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

Kenneth G. Miller and Brian E. Tucholke

Woods Hole Oceanographic Institution/Massachusetts
Institute of Technology Joint Program in Oceanography
Woods Hole, MA USA 02543, and Woods Hole Oceanographic
Institution, Woods Hole, MA USA 02543.

ABSTRACT

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

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

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

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

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

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

Additional evidence for changes in abyssal circulation also is found in shifts in lithology (e.g. texture and composition) and
Figure 1. Bathymetric location map of the northern North Atlantic with locations of DSDP sites and major sedimentary drifts. Contour interval 1000 m.
Figure 2. Summary of development of unconformities and current-controlled sedimentation in the North Atlantic, based on existing literature. Black bars - approximate time range of hiatuses, and open bars - current-controlled sedimentation (wider during major development of drifts); both dashed where time limits uncertain. Distribution of bio-siliceous sediments based on DSDP data. References: 1) Tucholke and Ewing (1974); 2) Sheridan et al. (1974); 3) Ewing and Hollister (1972); 4) Tucholke and Laine (1981); 5) Eglöf and Johnson (1975); 6) Ruddiman (1972); 7) Roberts (1975); 8) Jones et al. (1970); 9) Montadert and Roberts (1979). Tectonic events from Ruddiman (1972) and Talwani and Eldholm (1977).
benthic fauna (taxonomic and isotopic composition). These effects in themselves rarely are diagnostic of the nature of the circulation change, but they can provide important supporting evidence. For example, Berggren and Hollister (1974) suggested that increased early to middle Eocene deposition of biosiliceous sediments in the North Atlantic (Figure 2) correlates with the first formation, sinking, and upwelling of cold, deep waters in the northern North Atlantic. This correlation is reasonable, but it is possible that the biosiliceous sedimentation was stimulated by other mechanisms. Therefore, the lithologic change by itself is not definitive.

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

PREVIOUS WORK

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

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

Stratigraphic relations show that by the early Miocene, current-controlled deposition was well established in the Rockall Trough (Peni Drift; Roberts, 1975) and in the western North Atlantic (Blake Ridge; Ewing and Hollister, 1972; Tucholke and Mountain, 1979; Figure 2). By the middle to late Miocene, coherently deposited sediments also accumulated at significantly increased rates on such sedimentary drifts as the Greater Antilles, Bahamas, Matteres, and Gardar Ridges (Figure 2). Based on this observation and on the widespread distribution of a middle Miocene hiatus, Shor and Poore (1979) suggested that the deep circulation of the northeast Atlantic did not attain its present configuration until the early to middle Miocene; they related this to an inferred late Oligocene to early Miocene marine connection across the Iceland-Faeroe Ridge. Vogt (1972) also suggested a significant deepening of this marine connection in the middle to late Miocene (ca. 10-15 Ma) based upon a subsidence model; he related the presumed increase in Norwegian Sea overflow to a middle Miocene change in abyssal circulation noted by Ruddiman (1972) on the Gardar Ridge. Although there is some coherence in the sequence of events outlined above, the existing studies still show a general lack of agreement not only on the timing of initial development of the abyssal circulation, but also on the timing and influence of subsidence of the Greenland-Scotland Ridge.

