

32. BENTHIC FORAMINIFERAL CARBON ISOTOPIC RECORDS AND THE DEVELOPMENT OF ABYSSAL CIRCULATION IN THE EASTERN NORTH ATLANTIC¹

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ABSTRACT

The North Atlantic at present is ventilated by overflow of the Denmark Strait, Iceland-Faeroe Ridge, Faeroe Bank Channel, and Wyville-Thompson Ridge. The evolution of Cenozoic abyssal circulation of this region was related to tectonic opening and subsidence of these sills. We used $\delta^{13}\text{C}$ records of the benthic foraminifer *Cibicides* to decipher the timing of tectonically controlled changes in bottom-water circulation in the eastern basins (Biscay and Iberian) of the northern North Atlantic. Records from Site 608 (Kings Trough, northeastern North Atlantic) show that from about 24 to 15 Ma (early to early middle Miocene), $\delta^{13}\text{C}$ values in the Kings Trough area were depleted relative to western North Atlantic values and were more similar to Pacific $\delta^{13}\text{C}$ values. This reflects less ventilation of the Kings Trough region as compared to the well-oxygenated western North Atlantic. Comparison of Oligocene $\delta^{13}\text{C}$ records from Site 119 (Bay of Biscay) with western North Atlantic records suggests that the eastern basin was also relatively isolated prior to 24 Ma. At about 15 Ma, $\delta^{13}\text{C}$ values at Site 608 attained values similar to the western North Atlantic, indicating increased eastern basin ventilation in the middle Miocene. This increased advection into the eastern basin predated a major $\delta^{18}\text{O}$ increase which occurred at about 14.6 Ma. Subsidence estimates of the Greenland-Scotland Ridge indicate that the deepening of the Iceland-Faeroe Ridge was coincident with the marked change in eastern basin deep-water ventilation.

INTRODUCTION AND PREVIOUS WORK

In the Quaternary, high-frequency (10^4 – 10^5 yr.) abyssal circulation changes were climatically controlled (e.g., Curry and Lohmann, 1983, 1985). Cenozoic abyssal circulation changes on the 10^6 - to 10^7 -yr. scale, however, may be related either to long-term climatic or to tectonically controlled changes (e.g., Miller and Tucholke, 1983). Reconstructions of tectonic passageways allow evaluation of the causes of abyssal circulation changes. The history of basin development in the Cenozoic North Atlantic is a history of progressive tectonic enlargement and opening of passages that apparently resulted in increased bottom-water circulation (Berggren and Hollister, 1972; Miller and Tucholke, 1983). These tectonic changes had a profound effect upon global abyssal circulation and ocean chemistry, including loci of deposition of organic carbon, silica, and carbonate (Berger, 1970).

The history of Cenozoic bottom-water formation in the North Atlantic and its marginal seas has been controversial (cf., Schnitker, 1979, 1980a, b, with Miller and Tucholke, 1983). Today, North Atlantic Deep Water (NADW) is formed by a mixture of Norwegian-Greenland Sea Overflow Water, Labrador Sea Water, and entrained North Atlantic Water (Worthington, 1976; Broecker and Peng, 1983). Based upon evidence of changes in composition of benthic foraminiferal faunas in the eastern North Atlantic, Schnitker (1979) suggested that the first analogue of NADW formed in the middle Mio-

cene, subsequent to the subsidence of the Iceland-Faeroe Ridge. Miller and Tucholke (1983) presented seismic stratigraphic evidence from the eastern and western North Atlantic, which suggested that an analogue to NADW first formed near the end of the Eocene, and this has been substantiated for the western North Atlantic by carbon isotopic studies (Miller and Fairbanks, 1983, 1985). They suggested that Oligocene-Miocene bottom waters formed in the Arctic/Norwegian-Greenland Sea and flowed over sills in the proto-Denmark Straits and Faeroe Bank Channel, although they acknowledged that the Labrador Sea and northern North Atlantic were also possible sources (Figs. 1, 2) (Miller and Tucholke, 1983). We used carbon isotopes to evaluate possible sources of bottom-water supply to the eastern basins of the northern North Atlantic, and specifically to evaluate whether the deep eastern basins were isolated from the western basins.

Carbon isotopic comparisons provide strong evidence for the nature and timing of changes in North Atlantic bottom waters. Although changes in lithology and benthic foraminiferal fauna provide supportive evidence for abyssal circulation changes, they are rarely diagnostic by themselves. Benthic foraminiferal $\delta^{13}\text{C}$ analyses have proven to be useful in reconstructing the history of Quaternary abyssal circulation changes (Curry and Lohmann, 1982, 1983, 1985; Boyle and Keigwin, 1982; Shackleton et al., 1983; Mix and Fairbanks, in press; Fairbanks and Mix, in press). Curry and Lohmann (1985) demonstrated that the deep eastern equatorial Atlantic basin was isolated from the western Atlantic during Pleistocene ice ages, but was well mixed during interglacial periods. We have previously applied such tactics to Quaternary $\delta^{13}\text{C}$ records to decipher 10^6 - to 10^7 - yr. abyssal circulation changes for the western North Atlantic during the Oli-

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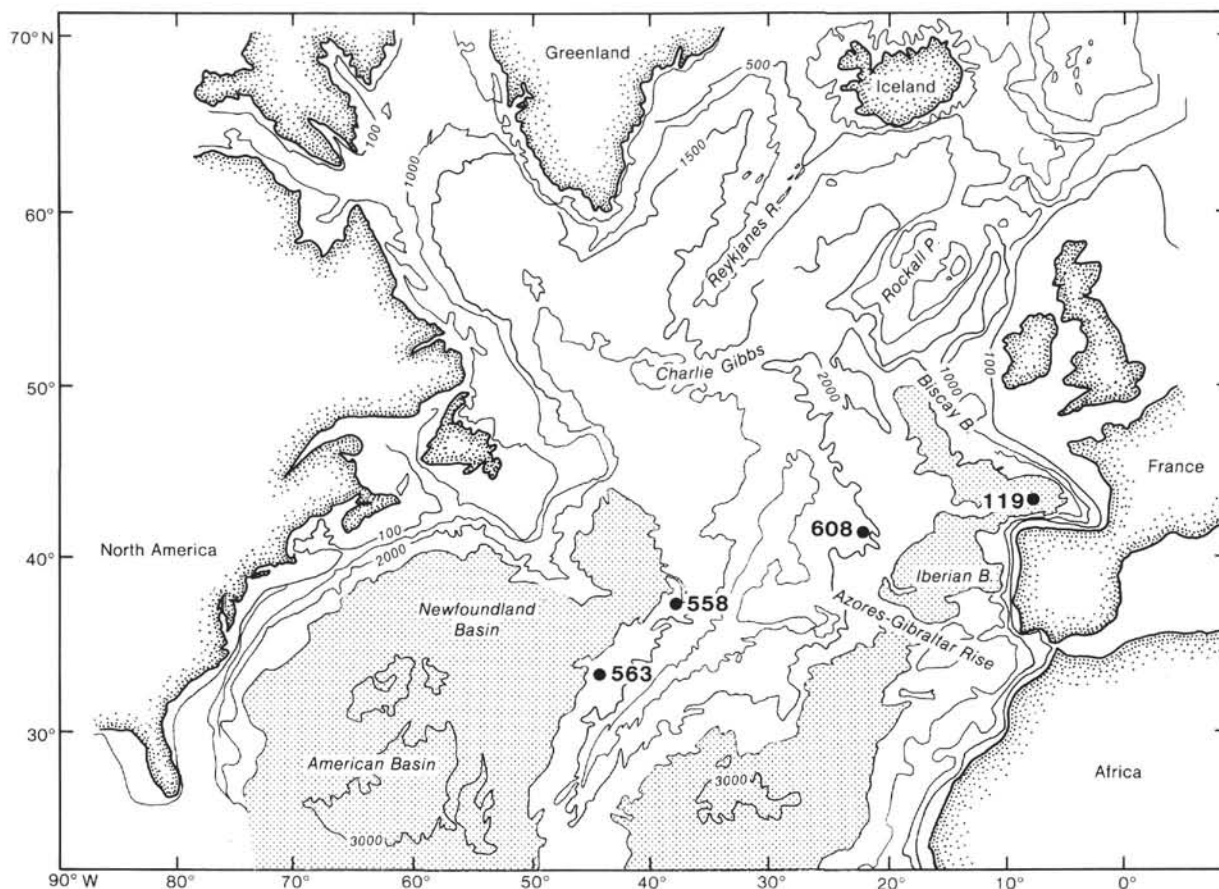


Figure 1. Bathymetric location map of the northern North Atlantic in fathoms. Modern basins deeper than 2500 fm are stippled; 1500-fm contour (heavy line) included to outline northern basins (after U.S. Naval Oceanographic Office, 1964).

gocene-Miocene (Miller and Fairbanks, 1983, 1985; Miller and Thomas, 1985).

Various investigators have delineated aperiodic Oligocene-Miocene global cycles in $\delta^{13}\text{C}$ of benthic and planktonic foraminifers, with peak values at ca. 36, 25, 16, and 9 Ma (Miller and Fairbanks, 1983, 1985; Shackleton et al., 1984; Miller and Thomas, 1985; Woodruff and Savin, 1985; Vincent and Berger, 1985; Vincent et al., 1985). These global Cenozoic cycles were related to changes in the input or output ratio of organic carbon to carbonate carbon (Miller and Fairbanks, 1985).

Differences between the global isotopic signal and regional signals provide evidence of regional abyssal circulation changes (Curry and Lohmann, 1982; Shackleton et al., 1983; Mix and Fairbanks, in press; Fairbanks and Mix, in press). Because Pacific deep waters compose the bulk of the oceanic reservoir (Baumgartner and Reichel, 1975), Pacific deep-water $\delta^{13}\text{C}$ records can be used to approximate the global $\delta^{13}\text{C}$ signal. Synoptic time series offsets between Atlantic $\delta^{13}\text{C}$ records and Pacific records can be compared with the modern ocean offsets: present-day Atlantic bottom waters are higher in $\delta^{13}\text{C}$ than those of the Pacific, reflecting production of nutrient-depleted and ^{13}C -enriched NADW (Kroopnick et al., 1972; Kroopnick, 1974, 1980, 1985). A similar carbon isotopic difference was noted to exist between the western North Atlantic (American-Newfoundland ba-

sins) and the Pacific in the early Oligocene and Miocene, suggesting the supply of "young," nutrient-depleted and ^{13}C -enriched bottom water in a mode similar to that of today (Miller and Fairbanks, 1983, 1985; Miller and Thomas, 1985).

