history (McDowell et al., this volume) and the subsidence history (Thiede, in press) of the rise have been presented elsewhere. Only those aspects of the origin and history of the rise for which Site 357 data have significance will be discussed here.

Depth-diagnostic sedimentary facies of known age can provide an insight into the vertical tectonic history of the Rio Grande Rise. Two such data points were derived from Site 357 drilling results. The first is the occurrence of abundant *Inoceramus* fragments in Santonian marly limestones about 2900 meters below present sea level. Thiede and Dinkelman (this volume) suggest that two findings oppose the possibility that the *Inoceramus* fragments are allochthonous. First, the fragments are present in finely laminated (reduced) sediments; introduction from outside the area would have disrupted the laminae, and no evidence of disruption was observed. Second, the *Inoceramus* fragments are the sole megafossil components. If introduced from a shallower water area outside the depositional site, they would be expected to be accompanied by other macrofaunal elements. The *Inoceramus* fragments are thus interpreted by Thiede and Dinkleman (this volume) as the remains of epibenthos living at the deposition site. These authors cite occurrences of *Inoceramus* remains from other DSDP sites and earlier studies of upper Mesozoic epicontinental sediments. They conclude that finding remains of autochthonous *Inoceramus* is significant as to depth, in that the organisms lived in lower neritic to upper bathyal depths; in this case, a depth of 300 to 500 meters is assumed. They therefore judge that the Santonian marly limestones at Site 357 were deposited in water of about this depth.

Earlier drilling results may be used as an independent check on this depth if we assume that the entire east-west portion of the rise belongs to the same structural province, i.e., was subject to the same motion over time. The basal sediment at Site 21 on the eastern Rio Grande Rise is a coquina overlain by *Inoceramus*-bearing Campanian sediments of Campanian (or pre-Campanian?) age, (Maxwell, Von Herzen, et al., 1970), and the presence of very shallow water indicators, including red algae, indicate that it formed at or only slightly below sea level. It is at a present depth of 2.2 km below sea level. Campanian sediments at Site 357 are at a present depth of 2.7 km. The resulting 500-meter difference between the two sites indicates a Campanian depth of this magnitude at Site 357. The underlying Santonian sediments were probably deposited in about the same water depth.

It should be noted here that there are paleontological arguments against the depth significance of the *Inoceramus* tests. The first concerns the laminated nature of the sediment. Although the laminae are preserved as discrete units, examination of the foraminiferal microfossil assemblages indicates a significant portion of older genera of mixed assemblages. Premoli-Silva and Boersma (this volume) believe this indicates introduction from outside the depositional site, perhaps as fine-grained or distal turbidites, or by downslope tractive creep. They argue that if microfossil elements are allochthonous, so might be the *Inoceramus* fragments. They also point out that the foraminifer fauna shows a high degree of dissolution, indicating deposition at or below the lysocline, which may have been at about 3000 meters in the Santonian. These factors will be discussed later.

A second and more problematic data point is the middle Eocene volcanic-sedimentary breccia, 2.4 km below present sea level and overlain and underlain by pelagic ooze facies of the same age. We are not sure whether the vulcanism that produced the breccia fragments occurred in the Eocene or occurred earlier, although McCoy and Zimmerman (this volume) indicate that the Eocene was a period of extensive vulcanism in the South Atlantic. The sedimentary particles in the breccia (which comprise only 5% of the total) contain shallow water indicators, notably red algae and nummulites, and the common size grading of the volcanic and sedimentary clasts indicates that both types of components originated in the same source area, which was in shallow water. The present depth of this middle Eocene "shoreline" is unknown since the erosional products from it moved downslope, but Thiede (in press) estimates a present depth of around 1000 to 1300 meters, corresponding to the depths of the heads of erosional canyons in the vicinity of Site 357.

