52. PALEOCIRCULATION OF THE SOUTHWESTERN ATLANTIC[1]

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ABSTRACT

The Cenozoic history of the abyssal circulation of the southwestern Atlantic is punctuated by tectonically or climatically controlled "threshold" events, triggered by the changing geophysical boundary conditions associated with plate tectonic development of the basins. The east-west volcanic lineaments of the Rio Grande Rise/Walvis Ridge and the Ceara-Sierra Leone rises effectively blocked the meridional exchange of deep waters for at least the first 20–30 Ma of South Atlantic history. Pelagic biogenic sedimentation persisted through the equable climates of the Late Cretaceous and early Paleogene, with no evidence of significant bottom current activity. The aseismic Rio Grande Rise extended above sea level during the initial episodes of volcanism in the Santonian-Coniacian and subsided at near-normal crustal subsidence rates for the first 35 Ma of its history. A major thermal episode during the middle Eocene (47 Ma) re-elevated large areas of the Rise to sea level; the shallowest guyots of the Rise subsided below sea level in the Oligocene, and the Rise has continued to subside since then.

Pelagic sedimentation in the deep basins was interrupted by a major erosional episode that began in the late Eocene to early Oligocene. High-salinity water (which was advected into the South Atlantic after the spreading of the Greenland-Scotland Ridge) extended to the perimeter of the Antarctic continent. This "proto-North Atlantic Deep Water" mixed with ambient cold surface waters to form a "proto-Antarctic Bottom Water." During the latest Miocene, the closure of the Gibraltar Sill stopped the high-salinity Mediterranean outflow, but the outflow started again in the early Pliocene, and the subsequent reinitiation of North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW) produced extensive erosional hiatuses below 2000 m on the Rio Grande Rise. The interval between 4.0 and 3.8 Ma is associated with a major cooling trend in the Southern Ocean.

Proto-NADW and proto-AABW production decreased between 3.8 and 3.2 Ma; circumpolar flow intensified, resulting in widespread erosion over the Maurice Ewing Bank. At 3.2 Ma, the growth of northern hemisphere ice sheets and the intensification of both NADW and AABW caused the Pliocene climates to deteriorate rapidly. The closure of the Panama isthmus probably contributed to these events by blocking the westward Tethyan flow in the tropical Atlantic and strengthening the northward transport of the shallow Western Boundary Current (which, in turn, advected warm surface water adjacent to eastern Canada to provide a source for evaporation to build the Arctic ice sheets and advected high-salinity subtropical waters into the northeastern Atlantic). The latest Pliocene and Quaternary were marked by co-occurring variations in NADW and AABW intensity, with the flow of NADW perhaps strongest at 40,000-yr. intervals, corresponding to maxima in summer insolation and tilt.

INTRODUCTION

The southwestern Atlantic is one of the most critical regions for examining the history of the major thermohaline current systems, because four of these deep currents are clearly identifiable in the modern hydrography of the region, and they intersect the seafloor on the flanks of the Rio Grande Rise and leave distinct "fingerprints" in the geologic record. Deep thermohaline flow of Antarctic Bottom Water (AABW) extends northward through the western basins of the Atlantic (e.g., Wüst, 1935; Wright, 1970; Gordon, 1971; Broecker et al., 1976), with an identifiable chemical signature as far north as the continental margin of New England and Nova Scotia (Needell, 1980; Richardson et al., 1981). Overlying the Antarctic-source water is southward-flowing high-salinity overflow from the Norwegian-Greenland Seas, commonly designated as North Atlantic Deep Water (NADW) (Broecker and Takahashi, 1980). This water mass can be traced as a high-salinity tongue into the southwestern Atlantic (Reid et al., 1977), eastward within the circumpolar flow (Reid and Lynn, 1971), and still farther south to the continental slopes bordering Antarctica (e.g., Foster and Middleton, 1980). The eastward-flowing Circumpolar Deep Water (CPDW) is of comparable density to NADW and is split into upper and lower branches that appear as oxygen minima on hydrographic sections (Reid et al., 1977). Antarctic Intermediate Water (AAIW) sinks at the polar front near 50°S, and can be traced northward as a subsurface salinity minimum centered near 1000 m depth (e.g., Buscaglia, 1971). Each of these four water masses intersects the flanks of the Rio Grande Rise near 30°S; thus, strategically located coring sites on the Rise offer the opportunity to examine the paleocirculation history of the region.

Leg 72 of the Deep Sea Drilling project provided an excellent suite of cores through which we may examine the Cenozoic history of these major water masses. Four sites were cored with the hydraulical piston corer (HPC) on the flanks of the Rio Grande Rise, at depths ranging from 1313 to 4250 m, such that each site intersects one of the principal water masses. In a separate report, Johnson (this volume) reviews the modern hydrography and circulation in the southwestern Atlantic. This report is a synthesis of the various paleoceanographic studies presented in this volume, plus previously published observations that have a direct bearing on southwestern Atlantic paleocirculation.

[1] Barker, P. F., Carlson, R. L., Johnson, D. A., et al., *Init. Repts. DSDP*, 72: Washington (U.S. Govt. Printing Office).

A proper assessment of the circulation history of the southwestern Atlantic requires the use of appropriate boundary conditions imposed by the spreading and subsidence history of the underlying ocean crust. Barker and Carlson (this volume) summarize the geophysical constraints associated with seafloor spreading, midplate volcanism, and crustal subsidence in the late Mesozoic and Cenozoic history of the southwestern Atlantic. These processes are reflected in gradual changes of the geometry of major basins and their associated sill depths. Even though the boundary conditions may evolve gradually, the adjustments in circulation patterns are most likely to be catastrophic, because a particular barrier to circulation may be abruptly opened or closed when a certain threshold sill depth is exceeded at an important passage. The geologic record supports the notion that paleocirculation "events" have indeed occurred abruptly, in a geologic sense. Some of these events may be climatically controlled (e.g., caused by the earth's energy balance and mechanisms of heat transfer between equator and poles), and others may be tectonically controlled.

The results and interpretations presented in this synthesis should be viewed in a global context and not simply as a regional study, because each of the water masses represented in the southwestern Atlantic can be traced laterally over thousands of kilometers (Reid et al., 1977; Johnson, 1982). The HPC cores from the depth transect on Leg 72 complement a transect of five sites on the Walvis Ridge in the southeastern Atlantic (Moore, Rabinowitz, et al., in press). Future HPC transects of Neogene sediments are planned for DSDP Leg 90 in the southwestern Pacific (early 1983) and DSDP Leg 94 in the northeastern Atlantic (July–August 1983). Each of these four transects will contribute toward deciphering the history and mechanisms of formation of the major deep ocean currents. The interpretations derived from the Leg 72 cores should be therefore viewed as preliminary and will certainly be revised when additional information is reported from the three subsequent drilling transects.

OBJECTIVES AND SCOPE OF THIS SYNTHESIS

In this synthesis, we shall address the following questions pertinent to the circulation of the southwestern Atlantic:

1) What were the first-order threshold events in the initiation or cessation of deep thermohaline circulation during the evolution of the Rio Grande Rise region of the southwestern Atlantic? Which water masses/current systems were involved? Did these water masses have properties and spatial extents comparable to their equivalents in the modern ocean? How precisely can these threshold events be dated? What tectonic and/or climatic conditions may have been responsible for these events?

2) Did these events extend beyond the southwestern Atlantic to regions farther "upstream" and "downstream" in each of the major water masses?

3) What ambiguities and uncertainties are associated with the interpretations from the Leg 72 cores, because of poor core recovery; poor stratigraphic control; or uncertainties in applying the geological fingerprints of paleocirculation?

4) Based on these uncertainties, can we designate high-priority target areas for future drilling for paleoceanographic objectives in the southwestern Atlantic?

There are a number of aspects of southwestern Atlantic paleocirculation that will *not* be considered in this synthesis because of time and space limitations, specifically:

1) We are concerned principally with the *deep* thermohaline flow and not surface currents. The mapping and interpretation of shallow water masses requires extensive geographic sample control in order to identify the principal fronts and circulation gyres from faunal and floral biogeographic patterns. Regional studies of this type have been carried out for Tertiary sediments (e.g., Haq and Lohmann, 1976; Kennett, 1980) and for the Quaternary (e.g., CLIMAP, 1976; Cline and Hays, 1976). The DSDP Leg 72 sites contribute relatively little new information to previous studies of this type in the South Atlantic.

2) We will not attempt the mapping and interpretation of lithofacies for various time slices on an ocean-wide scale. A number of previous publications contain lithologic syntheses of this type (e.g., McCoy and Zimmerman, 1977; van Andel et al., 1977; Melguen et al., 1978). The lithologic results obtained during Leg 72 are consistent with the first-order interpretations discussed in these earlier papers and thus do not warrant reinterpretation.

3) We will not consider the results from the Leg 74 coring transect on the Walvis Ridge, because these will be treated thoroughly in a subsequent volume (Moore, Rabinowitz, et al., in press). We might expect to find close parallels in the thermohaline circulation histories of the southeastern and southwestern Atlantic, especially for the water masses at intermediate depths (e.g., AAIW, CPDW, NADW). However, the sill depths of the Mid-Atlantic Ridge are such that the abyssal hydrography and circulation (i.e., below ~3500 m) are distinctly different east and west of the ridge crest, and this difference is reflected in the underlying sediments (e.g., Bremer and Lohmann, 1982; Thunell, 1982). We anticipate that the geologic effects of AABW should be much less clearly developed on the Walvis Ridge than on the Rio Grande Rise because of the east-west asymmetry in deep thermohaline flow. The shallower water masses, however, might be expected to show similar histories in the depositional records from the two regions.

4) We will not be considering high-resolution interpretations of the paleocirculation during the latest Pleistocene and Holocene, because these kinds of studies are addressed most satisfactorily with large-diameter gravity cores or other sampling devices that can recover the uppermost few centimeters to tens of centimeters in a relatively undisturbed form (e.g., Jones et al., 1981; Jones and Ruddiman, 1982). We shall discuss the nature of the problems that are currently being addressed by

such studies, but the specific results and interpretations will be published separately.