We attempted to isolate Miocene abyssal circulation changes by making interbasinal (eastern North Atlantic versus Pacific and versus western North Atlantic) comparisons of $\delta^{13}\text{C}$ records (Table 1). Our comparisons suggest that the eastern (Biscay-Iberian) and western (American-Newfoundland) basins of the northern North Atlantic had different circulation histories until the middle Miocene, but that bottom-water changes have been similar in these basins since about 15 Ma.

METHODS

All isotopic data considered here were performed on mixed species of the benthic foraminiferal taxon *Cibicides* (Table 2). This taxon secretes calcite tests that are offset from $\delta^{18}\text{O}$ equilibrium by 0.65‰ and that reflect the distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the modern ocean (Shackleton and Opdyke, 1973; Duplessy et al., 1980; Woodruff et al., 1980; Belanger et al., 1981; Graham et al., 1981). Sample preparation and isotopic procedures are given by Miller and Thomas (1985). Ages are reported using the Berggren et al. (in press a, b) time scale by interpolating between magnetostratigraphic (Site 608: Clement and Robinson, this volume; Site 563: Miller, Aubry, et al., 1985) and biostratigraphic (Sites 77, 119, 289, and 563) datum levels (Table 3; Figs. 4-6).

Our sampling interval at Site 608 averages 0.4 m.y. We believe that our record resolves changes on the million-year scale fairly well. For

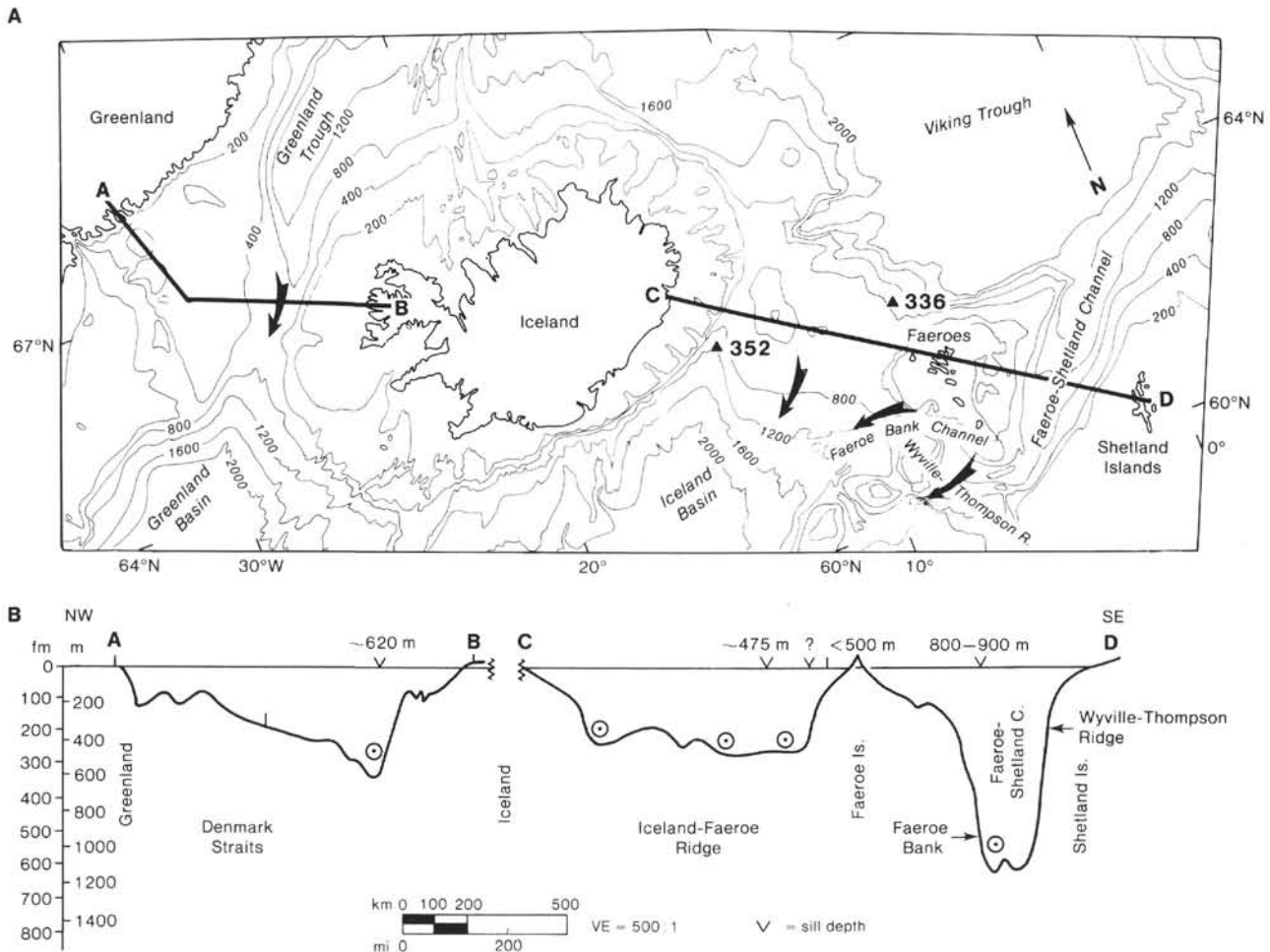


Figure 2. A. Bathymetry of the Greenland-Scotland Ridge, in meters. Arrows indicate modern overflow routes (after Miller and Tucholke, 1983). B. Bathymetric cross section of the Greenland-Scotland Ridge (located in Fig. 2A) showing maximum sill depths (after Miller and Tucholke, 1983). Section does not cross sill depths of Faeroe Bank Channel and Wyville-Thompson Ridge, which are indicated with arrows. (See text and Figure 7 for explanation of lines A-B and C-D.)

Table 1. Oxygen and carbon isotopic records.

DSDP site	Location	Present water depth (m)	Age	Reference
608	Kings Trough, eastern North Atlantic (42°50.21' N, 23°05.25' W)	3526	Miocene	This chapter
77	Eastern Equatorial Pacific (00°28.90' N, 133°13.70' W)	4291	Miocene	Savin et al., 1981
289	Western Equatorial Pacific (00°29.29' S, 15°30.69' E)	2206	Miocene	Savin et al., 1981
119	Bay of Biscay (45°01.90' N, 07°58.49' W)	4447	Oligocene	Miller and Curry, 1982
563	Western North Atlantic (33°38.53' N, 43°46.04' W)	3793	Miocene-Oligocene	Miller and Fairbanks, 1985

example, the $\delta^{13}\text{C}$ record at Site 608 adequately represents the global carbon isotopic record noted at more closely sampled Sites 77, 289, and 563 (Figs. 4, 5).

Diagenesis may alter the isotopic record, especially at deeply buried (greater than 400-m) sites (Miller and Curry, 1982; Killingley, 1983). We argue that the Miocene of Site 608 (Fig. 3) is substantially unaltered. Oxygen isotopic values at Site 608 mimic the patterns found at Site 563 (Fig. 5), even in the finer-scaled variations noted between 25 and 15 Ma (Fig. 5). This similarity occurs between sections with substantially different burial depths (240–300 m sub-bottom at Site 563 versus 300–400 m at Site 608), and argues against extensive diagenetic recrystallization of the calcite tests.

In addition, samples from Miocene sections with burial depths of less than 100 m to greater than 600 m (including those from Site 608) show no decrease in $\delta^{18}\text{O}$ values with burial depth as postulated by Killingley (1983), suggesting that the sections described here are not substantially altered.

HYDROGRAPHIC AND BATHYMETRIC SETTING

The modern bottom-water temperature and salinity characteristics at Sites 608 and 563 fall within the defini-

Table 2. Oxygen and carbon isotopic values for mixed species of *Cibicidoides*, Hole 608 (reported to PDB standard).

Sample (core-section, interval in cm)	Sub-bottom depth (m)	Age ^a (Ma)	$\delta^{18}\text{O}$	$\delta^{13}\text{C}$
17,CC	156.40	6.0	2.29	1.07
18,CC	164.80	6.5	2.01	1.09
21,CC	192.45	8.0	2.01	1.09
22-2, 86-92	197.26	8.3	2.00	1.25
22,CC	202.90	8.7	1.87	0.87
23-3, 88-93	208.38	9.0	2.02	0.77
23,CC	214.10	9.2	1.72	0.94
24-3, 74-79	217.84	9.3	1.25	0.94
24,CC	223.55	9.5	1.61	1.08
25,CC	233.30	9.9	1.71	0.75
26-3, 73-78	237.03	10.0	1.83	0.90
26,CC	241.23	10.2	1.85	1.20
27-3, 47-52	246.37	10.4	1.69	0.79
27-6, 47-53	250.87	10.6	1.57	1.11
28,CC	261.20	11.3	1.92	1.05
29,CC	271.00	11.9	1.87	1.16
30-3, 15-20	274.85	12.2	1.14	0.86
31-1, 85-90	282.15	12.7	1.69	0.91
32-3, 38-40	294.30	13.5	1.40	1.26
32-5, 80-85	297.70	13.7	1.54	1.60
33-1, 25-27	300.75	13.9	1.08	1.57
33-1, 52-56	301.00	13.9	1.31	1.70
34-1, 62-63	310.72	14.6	0.81	1.28
34-3, 28-30	313.38	14.7	0.82	1.07
34-4, 90-92	315.50	14.9	0.95	1.22
35-3, 90-92	323.60	15.5	0.97	1.51
36-3, 38-40	332.68	16.4	0.97	1.30
36-5, 38-40	335.68	16.7	0.57	0.92
38-6, 38-40	356.38	18.8	1.00	0.68
39-3, 38-40	361.48	19.3	0.76	0.60
40-2, 85-90	370.05	20.0	1.00	0.24
40,CC	376.01	20.5	1.20	0.67
41-4, 107-112	382.87	21.0	1.08	0.54
41-6, 86-91	385.66	21.2	0.95	0.28
42-3, 100-105	390.90	21.6	0.97	0.67
42-5, 86-91	393.76	21.8	1.03	0.59
43-3, 110-115	400.06	22.3	0.77	0.81
43-5, 10-15	402.60	22.5	0.70	1.06
44-2, 10-15	407.70	22.9	0.98	1.75
44-4, 38-40	410.98	23.2	0.98	1.20
45-3, 38-40	419.08	23.8	0.94	1.06
45-4, 38-40	420.58	23.9	0.94	0.71
45-5, 38-40	422.08	24.0	1.21	0.55
46-3, 24-26	428.54	24.5	1.14	0.61

^a Ages computed using age model 1, Table 3A.