Thiede (in press) uses two additional data points in attempting to determine the history of vertical displacement of the rise. One is the Campanian coquina at Site 21, now at a depth below sea level of 2.2 km, and which presumably formed at or near sea level. The other is the shallowest of a series of dredgings of outcrops of shallow water limestone (the shallowest one is used to minimize any effect of downslope movement); an Oligocene limestone was dredged at a depth of 1.4 km (Melguen, personal communication), but because this represents the depth at which the upslope dredging operation was started, and there is still the possibility of downslope movement, Thiede (in press) has accepted 1000 meters as the depth for this time point. Since the limestone is oolitic, it presumably formed in shallow water, probably less than 100 meters.

The left side of Figure 18 is a generalized north-south section through the Rio Grande Rise, showing the present depths below sea level at which the samples discussed above were collected. The depth points are then translated to a time-depth plot (right side of Figure 18). In order to determine the present true depths of the shorelines representing these various times, two corrections are necessary.

First a data point must be moved up (shallower) a distance equivalent to the depth below sea level at which the material accumulated (Santonian *Inoceramus*-bearing sediments, 300-500 m; Campanian coquina, sea level to 100 m; Eocene volcanic breccia ~1 km—because of transport; Oligocene limestone, sea level to 100 m). A second correction involves taking into account depression by the overlying sediment column, and is calculated by the method outlined by Berger and von Rad (1972). The four data points plotted in Figure 19 (a fifth point is the present shallowest part of the rise, some 650 meters below sea level, according to Dietrich and Ulrich, 1968) indicate a general submergence of the Rio Grande Rise in a manner and at a rate similar to that observed for normal oceanic crust (Sclater et al., 1971), as has been suggested for the Ninetyeast Ridge by Sclater and Fisher (1974) and more recently shown to apply to oceanic aseismic ridges in general (Detrick et al., in press).

Note that if the "paleo-shoreline" position points, as plotted on the right side of Figure 18, lie on a single subsidence curve of the Sclater type, the entire section of the rise over which the data were collected would have had to form at about the same time and subside as a unit. Thiede (in press) considers this assumption valid, and uses families of subsidence curves to estimate the age of basement under Site 357 as about 95 m.y. and the present depth as about 3700 meters below sea level (see Summary and Conclusions, Site 357, this volume).

Because of the very large unceratinties associated with paleontologic dating by epoch only (the "Oligocene" shallow water limestone could have an age uncertainty of over 10 million years; the "Campanian" age of the Site 21 coquina is based on the age of the overlying sediment, and a hiatus is possible), and the equal uncertainty of depth of formation or of collection of some samples (Eocene breccia, Oligocene dredged limestone), we choose to be slightly more conservative here. We do not believe that the data points on the right of Figure 18 necessarily lie on a unique subsidence curve; we do believe, however, that taken together, they indicate subsidence of the rise in a manner similar to that of oceanic crust.

Figure 19 is a subsidence curve for Site 357 based on two age-depth points for a single depth-indicative sedimentary indicator at Site 357. Point A is the present depth below sea level of the lower Santonian marly *Inoceramus*-bearing limestone, about 2500 meters after correction for sediment load according to the method of Berger and von Rad (1972). Point B is the depth of this sediment below sea level at the time of deposition, which we will accept as 500 meters. The Sclater-type subsidence curve passing through these points is unique, and requires that Site 357 began its subsidence at the point of origin 0 of the curve, i.e., 92 m.y.B.P. This age is consistent with the sea-floor age of the Brazil Basin to the north, based on magnetic anomalies (Ladd, 1974) and extrapolated sea floor spreading rates (Larson and Pitman, 1972).

Figure 19 does not indicate the depth of basement below the lowest sediments reached. McDowell et al. (this volume) report a "middle layer" (velocity 3.5-3.6 km/sec), 850 to 900 meters thick, lying between a sedimentary layer (2.3 km/sec) and what is presumably basement (4.3-4.7 km/sec). Deposition of the 850-meter layer would require accumulation rates of more than 100 m/m.y. Although this rate is very high for sediments, these authors suggest that the "middle layer" may be composed of volcanics, perhaps with interlayered sediments. Such a composition would explain both the required high rate of accumulation and the measured interval velocity. Such lenses of volcanics overlying ocean crust may be typical of certain aseismic ridges, as was previously noted in the discussion of the origin of the Ceará Rise (see Houtz et al., in press).