SOURCES OF DATA

In Table 1 we have listed the principal publications from this volume upon which we have based the interpretations in this chapter. In addition to the specialty chapters in the Leg 72 report, we have used selected results from the Leg 36 and Leg 39 *Initial Reports* (Barker, Dalziel, et al., 1977; Supko, Perch-Nielsen, et al., 1977), from the Leg 71 summary article in the *Geological Society of America Bulletin* (Ludwig, Krasheninnikov, et al., 1980), and from numerous studies of conventional piston and gravity cores in the southwestern Atlantic (e.g., cruises of *Chain* and *Islas Orcadas:* see Johnson et al., 1977; Ledbetter, 1979; Ciesielski et al., 1982). We have included several regional geophysical studies of seismic profiles when it has been possible to identify and date major acoustic horizons with drill information, and we have referred to those regional studies of sediment lithofacies that provide a first-order perspective on ocean basin geometry and depositional provinces for particular time slices in the Late Cretaceous and Cenozoic. Our emphasis is upon the four sites drilled on Leg 72; data from other cores and drill sites are included to provide corroboration or refinement of the Leg 72 observational data.

THE CRETACEOUS SOUTH ATLANTIC OCEAN

Tectonics and Physiography

The South Atlantic began to open in the Early Cretaceous, approximately 130 Ma (Larson and Ladd, 1973). The magnetic anomaly pattern and the locations of rotation poles (Rabinowitz and LaBrecque, 1979) show that the rate of spreading must have increased toward the south. The configuration of the opening basins was such that a continuous barrier was formed by the east-west trending São Paulo Plateau/Walvis Ridge complex. Parts of the São Paulo Plateau extended to or above sea level for much of the Cretaceous (Kumar et al., 1977), and the same is true of the Walvis Ridge until at least the middle Cretaceous (Bolli, Ryan, et al., 1978). This topographic barrier was crucial in controlling the lithofacies distribution patterns and the circulation of shallow and intermediate water masses.

The early South Atlantic was thus divided into the Cape-Argentine and the Angola-Brazil basins by an east-west lineament (São Paulo Plateau/Walvis Ridge complex). Both basins subsided to depths of perhaps

Table 1. Chapters in Leg 72 *Initial Reports* (this volume) that present observations and interpretations incorporated in this synthesis.

Author(s)	Title	Material studied
Barash, Oskina, Blyum	Quaternary biostratigraphy and surface paleotemperatures based on planktonic foraminifers	Quaternary biogeography and dissolution indices
Barker, Buffler, Gamboa	A seismic reflection study of the Rio Grande Rise	Single-channel and multichannel profiles on crest and upper flanks of Rise
Barker	Tectonic evolution and subsidence history of the Rio Grande Rise	Model for thermal subsidence of Rise, using paleobathymetric constraints at Sites 357 and 516.
Benson, Peypouquet	The upper and mid-bathyal Cenozoic ostracode faunas of the Rio Grande Rise found on Leg 72 DSDP	Ostracodes, Maestrichtian to Recent (Site 516); late Neogene (Sites 517 and 518)
Berggren, Hamilton, Johnson, Pujol, Weiss, Čepek, Gombos	Magnetobiostratigraphy of DSDP Leg 72, Sites 515–518, Rio Grande Rise (South Atlantic)	Magnetostratigraphy and biostratigraphy of all Leg 72 sites
Bryan, Duncan	Age and provenance of clastic horizons from Hole 516F	Petrography and age of ashes in middle Eocene volcanogenic turbidites
Dailey	Late Cretaceous and Paleocene benthic foraminifers from DSDP Site 516, Rio Grande Rise, western South Atlantic Ocean	Coniacian-Paleocene benthic foraminifers: paleobathymetry and paleoecology
Vergnaud Grazzini, Grably, Pujol, Duprat	Oxygen isotope stratigraphy and paleoclimatology of southwestern Atlantic Quaternary sediments (Rio Grande Rise) at DSDP Site 517	Pleistocene and late Pliocene stable isotopes in benthic and planktonic foraminifers, Site 517
Hodell, Kennett, Leonard	Climatically induced changes in vertical water mass structure of the Vema Channel during the Pliocene: evidence from DSDP Holes 516A, 517, and 518	Benthic foraminiferal paleoecology and stable isotopes, and dissolution indices: late Miocene to late Pliocene
Johnson	Cenozoic radiolarians from the Brazil Basin and Rio Grande Rise	Oligocene and Miocene radiolarians, Sites 515 and 516
Leonard, Williams, Thunell	Pliocene paleoclimatic and paleoceanographic history of the South Atlantic Ocean: stable isotopic records from Leg 72 DSDP Holes 516A and 517	Stable isotopes: Site 516, 2.7–4.6 Ma; Site 517, 1.8–3.2 Ma
Milliman	Coniacian/Santonian depositional environment on the Rio Grande Rise as evidenced from carbonate sediments at Hole 516F	Paleoecology of carbonate lithofacies in basal sediment sequence on Rio Grande Rise
Mussett, Barker	$^{40}Ar/^{39}Ar$ age spectra of basalts, DSDP Site 516	Argon-argon geochronology of basement rocks, Rio Grande Rise
Pujol	Cenozoic planktonic foraminiferal biostratigraphy of the southwestern Atlantic (Rio Grande Rise): DSDP Leg 72	Stratigraphy, ecology, dissolution of Cenozoic planktonic foraminifers
Shor, Jones, Rasmussen, Burckle	Carbonate spikes and displaced components at DSDP Site 515: Pliocene/Pleistocene depositional processes in the southern Brazil Basin	Displaced diatoms; carbonate analyses; benthic foraminiferal assemblages and $\delta^{18}O$
Tjalsma	Eocene to Miocene benthic foraminifers from DSDP Site 516, Rio Grande Rise, South Atlantic	Benthic foraminiferal paleobathymetry and paleoecology
Williams, Healy-Williams Thunell, Leventer	Detailed stable isotope and carbonate records from the late Maestrichtian-early Paleocene section of Hole 516F, including the Cretaceous/Tertiary boundary	Total $CaCO_3$; stable oxygen and carbon isotopes on bulk samples
Zimmerman	Clay mineral stratigraphy of the Rio Grande Rise and southern Brazil Basin, western South Atlantic Ocean	Sites 515 and 516, X-ray mineralogy

~3 km by middle Cretaceous (approximately Aptian) time; evaporite deposition then began in each and filled them with up to 3 km of salt deposits (Kumar and Gamboa, 1979).

The age of the volcanism associated with the São Paulo Plateau/Walvis Ridge complex cannot be determined with a high degree of certainty. Drilling at Site 516 near the summit of the Rio Grande Rise reached volcanic basement, which has a radiometric age of approximately 86 Ma (Mussett and Barker, this volume). Much of the volcanism associated with this east-west tectonic barrier, however, may have been considerably older. The lithologic evidence for restricted circulation (see next section) requires that a tectonic barrier existed along the trend of the São Paulo Plateau/Walvis Ridge complex before and during the stages of evaporite deposition (i.e., before 100 Ma). Thus, the volcanic basement to the west of Site 516 probably is considerably older than the Coniacian/Santonian age determined at Site 516, and the basement age along the São Paulo Plateau/Walvis Ridge complex may increase progressively away from the spreading center.

After the Late Cretaceous volcanism on the Rio Grande Rise near Site 516, volcanic basement subsided very rapidly. Volcanism and associated calcareous sedimentation occurred in shallow-water environments (several meters or less in depth) on a ridge or platform that was near sea level in the Coniacian-Santonian (Thiede and Dinkelman, 1977; Milliman, this volume). By the early Campanian, the ridge had subsided to middle slope depths (Dailey, this volume).

Lithofacies

The initial rifting of the South Atlantic created a narrow oceanic gulf comparable in some respects to the Gulf of California. Most of the pre-Aptian deposits filling these basins were turbidites (south of the São Paulo Plateau/Walvis Ridge complex) or thick fluvial and lacustrine deposits (north of the ridge). This ridge formed an effective barrier to seawater penetration from the south during most of the Early Cretaceous (about 100–130 Ma). This stage in the opening of the South Atlantic has been referred to as the "rift valley" stage (Kumar and Gamboa, 1979).

Beginning in the Aptian, extensive deposits of evaporites began to accumulate in the deepening Brazil and Angola basins, to the north of the São Paulo Plateau/Walvis Ridge complex. The evaporites probably rest upon continental sediments, thus the marine depositional history of the South Atlantic probably begins with this evaporite sequence. The only basins that were sufficiently isolated and restricted to allow evaporite deposition are those that extended northward from the São Paulo Plateau/Walvis Ridge complex to the African continental margin in the Gulf of Guinea. The evaporite sequence, with a thickness of up to 3 km (Kumar and Gamboa, 1979), suggests a dry and warm climate for the central basin, with intermittent spillover of ocean water across the São Paulo Plateau/Walvis Ridge barrier to the south. This barrier was effective in restricting south-to-north surface circulation, as indicated by the separation of temperate planktonic assemblages in the Cape Basin from the tropical to subtropical ones in the Angola Basin. The evaporitic depositional environment in these basins was terminated by the early Albian; some authors have suggested that the entire interval of salt deposition required only about ~3 m.y. during the late Aptian (Kumar and Gamboa, 1979).

Albian sediments overlying the evaporites of the Brazil and Angola basins include shales with a high content of organic carbon (up to 5%), indicating an oxygen-deficient environment of deposition (Thierstein, 1979). This facies is commonly interpreted as indicative of stagnant conditions (Thiede and van Andel, 1977), and supports other evidence that a vigorous intermediate and deep thermohaline circulation did not exist in the Late Cretaceous. The organic matter that formed these shales was not necessarily of marine origin. An increased influx of land-derived organic matter, coupled with restricted deep circulation and an oxygen minimum below sill depth, would be sufficient to create the black shales. The global extent of the mid-Cretaceous shales (Arthur and Natland, 1979), however, would argue that the conditions favorable to supply and preservation of organic matter in sediments were global in extent, not simply a consequence of tectonic and oceanographic conditions that were unique to the South Atlantic.