Table 3A. Site 608 age model 1.

Criterion	Sub-bottom depth (m)	Age (Ma)
Gauss/Gilbert boundary	109.31	3.40
Chron 10/C5 boundary	206.99	8.92
base Chron C5n	248.00	10.42
base Chron C5Bn	321.60	15.27
top Chron 6C	361.60	19.35
base Chron C6Cn	424.44	24.20
Chron C8/C9	448.00	28.15

Note: Table 3A-G age model parameters were obtained by linear interpolation. Table 3A was used to construct Figures 4 and 5A, and employs magnetostratigraphy of Clement and Robinson (this volume) and Figure 3 by parsimoniously selecting prominent, unambiguous chron boundaries (Fig. 3). Interpretations of base Chron C6Cn differ from Clement and Robinson (this volume); see Figure 3.

Table 3B. Site 608 age model 2.

Criterion	Sub-bottom depth (m)	Age (Ma)
Gauss/Gilbert	3.4	109.31
Chron 6/7	6.7	170.09
Chron 9/10	8.5	199.66
Chron 10/C5	8.92	206.99
Chron C5/C5A	11.55	263.00
Chron C5A/C5AA	13.20	297.97
Chron C5AD/C5B	14.87	316.99
Chron C5B/C5C	16.22	332.09
Chron C5C/C5D	17.57	344.37
Chron C5D/C5E	18.56	353.26
Chron C5E/C6	19.35	361.61
Chron C6/C6A	20.88	388.73
Chron C6A/C6AA	21.90	391.57
Chron C6AA/C6B	22.60	407.85
Chron C6B/C6C	23.30	424.44

Note: This table used to construct Figure 5B, and employs magnetostratigraphy of Clement and Robinson (this volume) by interpolating between most chron boundaries.

Table 3C. Site 563 age model 1.

Criterion	Sub-bottom depth (m)	Age (Ma)
Base Chron C4An	159.5	8.2
Base Chron C5n	188.0	10.4
Top Chron C5An	204.5	11.9
Base Chron C5Bn	242.5	15.3
Top Chron C5Cn	251.0	16.2
Top Chron C5Dn	260.0	17.6
Base Chron C5Dn	267.0	18.1
Unconformity 270.5 From 18.4 to 19.4		
Base Chron C6n	277.5	20.5
Top Chron C6c	296.5	23.27
Top Chron C8	315.5	26.86
Base Chron C8n	323.5	27.74
Top Chron C12	337.0	32.46
Base C12n	339.2	32.90
Earliest Chron C12	360.1	35.28

Note: Used to construct Figure 5A; magnetostratigraphy after Miller, Aubry, et al. (1985).

tion of NADW (Worthington, 1976). Site 608 lies near Discovery Stations 3891 and 3889 (potential temperatures 2.69°C and 2.75°C, salinities 34.931‰ and 34.945‰, respectively), and Site 563 lies near Discovery Station 3628 (2.37°C, 34.905‰) (Fuglister, 1960).

Bottom waters in the eastern basins north of the Azores-Gibraltar Rise (Fig. 1) today reflect several sources. Iceland-Faeroe Ridge, Faeroe Bank Channel, and Wyville-Thompson Ridge overflow waters (Fig. 2) are major constituents of NADW in the eastern basins, termed Northeast Atlantic Deep Water (NEADW) (Lee and Ellett, 1965). The eastern basins, however, export most of the overflow waters to the western basins through the Charlie Gibbs Fracture Zone, and the overflows apparently do not directly contribute to the ventilation of the Iberian or Biscay basins below the 2.4°C potential temperature surface (Fig. 1) (Broecker et al., 1985). At present, the major source of ventilation for the deep basins (temperatures below 2.4°C; below about 3250 m) is advection from the south (Broecker and Peng, 1983; Broecker

Table 3D. Site 563 age model 2.

Sub-bottom depth (m)	Age (Ma)
157.69	8.15
189.90	10.98
197.30	11.46
206.80	12.01
211.21	12.23
216.31	12.65
235.14	13.47
236.54	14.32
254.00	16.61
268.24	18.91
273.24	19.34
282.79	20.16
295.10	21.99
301.67	22.59

Note: Determined using correlation lines shown in Figure 5A; assumes that the fine-scale features of Site 608 (age model 2) and Site 563 oxygen isotopic records correlate; used to construct Figure 5B.

Table 3E. Site 77 age model.

Criterion	Sub-bottom depth (m)	Age (Ma)
FO <i>Globigerina nepenthes</i>	178.	11.3
FO <i>Praeorbulina</i> spp.	241.	16.6
FO <i>Globorotalia kugleri</i>	317.	23.6
LO <i>Chiloguembelina</i> spp.	375.	30.0

Note: First occurrence = FO; last occurrence = LO. Biostratigraphic data summarized by Miller and Fairbanks (1985). Age models differ slightly from Miller and Fairbanks (1985) because of reassignment of the *G. nepenthes* datum from 12 Ma (W. A. Berggren, personal communication, 1983) to 11.3 Ma (Miller, Aubry, et al., 1985; Berggren et al., in press b).

Table 3F. Site 289 age model.

Criterion	Sub-bottom depth (m)	Age (Ma)
FO <i>Globorotalia margaritae</i>	156.5	5.6
FO <i>Globigerina nepenthes</i>	348.0	11.3
FO <i>Praeorbulina</i> spp.	510.0	16.0
FO <i>Globoquadrina dehiscentes</i>	654.0	23.2

Note: See Table 3E for explanation of FO and LO. Biostratigraphic data summarized by Miller and Fairbanks (1985). Age models differ slightly from Miller and Fairbanks (1985) as a result of reassignment of the *G. nepenthes* datum from 12 Ma (W. A. Berggren, personal communication, 1983) to 11.3 Ma (Miller, Aubry, et al., 1985; Berggren et al., in press b).

Table 3G. Site 119 age model.

Criterion	Sub-bottom depth (m)	Age (Ma)
Oligocene/Miocene boundary	293.0	23.7
LO <i>Globorotalia opima opima</i>	315.5	28.2
Top Zone NP23	330.5	30.3
Top Zone NP22	349.9	34.6
LO <i>Ericsonia formosa</i>	356.0	35.1

Note: See Table 3E for explanation of LO. Biostratigraphic data summarized by Miller (1982).

er et al., 1985). This high-silicate bottom water of southern origin has been attributed to Antarctic Bottom Water (AABW) (Lonsdale and Hollister, 1979), but is actually NADW mixed with a small amount of AABW that has passed through the Romanche Fracture Zone and returned north (Broecker and Peng, 1983; Broecker et al., 1985). At lesser depths (2.4–4.0°C; <3000 m), the eastern basin apparently is ventilated by a supply of NEADW and Labrador Sea Water.

In the Oligocene, flow from the south probably was impeded by the Azores-Gibraltar Rise (Fig. 1) and Azores-Biscay Rise, which had sill depths of 3000 to 3500 m (Miller and Tucholke, 1983). These barriers probably contributed to the isolation of the Iberian-Biscay basins, although today they are not effective barriers to bottom-water flow. The Charlie Gibbs Fracture Zone and numerous small fracture zones in the mid-ocean ridge provided conduits for east-west flow in the Oligocene to early Miocene (Miller and Tucholke, 1983). The subsidence of the other tectonic passage into the eastern North Atlantic, the Greenland-Scotland Ridge, is addressed in the Appendix.

All sites compared here (Table 1) were located at abyssal depths (greater than 2000 m) during the Oligocene and Miocene. Site 608 can be "backtracked" (Sclater et al., 1971; Berger and Winterer, 1974; Sclater et al., 1977; Parsons and Sclater, 1977) using the equation:

$$Pd = Id + kt^{1/2} - S \quad (1)$$

where Pd = paleodepth, Id = initial depth, t = (age of basement – age of level considered), S = sediment correction of $0.66 \times$ (basement depth sub-bottom – depth sub-bottom of level considered), and k = a constant derived from empirical age versus subsidence curves. Knowing k and present depth (Pr), initial depth for a site can be computed:

$$Id = Pr - kt^{1/2} + S \quad (2)$$

For the western North Atlantic, k has a value of 300 (computed from data of Tucholke and Vogt, 1979, solving equation 1 for k , assuming $Id = 2650$ m); for the Pacific, it has a value of 350 (Sclater et al., 1977). These equations are valid for crustal ages less than about 80 Ma (Parsons and Sclater, 1977), which includes Sites 563, 608, and 77. For crustal ages greater than this (e.g., Site 289), subsidence follows the exponential curve:

$$Pd = Id - A + A \times e^{(-t/\tau)} - s \quad (3)$$

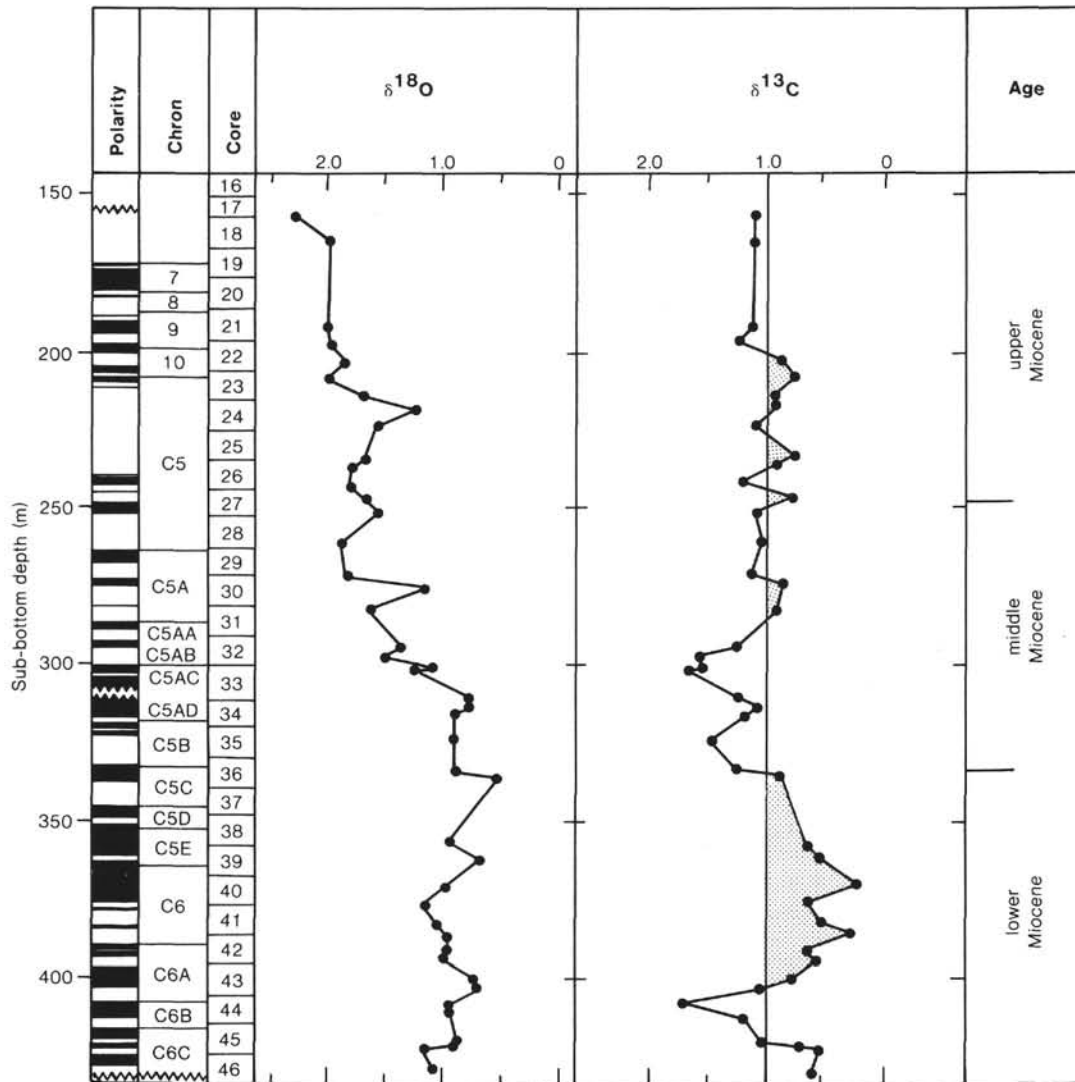


Figure 3. Oxygen and carbon isotopic values for the benthic foraminiferal taxon *Cibicidoides* spp. at Site 608 reported in parts per mille to PDB standard. The vertical line is drawn through the mean of $\delta^{13}\text{C}$ values (1.0‰), and the depleted intervals are indicated with stipples. Magnetostratigraphic polarity patterns after Clement and Robinson (this volume); chron interpretations follow Clement and Robinson except for Chrons C6B to C6C. Chronostratigraphic subdivisions were drawn on the basis of magnetostratigraphy and the time scale of Berggren et al. (in press b).

where $\tau = 62.5$ and A is an empirical constant of about -3650 for Pacific crust (computed using empirical data of Parsons and Sclater, 1977, table 2, solving equation 3 for A , assuming $l_d = 2700$ m).

Assuming simple thermal subsidence (equation 1), depths at Site 608 for the Miocene intervals investigated range from 3100 to 3500 m; we assume that Site 608 was not uplifted significantly. This might not be the case, for Kings Trough is known to have been tectonically active in the Eocene-Oligocene (Kidd et al., 1982; Kidd and Ramsay, this volume). Nevertheless, Eocene-Oligocene uplift values for nearby Cantabria Seamount (Bay of Biscay) were estimated to be less than 1000 m (Miller, 1982). It is reasonable to assume, then, that Site 608 lay at depths greater than 2000 m in the Miocene, and benthic foraminifers are indicative of oceanic depths (Thomas, this volume). At Site 563, Miocene depths ranged from 3200 to 3600 m, whereas Oligocene depths ranged

from 2200 to 3200 m. Miocene depths of 3600–4000 and 2200–2300 m are computed for Sites 77 and 289, respectively. Based upon geological and thermal subsidence constraints, the Oligocene depth at Site 119 was estimated as approximately 4050 m (Miller, 1982).

RESULTS AND DISCUSSION

Oxygen Isotopic Stratigraphy

Oxygen isotopic values at Site 608 are relatively constant between 310 and 420 m sub-bottom (Fig. 3) (lower to lower middle Miocene). The $\delta^{18}\text{O}$ values for this interval at Site 608 are slightly lower (approximately 0.3‰) than at Pacific sites at this time (Fig. 4), and are much lower (approximately 0.6‰) than at the western North Atlantic site (Fig. 5). This may be the result of differences in bottom-water temperature and/or seawater $\delta^{18}\text{O}$ between basins. If ascribed entirely to temperature dif-

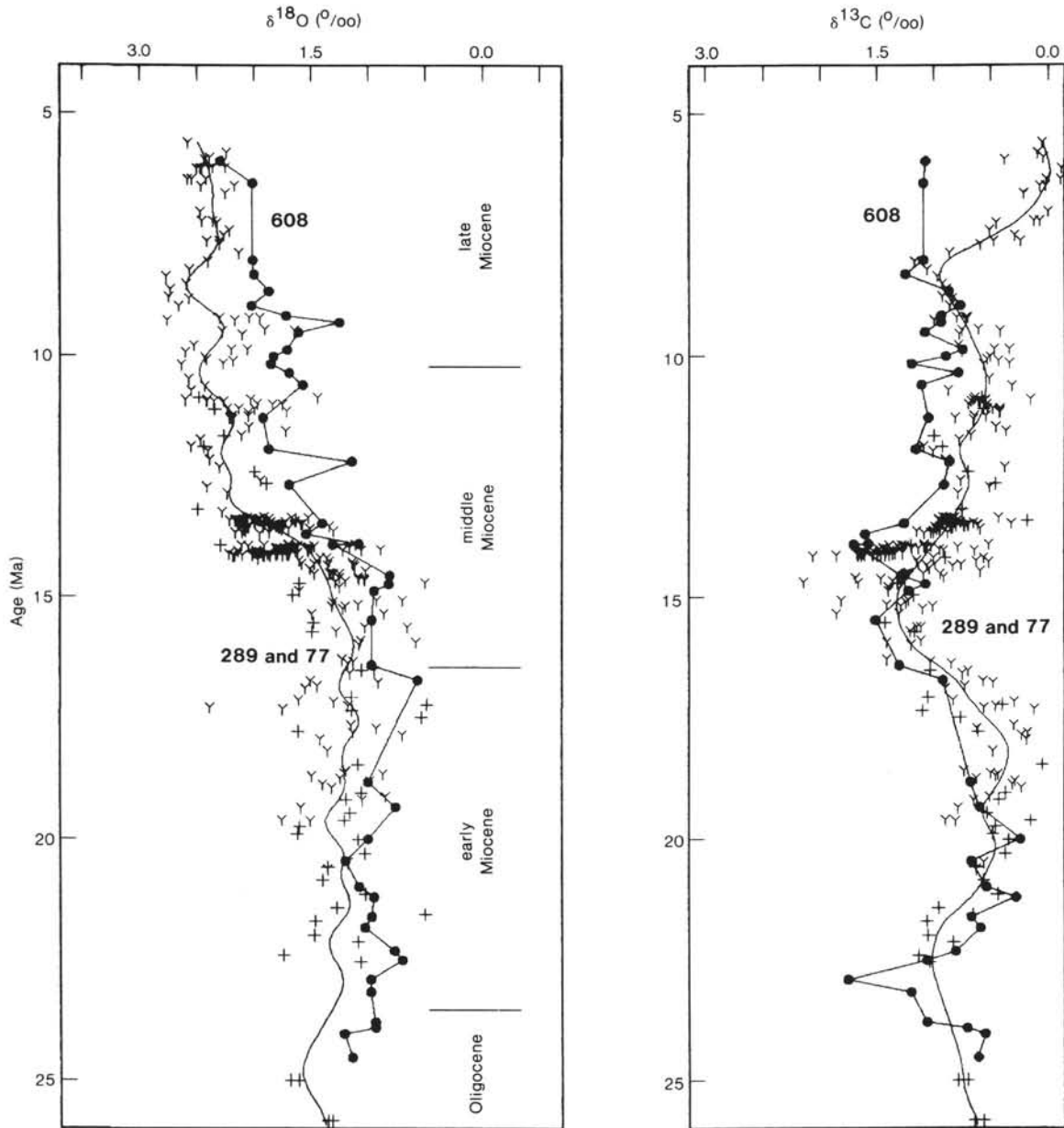


Figure 4. Age comparison of isotopic records from Pacific Sites 289 (Y) and 77 (+) with eastern North Atlantic 608 (closed circles). Sites 289 and 77 data after Savin et al. (1981). Curve drawn through data from Sites 289 and 77 was obtained by interpolating data from both sites to a constant time interval (0.1 m.y.), and smoothing with a 21-point Gaussian convolution filter, eliminating frequencies less than 1/m.y. This filter is essentially similar to a 1-m.y. running mean. Age models for Sites 289 and 77 obtained by interpolating between biostratigraphic datum levels (Table 3E, F). Age model for Site 608 obtained from magnetostratigraphy (Fig. 3, Table 3A). Chronostratigraphic subdivisions drawn using the time scale of Berggren et al. (in press b).

ferences, these data indicate that the eastern North Atlantic was approximately 3°C warmer than the western basin and 1 to 2°C warmer than the Pacific.

