Late Oligocene to early Miocene geochronology and paleoceanography from the subantarctic South Atlantic

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[1] At Ocean Drilling Program (ODP) Site 1090 on the Agulhas Ridge (subantarctic South Atlantic) benthic foraminiferal stable isotope records span the late Oligocene through the early Miocene (25–16 Ma) at a temporal resolution of ~10 kyr. In the same time interval a magnetic polarity stratigraphy can be unequivocally correlated to the geomagnetic polarity timescale (GPTS), thereby providing secure correlation of the isotope record to the GPTS. On the basis of the isotope-magnetostratigraphic correlation we provide refined age calibration of established oxygen isotope events Mi1 through Mi2 as well as several other distinctive isotope events. Our data suggest that the $\delta^{18}O$ maximum commonly associated with the Oligocene/Miocene (O/M) boundary falls within C6Cn.2r (23.86 Ma). The $\delta^{13}C$ maximum coincides, within the temporal resolution of our record, with C6Cn.2n/r boundary and hence to the O/M boundary. Comparison of the stable isotope record from Site 1090 to the orbitally tuned stable isotope record from ODP Site 929 across the O/M boundary shows that variability in the two records is very similar and can be correlated at and below the O/M boundary. Site 1090 stable isotope records also provide the first deep Southern Ocean end-member for reconstructions of circulation patterns and late Oligocene to early Miocene climate change. Comparison to previously published records suggests that basin to basin carbon isotope gradients were small or nonexistent and are inconclusive with respect to the direction of deep water flow. Oxygen isotope gradients between sites suggest that the deep Southern Ocean was cold in comparison to the North Atlantic, Indian, and the Pacific Oceans. Dominance of cold Southern Component Deep Water at Site 1090, at least until 17 Ma, suggests that relatively cold circumpolar climatic conditions prevailed during the late Oligocene and early Miocene. We believe that a relatively cold Southern Ocean reflects unrestricted circumpolar flow through the Drake Passage in agreement with bathymetric reconstructions. *INDEX TERMS:* 4267 Oceanography: General: Paleoceanography; 1035 Geochemistry: Geochronology; 4870 Oceanography: Biological and Chemical: Stable isotopes; *KEYWORDS:* Paleoceanography; 1035 geochronology, stable isotopes

1. Introduction

[2] Climate reconstructions based on deep-sea benthic foraminiferal oxygen isotope records indicate that the Antarctic cryosphere, as symbolized by massive ice sheets, developed during the early Oligocene [e.g., *Miller et al.*, 1987] and remained a substantial feature of the global climate system until the latest Oligocene [*Zachos et al.*, 1994]. At that time, the ice sheets waned as the climate warmed, a pattern that was sustained for much of the early Miocene. This period of warmth was punctuated by several brief glacial events [*Miller et al.*, 1987] characterized by large $\delta^{18}O$ increases of >0.5‰ on million year timescales (Figure 1). The largest of these isotope events have been designated as isotope zones, numbered with a M (Miocene) or O (Oligocene) prefix, and correlated to the geomagnetic polarity timescale (GPTS) [*Miller and Fairbanks*, 1985; *Miller et al.*, 1991]. The Oligocene/Miocene (O/M) boundary is associated with the first and largest of a series of early Miocene glaciations characterized by $\delta^{18}O$ maxima. A corresponding but slightly out of phase $\delta^{13}C$ maximum suggests that a major shift in global carbon cycling took place during this event, possibly in the form of increased organic carbon burial that may be related to global climate cooling [*Miller and Fairbanks*, 1985]. High-resolution stable isotope records from the tropical Atlantic illustrate that late Oligocene to earliest Miocene climate change was tightly coupled to orbital forcing [*Zachos et al.*, 2001a].