The critical question whether lower Santonian marly chalks were deposited in deep or shallow water is still open to doubt. The argument that *Inoceramus* may have been introduced from shallower depths is substantial, on the basis of the allochthonous foraminifer assemblages and the high (~30 m/m.y.) Santonian sediment accumulation rate. The dissolution of planktonic foraminifers in the upper Santonian and Campanian (where it was particularly severe) is more equivocal. The dissolution may have been a local current effect, or may have been caused by acidity, as suggested by retalively high organic contents. If it was a CCD effect, it may not necessarily indicate great depth.

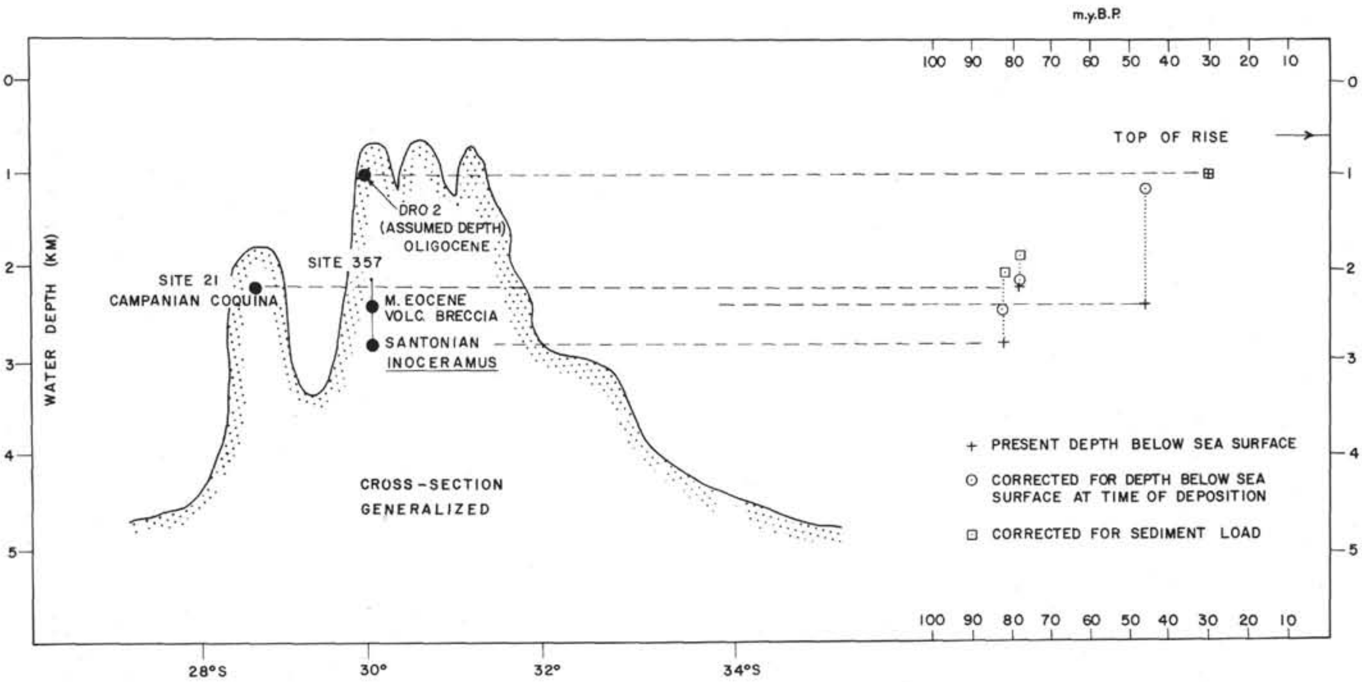


Figure 18. *Left, generalized cross section of the Rio Grande Rise showing the present depths of shallow water indicators of known age. Right, time-depth plot of former shoreline positions of the rise. The path and rate of subsidence is similar to the age-depth constancy associated with normal oceanic crust (Sclater et al., 1971).*