Between 310 and 294 m sub-bottom (14.6–13.5 Ma), Site 608 $\delta^{18}\text{O}$ values increase to approximately 1.9‰ (Fig. 3). This increase correlates with the well-known early middle Miocene $\delta^{18}\text{O}$ increase (Shackleton and Kennett, 1975; Savin, Douglas, et al., 1975; Savin, Douglas et al., 1981; Savin, Abel, et al., in press; Woodruff et al., 1981; among others). The increase at Site 608 appears to have been more gradual than that noted at Site 563 (Fig. 5A). The very sharp increase noted at Site 563 (Fig. 5A)

may be due to a short (1-m.y.) hiatus, because Chrons C5B and C5AD are concatenated (Miller, Aubry, et al., 1985).

Authors of most oxygen isotopic studies have assumed that the earth was ice-free prior to the middle Miocene and that the middle Miocene $\delta^{18}\text{O}$ increase represented the first Antarctic glaciation (e.g., Shackleton and Kennett, 1975; Savin, Douglas, et al., 1975; Savin, Douglas, et al., 1981; among others). Recent studies indicate periods of Oligocene glaciation (Miller and Fairbanks, 1983, 1985; Keigwin and Keller, 1984; Shackleton et al., 1984; Miller and Thomas, 1985; Poore and Matthews, 1984),

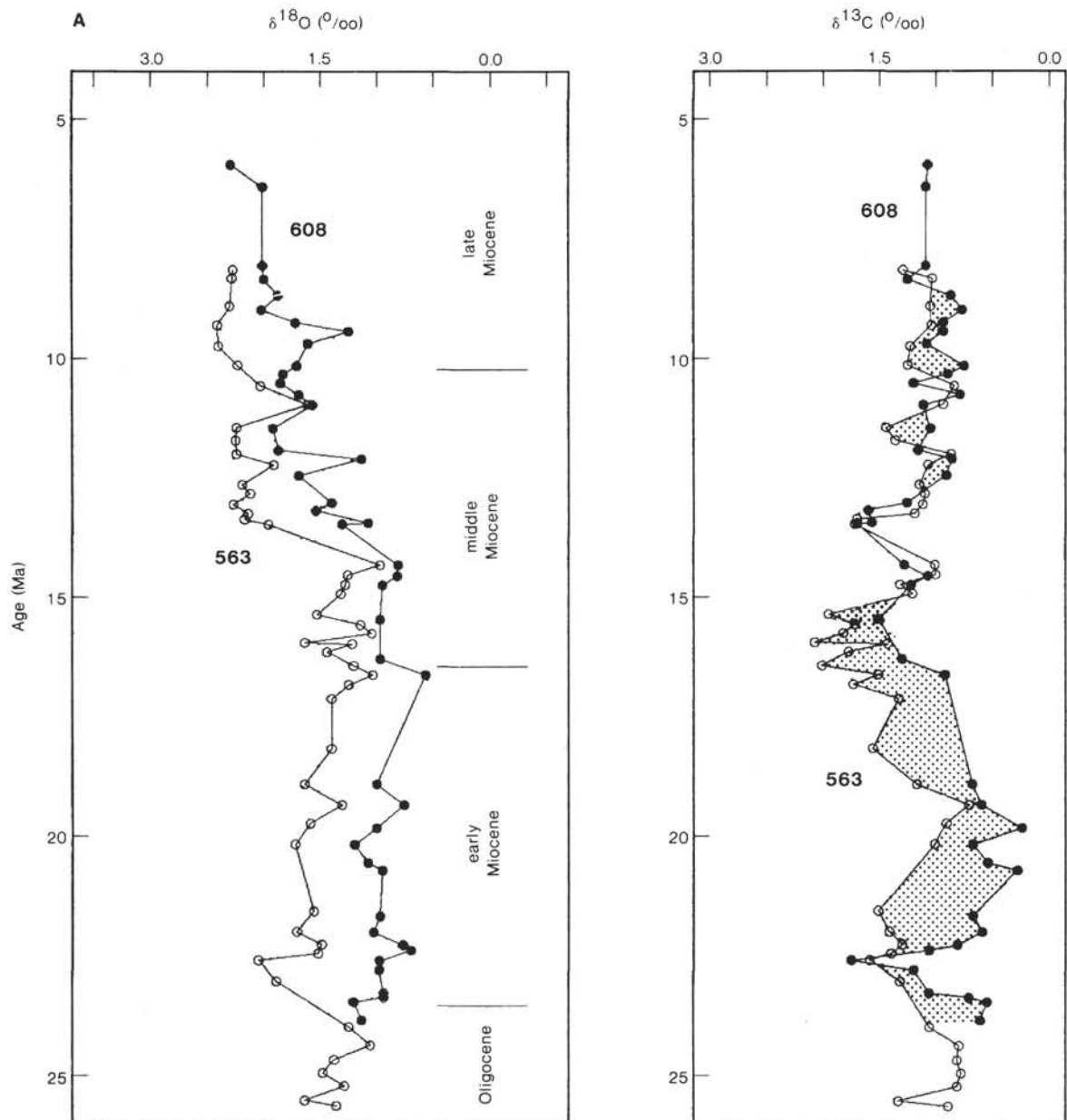


Figure 5. A. Age comparison of isotopic records from western North Atlantic Site 563 (open circles) (Miller and Fairbanks, 1985) and eastern North Atlantic Site 608 (closed circles). Correlation lines drawn between Site 563 and Site 608 oxygen isotopic records were used in deriving the age model for Fig. 5B. Age model for Site 563 derived by interpolating between magnetostratigraphic chron boundaries (Table 3C). Age model for Site 608 is that of Table 3A. Chronostratigraphic subdivisions drawn using the time scale of Berggren et al. (in press b). B. Age comparison assuming correlation between fine-scale $\delta^{18}\text{O}$ structure between Sites 563 (open circles) (Miller and Fairbanks, 1985) and 608 (closed circles). Age models (Table 3B, D) differ from those in Figure 5A.

so the middle Miocene $\delta^{18}\text{O}$ increase probably cannot be ascribed entirely to ice growth. Because the increase is greater than 1‰, it must represent either development of ice caps that exceeded present-day ice volume (which we regard as unlikely) or a combination of ice growth and bottom-water temperature decrease.

Detailed carbon isotopic comparisons between sites require better correlation control than that afforded by biostratigraphy. Magnetostratigraphy potentially provides such control; unfortunately, drilling disturbance and uncertainties in magnetostratigraphic identifications render correlations somewhat uncertain (Miller, Aubry, et al.,

1985; Clement and Robinson, this volume). Isotopic records provide a means for improving correlations between sites. We (1) assumed an age model for Site 608 based upon detailed magnetostratigraphy (Fig. 5B); (2) assumed that fine-scale Miocene $\delta^{18}\text{O}$ changes at Site 563 correlate with Site 608 (viz., that the small variations noted on Fig. 5A correlate between sites); (3) used these changes as interpolation points for computing a different age model for Site 563 (Fig. 5B; Table 3D). The resulting comparison (Fig. 5B) shows improved correlation of the carbon isotopic records, although the major features are unchanged from the magnetostratigraphically derived age

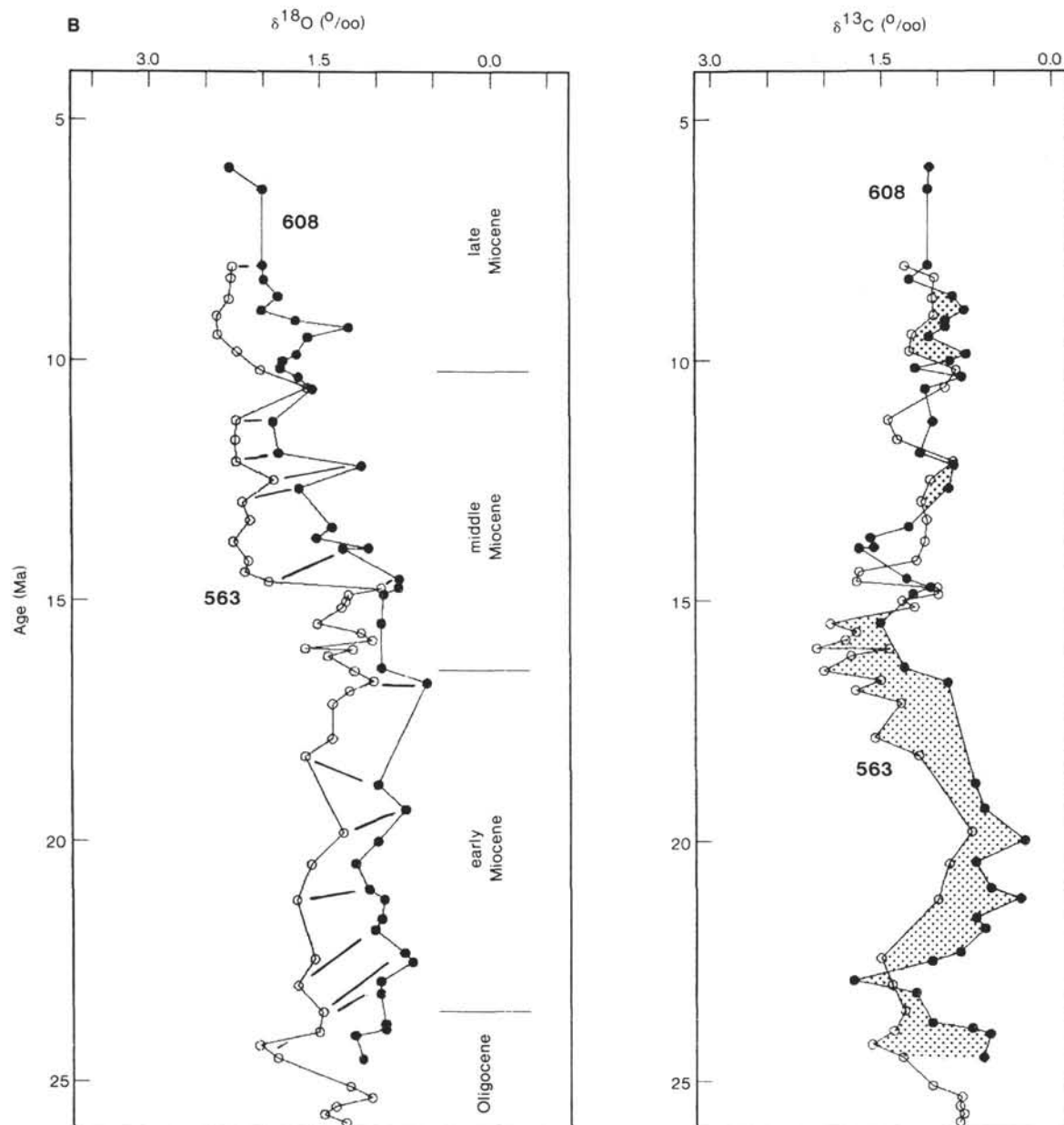


Figure 5 (continued).

model (Fig. 5B). Our data sets are not sufficient to "tune" the chronostratigraphic records with certainty; however, this exercise illustrates the potential for using oxygen isotopes for temporal control of carbon isotopic records. Another approach, using carbon isotopes to "tune" the oxygen isotopic record, also is a potential stratigraphic tool (e.g., Loutit et al., 1983).