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

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

NORTHERN NORTH ATLANTIC OBSERVATIONAL DATA

Lithofacies/Age Distribution – DSDP Boreholes

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

At Site 117, drilled along the eastern edge of the Hatton–Rockall Basin, Berggren and Aubert (1976) revised the original biostratigraphic zonation by removing core 1 from the early Miocene and placing it in the late Oligocene (Figure 4). This greatly improves correlations between Sites 116 and 117 as suggested by the seismic stratigraphy (Figure 7). We also suggest a revision to the original zonation at Site 116, drilled 40 km west of Site 117. An intra-Oligocene hiatus was inferred between cores 22 and 23 on the basis of the assignment of core 22 to the Sphenolithus ciperoensis (NP25) Zone and cores 23 to 25 to the Erionota obtusa (NP21) Zone (Perch-Nielsen, 1972). Bukry (1972), on the other hand, placed core 25 in the Helicophea reticulata (NP22) Zone. Roberts (1975) noted this discrepancy and used Bukry's assignment to suggest that sedimentation was slow but continuous in the Oligocene section. Examination of the planktonic foraminiferal data supports the contention of continuous sedimentation. In addition, Berggren (1972) noted the last appearances of Globigerina ampliapertura and Chiloquembelina in cores 24 and 23, respectively, but he assumed that Perch-Nielsen's nanoplankton assignment was correct and therefore thought that the ranges of these taxa were truncated by an intra-Oligocene unconformity. However, if Bukry's assignment is used, then these last appearances may be taken as the G. ampliapertura (top of Zone P20) and Chiloquembelina (lower Zone P21) datums. Thus, cores 19 to 22 may represent the late Oligocene, cores 23 and 24 the middle Oligocene, and core 25 the early Oligocene. The average Oligocene sedimentation rate, if interpreted in this fashion, is continuous at 9.5 m/Ma (Figure 4).
Figure 3. Age versus lithofacies for Tertiary sediments recovered at DSDP sites in the Bay of Biscay. Site locations in Figure 1. Lithologic symbols follow standard "Initial Reports" format. To left of columns are subbottom depths in meters (T.D.=total depth); to right are indicated core number and fossil group on which age is based (N=nanoplankton, F=foraminifers, R= radiolaria, P=palynomorphs, BF=benthic foraminifers, D=diatoms, X/S =extrapolated by sedimentation rate). Heavy wavy lines show hiatuses, dashed lines show inferred continuous sedimentation.
Figure 4. Age versus lithofacies for DSDP sites in Rockall region with correlation of major reflectors. Explanation in Figure 3. Numbers in quotes at left of columns are reflector numbers used in individual site reports by Montadert and Roberts (1979).
Figure 5. Age versus lithofacies for DSDP sites on Orphan Knoll and Reykjanes Ridge in western Northern Atlantic and on Iceland-Faeroe Ridge. Explanation in Figure 3.
Figure 6. Summary of lithology and foraminiferal data at DSDP Site 112 in Glaia Drift, Labrador Sea. Isotopic values from Miller and Curry (in prep.). Explanation in text.
Figure 7. Seismic reflection profile and interpretation across DSDP Sites 116 and 117, Hatton-Rockall Basin.

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

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

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

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

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

Reflector R4 is a continuous, interbasinal horizon which can be traced throughout the Rockall region (Roberts, 1975), the Iceland Basin (Ruddiman, 1972), and the Labrador Sea (Egloff and Johnson, 1975). The change in sedimentary regime associated with this reflector therefore must have a regional cause. Like Horizon A1 in the western North Atlantic, reflector R4 may truncate deeper horizons (Roberts, 1975), but the exact relation of reflector R4 to the Horizon-A complex in the western North Atlantic has not been established. The lack of a significant horizon of similar stratigraphic position in the South Atlantic indicates that the paleoceanographic event associated with reflector R4 and Horizon A1 had a North Atlantic genesis. Jones et al. (1970) determined that reflector R4 is pre-late Oligocene in age, and more recent reflector/borehole correlations indicate that it dates to the latest Eocene to early Oligocene (Figures 4, 6; Laughton and Berggren, 1972; Roberts, 1975; Montadert and Roberts, 1979; Miller et al., in press).

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

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

We have traced both R4 and the overlying reflectors and tied them to boreholes using seismic reflection data of Lamont-Doherty Geological Observatory and Woods Hole Oceanographic Institution (cruises V28-04, V27-06, C 09-13, and KN51), Deep Sea Drilling Project Legs 12 and 48, and previously published profiles and picks (Roberts, 1975; Montadert and Roberts, 1979). A convenient method for initial determination of subbottom depth of reflectors is to use velocity-regression equations derived from wide-angle sonobuoy measurements. These equations are of the form \( V = V_0 + KT \), where \( V_0 \) is the initial velocity, \( K \) is the acceleration term, and \( T \) is half travel-time in seconds. This can be integrated to determine depth, \( H = V_0 T + KT^2/2 \). We used constants derived by Houtz (1980, and unpublished) for the Rockall area, and then modified the depth determination if correlation of the reflector to a nearby lithologic break could be improved, or if auxiliary velocity/density data from the sediments or borehole were available and justified a modification. In either case we restricted the resulting interval velocities to geologically reasonable values. In order to clarify some previously published correlations and to document our new correlations, distribution of reflectors at several drill sites is discussed below.