[3] The late Oligocene to early Miocene timescale is currently undergoing significant revision. In recent timescales [*Berggren et al.*, 1995; *Cande and Kent*, 1992, 1995] the O/M boundary is placed at polarity chron boundary C6Cn.2n/r with an age of 23.8 Ma. More recently, *Shackleton et al.* [2000] provided an astronomically calibrated age for C6Cn.2n/r of 22.9 Ma. This age is indirectly derived from correlation of astronomically tuned biostratigraphic datums and stable isotope records from Ocean Drilling Program (ODP) Site 929 to the biostratigraphic and paleomagnetic record at Deep Sea Drilling Project (DSDP) Holes 522 and 522A. The indirect approach was necessary because a magnetic polarity stratigraphy was not recoverable at Site 929 (and other ODP Leg 154 sites) for direct correlation to the astrochronology. To achieve its full potential, the ODP Leg 154 astrochronology must be correlated to the GPTS, and one way to achieve this is by constructing

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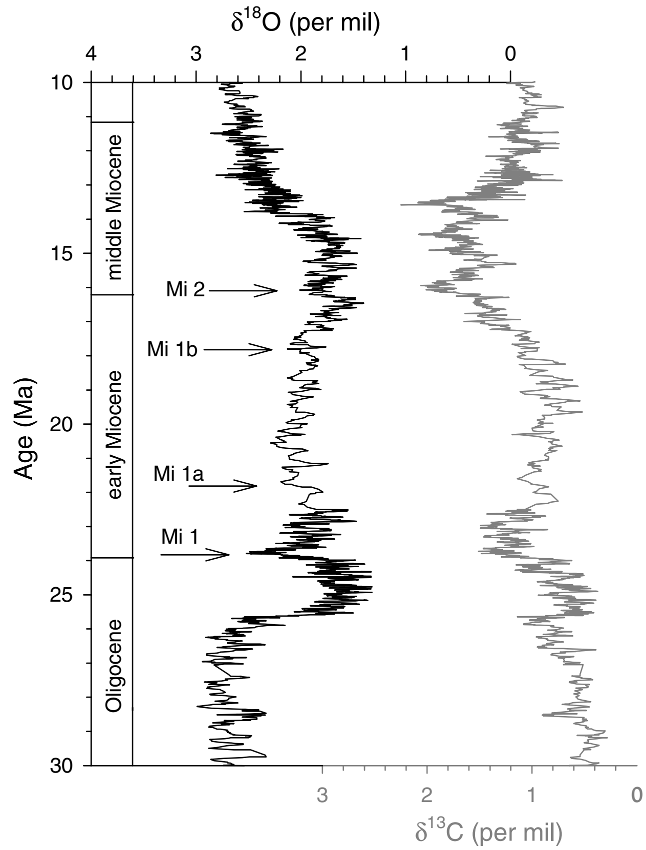

Figure 1. Composite benthic foraminiferal $\delta^{18}O$ and $\delta^{13}C$ records for the late Oligocene through middle Miocene on the *Cande and Kent* [1995] timescale (from *Zachos et al.* [2001b]). Mi1 through Mi2 denote increases in $\delta^{18}O$ values and previously established oxygen isotope zones in the early Miocene [*Miller et al.*, 1991]. The $\delta^{18}O$ values have been adjusted by +0.64 for disequilibrium vital effects [*Shackleton*, 1974].

high-resolution stable isotope records at sites with well-defined magnetostratigraphies.

[4] The sediment cores recovered at ODP Leg 177 Site 1090 yielded an excellent magnetic polarity record. On the basis of shipboard half core measurements, discrete samples, and continuous u channel samples, it was possible to identify all polarity chrons in the C6Cr to C5Cn (25–16 Ma) interval. In an initial comparison of Site 1090 and 929 stable isotope records we note a constant offset of ~0.9 Ma between the Site 929 astrochronolgy [*Shackleton et al.*, 1999] and the Site 1090 age model, which is based on correlation of the polarity stratigraphy to the GPTS of *Cande and Kent* [1995]. This offset is equivalent to the difference in the age of the O/M boundary given by *Cande and Kent* [1995] and *Shackleton et al.* [2000].