After the middle Cretaceous interval of black shale accumulation, biogenic pelagic sedimentation under normal oxygenated conditions predominated in the South Atlantic for the remainder of the Cretaceous and into the early Paleogene. Melguen and others (1978) present maps of Cretaceous lithofacies in the South Atlantic to show the contrasting depositional conditions in the middle Cretaceous (Anomaly M-0, 110 Ma, fig. 15) and Late Cretaceous (Anomaly 34, 79 Ma, fig. 17).

Although portions of the São Paulo Plateau/Rio Grande Rise/Walvis Ridge barrier remained close to sea level for much of the Cretaceous, individual segments apparently experienced rapid subsidence from shallow to bathyal paleodepths over a time span of a very few million years. At Site 516, for example, the paleodepth indicators for the Late Cretaceous indicate very rapid subsidence (Table 2) (see Dailey, this volume; Milliman, this volume).

We believe that this history of rapid initial subsidence is representative of that which might be expected for other segments of the São Paulo Plateau/Rio Grande Rise/Walvis Ridge complex. Presumably the sequence of volcanic and subsidence episodes becomes progressively younger in the direction of the Mid-Atlantic Ridge, if the volcanism that created the São Paulo Plateau/Walvis Ridge complex was associated with plate motion over one or more mantle hot spots.

Paleoclimatology and Paleocirculation

The degree of interaction between the emerging South Atlantic Ocean and the global circulation system depended on the opening and deepening of north-south passages for surface and deep circulation through the three transverse barriers in southern (~50°S), central (~30°S), and equatorial latitudes. A surface connec-

Table 2. Paleodepth indicators for the Late Cretaceous at Hole 516F.

Core	Age	Lithology	Paleodepth
125	Coniacian	Miliolids, ophthalmidiids, ostracodes in well-sorted skeletal grainstone	< 20 m, high-energy environment
124	Coniacian	*Inoceramus* shell fragments, calcispheres, coralline algae, verneuilinids, miliolids, heterohelicid, and globotruncanid planktonic foraminifers; iron-strained mud matrix	20–150 m, open marine shelf conditions
123	Coniacian	Alternating layers of claystone and fine-grained skeletal debris in a claystone matrix; common *Inoceramus* prisms, ostracodes, heterohelicid and hedbergellid planktonic foraminifers	200–500 m, deeper neritic or shallow bathyal depths
119 through 116	Santonian	Abundant deep water benthic foraminifers (including *Gavelinella beccariiformis*); no shallow indicators; high planktonic/benthic foraminiferal ratio (P/B ~ 10:1); evidence of selective dissolution	500–1000 m, upper to middle bathyal depths
113 through 89	Campanian/ Maestrichtian	Diverse and well-preserved benthic assemblages; high P/B ratio; diverse gyroidnids and agglutinated benthic foraminifers	1000–1500 m, middle bathyal to shallow lower bathyal depths
89 through 83	Paleocene	Benthic foraminifers dominated by a deep-water association; wide fluctuations in abundance in lowest Paleocene; pelagic character	1500–2000 m, lower bathyal depths

tion between the Southern Ocean and the South Atlantic was established as soon as initial rifting began (~ 130 Ma), but a continuous passage between the North and South Atlantic did not develop until at least the Albian, based on ammonite distribution patterns (Reyment and Tait, 1972). Before this time, the South Atlantic was a narrow Mediterranean-type sea with no connection to the North Atlantic nor to the Tethys across Africa. The São Paulo/Rio Grande/Walvis Ridge restricted north-south circulation to very shallow depths (< 1 km) at ~ 30°S, and the Falkland Plateau provided a similar barrier near 50°S (van Andel et al., 1977). At the north end of the South Atlantic, the bulge of West Africa and the coast of Brazil prevented major exchange with the North Atlantic until the late Albian (Evans, 1978; Arthur and Natland, 1979; Reyment, 1980). Only cold-water planktonic and benthic foraminifers reached the tropical South Atlantic before that time, and these came in from the south (Premoli Silva and Boersma, 1977). The climate in the equatorial regions was harsh and arid, based on the low diversity of palynomorphs (Morgan, 1978) and the presence of evaporites in coastal basins (Evans, 1978).

During the Early to middle Cretaceous, temperature gradients were much lower than modern values, and climates in tropical to temperate latitudes were fairly cool and dry. No definitive evidence for Mesozoic glaciations has been found; thus we may infer essentially ice-free polar regions. This condition has important implications for atmospheric and oceanic circulation during the Cretaceous. Lower north-south temperature gradients would lead to more sluggish atmospheric circulation and surface current flow. The lack of polar ice caps may have prevented the freezing of surficial seawater on polar ice shelves and the initiation of thermohaline flow in high latitudes. Instead, thermohaline flow (if it existed at all) must have been driven by excess evaporation at low and middle latitudes within marginal seas that had a sufficiently deep sill to allow thermohaline outflow to the global ocean (Roth and Bowdler, 1981). Although such scenarios of haline-generated flow in middle latitudes are quite plausible, they are difficult to document in a convincing way because of the lack of reliable indices in the geologic record (micropaleontologic, geochemical, or isotopic) of hypersaline current flow.

The middle Cretaceous shales and subsequent pelagic biogenic sediments indicate the gradual ventilation of intermediate and deep water. As the South Atlantic widened, the degree of circulation in the Cape Basin kept one step ahead of that in the Angola Basin to the north. When the Cape Basin was anoxic, the Angola Basin was evaporitic (Arthur and Natland, 1979). When anoxic conditions disappeared from the Cape Basin, the Angola Basin became anoxic without accompanying evaporite deposition. Finally, during the Late Cretaceous, the Angola Basin achieved sufficient circulation to become permanently oxygenated, when deep water connections became established to the north and south (Arthur and Natland, 1979, fig. 2). Throughout this gradual ventilation process from south to north, a form of thermohaline circulation may have developed in response to the sinking of hypersaline surface waters. The limiting factor, however, was the deepening of sills to the point where the dense saline waters no longer became trapped within isolated basins but were able to exchange freely with adjacent basins.

The Cretaceous/Tertiary Boundary

The Cretaceous/Tertiary boundary was recovered at seven drill sites during recent paleoceanographic drilling legs in the South Atlantic: Hole 356 on the São Paulo Plateau (Thierstein, 1981); Hole 516F on the Rio Grande Rise (Hamilton, this volume); Site 524 on the Walvis Ridge (Hsü, 1982); and Sites 525, 527, 528, and 529 on the flank of the Walvis Ridge (Moore, Rabinowitz, et al., in press). We anticipate that this suite of cores will contribute substantially to the explanation of the faunal and floral extinctions associated with this punctuation mark in the geologic record (e.g., Thierstein, 1980). Because of the extensive observational data near the Cretaceous/Tertiary boundary on the Rio Grande Rise, we have prepared a separate synthesis chapter that incorporates all the interpretations derived from the Site 516F core material near the boundary (Hamilton, this volume). Below we shall summarize only those interpretations pertinent to the oceanographic environment near the boundary.

An extensive erosional unconformity is associated with the Cretaceous/Tertiary boundary on the Falkland Plateau (Ludwig et al., 1980). At the latitude of the Rio Grande Rise–Walvis Ridge, however, the Cretaceous/Tertiary boundary is remarkably complete in virtually all sites cored. Thus, we believe it is unlikely that the boundary event coincided with a significant global intensification in deep current flow, although such may have been true on a local scale.

Faunal, floral, and stable isotopic analyses on the São Paulo Plateau (Site 356) and Rio Grande Rise (Site 516F) demonstrate that the Cretaceous/Tertiary boundary corresponded with a dramatic shift in the properties and perhaps the circulation of the surficial oceans, yet there was relatively little change in the abyssal ocean characteristics. The Maestrichtian and Paleocene benthic foraminiferal assemblages are essentially identical at the generic level, with some minor differences at the species level (Douglas and Woodruff, 1981; Dailey, this volume). Planktonic organisms, by contrast, show dramatic extinctions and appearances over a limited stratigraphic interval (Thierstein, 1981). The stable isotope geochemistry of the fine fraction and planktonic foraminifers shows a dramatic $\delta^{13}C$ depletion of 0.5–2.0‰ at the Cretaceous/Tertiary boundary (Zachos et al., 1982; Williams et al., this volume). Yet there is essentially no change in $\delta^{13}C$ values of benthic foraminifers alone across the boundary. This pattern further supports the notion that the Cretaceous/Tertiary boundary "catastrophe" had a significant effect only upon the surficial oceans and not upon deeper waters. The catastrophic scenario invoking extraterrestrial asteroidal or cometary impact upon the earth's surface (e.g., Alvarez et al., 1980; Ganapathy, 1980; Hsü et al., 1982) is consistent with the geochemical observations showing an iridium anomaly at Hole 516F (Michel and others, this volume). The faunal, floral, and stable isotopic data provide further control on the oceanographic response to such an impact.

The general scenario in the South Atlantic for the oceanic response to the Cretaceous/Tertiary event may be summarized as follows (also, see Thierstein, 1981; Hsü et al., 1982; Williams et al., this volume):

1) The last several million years of the Cretaceous are marked by stasis of nannofossil assemblages, suggesting stable oceanic surface environments leading up to the boundary event. The South Atlantic was characterized by a relatively deep carbonate compensation depth (CCD) and a diverse nannoplankton flora with typical $\delta^{13}C$ values of +2.5‰. The late Maestrichtian $\delta^{18}O$ values, if interpreted as denoting isotopic paleotemperatures (e.g., Epstein et al., 1953), indicate a warm South Atlantic with temperature extremes ranging from 18 to 25°C (Williams et al., this volume).

2) The replacement of Latest Cretaceous nannofossil assemblages by "new" earliest Tertiary taxa typically occurs over sediment thicknesses of a few tens of centimeters. This transition can be deconvoluted and reinterpreted as an instantaneous event, using bioturbation parameters derived in studies of Holocene sediment mixing (Thierstein, 1981). Geochemical anomalies associated with the boundary (Michel et al., this volume) also require that we interpret the data in terms of realistic mixing models, instead of presuming that the preserved geochemical record is an accurate representation of the original flux of material to the sediment-water interface.