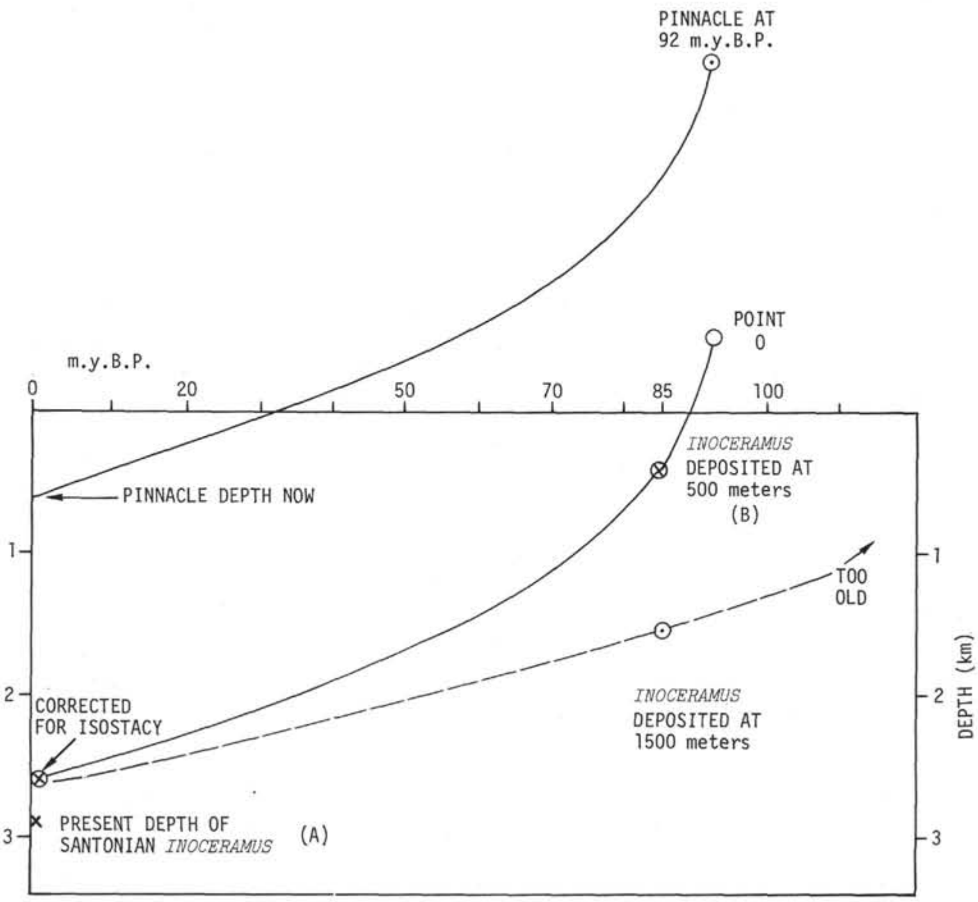


Figure 19. *Probable subsidence path for the Rio Grande Rise at Site 357 with a resulting basement age of ~92 m.y.B.P. See text for discussion.*