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

At Site 117 on the margin of Hatton-Rockall Basin, reflector R4 lies at 0.17 seconds and reflector R3 at 0.09 seconds subbottom. Reflectors R1 and R2 pinch out between Sites 117 and 116 (Figures
4, 7). Using the velocity-regression equation with $V_0=1.61$ and $K=2.28$, reflector R4 lies at 147m, which correlates with the unconformity between the early Eocene and late Oligocene. Reflectors R3 is close to 75m subbottom, correlating with an uncored interval between cores 1 and 2 (late Oligocene cherty limestones and oozes).

At Site 116, farther from the margin of the basin, four major sedimentary reflectors are present (Figs. 4, 7). Here, reflector R4 was first correlated with increased lithification (oozes to chalks) at 700m subbottom, corresponding to a late Oligocene to early Miocene age (Laughton and Berggren, 1972). Roberts (1975) recorrelated this horizon to the boundary between upper Eocene ooze/chalks and lower Oligocene cherts. At Site 116 reflector R4 lies at 0.79 to 0.81 seconds subbottom, reflector R3 at 0.70 seconds, reflector R2 at 0.60 seconds, and reflector R1 at 0.31 seconds. Although the constants in Houts' velocity-regression equation yield good correlations of reflectors to lithostratigraphic breaks in the thin sedimentary section at Site 117, this is not the case for the thicker section at Site 116. The high acceleration term (2.28 km/s²) for the Rockall region results in the placement of horizons below 0.3 to 0.4 seconds at anomalously great depths. We adjusted the acceleration term downward to 2.09 km/s² based on the sonic logs of the very similar section at Site 406 (south flank Rockall Plateau); this is within the standard error of estimate in the measurement of the acceleration term. As a result, depths computed using either acceleration term are very similar above 0.3 seconds (less than 8m difference) but diverge significantly at greater reflection times (71m difference at 0.8 seconds). The computed depth of reflector R4 then is 800–825m (late Eocene to early Oligocene) in agreement with Roberts' (1975) interpretation. Reflectors R3 falls at 692m and correlates with the latest Oligocene upward change from cherty chalks to oozes, reflector R2 at 577m matches the top of early Miocene siliceous sediments, and reflector R1 at 284m falls within calcareous oozes in the lower upper Miocene section. This chronostratigraphic placement of horizons at Site 116 correlates well with their placement at Site 406 (Figure 4) where depths were estimated using synthetic seismograms computed from sonic logs (Montadert and Roberts, 1979). It also agrees with the late early Miocene age of reflector R2 (or IR) suggested by Ruddiman (1972).

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

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

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

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

Reflector R4 at Site 403 can be traced to a position 0.23 to 0.25 seconds subbottom (199-218m) at Site 404. This also correlates with the unconformity separating Miocene from Eocene strata as suggested by Montadert and Roberts (1979).
The foregoing correlations suggest that we can characterize each of the major reflectors by rather specific lithologic and age limits. Reflector R4 in the Rockall area shows a strong correlation to the widely distributed unconformity straddling the Eocene-Oligocene boundary (Figures 3-6; see also Roberts et al., 1979), except possibly at Site 406 where it may underlie the unconformity by up to 45m. Away from basin margins the unconformity is not present (e.g. Site 116; Roberts, 1975), and in those regions reflector R4 appears to correlate with the uppermost Eocene to lower Oligocene section. Similarly, the apparently correlative reflector in the Labrador Basin, which we also term R4, dates to the lower part of the lower Oligocene in an apparently continuously deposited hemipelagic clay sequence at Site 112. A reflector similar to R4 in stratigraphic position has been noted in the Bay of Biscay, where it is thought to correlate with the middle Eocene to early Oligocene unconformity (Montadert and Roberts, 1979); however, reflector R4 has not been traced seismically from the Rockall region into the Bay of Biscay.