Carbon Isotopic Records and Abyssal Circulation

The carbon isotopic record at Site 608 shows fluctuations with minima ranging from 340 to 400 m sub-bottom (17–22 Ma) and less prominent minima from 280 to 200 m (about 12.5–8.5 Ma). These fluctuations correlate with the global $\delta^{13}\text{C}$ reported by Miller and Fairbanks (1983, 1985) (Figs. 4, 5). The global maximum noted at ca. 8–9 Ma in Pacific data (Fig. 4; fig. 1 in Woodruff and Savin, 1985) is poorly resolved at Site 608 because

of limited sample resolution and "overprinting" by abyssal circulation effects.

Comparison of the carbon isotopic records from Site 608 with those from Pacific Sites 77 and 289 (Fig. 4) shows little difference prior to 15 Ma (early Miocene). Between approximately 14 and 10.5 Ma (middle Miocene) the eastern North Atlantic sites were generally higher in $\delta^{13}\text{C}$ than the Pacific sites. From approximately 10.5 to 8 Ma, Pacific locations and Site 608 again were similar, but after this the eastern North Atlantic was again higher in $\delta^{13}\text{C}$ than the Pacific. More data from Site 608 are needed to validate these differences, although we believe that the overall pattern will remain unchanged.

This comparison indicates supply of high $\delta^{13}\text{C}$ bottom waters to the region of Site 608 (Kings Trough area) in the middle Miocene. If Kings Trough can be taken as representative of the eastern North Atlantic, this sug-

gests supply of "younger" (higher oxygen, lower CO₂, lower nutrients) bottom waters to this basin from 14 to 10.5 Ma (middle Miocene) and 8 to 6 Ma (late late Miocene), analogous to modern NADW. However, the lack of a Pacific-eastern Atlantic $\delta^{13}\text{C}$ difference from 24 to 15 Ma and 10.5 to 8 Ma does not unequivocally indicate lack of bottom-water supply to the eastern basin. The similarity between Pacific records and Site 608 at these times may be due to absence or reduction of supply of young bottom waters to the eastern Atlantic basin. Alternatively, this similarity may be due to increased amounts of preformed nutrients in the source region for eastern North Atlantic bottom waters, as suggested for the glacial Pleistocene North Atlantic (Mix and Fairbanks, in press).

Comparisons of western and eastern North Atlantic records can be used to constrain the relative isolation-ventilation of the eastern basins. Basin-basin $\delta^{13}\text{C}$ differences within the North Atlantic suggest that the abyssal circulation in Iberian and Biscay basins was relatively restricted prior to the middle Miocene. Comparison of the Site 608 $\delta^{13}\text{C}$ record with the western North Atlantic (Site 563) (Fig. 5A,B) record shows that the eastern location was lower in $\delta^{13}\text{C}$ from 23 to 15 Ma (early Miocene). After 15 Ma there was little difference between eastern and western locations (Fig. 5). We suggest that Site 608 was more isolated than the western basin in the early Miocene. We interpret the rapid convergence of the Site 608 and Site 563 $\delta^{13}\text{C}$ records as reflecting increased ventilation of the Kings Trough area at ca. 15 Ma. This circulation change preceded the middle Miocene $\delta^{18}\text{O}$ increase (which occurred at about 14.6 Ma) by about 0.4 m.y (Figs. 5A, B). We propose four possible hypotheses to explain the apparent increase in ventilation at Site 608 in the middle Miocene:

1. Eocene-Oligocene uplift in the Kings Trough vicinity resulted in tectonic isolation and local stagnation at Site 608 until the middle Miocene.

2. Climatically induced hydrographic changes (temperature decrease or salinity increase) at high latitudes in the North Atlantic and/or Norwegian-Greenland Sea resulted in increased bottom-water production. This hypothesis requires that surface hydrographic changes in these areas preceded the middle Miocene global benthic $\delta^{18}\text{O}$ increase by about 0.4 m.y.

3. Assuming that the source of high-oxygen bottom waters was advection from the western basin (including Labrador Sea and/or Denmark Strait Overflow), Site 608 may have been situated below the sill depth in the mid-ocean ridge (Charlie Gibbs, Romanche fracture zones, or Azores-Gibraltar Rise) (Fig. 1). This situation would be analogous to the modern eastern equatorial basins, which are ventilated from the western equatorial Atlantic. In the Pleistocene, reduced flow of "NADW" resulted in decreased ventilation of the eastern equatorial Atlantic basins (Curry and Lohmann, 1983, 1985). By analogy, the middle Miocene increase in ventilation may be attributable either to an increase in the sill depths or to increased western basin ventilation and increased advection into the eastern basin.

4. Subsidence of the portion of the Greenland-Scotland Ridge between Iceland and the Faeroes in the middle Miocene resulted in increased supply of bottom water to the eastern basin.

We regard the local tectonic hypothesis (1) as unlikely. Oligocene carbon isotopic evidence from the Bay of Biscay (Site 119; Miller and Curry, 1982) indicates that the eastern basins were significantly less ventilated than the western basins, in agreement with the Site 608 record. Comparison of Site 119 with Site 563 (Fig. 6) clearly shows a dramatic difference, with very low $\delta^{13}\text{C}$ values in the Bay of Biscay from approximately 33 to 30 Ma ("middle" Oligocene). This period of inferred reduced advection correlates with an acme of *Nuttallides umbo-nifera* (Miller, 1982). Because this taxon is associated with low carbonate saturation (Bremer and Lohmann, 1982), its high abundance may be interpreted as indicating the presence of older, more corrosive water in the "middle" Oligocene of the Bay of Biscay (Miller, 1982) consistent with carbon isotopic comparisons.

The possibility that surface hydrographic changes in "NADW" source regions (hypothesis 2) caused the abyssal circulation change cannot be ruled out with the available data. Planktonic foraminiferal isotopic data from these potential source regions are needed to test this hypothesis (Mix and Fairbanks, in press). We regard this hypothesis as unlikely, because the $\delta^{13}\text{C}$ convergence apparently preceded the $\delta^{18}\text{O}$ increase (Fig. 5).

We also argue against the supply of bottom water solely from the western basin (hypothesis 3). Numerous small fracture zones in the mid-ocean ridge that were present in the Oligocene-early Miocene closed in the middle Miocene (Miller and Tucholke, 1983), suggesting that the middle Miocene circulation change was not due to increased advection across the mid-ocean ridge. However, because of possible tectonic complexities, backtracking estimates of the Romanche Fracture Zone (Bonatti and Chermack, 1982) and Azores-Gibraltar Rise (B. E. Tucholke, personal communication, 1985) may not be valid. Therefore, it is possible that increased advection into the Iberian-Biscay basins was due to subsidence of these sills. The hypothesis that bottom-waters were supplied to the Iberian-Biscay basins from the western basin via the Romanche Fracture Zone is testable with data from the Sierra Leone Rise (e.g., Curry and Lohmann, 1983).

We suggest that increased advection into the Iberian-Biscay basins was linked with a northern bottom-water source (hypothesis 4). The eastern basins of the northern North Atlantic probably had a bottom-water source from the Norwegian-Greenland Sea through the Faeroe Bank Channel (Fig. 7) during the Oligocene-early Miocene, albeit at a lower rate of supply than after 15 Ma. Seismic stratigraphic evidence indicates that widespread erosion occurred in the eastern and western basins of the northern North Atlantic in the Oligocene (Miller and Tucholke, 1983). Margin-intensified erosion caused by contour-following currents may be traced to the foot of the Wyville-Thompson Ridge/Faeroe-Shetland Channel system (Figs. 1, 2). Backtracking of the Faeroe Bank