On the basis of the few data points available, van Andel (1975) takes the Campanian CCD in the South Atlantic to be about 3500 meters. Our data indicate a Campanian CCD shallower than this. The depth of the corroded Campanian foraminifers is only 2400 meters (corrected for sediment load) at present; assuming that no uplift has occurred, the Campanian lysocline had to be at a depth of 2400 meters at most, and had to be shallower than this if the site has subsided since the Campanian. Recognition of variations in CCD in space and time remains a great problem. Finally, we must realize that if the deposition site in the Santonian was significantly deeper than 500 meters, and if our assumption of subsidence along the Sclater curve is valid, the required basement age would be improbably old (e.g., the dashed line in Figure 19, which assumes deposition of the Santonian limestones at 1500 meters, and which would require an unreasonably old basement). Thiede (in press), assuming a deposition depth of 500 meters for the Santonian *Inoceramus*-bearing marly limestones, and the applicability of the subsidence curve, calculates the deposition depth of the facies recovered at Site 357. This is reported in Table 5 of the Site 357 chapter (this volume).

The oldest cored sediments at Site 357 and the dredgings on the top of Rio Grande Rise indicate something of the early nature of the feature. The Santonian *Inoceramus*-bearing marly limestones, possibly deposited originally in water no more than a few hundred meters deep, now lie almost 3 km below sea level. McDowell et al. (this volume) report that igneous rocks have been dredged from depths of about 1000 meters. This indicates that in early Santonian time, portions of the Rio Grande Rise stood as an island or islands 2000 meters or more above sea level, an elevation similar to the present elevation of Tristan da Cunha (2060 m; see also the upper curve in Figure 19). Analyses of the basalts from the Eocene breccia at Site 357 (Fodor and Thiede, this volume) and the Lamont dredgings (Fodor and Husler, 1976) indicate that they are alkalic rather than tholeiitic, i.e., similar to the rocks of the high igneous islands of the central South Atlantic (Baker, 1973). Similar results have been reported for the eastern Walvis Ridge, where dredgings of Albian-Cenomanian shallow water macrofossils at present depths of 2.7 km, and the shallowest igneous rocks at 1.2 km, indicate that this feature too, at least locally, may have stood some 1500 meters above sea level early in its history (Pastouret and Goslin, 1974; Goslin et al., 1974). The igneous rocks dredged from the Walvis Ridge are also alkalic (Hekinian, 1972).

Data from Site 357 and other studies cited above indicate that the east-west portion of the Rio Grande may have formed as an island or islands in the vicinity of the spreading center, about 90-95 m.y. ago at Sites 357 and 21, perhpas earlier to the west and later to the east. The island(s), up to 2 km above sea level, were at least capped by alkalic basalts (Fodor and Thiede, this volume; Fodor and Husler, 1976). Gravity data indicate some compensation (McDowell et al., this volume), which was probably local and accomplished soon after formation of the feature (Detrick et al., in press). Subsidence from formation until present followed a path parallel to that typical of oceanic crust (Thiede, in press; Detrick et al., in press; Site 357 chapter, this volume); the main plateau of the rise subsided below sea level in the Campanian (Perch-Nielsen, Supko, et al., 1975), and the uppermost peaks did so in the Oligocene-Miocene. Sedimentary conditions at Site 357 were restricted in the Santonian, became open-marine in the Campanian, and oxygenated conditions have prevailed ever since. Sedimentation has been almost wholly biogenic (except for terrigenous input in the Santonian and Campanian) and continuous (apart from several local or regional hiatuses or periods of slow accumulation—see earlier section), and in an increasingly deepening environment as subsidence proceeded. Geophysical results (McDowell et al., this volume) indicate true igneous oceanic basement may lie some 850 meters below the Santonian marly limestone, the top of which has now been shown to correlate with the deepest regionally recognized seismic reflector (Site 357 chapter, this volume).

Walvis "Seamount"

Site 359 was drilled in the southwestern, or "seamount," province of the Walvis Ridge; water depth at the site is 1658 meters. Soviet bathymetry data collected after the drilling show that the site lies on a lineament running northeast for an indeterminate distance (see Neprochnov et al., this volume). To the south and west of Site 359, the Walvis Ridge consists of individual seamounts; the islands of Tristan de Cunha and Gough, to the northeast, are discontinuous narrow elongate positive features, presumably chains of closely spaced seamounts (Dingle and Simpson, manuscript).