3) At Hole 516F, there appear to be significant lead-lag relationships among the various geologic parameters undergoing change. These lead-lag relationships do not appear to be artifacts of bioturbation, core disturbance, or sample spacing, and thus they may be significant in terms of unraveling the sequential responses of the earth to the boundary event. Approximately 40 cm below the boundary (based on nannofossils), the $\delta^{18}O$ record of bulk samples begins a decrease of 1.2‰; 20 cm below the boundary, the $\delta^{13}C$ record begins a depletion of 1.8‰ (Williams et al., this volume). Both of these isotopic depletions occur within a broad $CaCO_3$ maximum that spans the boundary. If Maestrichtian accumulation rates are used for Hole 516F, the onset of the oxygen and carbon isotopic shifts preceded the boundary event (determined stratigraphically) by 50,000 and 20,000 yr., respectively.

4) The oxygen isotopic shift across the boundary (−1.2‰) indicates a rapid temperature increase of 5 or 6°C, lasting approximately 100,000 yr. (Williams et al., this volume).

5) Beginning with the iridium anomaly at the Cretaceous/Tertiary boundary, the $\delta^{13}C$ record undergoes a depletion of 1.8‰ (Williams et al., this volume), and the Maestrichtian nannofossil population becomes extinct. This correspondence suggests a near-total collapse in the primary productivity of the South Atlantic and is in agreement with carbon isotopic data at other South Atlantic sites that span the boundary (Hsü et al., 1982).

6) The biostratigraphically determined Cretaceous/Tertiary boundary at Hole 516F is perhaps 4–8 cm higher in the core for planktonic foraminifers (Pujol, this volume) than for calcareous nannofossils. This apparent difference may be an artifact of sample spacing and/or bioturbation effects upon the original depositional sequence. Alternatively, it may point to a measurable lag between the phytoplankton extinctions and those of the zooplankton.

7) As the Paleocene planktonic assemblages became established, the $\delta^{13}C$ values became more positive, and by 63 Ma (Zone P2) they became stable at levels similar to those before the Cretaceous/Tertiary boundary (Williams et al., this volume).

THE PALEOGENE SOUTH ATLANTIC OCEAN

Basin Development and Hydrography

By the early Cenozoic, the South Atlantic had opened and deepened sufficiently to become an integral part of the world ocean. The paleoceanographic events of this time in the South Atlantic more closely paralleled comparable events that occurred in the Pacific (e.g., van Andel et al., 1975) and Indian oceans (e.g., Sclater et al., 1977). Global-scale processes (such as those controlling the CCD, the poleward transfer of heat, the surface circulation and productivity, and the presence of erosional unconformities at the seafloor) dominated the sedimentation pattern of the South Atlantic starting at the early Cenozoic. Thus the Cenozoic South Atlantic must be interpreted in the context of global current systems and not as localized events within an isolated set of basins. These current systems need not have followed the same pattern as those of the modern oceans (e.g., Warren, 1981); nevertheless, the tectonic barriers to deep meridional flow in the Atlantic had subsided to very near their present depths, and relatively free exchange was possible between the South and North Atlantic throughout most of the Cenozoic.

The early Paleogene was characterized by relatively mild climates, a highly stratified water column, and relatively little thermohaline flow at intermediate to abyssal depths. Temperate faunal and floral assemblages extended to high latitudes (e.g., Haq et al., 1976), where diverse calcareous and siliceous microfossil groups formed widespread biogenic deposits at high accumulation rates (>30 m/Ma) (e.g., Ludwig et al., 1980). The zonal circulation in the circumglobal Tethyan seaway moderated polar influences. On the shallow Rio Grande Rise, ostracodes are notably rare in lower Paleogene sediments, and one important genus (*Krithe*) is missing entirely, suggesting a sharply defined oxygen minimum below the main thermocline (Benson, this volume). At the same time, much more diverse ostracode faunas were present at greater depths on the Rise, indicating a highly stratified water column. Bottom water temperatures were in the range of 10–12°C, with surface water temperatures of 12–20°C (Shackleton and Boersma, 1981). A strong east-west gradient in surface temperature was present in the South Atlantic, indicating that the anticyclonic subtropical gyres and the associated poleward thermal gradients were well developed (Shackleton and Boersma, 1981).

By the early Cenozoic, deep and intermediate water exchange was possible from the Southern Ocean to the North Atlantic via sills and channels through the major topographic barriers. Many of the critical sills are on middle or Lower Cretaceous crust, and hence they had subsided to nearly their present depth by the late Paleogene (e.g., van Andel et al., 1977). The Vema Channel, for example, had subsided to very near its present sill depth by the end of the Eocene (fig. 4 of van Andel et al., 1977). Thus subsidence of the Vema Channel below a critical threshold depth was probably not the limiting factor in allowing deep thermohaline flow to enter the central and northern Atlantic. A coherent temporal and spatial pattern of erosional events occurred throughout much of the abyssal Pacific, Indian, and Atlantic oceans. Apparently there was no coherent pattern of vigorous deep water flow in the world ocean until the early Cenozoic, and when it finally developed, the northward routes through the South Atlantic were already open in a configuration not greatly different from that of the modern ocean (van Andel et al., 1977, p. 694).

The stable climatic conditions of the Paleogene were interrupted by a major thermal event in the southwestern Atlantic during the middle Eocene (approximately 47 Ma), during which there was considerable intrusive and extrusive volcanism, faulting, slumping, and sediment reworking on the upper flanks of the Rio Grande Rise (Bryan and Duncan, this volume; Barker et al., this volume). This episode was evidently of relatively short duration (no more than 2–3 Ma) and was probably confined to the shallower portions of the Rise where steep-flanked, flat-topped banks are common (30–31°S; 35–37°W). The episode produced considerable uplift around the central portion of the Rise, and "normal" crustal subsidence near Site 516 was interrupted by this major reheating event (Barker, this volume). We do not believe, however, that this event significantly altered the paleobathymetry and paleocirculation except within perhaps 100–200 km of the summit of the Rio Grande Rise. The Vema Channel and Hunter Channel remained open to intermediate and deep water exchange.

There is some evidence for lateral sediment transport associated with the middle Eocene volcanic episode, perhaps (though not necessarily) indicating the presence of intermediate and abyssal current flow. Downslope transport of reefal debris (calcareous algae, bryozoa, larger foraminifers) is evident in the middle Eocene (Zone P10) of Cores 516F-50 and 516F-63 (Tjalsma, this volume). These displaced components are within graded turbidite sequences, and thus do not require the presence of thermohaline currents. The clay minerals in the middle Eocene at Site 516 are dominated by smectite. The mineralogy has a high total clay index (TCI) and a high TCI/quartz ratio, and the smectites are closely associated with volcanogenic sediment clasts (Zimmerman, this volume). The smectite-rich clays were perhaps formed on a subaerial volcanic island or in a shallow submarine environment on a basaltic platform. These platforms were eroded near sea level to produce flat summits, and the smectite rich minerals were transported downslope to the upper flanks of the Rise (e.g., Site 516). The displaced calcareous and clay minerals from the Rise summits could have been the result of episodic gravity-driven debris flows and would not require the presence of steady thermohaline currents.

In the Brazil Basin, there is some sparse yet unavoidable evidence that deep thermohaline flow may have begun as early as the early Eocene. This evidence includes:

1) A seismic discontinuity in sediments lower than lower Eocene near Site 515 in the Brazil Basin, designated as Unconformity 1 by Gamboa and others (this volume), was interpreted as denoting bottom current activity, although a diagenetic origin cannot be ruled out because drilling at Site 515 did not reach the unconformity (approximately 1.0 s).

2) The high smectite content of lower Eocene to Recent sediments in Brazil Basin DSDP Sites 355 and 515 (Zimmerman, this volume) suggests the initiation of volcanic activity on the Rio Grande Rise in the early Eocene and the subsequent lateral dispersal of associated

clay minerals (e.g., smectite) by a northward-flowing bottom current into the Brazil Basin.

3) The presence of thin parallel laminations and cross-bedding structures in the lower Eocene pelagic carbonates at Site 515 (Barker et al., 1981; site chapter, Site 515, this volume) suggest that thermohaline flow may have begun substantially before the late Eocene. Our present assessment, however, is that the overwhelming body of evidence points to the major initiation of thermohaline flow in the latest Eocene (see next section).

Eocene/Oligocene Boundary "Event"

Before Leg 72 drilling, a considerable body of evidence pointed to the initiation of global deep thermohaline flow approximately at the Eocene/Oligocene boundary (e.g., Benson, 1975; Kennett and Shackleton, 1976). One of the primary objectives of Leg 72 was to determine the timing of the initiation of AABW flow into the Brazil Basin. We also examined whether this deep circulation event was reflected in shallower depths on the nearby Rio Grande Rise, as suggested by stable isotopic analysis of foraminifers at Site 357 (Boersma and Shackleton, 1977). We were only partially successful in achieving these objectives. In the Brazil Basin, an erosional unconformity brackets the Eocene/Oligocene boundary; middle Oligocene sediments (~30 Ma, top of Magnetic Anomaly 10) unconformably overlie lower Eocene sediments (~52 Ma, base of Magnetic Anomaly 23). The 22-Ma hiatus does not allow us to date precisely the onset of deep thermohaline flow, except between the limits of 52 and 30 Ma.

At Site 516, on the upper flanks of the Rio Grande Rise, the Eocene/Oligocene boundary occurs in an interval of strong dissolution in Cores 516F-38 and 516F-39. Tjalsma (this volume) examined benthic foraminiferal assemblages on either side of the boundary and found that neither the generic frequency distribution nor the stratigraphic ranges of species indicates any significant faunal turnover near the boundary. This uniformity may reflect the fact that the paleodepth of Site 516 at the Eocene/Oligocene boundary was relatively shallow (~800 m; see Barker, this volume), and perhaps well above depths where deep thermohaline flow became important. Ostracode evidence from Sites 356, 357, and 516 indicates that the psychrosphere progressed from deeper to shallower regions on the Rio Grande Rise. At Sites 356 and 357, the ostracode genus *Krithe* first appeared in the latest Eocene, indicating an increase in dissolved oxygen to 4.5–5.0 ml/l. At Site 516, however, the "psychrospheric" ostracodes did not appear until the late Oligocene to early Miocene (Benson and Peypouquet, this volume), thereby suggesting a gradual thickening and upward expansion of the psychrosphere after its initiation. The Leg 72 data across the Eocene/Oligocene boundary are consistent with Corliss's (1981) interpretation of the rate of benthic foraminiferal faunal turnover: the Eocene/Oligocene growth of the psychrosphere appears to have been gradual in time and space, rather than catastrophic.