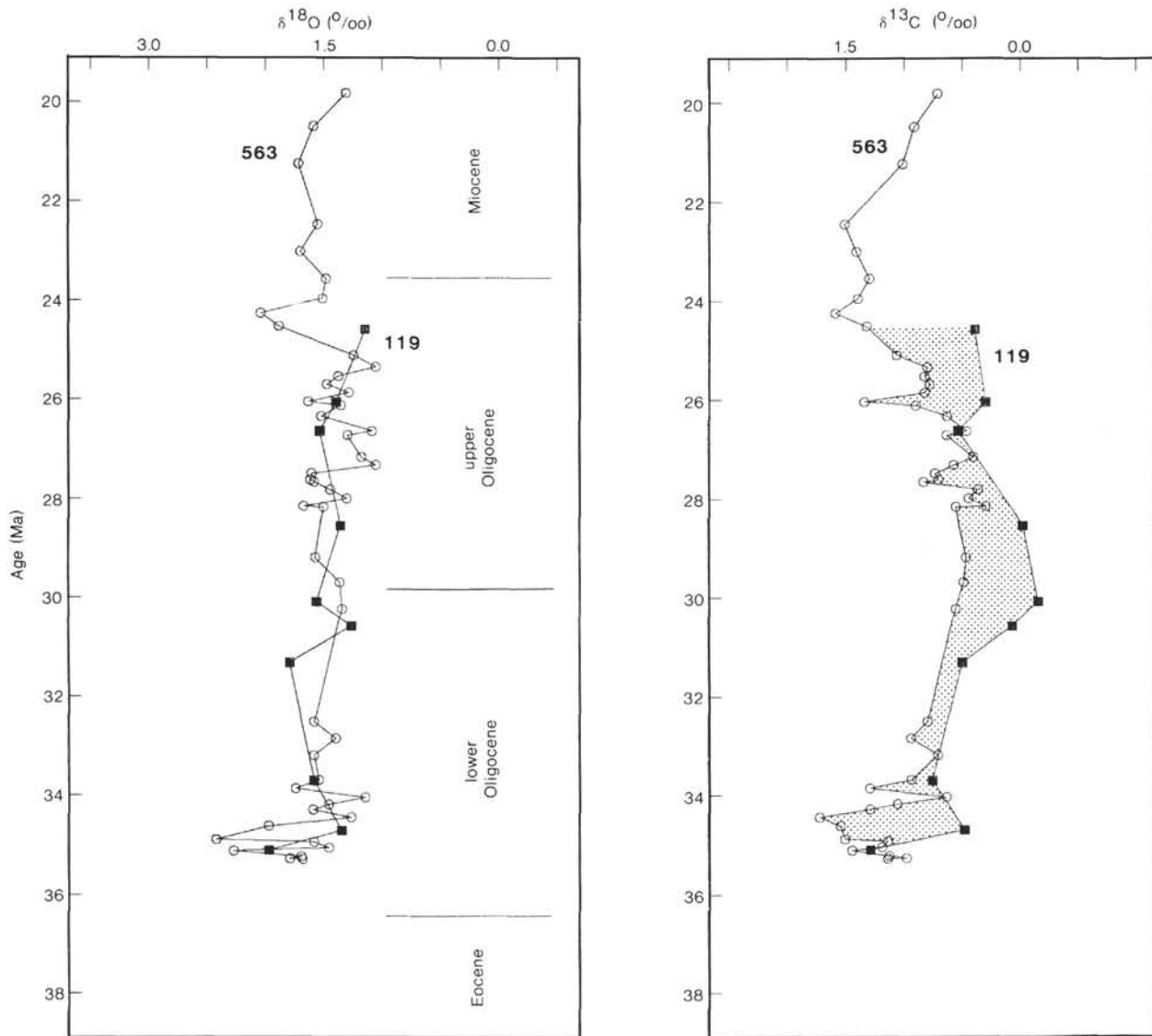


Figure 6. Age comparison of isotopic records from western North Atlantic Site 563 (open circles) (Miller and Fairbanks, 1985) and Bay of Biscay Site 119 (closed squares) (Miller and Curry, 1982). Chronostratigraphic subdivisions drawn using the time scale of Berggren et al. (in press a). Age model for Site 119 given in Table 3G.

Channel (modern sill depth about 900 m; backtracked using equation 1, basement age = Anomaly 24, 56 Ma, $k = 300$) suggests that it subsided below sea level by 36 Ma (earliest Oligocene) (Fig. 7). The Iceland-Faeroe Ridge (Figs. 2, 7) remained above sea level until about 21 Ma (earliest Miocene) (on the basis of the same assumptions as above and a sill depth of about 475 m; see also Thiede and Eldholm, 1983). By 15 Ma, the Iceland-Faeroe Ridge had subsided to about 150 m (Fig. 7) (see the Appendix for full discussion of backtracking of the Greenland-Scotland Ridge).

The correlation of the subsidence of the Iceland-Faeroe Ridge with the abyssal circulation change in the middle Miocene suggests a causal relationship. We favor the following scenario: (1) subsidence of the Iceland-Faeroe Ridge below a given threshold resulted in a rapid increase of advection (Fig. 5) into the eastern basin at 15 Ma; (2) the resulting increase in "NADW" may have

triggered the global climate change at about 14.6 Ma (see also Schnitker, 1980a, b). This scenario is most satisfying in that it emphasizes the importance of "NADW" in modulating climate (Broecker et al., 1985).

Seismic and lithostratigraphic evidence indicates that other abyssal circulation changes predated increased advection into the eastern basins. The top of biosiliceous sediments in the northeastern North Atlantic correlates with Reflector R2 at about 17 Ma (Miller and Tucholke, 1983). The association of Reflector R2 with several unconformities led Miller and Tucholke to conclude that this horizon represented an erosional pulse of bottom water (1983). These unconformities are best developed at western basin sites (563, 558, 407) and eastern basin sites shallower than 3 km (406, 555). Miller and Fairbanks (1985) noted a peak difference between Pacific and western North Atlantic $\delta^{13}\text{C}$ records at about 17 Ma, which they interpreted as increased bottom-water supply

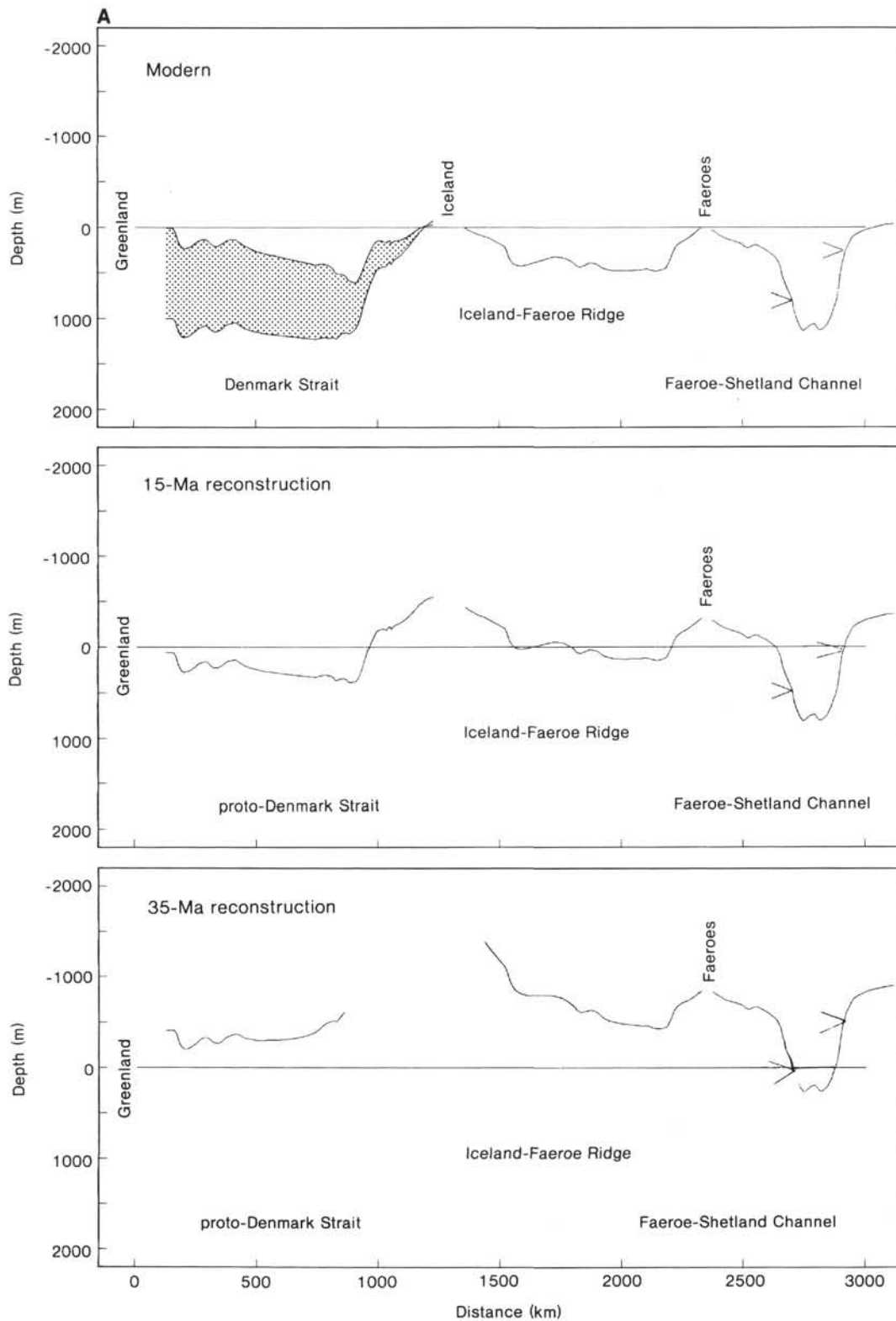


Figure 7. A. Tectonic reconstructions of the Greenland-Scotland Ridge across-sections A-B and C-D (Fig. 2) assuming simple thermal subsidence (equation 1), age models derived from Nunns (1983), and a sediment thickness in Denmark Strait of about 1 km (stippled area) (Larsen, 1983). See the Appendix for discussion. Portion between the Faeroe and Shetland islands does not cross sill depths; arrow points indicate sill depths for the Faeroe Bank Channel (deeper arrow) and Wyville-Thompson Ridge (shallower arrow). B. As in Figure 7A, but assuming 2 km of sediment in Denmark Strait (stippled area).