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

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

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

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

Summary of Evidence for Abyssal Circulation

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

The late Eocene to early Oligocene timing for development of the abyssal circulation is supported indirectly by a major faunal change at the end of the Eocene in the deep Labrador Sea (DSDP Site 112) where predominantly agglutinated Eocene benthic foraminiferal assemblages were replaced by an Oligocene calcareous fauna (Figure 6; Miller et al., in press). It is possible that reflector R4 at Site 112 postdates the faunal exit by up to 2 Ma, but Miller et al. suggest that the changes in fauna and depositional conditions both are expressions of changes in hydrographic regime and development of vigorous abyssal circulation. Miller and Curry (in prep.; Figure 6) also note that a 4°/oo shift in oxygen isotopic composition of benthic foraminifera coincides with the faunal change. Some of this isotopic signal may be due to diagenesis and perhaps to ice volume, but part of the shift probably represents a temperature drop which in turn can be attributed to the initiation of northern sources of bottom water. Similar supporting evidence is found in the Bay of Biscay (Site 400A) where a change of benthic
assemblages also occurs between the middle Eocene and the early Oligocene (Schnitker, 1979; personal observation). An associated shift in oxygen isotopes also occurs in this region (Site 400A, Vergnaud-Grazzini et al., 1978, 1979; Site 119, Miller and Curry, in prep.); at Site 119 this shift represents a temperature drop of at least 20°C but less than 60°C.

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

POTENTIAL BOTTOM-WATER SOURCES

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

Greenland–Scotland Ridge Overflow

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

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

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

There also is a great deal of uncertainty about the Paleogene configuration and depth of the ridge between Iceland and Greenland (Figures 10, 11). On the edge of the Greenland margin just northeast of the ridge, prograded sediments at least 1 to 2 seconds thick occur. The east Greenland margin is still not well surveyed by seismic data, but recent work by H.C. Larsen, B. Larson (both this volume), and Hinz and Schlüter (1978) indicates that thick prograding sediments have caused both narrowing and shoaling of the Denmark Straits. Accurate age assignments for these sediments and "backstripping" in conjunction with a subsidence model will be necessary to fully evaluate the possibility of a Paleogene marine passage in this area, but it presently seems possible that a significant passage did exist. The area also has an uncertain plate tectonic history. Talwani and Eldholm (1977) suggested that at anomaly 7 time (27 Ma) a westward jump of the spreading center occurred on the Greenland-Scotland Ridge, from a central position on the Ridge to a position near the Greenland margin. This particular jump disagrees with magnetic-anomaly identifications on the Ridge by Vogt et al. (1980), but a jump just north of the Ridge still is required. The depth and morphology of any existing passage adjacent to Greenland therefore may have been substantially changed in the late Oligocene by this spreading-center jump and by the subsequent formation of the Iceland Plateau.
Figure 8. Summary of evidence for marine connection between North Atlantic and Norwegian-Greenland Sea, with reference to Greenland-Scotland Ridge.
Figure 10. Bathymetry of Greenland-Scotland Ridge, in meters (from Uchupi and Hays, unpublished data). Arrows indicate modern overflow routes. Drill sites shown by triangles. Line A-D shows location of cross-section in Figure 11.
All the above evidence for a Paleogene seaway between the North Atlantic and the Norwegian-Greenland Sea is in marked contrast with interpretation of DSDP Leg 38 drilling results (Talwani and Udintsev, 1976) which suggested isolation of the Norwegian-Greenland Sea from the North Atlantic during the late Oligocene (Sites 336, 352; Figure 10). This separation was invoked because of differences in the upper Oligocene benthic foraminiferal assemblages north and south of the Greenland-Scotland Ridge (van Hinte: in Talwani and Udintsev, 1976; Figure 8); however, this difference may be due simply to different paleobathymetry at the two sites (Berggren and Schnitker, this volume). The appearance of the radiolarian species Velicucullus goodgurneri (late Oligocene) and Cyrtocapsella tetrapera (early Miocene) in the Norwegian-Greenland Sea (Schrader et al., 1976) indicate a definite connection existed in the early Miocene. Because there also are faunal preservational problems in the Norwegian-Greenland Sea, the suggested late Oligocene isolation from the North Atlantic may be more apparent than real, or if real, it may have been a transient event.