Seismic profiles from the axis of the Vema Channel point to the presence of a buried paleochannel beneath and slightly to the west of the modern Vema Channel axis (Gamboa et al., this volume). This buried channel, with 50–100 m of relief, is represented as a strong reflecting horizon that can be traced northward into the Brazil Basin, where it was cored at DSDP Site 515 (Unconformity 2 of Gamboa et al., this volume; boundary between Lithologic Units 2 and 3 at Site 515). To the south of the Vema Channel, in the northern Argentine Basin, the most prominent midsection reflector is evidently an upper Oligocene diagenetic horizon, with no evidence of erosional unconformities or strong acoustic reflectors near the Eocene/Oligocene boundary (Zimmerman et al., 1979). Thus the seismic evidence is consistent with thermohaline strong flow within the north of the Vema Channel at the Eocene/Oligocene boundary. Farther upstream in the Argentine Basin, however, the flow was sufficiently broad and unconstrained topographically such that velocities were low, and, as a result, no striking unconformities mark the onset of thermohaline flow.

Even though the drilling record at Site 515 is marked by a 22-Ma unconformity spanning the initiation of deep thermohaline flow, data from elsewhere in the Atlantic may allow us to date the initiation of AABW flow more precisely. Following the arguments of Johnson (1982), we presume that the global deep thermohaline flow can be described as a set of teleconnective linkages between the major current systems. Furthermore, we assume that the generation of a southern source of bottom water around the Antarctic perimeter requires not only cold temperatures in surface waters, but the advection of high-salinity waters in the subsurface from the north (e.g., Foster and Middleton, 1980). In other words, high-salinity NADW (or its equivalent) is a necessary condition to the generation of AABW. Thus, the limiting factors that govern AABW formation may include conditions in the North Atlantic, and not simply the oceanographic parameters in the regions of AABW generation around the Antarctic continent. Recent evidence from the North Atlantic points to the inititation of deep thermohaline circulation and the gradual development of faunal and geochemical changes over a period of 3–5 Ma near the Eocene/Oligocene boundary. The sequence of events includes the following (also see Miller and Curry, 1982; Miller et al., in press; Miller and Tucholke, in press):

1) In the middle Eocene (more than >40 Ma), bottom waters in the North Atlantic were warm, old, and chemically corrosive. Benthic foraminiferal assemblages were principally agglutinant forms (e.g., Labrador Sea) or calcareous assemblages dominated by *Nuttalides truempyi* (circum-Atlantic), with depleted $\delta^{13}C$. Deep circulation was nonexistent or very sluggish.

2) In the early late Eocene (40 to 38.5 Ma), there were several first and last appearance datum levels of benthic foraminiferal taxa, and the $\delta^{13}C$ of benthic foraminifers became heavier by ~0.6‰. The major benthic faunal abundance change, i.e., the replacement of the

N. truempyi assemblage by stratigraphically long-ranging and bathymetrically wide-ranging taxa, occurred at this time.

3) Near the Eocene/Oligocene boundary, a major increase in $\delta^{18}O$ of benthic foraminifers began at 38 Ma, culminating in a rapid increase in $\delta^{18}O$ just above the boundary at about 36.5 Ma. We presume that this "permanent" $\delta^{18}O$ increase in synchronous with, and equivalent to, the isotopic event identified in numerous other regions (e.g., Savin et al., 1975; Shackleton and Kennett, 1975; Corliss, 1981).

4) A major seismic reflector, designated as R4 by Miller and Tucholke (in press), becomes widespread over the North Atlantic between the Azores and the Greenland-Iceland-Scotland Ridge. This reflector spans the Eocene/Oligocene boundary (38–36 Ma) and marks the initiation of overflow from the Norwegian Sea in response to the subsidence of the Faeroe-Shetland Sill to a depth sufficient to allow thermohaline overflow.

5) In the middle Oligocene, about 32 to 28 Ma, the benthic foraminifer *Nuttalides umbonifera* increased in abundance in the North Atlantic. *N. umbonifera* appears to be diagnostic of markedly undersaturated, corrosive bottom water (Bremer and Lohmann, 1982). Moreover, the $\delta^{13}C$ of benthic assemblages decreased by $\sim 0.5‰$ in the middle Oligocene. This evidence suggests the introduction of an older and more corrosive bottom water into the North Atlantic during the middle Oligocene.

We believe that this scenario for abyssal flow in the North Atlantic is applicable to the western South Atlantic as well. In particular, the initiation of Norwegian Sea overflow (proto-NADW) in the latest Eocene was crucial in providing high-salinity flow at mid-depths into the central and South Atlantic. This flow ultimately extended southward to the Antarctic continental margins, where it elevated the salinity to the point at which mixing with the ambient waters produced thermohaline convection (e.g., Foster and Middleton, 1980) and a return flow to the north of proto-AABW. By the middle Oligocene, this northward flow had reached the North Atlantic; these older and more corrosive waters (proto-AABW) replaced the younger proto-NADW that had initially filled the entire deep North Atlantic without competition from the south. For the Vema Channel region, we believe that the strong southward flow of proto-NADW through the abyssal Brazil Basin began near the Eocene/Oligocene boundary, and, within a short time, an underlying return flow to the north of proto-AABW was initiated. Since the mixing time of the oceans is very short (10^2–10^4 yr.), we would not expect a measurable lag between the initiation of proto-NADW and that of proto-AABW. Nevertheless, the middle Oligocene benthic and stable isotopic evidence of Miller and others (in press) points to the entry of older and more corrosive bottom waters into the northernmost Atlantic, at a time when the overflow of proto-NADW remained strong. We thus believe that the southwestern Atlantic deep circulation can be explained from evidence in the northernmost Atlantic, and that this evidence points to the presence of proto-NADW as a precondition for the formation of proto-AABW.

Oligocene Sedimentation

The Oligocene is marked by notably high sediment accumulation rates (>20 m/Ma) in the northern Argentine Basin (Site 358; see Supko, Perch-Nielsen, et al., 1977) and in the southern Brazil Basin (site chapter, Site 515, this volume). These observations can be explained in terms of a major sea-level regression identified by Vail and others (1977) in middle Oligocene sedimentary sequences from continental margins, although some authors believe the regression was of the earliest Oligocene rather than middle Oligocene (Olsson et al., 1980). This regression allowed terrigenous sediments to bypass the continental shelves, resulting in a marked increase in sediment flux to the deep oceans. The northward flow of AABW through the Argentine and Brazil Basins was well developed by the time of this regression; thus the thick Oligocene deposits reflect a high suspended sediment load that was transported and redeposited by the existing thermohaline flow. During this same period of time, extensive sediment drift deposits grew along the lower continental rises of the northern and western Atlantic, marking the coincidence of strong thermohaline flow from the Norwegian Sea and massive influxes of terrigenous sediments during the Oligocene regression (Miller et al., in press).

Toward the end of the Oligocene, the widening and deepening of the Drake Passage (e.g., Barker and Burrell, 1977) may have been sufficient to allow strong circumpolar flow to begin in the South Atlantic. A dramatic lithologic change from calcareous to siliceous deposition occurred in the southeastern Argentine Basin at this time (Ludwig et al., 1980), perhaps pointing to the establishment of strong zonal flow and sharp latitudinal frontal zones in the surface and subsurface circumpolar waters. This eastward circumpolar flow, however, need not have altered the flow of the underlying proto-NADW and proto-AABW appreciably, because circumpolar flow was presumably confined to shallow depths upon first breaking through the Drake Passage. There is no evidence from Site 515 or elsewhere in the western Atlantic that the deep thermohaline flow (i.e., below 2 km) changed markedly at the end of the Paleogene.

MIOCENE

Middle Miocene: Antarctic Ice Growth

It is generally accepted that the rapid global increase in $\delta^{18}O$ of planktonic and benthic foraminifers during the middle Miocene (e.g., Savin et al., 1975) can best be explained in terms of a major expansion of the ice sheet covering eastern Antarctica. Recent detailed work by the Cenozoic Paleoceanography (CENOP) project documents that the most rapid phase of ice growth occurred between 14.8 and 14.0 Ma (Woodruff et al., 1981). This event was accompanied by a pronounced equatorward migration of floral and faunal biogeographic provinces, steeper poleward temperature gradients, and an increased abundance of diatom species indicative of upwelling in the equatorial and northeastern Pacific (e.g., Barron, 1982). Antarctic glaciation and the associated cooling of the circum-Antarctic ocean increased the poleward temperature gradient, requiring an intensification of the atmospheric and surface circulation to maintain the global heat balance. Such a major climatic event might be reflected in the deep thermohaline circulation. Accordingly, we examined the Leg 72 cores to determine if a paleocirculation event might be observed in the southwestern Atlantic.

On the shallow Rio Grande Rise at DSDP Site 516, the middle Miocene is marked by the reappearance of a particular ostracode (*Krithe* Type D), which Benson and Peypouquet (this volume) interpret as indicative of the return of well-oxygenated waters from the Antarctic and an end to a stagnation phase in circulation at intermediate depths (1–2 km). This event is the earliest evidence for the presence of a water mass during the Neogene equivalent to Antarctic Intermediate Water (AAIW). Farther down the Rise flanks, the middle Miocene at Site 518 (3950 m) is marked by one or more erosional unconformities (Berggren et al., biostratigraphic synthesis chapter, this volume). In the Brazil Basin, the mineralogy at Site 515 shows a persistence of the smectite-dominant clay mineral assemblage throughout the Neogene, with no noticeable anomalies (Zimmerman, this volume). A lithologic boundary is present at ~180 m sub-bottom at Site 515 and may represent an unconformity between approximately 12 and 8 Ma (Berggren et al., biostratigraphic synthesis, this volume). Lithologically the boundary corresponds with an increase in the total clay index (TCI) and TCI/quartz ratio, indicating an increased supply of fine-grained sediment from terrigenous sources (Zimmerman, this volume). This could be explained by an increased glacial runoff from South America, an increased transport of AABW, or both. This lithologic boundary also corresponds with a seismic discontinuity (Unconformity IV of Gamboa et al., this volume), which might be interpreted as an erosional episode. The evidence for significant AABW intensification following the middle Miocene Antarctic ice growth is meager, suggesting that climate threshold events in the Antarctic are not necessarily limiting factors in governing AABW production. If NADW production remained persistently strong through the middle Miocene event, then we might expect AABW flow to persist as well.