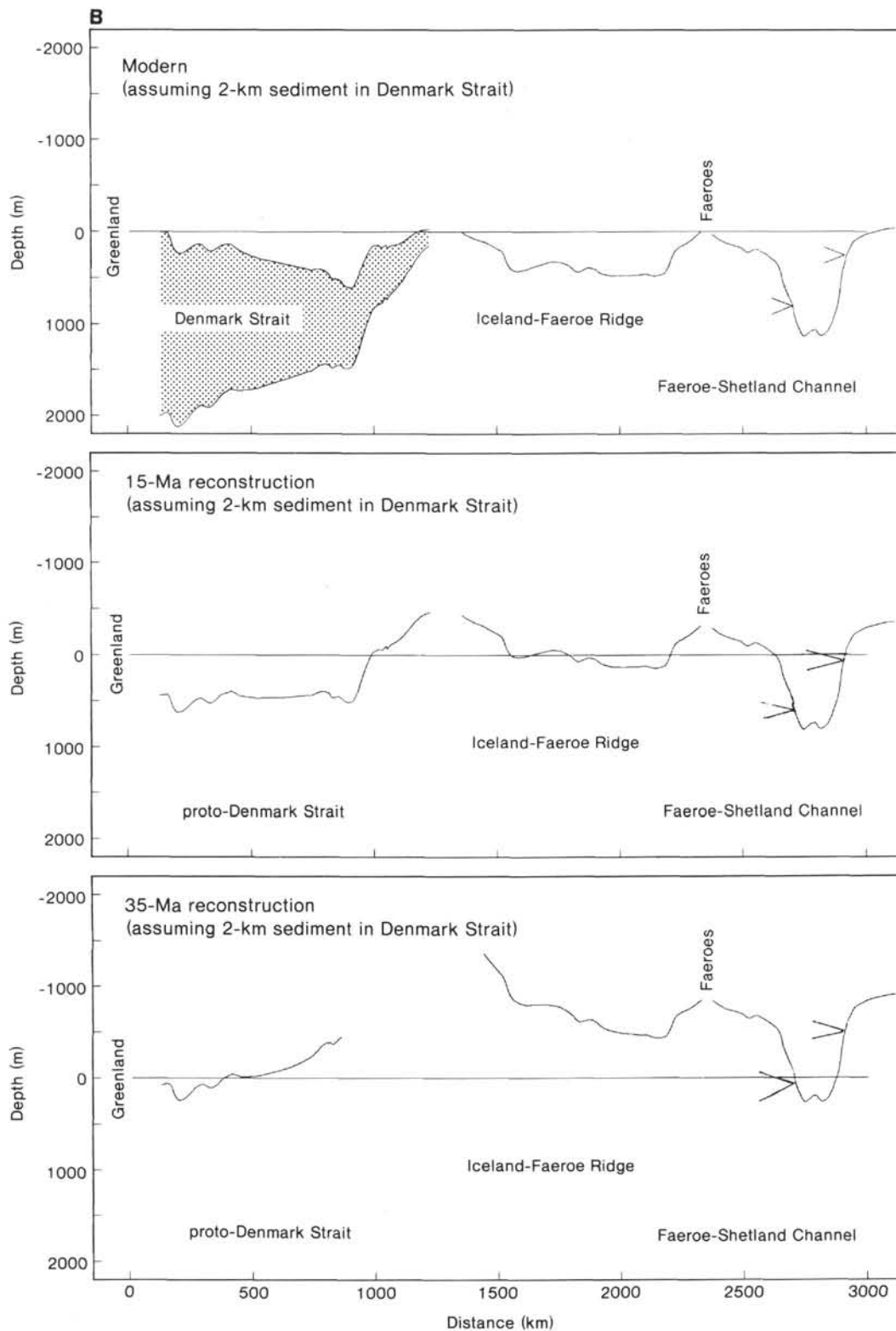


Figure 7 (continued).

associated with Reflector R2 (Miller and Fairbanks, 1985). However, Reflector R2 apparently correlates in time with little difference between Pacific and eastern basin $\delta^{13}\text{C}$ values (Fig. 4) and continuous sedimentation at Site 608 (Clement and Robinson, this volume). We speculate that

erosion associated with Reflector R2 was due to increased influx of bottom water to the western basin, but that advection into the deep eastern basin below 2 to 3 km remained relatively low. Erosion at sites shallower than 2 to 3 km in the eastern basin suggests the advection of

western deep water into the eastern basin above this depth, perhaps analogous to the influx of modern Labrador Sea Water at similar water depths in the modern ocean (McCartney and Talley, 1982; Talley and McCartney, 1982).

The ventilation of the eastern basin may be sensitive to sea-level changes. From about 10.5 to 9 Ma, the Pacific-eastern Atlantic $\delta^{13}\text{C}$ difference was reduced (Fig. 4), suggesting reduced advection into the eastern basin. A sea-level lowering has been suggested for ca. 10 Ma (Vail et al., 1977; Farre, 1985). Lowered sea level may have had an impact on the shallow Iceland-Faeroe sill (Fig. 7). Similar reduced advection into the eastern basin in the "middle" Oligocene (Fig. 6) was also linked with a sea-level lowering (Vail et al., 1977; Miller, Mountain, et al., 1985).

CONCLUSIONS

Abyssal circulation was quite different in the eastern and western northern North Atlantic basins in the Oligocene-early Miocene, in contrast to the situation in the modern ocean. The western basin had a ^{13}C -enriched bottom-water source; however, the eastern basin was depleted in ^{13}C and presumably less ventilated. This difference between the Atlantic basins disappeared at about 15 Ma, correlating with subsidence of the part of the Greenland-Scotland Ridge between Iceland and the Faeroes (Fig. 7). Although bottom waters probably had entered the North Atlantic across portions of the Greenland-Scotland Ridge since the earliest Oligocene (Fig. 7), subsidence of the Iceland-Faeroe Ridge marked a breaking of a major tectonic threshold to the development of North Atlantic abyssal circulation. The general pattern of abyssal circulation in the North Atlantic has not changed since the middle Miocene, although it has been modulated by high-frequency climatic events (e.g. Fairbanks and Mix, in press).

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APPENDIX
Backtracking of Greenland-Scotland Ridge³

The aseismic Greenland-Scotland Ridge is divided into three segments: the Denmark Straits between Greenland and Iceland, the Iceland-Faeroe Ridge, and the Faeroe-Shetland Channel/Faeroe Bank Channel/Wyville-Thompson Ridge system (Figs. 1, 2). We consider each segment individually.

Iceland-Faeroe Ridge

The backtracking of the Iceland-Faeroe Ridge is the best constrained of the segments. This portion of the ridge is oceanic crust; no sediment corrections need be made because the ridge crest is devoid of sediment (Miller and Tucholke, 1983). We assumed that Anomaly 13 (35.87 Ma) lay about 1429 km (Fig. 7) and interpolated to Anomaly 24 at about 2279 km (Fig. 7); these age constraints are tentative, because magnetic anomalies are poorly defined on the ridge crest (Nunns, 1983). Assuming simple thermal subsidence (equation 1), we obtained subsidence estimates for this portion of the ridge very similar to those of previous studies (e.g., Thiede and Eldholm, 1983; among others). We caution that the assumption of simple thermal subsidence may not be valid. If the tectonic change associated with the middle Miocene closing of Reykjanes Ridge fracture zones (Miller and Tucholke, 1983) was due to initiation or rejuvenation of the Iceland hot spot, then pre-middle Miocene depths on the ridge may have been deeper.

Faeroe-Shetland Channel/Faeroe Bank Channel/Wyville-Thompson Ridge System

This ridge system, the deepest of modern sills, is apparently floored by continental crust of unknown age (see discussion in Miller and Tucholke, 1983). Acoustic basement in this region consists of flat-lying basalts of Anomaly 24 (56 Ma) age. Assuming complete thermal "resetting" of subsidence at 56 Ma (Miller and Tucholke, 1983) as well as subsidence along an oceanic thermal subsidence curve, and neglecting sediments (again, allowed by the absence of sediments in the sills; Miller and Tucholke, 1983), provides a minimum estimate of subsidence (Fig. 7). If thermal subsidence was not completely reset, then the channel may have been deeper. Our estimate suggests that a connection through the Faeroe Bank Channel has existed since the end of

the Eocene (Fig. 7) (see also Miller and Tucholke, 1983) but that a connection across the Wyville-Thompson Ridge did not develop until perhaps as late as 15 Ma (Fig. 7).

Denmark Straits

The spreading history of this portion of the ridge is the least known. A spreading ridge jump may have occurred on this portion of the aseismic ridge (see discussion in Miller and Tucholke, 1983). However, we assumed the following: Anomaly 24, near the Greenland side about 230 km (Fig. 7); Anomaly 21, about 410 km; Anomaly 20, about 530 km; Anomaly 18, about 660 km; Anomaly 15, about 790 km; and extrapolation of the age to 1226 km.

These are reasonable interpretations of the magnetics of Nunns (1983); however, other interpretations are possible.

Larsen (1983) noted that the Denmark Strait is filled by 1 to 2 km of sediment. His limited sonobuoy coverage suggests that the best estimate near minimum width of the strait is about 1 km (his fig. 3). We used the following sedimentary basin (stippled area in Fig. 7A): thinning from 1 km to 0.8 km between km 129 and 800 (Fig. 7A); thinning from 0.8 km to 0.4 km between km 800 and 950; thinning from 0.4 to 0 km from km 950 to 1200 (Fig. 7A, stippled area).

We assumed that two-thirds of the sediment was younger than 15 Ma (i.e., consistent with Larsen's [1983] estimate that most of the sediments were due to glacial input); we also assumed a linear sedimentation rate between basement age and 15 Ma for the lower one-third of the sediment.

The results suggest that the Denmark Strait did not subside below sea level until after 35 Ma. However, assuming 2 km of sediment, the Denmark Straits subsided below sea level by 35 Ma (Fig. 7B). We assumed: thinning from 2.0 to 1.0 km from 129 km to 800 km; from 1.0 to 0.8 km from 800 km to 950 km; and from 0.8 km to 0 from 950 km to 1200 km (Fig. 7B, stippled area).

If the Denmark Straits were closed in the early Oligocene as suggested by the model using 1 km of sediment (Fig. 7A), then the erosional event associated with Reflector R4 at the end of the Eocene must be attributed to formation of bottom water in the Labrador Sea/Baffin Bay and/or northern North Atlantic (Miller and Tucholke, 1983). If a connection existed through the Denmark Straits in the Oligocene as suggested by the model using 2 km of sediments (Fig. 7B), then the erosional event associated with Reflector R4 may be attributed to influx of water from the Norwegian-Greenland Sea/Arctic Ocean, as postulated by Miller and Tucholke (1983).

³ See chapter proper for all equations, figures, and bibliographic references cited here.