The top sedimentary unit cored consists of 57 meters of Pleistocene to Miocene and upper Eocene foraminifer-nannofossil ooze. A major hiatus is present from the upper Eocene to the mid-Miocene, and portions of the Pliocene and Pleistocene are also missing; it is not known whether these gaps are of regional or only local importance. The biogenic oozes overlie 29 meters of calcareous volcanic mud with pumice layers, which in turn overlies at least 10 meters of a gray to black volcanic ash flow tuff, in which the hole bottomed. The tuff has a trachytic bulk composition, similar to ashflow tuffs occurring on several Atlantic islands, including Tristan da Cunha and Gough (Fodor et al., this volume). The tuff at Site 359 is presumed to have been subaerially emplaced, on the basis of its lack of graded bedding, lack of evidence of reworking by marine organisms, and its lack of included shell material or marine sediment clasts. A K/Ar date obtained on the feldspars in the tuff indicates an age of 40.7 ± 1 m.y. That the seamount subsided below sea level just after the tuff was deposited is indicated by the fact that the overlying calcareous volcanic mud is of roughly the same age (late Eocene, Zone P15) and contains marine fossils, both open sea pelagic nannoflora and foraminifers and indicators of relatively shallow water, including benthic foraminifers, larger mollusc fragments, bryozoans, and brachiopods.

We may use sea-floor spreading and subsidence theory to make some observations regarding the history of the seamount. Site 359 is at about the location of anomalies 23 to 25 (Ladd, 1974; Larson, personal communication). The age range for these anomalies is 47 to 53 m.y., according to Tarling and Mitchell (1976). We use an age of 50 m.y. for the subsidence calculations below, although recent work indicates an age range of 54 to 59 m.y., which may be more appropriate for anomalies 23 to 25 (La Brecque et al., in press).

Detrick et al. (in press) show that aseismic ridges and oceanic islands subside at rates typical of normal oceanic crust. The solid line in Figure 20 indicates that if the Site 359 seamount was formed at the Mid-Atlantic Ridge 50 m.y. ago, that portion then at sea level would now be at a depth of 2400 meters. The isostatically corrected depth of the top of the tuff unit is about 1725 meters. The subaerial nature of the tuff indicates that either the seamount was once an island or at least 800 meters of upgrowth by volcanic processes had occurred between 50 and 40 m.y. ago.

The late Eocene, 40 m.y.B.P., was the time at which the tuff (i.e., the island) subsided below sea level and the time when igneous activity ceased, as indicated by the absence of volcanic components in the uppermost biogenic oozes of late Eocene age. Two factors support the argument that the tuff did not form at the spreading center 40 m.y.B.P. If it did, the dash-dot subsidence curve indicates that the tuff would now be at a depth of 2 km, 400 meters deeper than at present. Also, the present position of the site is about 1000 km from the spreading center; this indicates, on the basis of the average South Atlantic spreading rate of 2 cm/year, an age of about 50 m.y. (Larson and Pitman, 1972).

We believe that the top of the Site 359 seamount was just submerging below sea level at 40 m.y.B.P., and that it has been subsiding along with oceanic crust at least 10 m.y. older. Subsidence would then have been along the dashed line in Figure 28, and the top of the tuff should now lie at a depth of 1200 rather than 1625 meters. Even allowing for some isostatic compensation, the tuff is anomalously deep. An initial depth of 300 to 400 meters would be more in line with evidence from benthic foraminifers, which are interpreted as a slope fauna (Boersma, this volume). The larger megafossils (including an oyster-like bivalve) could be in place or transported from local shallower sources; the presence of open ocean pelagic microfossils is easily explained by the surrounding deep water. Fodor et al. (this volume) have suggested that the subaerial tuff subsided catastrophically by summit collapse of a shield-volcano, although a depth of only 300 to 400 meters and the uncertainties of radiometric and paleontologic dating techniques may not require a catastrophic event.