In a recent paper, Schnitker (1980a) proposed that the middle Miocene glaciation in eastern Antarctica is directly attributable to the initiation of Norwegian Sea overflow. He argues that this high-salinity proto-NADW extended to the subsurface waters in the Antarctic, where it upwelled and served as a moisture source for a growing ice sheet in eastern Antarctica. Blanc and Duplessy (in press) have also argued from stable isotopic analyses that there was no overflow across the Greenland-Iceland-Scotland Ridge before the middle Miocene. Although we concur with Schnitker in recognizing the role of "proto-NADW" in contributing to the oceanographic characteristics of the subsurface waters in the Antarctic, we see no evidence to associate a middle Miocene episode of Antarctic ice growth with the initiation of Norwegian Sea overflow. A diverse array of faunal, isotopic, and geophysical evidence points to the subsidence of the Iceland-Scotland Ridge and the initiation of Norwegian Sea overflow much earlier, near the Eocene/Oligocene boundary (e.g., Miller et al., in press). If, however, the interpretation of Matthews and Poore (1980) proves to be correct regarding the initial growth of Antarctic ice during the Paleogene rather than in the middle Miocene, then perhaps Schnitker's (1980a) explanation of the role of proto-NADW in Antarctic ice growth will be very important.

After the middle Miocene, more severe climatic conditions developed in temperate latitudes (e.g., Keller, 1980). The Maurice Ewing Bank (Site 512) was marked by very low species diversity of nannofossils and planktonic foraminifers between 12 and 5 Ma (Ludwig et al., 1980), attesting to the increasingly harsh climate. In the equatorial Pacific, climatic oscillations with a periodicity of ~400,000 yr dominated the late Miocene pelagic deposition (Moore et al., 1982), perhaps pointing to the natural period of the dominant glacial/interglacial oscillations when the only extensive ice sheet was that in the southern hemisphere. Deep thermohaline circulation during the late Miocene and early Pliocene may also have experienced a 400,000-yr. cyclicity. The Leg 72 sites are insufficient for resolution at this frequency, because of erosional unconformities and limited stratigraphic control at the two deep sites that reached middle Miocene sediments (Sites 515 and 518).

Late Miocene: the Messinian Salinity Crisis

The Messinian Stage (~6.2–5.0 Ma) is marked by a series of global climatic and oceanographic events that are widely recorded in the geologic record.

1) Global sea level fell by 50–70 m (Adams et al., 1977), producing regressive depositional facies in the lithologic and seismic stratigraphic sequences along continental margins (Vail et al., 1977).

2) There was major expansion of the Antarctic ice sheets (Mercer and Sutter, 1982; Mercer, 1983), and an associated global increase in $\delta^{18}O$ of planktonic and benthic foraminifers (e.g., Shackleton and Kennett, 1975).

3) The Antarctic Circumpolar Current and Antarctic surface waters expanded northward, leading to a 300-km displacement northward of the zone of biogenic silica accumulation around Antarctica (Kennett, 1977).

4) The Mediterranean Sea was isolated from the North Atlantic, and concurrently evaporitic facies up to several km in thickness were deposited throughout the western Mediterranean (e.g., Ryan, 1976; Ryan and Cita, 1977; Berggren and Haq, 1976).

5) A global decrease in $\delta^{13}C$ of planktonic and benthic foraminifers occurred, averaging ~0.8‰ (e.g., Keigwin, 1979; Bender and Keigwin, 1979; Vincent et al., 1980). This event, dated at approximately 6.2 Ma (Haq et al., 1980; Loutit, 1981), has been interpreted to be a result of a global increase in the supply of organic carbon from terrestrial lowlands and continental shelves, which were exposed to subaerial conditions by the global sea

level regression. Alternatively, the carbon shift may reflect a change in the deep thermohaline circulation intensity.

In the South Atlantic, Messinian cores from Leg 72 sites and elsewhere yield evidence concerning the response of the deep thermohaline circulation to the sequence of global climatic events listed above. The CCD abruptly shallowed to ~4000 m in the tropical Atlantic during the Messinian, then dropped to more than 4600 m in the earliest Pliocene (Thunell, 1981). This pattern of migration of the CCD is consistent with the complete cessation of NADW production during the Messinian, producing a dissolution "spike" and a shallowing of the CCD in Atlantic Ocean sediments lying within the pathway of NADW. The reinitiation of NADW flow in the earliest Pliocene would give better carbonate preservation and a deepening of the CCD because of the reintroduction of young, well-oxygenated thermohaline flow from the north. At DSDP Site 518, within the AABW/NADW transition zone on the Rio Grande Rise, the terminal Miocene (6.0 to 5.4 Ma) is marked by heavy values of $\delta^{13}C$, minimal carbonate dissolution, and a benthic foraminiferal assemblage (*Uvigerina peregrina/Globocassidulina subglobosa*) that Hodell and Kennett (this volume) interpret as indicative of a youthful, well-oxygenated Circumpolar Water (CPW) that expanded northward to the latitude of the Rio Grande Rise. Their benthic foraminiferal assemblage data suggest that there was no production of either NADW or AABW during the Messinian. As a result, CPW was able to expand northward to regions formerly occupied by NADW or AABW. This interpretation is consistent with evidence from cores on the Maurice Ewing Bank, indicating an intensification of circumpolar flow and strong erosion on the surface of the bank during the latest Miocene (Ciesielski et al., 1982). Stable isotopic evidence from the North Atlantic also indicates a cessation of NADW flow during the Messinian (Blanc and Duplessy, in press), strongly supporting the hypothesis that the Mediterranean outflow is an important, and perhaps necessary, precondition to the formation of NADW (Reid, 1979; Johnson. 1982).

Other investigators have interpreted the global carbon shift at 6.2 Ma as an indication that NADW production *intensified* after the shift, i.e., during the Messinian (e.g., Bender and Graham, 1981, p. 458). Rather than relying upon stable isotopic data alone to identify pulses in deep circulation, we need many geologic indices of paleocirculation to resolve this discrepancy. The Messinian and the Miocene/Pliocene boundary are generally missing or incomplete at shallow sites on the Rio Grande Rise (Site 357, Supko, Perch-Nielsen, et al., 1977; site chapter, Site 516, this volume). The extent of this unconformity is such that, if it were caused by an erosional episode, we could not ascertain whether this episode was Messinian or post-Messinian. Nevertheless, the terminal Miocene represents a time of changing flow regimes over the shallow Rio Grande Rise, perhaps denoting pulses in the production and northward transport of proto-AAIW over the upper flanks of the Rise.

At the top of the Miocene and into the Pliocene, a marked erosional unconformity is present from 5.4 to 4.0 Ma at Site 518 (Hodell and Kennett, this volume). This hiatus can be explained by the reintroduction of strong thermohaline flow of both NADW and AABW, after the reinitiation of Mediterranean outflow supplied high-salinity surface water to the Norwegian Sea. Moreover, it coincided with the deepening of the CCD in the tropical Atlantic (Thunell, 1981) as oxygenated NADW was reintroduced to abyssal depths.

PLIOCENE: OCEANOGRAPHIC RESPONSE TO NORTHERN HEMISPHERE GLACIATION

During the early Pliocene (~5.0–3.2 Ma), global ice volume and climatic conditions were relatively stable, and surface water temperatures were comparable to or warmer than those of today (e.g., Poore, 1981). Beginning in the middle Pliocene, a sequence of paleogeographic and paleoclimatic events left a profound signature upon the global geologic record. In many instances these events can be dated fairly precisely (i.e., ±0.1 m.y.); it is not clear, however, which of these events are abrupt and which developed gradually, which are synchronous and which are time transgressive. Therefore, it is not yet possible to ascertain cause-effect relationships among these numerous events that are closely associated over a 1-m.y. time span of the middle Pliocene. These events include the following:

1) The Panamanian seaway closed (~4.0 Ma), producing an increasing contrast in $\delta^{18}O$ between Caribbean and eastern Pacific calcareous microfossils (e.g., Keigwin, 1982).

2) The strength of the Gulf Stream system gradually increased, beginning in the latest Miocene and culminating in the mid-Pliocene, about 3.8 Ma (Kaneps, 1979).

3) Biogeographic provinces migrated towards the equator, and surface water temperatures decreased in temperate and subpolar latitudes. This major cooling event is observed at 3.2–3.0 Ma in the North Atlantic (Poore, 1981) and Mediterranean (Thunell, 1979; Keigwin and Thunell, 1979), and at 2.9–2.8 Ma in the South Atlantic (Ludwig et al., 1980).

4) The mean value of benthic foraminiferal $\delta^{18}O$ increased by 0.5‰ in the equatorial Pacific at ~3.2 Ma, followed by an increase in the amplitude of $\delta^{18}O$ excursions at 2.5 Ma (Shackleton and Opdyke, 1977). In the northeastern Atlantic at DSDP Site 397, there is an enrichment in benthic foraminiferal $\delta^{18}O$ of 1‰ near 2.5 Ma. This enrichment is the single most striking isotopic event in the Pliocene and is evidence for the onset or advancement of northern hemisphere glaciation (Shackleton and Cita, 1979).

5) Extensive continental ice sheets appeared in Europe about 2.5 Ma (e.g., Zagwijn, 1974), with perhaps a concurrent phase of ice growth in New Zealand (e.g., Kennett et al., 1971) and South America (Mercer, in press).

6) Ice-rafted debris occurred for the first time in Pliocene sediments from the northeastern Atlantic. The estimated age of this event varies from 2.4 Ma (e.g.,

Backman, 1979; Roberts et al., 1982) to 3.0 Ma (e.g., Poore, 1981). The event is almost certainly time transgressive over the sub-Arctic North Atlantic, based on our knowledge of the multiple advances and retreats of the polar front during the latest Pleistocene (e.g., Ruddiman and McIntyre, 1977).