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

**Baffin Bay - Davis Straits**

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

**Climatic Controls**

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

In the modern ocean Norwegian Sea Overflow Water and Labrador Sea Water (which form the bottom and intermediate waters, respectively, in the North Atlantic) form by winter convection of high-salinity subtropical water that has been advected northward and cooled (Worthington, 1976; McCartney and Talley, in press). The source of this low latitude water is thought to be either the Gulf Stream/North Atlantic Current (Worthington, 1976) or Mediterranean Outflow Water (Reid, 1979). A similar situation may have developed in the late Eocene. At that time, the configuration of continents
### TABLE 1. GLOBAL EVIDENCE OF MIDDLE EOCENE TO EARLY Oligocene CLIMATIC COOLING

<table>
<thead>
<tr>
<th>Author</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>Savin et al. (1975)</td>
<td>Equatorial Pacific, DSDP</td>
</tr>
<tr>
<td></td>
<td>Sites 44 (1428m, present depth of site), 167</td>
</tr>
<tr>
<td></td>
<td>(3166m), 171 (2283m).</td>
</tr>
<tr>
<td>Shackleton &amp; Kennett (1975)</td>
<td>Southern Ocean, DSDP Site 277 (1214m).</td>
</tr>
<tr>
<td>Kennett &amp; Shackleton (1976)</td>
<td>Mid-Latitude South Atlantic, DSDP Site 357 (2086m).</td>
</tr>
<tr>
<td>Buchardt (1978)</td>
<td>Eastern North Atlantic, DSDP Site 398 (3910m), 400A</td>
</tr>
<tr>
<td></td>
<td>401 (2495m).</td>
</tr>
</tbody>
</table>

**Distribution of Calcareous Nannoplankton**

- Edwards & Perch-Nielsen (1975) - Southern Ocean
- Haq et al. (1977; 1979) - Atlantic

**Distribution of Planktonic Foraminifera**

- Jenkins (1968, 1975) - New Zealand
- Hornibrook (1971) - Antarctic Ocean
- Kennett (1977, 1978) - North Atlantic
- Berggren (1978) - Labrador Sea
- Gradstein & Srivastava (1980) - Labrador Sea

### Paleobotanical Evidence

**Northern Hemisphere:**

- Dorf (1964) - Western United States
- Wolfe & Hopkins (1967) - Japan
- Wolfe (1978)
- Tanai & Huzioka (1967)

**Southern Hemisphere:**

- Kemp (1978) - Antarctic Ocean, Antarctica, and Australia
- Kemp & Barrett (1975)
was similar to that of today (Figure 12A), although the basins were slightly smaller and the continental latitudes were 5–10° lower (Scneider et al., 1977). Northward advection of warm, presumably high-salinity water is thought to have occurred in an anticyclonic subtropical gyre (proto-North Atlantic Current). Another possible low-latitude source of warm, high-salinity water was the proto-Mediterranean (Tethys) Sea which lay at latitudes 15° to 30°N in the Eocene/Oligocene (Berggren and Hollister, 1974); such latitudes are the present sites of formation of high-salinity surface and intermediate water in the North Atlantic (Worthington, 1976). Cooling of these high-salinity waters in the northern North Atlantic and/or Norwegian-Greenland Sea presumably occurred during the Eocene/Oligocene phase of climatic cooling and increased latitudinal thermal gradients; based upon the modern analog, the likely sites of most intense cooling (with possible convection and bottom-water formation) would have been in regions where polar air was delivered from the large land masses, i.e. east of North America and Greenland. Thus, bottom-water formation could have occurred not only in the Norwegian-Greenland Sea, but also south of the Greenland—Scotland Ridge and in the Labrador Sea. Unfortunately it is not presently possible to determine whether such northerly advection and cooling generated significant amounts of deep and bottom water in the northern Atlantic. It seems unlikely that the shift in climatic pattern alone was responsible for the dramatic development of abyssal circulation at the Eocene/Oligocene boundary. However, it may have accentuated bottom-water formation, especially in the Labrador—Baffin and Norwegian—Greenland Seas.