The main constructional phase of oceanic island building may be short, on the order of 1-2 million years in the Hawaiian Islands (Jackson et al., 1972). Activity at Site 359, in contrast, continued for at least 10 million years. Presumably, an entrained magma chamber was carried along with the island while it moved as much as 200 km away from the spreading center.

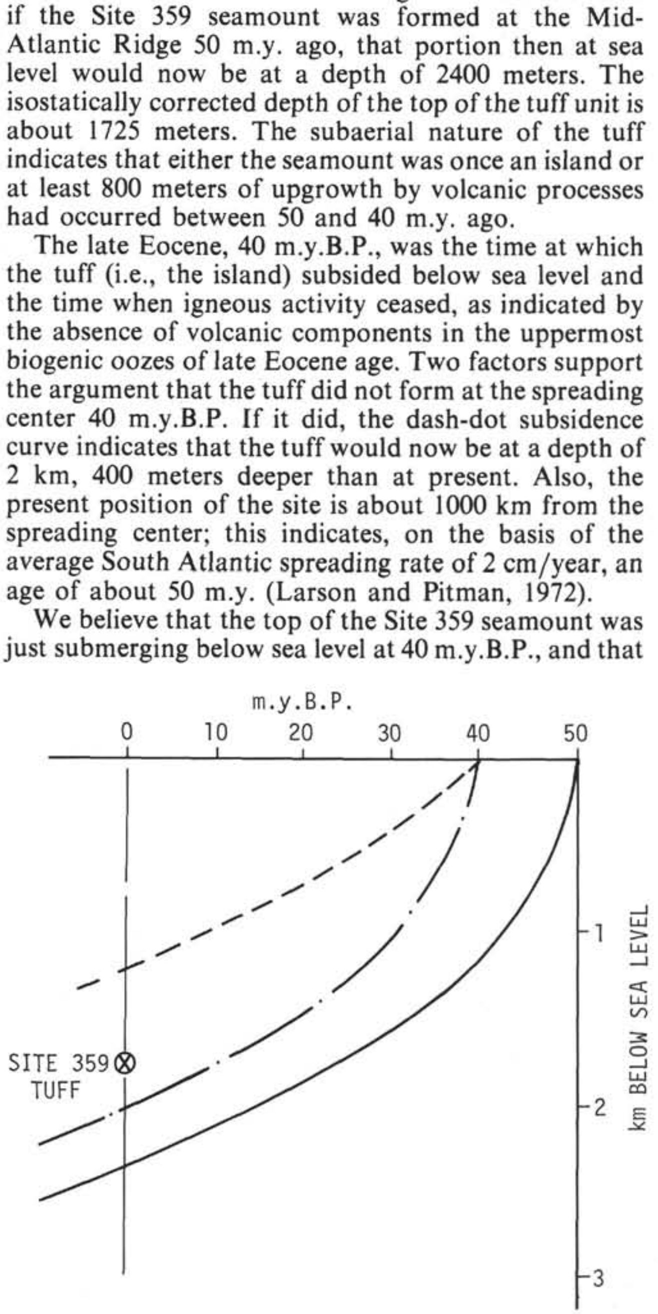


Figure 20. *Subsidence curves having bearing on the age and history of deposition at the Walvis Seamount Site 359. See text for discussion.*

ACKNOWLEDGMENTS

This chapter is a synthesis of some of the major geologic findings consequent to Leg 39 drilling, put into a larger context. Many of the ideas took form at sea and many issued from meetings after the cruise, as offspring of the studies of our Leg 39 colleagues. Although some specific ideas are cited here, in many cases it is hard to track down the beginning of a thought; we here express our deep gratitude to all of our shipboard colleagues for their many hours of hard work at sea and ashore. Similarly, we thank those shore laboratory investigators whose studies, cited in the text, provided valuable data.

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