One of the primary objectives of Leg 72 drilling was to determine if the sequence of Pliocene climatic events listed above influenced the global thermohaline circulation. Site 517 was unfortunately terminated prematurely at ~3.1 Ma; thus we are relying primarily upon observations from Sites 516 and 518. The following observations are pertinent to the interpretation of deep flow in the Vema Channel during the middle Pliocene:

1) The $\delta^{18}O$ of benthic foraminifers steadily increased between 4.1 and 2.8 Ma at Site 516, with a relatively sharp increase (0.5‰) occurring at 3.2 Ma (Leonard et al., this volume). These intervals of isotopic enrichment may reflect global sea-level regression (e.g., Vail et al., 1977) and a corresponding growth in northern hemisphere ice, as proposed by Shackleton and Opdyke (1977).

2) The planktonic foraminiferal $\delta^{18}O$ records of the period after the start of ice growth at 3.2 Ma contrast with the benthic foraminiferal $\delta^{18}O$ records of that same time at Site 516. *Globigerinoides sacculifer* became depleted in $\delta^{18}O$ with a dampened amplitude of fluctuations. Leonard and others (this volume) suggest that this may reflect an increase in surface water temperatures in the South Atlantic during glacial expansion in the northern hemisphere, with the temperature effect and ice-volume effect upon $\delta^{18}O$ essentially cancelling each other.

3) At Site 518, in the NADW/AABW transition zone, an erosional unconformity at 46 m sub-bottom spans the Miocene/Pliocene boundary and the lowest Pliocene (~5.4 to 4.0 Ma; see Berggren et al., biostratigraphic synthesis, this volume). Directly above the unconformity, lower Pliocene sediments (4.0-3.6 Ma) exhibit the simultaneous reappearance of benthic foraminifers characteristic of both AABW (*Nuttalides umbonifera*-rich assemblages) and NADW (*Oridorsalis umbonatus*-rich and *Epistominella exigua*-rich assemblages). During the same time, the benthic foraminiferal $\delta^{13}C$ gradient increases between Sites 516 and 518, and dissolution intensity also increases. Hodell and Kennett (this volume) suggest that these faunal, isotopic, and lithologic changes reflect the displacement of circumpolar waters by the reinitiation of NADW and AABW flow. This reactivation would explain the erosional unconformity spanning the Miocene/Pliocene boundary at Site 518. The faunal, isotopic, and lithologic changes suggest that strong flow may have begun as early as the earliest Pliocene (~5.4 Ma) and continued until 3.6 Ma. The data suggest that the reestablishment of Mediterranean outflow after the Messinian salinity "crisis" may have triggered the reinitiation of strong thermohaline flow from the North Atlantic, and thence from the South Atlantic.

4) Between 3.6 and 3.2 Ma, the production of both NADW and AABW diminished, and a youthful CPW expanded northward to the latitude of the Rio Grande Rise. This interpretation is based on the reintroduction of a *Globocassidulina subglobosa–Uvigerina peregrina* benthic assemblage, the lack of a strong vertical $\delta^{13}C$ gradient between Sites 516 and 518, and a decrease in dissolution at Site 518 (Hodell and Kennett, this volume). This interval of time corresponds with an intensification in the Antarctic Circumpolar Current at ~3.4 Ma (Ciesielski et al., 1982).

5) During the middle Pliocene (~3.2-2.7 Ma) at Site 518, the AABW and NADW faunal assemblages of benthic foraminifers are reintroduced, carbonate dissolution increases, and the vertical gradient of $\delta^{13}C$ in benthic foraminifers sharply increases on the Rio Grande Rise. The steepening of the $\delta^{13}C$ vertical gradient was largely caused by a 1.0‰ depletion in $\delta^{13}C$ at Site 518. Hodell and Kennett (this volume) interpret these changes as the results of marked intensifications of both NADW and AABW, coincident with the benthic enrichments in $\delta^{18}O$ at Sites 516 and 518, and the inferred initiation of ice growth in the northern hemisphere. Their evidence is consistent with the earlier interpretations of Ledbetter and others (1978) and Williams and Ledbetter (1979), pointing to the development of an erosional pavement below ~4000 m in the Vema Channel during the middle Pliocene about 3.3 to 2.9 Ma.

According to the interpretations of Hodell and Kennett (this volume), AABW and NADW intensifications co-occurred during the Pliocene. If further study supports their interpretation, it will provide more evidence that the advection of high-salinity water from the North Atlantic (e.g., NADW) is a necessary precondition to the formation of bottom water in high southern latitudes (e.g., AABW). By this theory, increases in AABW and NADW production should occur in parallel and might be reflected in a narrowing of the transition zone and a steepening of the property gradients between the cores of the two water masses (Johnson, 1982). Conversely, the cessation of strong thermohaline flow in the Vema Channel should result in a gradual thickening (by diffusion) of the transition zone, with relatively low velocity shear and small vertical gradients in water properties. A verification of Hodell and Kennett's interpretation will require time-series analyses of cores from within the "core regions" of both NADW and AABW to determine whether fluctuations in the intensity of these two major thermohaline flows are coherent and in phase.

Some recent observations (e.g., Prell, 1982) are questioning the widely accepted notion that northern hemisphere ice growth was a middle Pliocene event; instead, the ice may have become established by the late Miocene as suggested by Mercer and Sutter (1982). This observation is a reminder that sediment analyses from a limited geographic area must be used with caution in inferring oceanographic events of global significance.

QUATERNARY: ROLE OF ORBITAL PARAMETERS IN CONTROLLING THERMOHALINE FLOW

Among the most significant advancements in paleoceanographic research is the documentation of the role of the earth's orbital parameters in the determination of the earth's energy balance and thereby the control of the periodic warming and cooling of the ocean surface during the late Quaternary (e.g., Hays et al., 1976; Imbrie and Imbrie, 1980; Ruddiman and McIntyre, 1981). These studies imply that if the surface ocean circulation has responded to orbital forcing functions in a recurring and predictable fashion, the same may be true of the deep thermohaline circulation. Tests of this notion require (1) the selection of appropriate geologic fingerprints to monitor deep flow (2) a continuous depositional record, (3) high stratigraphic precision, and (4) time-series analyses of appropriate geologic indices of thermohaline flow.

In some respects, the Quaternary cores from Leg 72 are not ideally suited for this purpose, because useful time-series analyses require intact recovery of the uppermost sediment and continuous recovery of the underlying layers. Although the use of the hydraulic piston corer (HPC) has improved core recovery substantially, the continuity of recovery between consecutive cores is always uncertain. For this reason, the Leg 72 Quaternary cores may be of greatest value in documenting first-order trends in paleocirculation within the Quaternary, and less reliable in identifying higher-frequency fluctuations (i.e., periods of 10^2–10^5 yr.). Such events may be better documented in conventional gravity and piston cores.

The following interpretations of Quaternary deep circulation are based in part on the Leg 72 cores and also upon precisely dated piston and gravity cores from the Vema Channel and other regions of the western Atlantic where deep flow leaves a mark in the geologic record:

Pliocene/Pleistocene Boundary

No particular change in abyssal flow is evident in cores spanning the Pliocene/Pleistocene boundary in the Brazil Basin (Site 515) or on the Rio Grande Rise (Sites 516–518). The high accumulation rate and presence of advected clay minerals persist through this time interval in the southwestern Brazil Basin (Site 515). At all depths shallower than 3950 m on the east flank of the Vema Channel, depositional continuity is virtually complete for the Pleistocene and upper Pliocene. At depths below 3950 m, however, erosional unconformities in middle Pliocene to upper Pliocene sediments signify strong flow of AABW (e.g., Johnson et al., 1977; Ledbetter et al., 1978; Ledbetter, 1979). Farther south, in latitudes of strong circumpolar flow, hiatuses of at least 2-Ma duration commonly occur beginning in the middle Pliocene (e.g., Ledbetter, 1981; Ledbetter and Ciesielski, 1982). Thus the onset of strong bottom water flow appears to have preceded, and not coincided with, the Pliocene/Pleistocene boundary.

Middle Pleistocene

Displaced Antarctic diatoms may serve as semiquantitative indicators of the presence of AABW flow (Burckle and Stanton, 1975), provided that the cores are well within silica-rich AABW in order that the effects of postdepositional dissolution of the diatom tests are minimized. At Site 515, four major pulses of AABW flow during the Pleistocene were identified by Shor and others (this volume); these pulses occur at 1.35–1.32; 0.78–0.70; 0.43–0.37; and 0.27–0.04 Ma. Benthic foraminiferal assemblages analyzed by Peterson and Lohmann (1982) suggest an intensification of AABW flow through the Vema Channel near 0.7 Ma, with Circumpolar Deep Water accounting for the bulk of northward flow through the Vema Channel during the Matuyama Epoch (more than 0.7 Ma).

Glacial Stages

A number of lines of evidence have pointed to a reduction or a complete cessation of NADW flow during glacial maxima. These include:

1) *Studies of benthic foraminifers.* In the Norwegian and Greenland Seas, conditions during the past 140,000 years have been favorable to bottom water production only during the Holocene and during Isotopic Stage 5e; at other times, these seas were capped either by year-round ice cover (Stages 2 and 3) or by a low-salinity surface layer (Stages 4 and 5a–d), thereby making deep convective flow less likely (Belanger, 1982). The glacial North Atlantic above 4500 m was dominated by benthic genera suggesting relatively warm temperatures ($\sim 4°C$; see Streeter, 1973) and "old", poorly oxygenated water (Streeter and Shackleton, 1979; Schnitker, 1979). Even far "downstream" in the Circumpolar Current, benthic foraminifers of glacial age are dominated by taxa that suggest the shutting off of NADW contributions to CPW (e.g. Corliss, in press). Thus, overflow from the Norwegian-Greenland Sea appears to have been substantially reduced during glacial stages.

The termination of NADW flow from the north may or may not have been sufficient to shut off the flow of AABW from the south. Schnitker (1980b) finds the presence of *Nuttalides umbonifera* and *Epistominella exigua* during glacial stages in the abyssal South Atlantic (>4000 m) and North Atlantic (>4500 m), suggesting that a modified AABW continued to enter the central and North Atlantic even during glacial stages and that the cessation of NADW was not sufficient to end the production of AABW near Antarctica. The evidence of Lohmann (1978), however, suggests that the response of the deep oceans to the ice ages has been less regular than the response of the surface oceans, and that the cycles of glacial minima and maxima (as identified by $\delta^{18}O$) are not reflected in similar variations in the benthic foraminiferal assemblages.