MODEL OF CIRCULATION

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

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

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

Bottom water from the eastern basin probably entered the western basins through major fracture zones in the mid-ocean ridge (e.g. Charlie Gibbs F.Z.) as well as through numerous smaller fracture-zone conduits in the Reykjanes Ridge (Ruddiman, 1972; Figure 12A). Magnetic anomaly patterns suggest that these fracture-zone conduits did not develop until the late Eocene to early Oligocene (Figure 2; Ruddiman, 1972). The abyssal circulation west of the mid-ocean ridge may also have been augmented by overflow across the Greenland–Scotland Ridge adjacent to Greenland and possibly by bottom water derived from the area of Baffin Bay/Nares
Figure 12A. Model of abyssal circulation in the North Atlantic in the early Oligocene (Anomaly 13 time). Solid and open bold arrows - probable and possible bottom-water sources; solid arrows - inferred abyssal circulation (length and spacing approximately proportional to intensity of circulation). Active spreading centers indicated by open rectangles (i.e. with fracture-zone conduits) or continuous parallel lines (no conduits). Major fracture zones are solid lines. Extinct spreading axes are long dashed lines, and approximate ocean-continent boundaries are short dashed lines. Present 2000m bathymetric contour shown for reference. Tectonic data and plate geometries from Ruddiman (1972), Talwani and Eldholm (1977), Srivastava (1978), and Vogt et al. (1980).

Strait (Figure 12A). We suggest that the combined flow from these sources formed the intense abyssal boundary current that eroded the Horizon A₄ unconformity in the western North Atlantic, probably in the early to middle Oligocene.

By the late Oligocene (reflector R3), subsidence of the Faeroe Bank Channel permitted increased flow from the Norwegian-Greenland
Sea (Figure 12B) into the Rockall region, probably accounting for continued development of unconformities and chaotic current-controlled deposition. However, abyssal-current influence on sedimentation in the Bay of Biscay apparently declined as the Azores-Biscay Rise and Azores-Gibraltar Ridge subsided and failed to divert currents into the Bay of Biscay.

Keykianes Ridge fracture zones continued to provide conduits for bottom water to enter the western basins, but north of the Greenland-Scotland Ridge a spreading-ridge jump to the Greenland margin may have blocked any significant flow from the Norwegian-Greenland Sea directly into the northwestern basins (Figure 12B). The implied decrease in abyssal boundary flow in the western North Atlantic correlates with current-controlled deposition above the Horizon A\textsuperscript{U} unconformity in the late Oligocene to early Miocene (Figure 2).
During the early to middle Miocene the fracture zone conduits across the Reykjaness Ridge ceased to be developed, and, in agreement with Ruddiman (1972) and Shor and Poore (1979), we correlate this trend toward east-west basin isolation with the coherent development of sediment drifts and waves in the northern and western North Atlantic (Figures 2, 12C). However, we interpret this coherent development to be a result of a general decrease and stabilization of the abyssal circulation. This interpretation is based on the fact that the preceding sedimentary record, characterized by chaotic sedimentation and prominent unconformities, must have required much higher current velocities.

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

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

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

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

CONCLUSIONS

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

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

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

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

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

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

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STRUCTURE AND DEVELOPMENT OF THE GREENLAND–SCOTLAND RIDGE

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Edited by

Martin H. P. Bott
University of Durham
Durham, United Kingdom

Svend Saxov
Aarhus University
Aarhus, Denmark

Manik Talwani
Lamont-Doherty Geological Observatory of
Columbia University
Palisades, New York

and

Jörn Thiede
University of Oslo
Oslo, Norway

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