2) *Stable isotopic analyses.* The glacial ocean water chemistry was depleted in $\delta^{13}C$ by approximately 0.7‰ (Shackleton, 1977; Broecker, 1982). Furthermore, deep water was depleted in dissolved oxygen by an average of 120 μm/kg (Broecker, 1982). These changes in $\delta^{13}C$ and dissolved oxygen are consistent with the transfer of the global carbon and nutrients from the ocean to some other reservoir (e.g., Shackleton, 1977). Alternatively, they may reflect a major change in the global intensity of deep thermohaline convection. Kroopnick (1980) has shown that $\delta^{13}C$ closely follows the pathway of NADW in the Atlantic and may thus reflect the age of deep and abyssal waters. Certain species of benthic foraminifers,

such as *Cibicides wuellerstorfi*, show $\delta^{13}C$ values that vary systematically with apparent oxygen utilization, thereby making such species potentially valuable as monitors of thermohaline convection (Belanger et al., 1981). Recent work in the Vema Channel by Curry and Lohmann (1982) show depleted values of benthic foraminiferal $\delta^{13}C$ during glacial stages, with the steepest $\delta^{13}C$ gradient shallowing from 3.7 to 2.7 km during glacial stages. They interpret this evidence as indicating a reduced production or elimination of NADW during glacial stages.

Not all of the stable isotopic data suggest the cessation of NADW during glacial stages. Duplessy and others (1980) find an amplitude of 2.0‰ in $\delta^{18}O$ fluctuations of benthic foraminifers in the deep North Atlantic and conclude that glacial NADW continued to form and was actually 1.3°C colder than at present. Grazzini and others (this volume) suggest that the $\delta^{13}C$ in benthic foraminifers show maxima during glacial stages, suggesting the presence of local sources of NADW, the enrichment in $\delta^{13}C$ of surface waters where NADW is generated, or both. Stable isotopic data alone are perhaps not a satisfactory monitor of thermohaline flow, but in combination with other lines of evidence they may be reliable.

3) *Geochemical indices.* Boyle and Keigwin (1982) find that dissolved cadmium in NADW is inversely proportional to flow rate, and that $\delta^{13}C$ is highest in the core of NADW where flow is strongest. They find low values of Cd and high $\delta^{13}C$ in benthic foraminifers within the core of NADW during interglacial stages, indicating strong flow. During glacial stages, Cd is high and $\delta^{13}C$ is low, suggesting low flow. They suggest that doubling of the Cd content signifies halving of the NADW flow rate, all other parameters being constant, and therefore NADW production was no more than half its present value during glacial stages.

4) *Dissolution indices.* Barash and others (this volume) determined dissolution indices of planktonic foraminifers in Pleistocene sediments at DSDP Site 518, in the AABW/NADW transition zone. They find that strong dissolution corresponds with relatively cold surface water assemblages and suggest that AABW flow (and associated dissolution) was intensified during glacial stages. Because of the location of this core near the AABW/NADW transition zone, however, one could also argue that intervals of strong dissolution represented the relative cessation of all thermohaline flow, allowing poorly oxygenated, corrosive waters to fill the abyssal Atlantic.

Higher-Frequency Fluctuations

Even though the dominant period of glacial-interglacial cycles for the past 0.8 Ma is approximately 100,000 years, it is clear that the higher-frequency orbital parameters at periods of 23,000 and 40,000 yr. also control surface water paleotemperatures (e.g., Ruddiman and McIntyre, 1981) and perhaps thermohaline convection as well. Recent studies suggest that the 40,000-yr. tilt cycle may indeed be reflected in geologic indices of deep convection in the western Atlantic (e.g., Mix et al., 1982; Boyle, 1982). Further high-resolution studies should attempt to document the frequency response spectra of geologic indices within the flow paths of both NADW and AABW, and then to determine the coherence and phase between the two time-series. In this way, we can perhaps document whether or not NADW is in fact a necessary precondition to the production of AABW.

SUMMARY

The South Atlantic began to open during the Early Cretaceous (~130 Ma). The degree of interaction between the opening South Atlantic and the global circulation was controlled by the widening and deepening of north-south passages through the three transverse barriers in southern (50°S), temperate (30°S), and equatorial latitudes. Before the Albian, the South Atlantic was a Mediterranean-type sea without connections to either the North Atlantic or the Tethys across Africa. The Falkland Plateau near 50°S and the São Paulo Plateau/Walvis Ridge complex near 30°S restricted north-south circulation to very shallow depths (<1 km).

Early and middle Cretaceous climates were harsh and arid in equatorial latitudes, and cool and dry in temperate latitudes. Temperature gradients were much lower than modern values, and the polar regions were ice free. Atmospheric and surface water circulation was relatively weak. The lack of polar ice caps prevented freezing temperatures and the initiation of thermohaline flow in high latitudes. Thermohaline convection, if it existed at all, was driven by excess evaporation at low and middle latitudes within marginal seas that had a sufficiently deep sill to allow thermohaline outflow to the global ocean.

The middle Cretaceous shales and subsequent biogenic pelagic sedimentation over much of the South Atlantic indicate the gradual ventilation of intermediate and deep water. Portions of the São Paulo Plateau/Walvis Ridge complex remained at or near sea level for much of the Cretaceous; hence the Argentine and Cape basins became oxygenated somewhat earlier than the Brazil and Angola basins to the north.

The Cretaceous/Tertiary boundary (~65 Ma) corresponds with a dramatic shift in the properties and perhaps the circulation of the surficial oceans, yet there was relatively little change in the abyssal ocean characteristics. The last several million years of the Cretaceous are marked by stasis of nannofossil assemblages, suggesting stable oceanic surface environments leading up to the boundary event. Beginning with the iridium anomaly at the Cretaceous/Tertiary boundary, the $\delta^{13}C$ record undergoes a depletion of 1.8‰, and the Maestrichtian nannofossil population abruptly becomes extinct. This correspondence suggests a near total collapse in the primary productivity of the South Atlantic. As the Paleocene planktonic assemblages became established, the $\delta^{13}C$ values became more positive, and by 63 Ma (Zone P2) they reached levels similar to those of latest Maestrichtian sediments beneath the Cretaceous/Tertiary boundary.

The early Paleogene was characterized by relatively mild climates, a highly stratified water column, and rel-

atively little thermohaline flow at intermediate to abyssal depths. Bottom water temperatures were in the range of 10–12°C; surface water temperatures were between 12 and 20°C. A strong east-west surface temperature gradient indicated that the southern subtropical gyre and the associated poleward thermal gradients were becoming well developed.

The Vema Channel had subsided to very near its present depth by the Eocene, with only 100–200 m of additional subsidence between the late Eocene and the Recent. Thus it is unlikely that subsidence of the Vema Channel below a critical sill depth was the limiting factor in allowing deep thermohaline flow to enter the central and northern Atlantic.

The initiation of strong thermohaline flow in the Brazil Basin is marked by an erosional unconformity at Site 515 with middle Oligocene hemipelagic muds (~30 Ma) overlying early Eocene pelagic chalks (~52 Ma). The seismic discontinuity at this unconformity is correlative with a buried paleochannel beneath and slightly west of the modern Vema Channel axis, suggesting that strong northward flow through the Vema Channel began between 52 and 30 Ma.

The initiation of Norwegian Sea overflow (proto-NADW) during the latest Eocene was crucial in supplying high-salinity waters to intermediate and abyssal regions of the central and South Atlantic. This flow extended southward to the Antarctic margins, where it mixed with ambient, cold near-surface waters to produce thermohaline convection and a northward return flow along the bottom of proto-AABW. By the middle Oligocene, this return flow had reached the North Atlantic, replacing the younger proto-NADW with older and more corrosive waters of southern origin.

The flow of both NADW and AABW was strong during the Oligocene, as indicated by thick sediment drift deposits in the North Atlantic and high accumulation rates of reworked hemipelagic muds in the Argentine and Brazil Basins. The widening of the Drake Passage and the initiation of relatively shallow circumpolar flow in the late Paleogene did not alter the flow of NADW and AABW appreciably.

There is only sparse evidence for significant intensification in AABW flow associated with the growth of the Antarctic ice cap in the middle Miocene. This suggests that the climate in the Antarctic region may not always be a limiting factor in governing AABW production.

During the Messinian salinity crisis of the latest Miocene, the central and South Atlantic show an abrupt shallowing of the CCD to 4000 m. In the AABW/NADW transition zone, a benthic foraminiferal assemblage (*Uvigerina peregrina/Globocassidulina subglobosa*) displaces taxa indicative of NADW (*Epistominella exigua*) and AABW (*Nuttalides umbonifera*), suggesting the cessation of both NADW and AABW flow. On the other hand, the presence of heavy values of $\delta^{13}C$ in benthic foraminifers suggest well-oxygenated bottom waters in the abyssal Atlantic. If in fact NADW production was shut off due to the cessation of Mediterranean outflow, then another explanation is needed for the Messinian $\delta^{13}C$ enrichment in the abyssal Atlantic. The Messinian dissolution spike and shallow CCD argues strongly for the reduction of NADW flow; therefore, the $\delta^{13}C$ profiles may be best explained by changing global reservoirs of carbon rather than thermohaline convection.

The composition of Pliocene benthic foraminiferal assemblages deposited on the east flank of the Vema Channel suggest that intensification of AABW and NADW flow co-occurred. If this proves to be the case upon further study, it will support the concept that the advection of high-salinity water from the North Atlantic (i.e., NADW) is a necessary precondition to the formation of bottom water in high southern latitudes (i.e., AABW).

Several lines of evidence indicate that NADW production was significantly reduced or terminated entirely during glacial stages of the Pleistocene. We do not yet know if the same is true of AABW. Recent observations suggest that NADW production may be controlled largely by the 40,000-yr. tilt cycle, and not by the 100,000-yr. sea-level/ice volume/surface temperature cycle of the late Pleistocene.

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