[5] Site 1090 is located in the Atlantic sector of the Southern Ocean on the southern flank of the Agulhas Ridge (43°S, 9°E) at a water depth of 3699 m (Figures 2 and 3 and Table 1). Today, the site lies within the mixing zone of North Atlantic Deep Water (NADW) and lower Circumpolar Deep Water (CDW) (Figure 3) and may have been sensitive to change in relative water mass fluxes in the past. Five holes were drilled at this site, the deepest of which reached 397.5 m below seafloor (mbsf), stratigraphically spanning the Holocene to middle Eocene. The sedimentary sequence is truncated by a hiatus at ~70 m composite depth (mcd) representing the lower Pliocene through middle Miocene. Magnetic susceptibility data were used to construct a continuous composite section to 240 mcd [*Shipboard Scientific Party*, 1999]. Sedimentation rates average ~20 m/Myr during the late Oligocene to early Miocene climate transition (25–23 Ma) and decrease to 10 m/Myr during the early Miocene (23–16 Ma).

[6] Our ultimate objective is to provide an orbital-scale oxygen isotope record at Site 1090 that spans the latest Oligocene-early Miocene interval (25–16 Ma) (K. Billups et al, work in progress, 2001). Thus far we have generated a stable isotope record at Site 1090 with a temporal resolution of ~5–10 kyr. This resolution allows us to refine the timing of early Miocene oxygen isotope zones and make an initial comparison among sites from different oceans. We are currently doubling the resolution of the isotope record in order to provide an orbital-scale oxygen isotope record for magnetostratigraphic correlation with astrochronology.

[7] In addition to its stratigraphic value, Site 1090, at a water depth of 3699 m, presents the first opportunity to study late Oligocene to early Miocene deep water masses in the Southern Ocean. Intermediate depth and upper deep water masses in the Southern Ocean have been characterized previously at Sites 747 and 704 [*Wright et al.*, 1992] (Table 1). Site 1090 provides a deep ocean end-member to establish a depth and longitudinal transect within the Southern Ocean in addition to comparisons between the major ocean basins (Figure 2 and Table 1).

2. Methods

[8] Approximately 40 cm^3 of sediment were taken at 10 cm intervals from Site 1090D 8H-1 (72 mcd) to 1090E 21H-5 (220 mcd), spanning the middle Oligocene through early Miocene. The stable isotope record was generated to a depth of 160 mcd because below this level, there are few intervals containing benthic foraminifera. Bulk sediments were oven-dried at 40°C, soaked overnight in a buffered solution, and washed through a 63 μm sieve. In preparation for stable isotope analyses, all shells were cleaned ultrasonically in deionized water to remove adhering particles and broken into smaller pieces to ensure speedy reaction. Stable isotope analyses were conducted using a VG Prism instrument located at the University of Santa Cruz (UCSC) and a VG Optima at Harvard University (HU). Both mass spectrometers are equipped with a common acid bath, and samples were reacted at 90°C following standard procedures. The $\delta^{13}C$ and $\delta^{18}O$ values are calibrated to Peedee belemnite (PDB) via NBS-19 and in-house standards (Cararra Marble). Replicate analyses of standards in the size range of the samples suggest that our analytical precision is better than 0.07‰ for $\delta^{13}C$ and 0.08‰ for $\delta^{18}O$ (n = 90). On the basis of duplicate analyses (n = 34) we note a small offset (0.14 ± 0.23‰) between the oxygen isotope data first generated at UCSC and later at HU. Although the small offset is not statistically significant, we apply a correction of −0.14‰ to the record generated at HU. In this manner we account for any interlaboratory offsets between Sites 1090 and 929 and 926 from the western tropical Atlantic [*Zachos et al.*, 1997; *Flower et al.*, 1995; *Paul et al.*, 2000], which we use for basin-scale comparisons of late Oligocene and early Miocene stable isotopic gradients. Site 1090 data are available electronically at the World Data Center-A for Paleoclimatology.

[9] Owing to the scarcity of benthic foraminifera an almost continuous high-resolution record could only be constructed from *Cibicidoides* spp. together with *Oridorsalis* spp. *Oridorsalis* $\delta^{13}C$ values are generally not used for paleoceanographic reconstructions because they have an infaunal habitat and do not represent the $\delta^{13}C$ values of dissolved inorganic carbon at the sediment water interface. However, analysis of 79 *Cibicidoides* and *Oridorsalis* from the same intervals justifies a constant correction of *Oridor-*

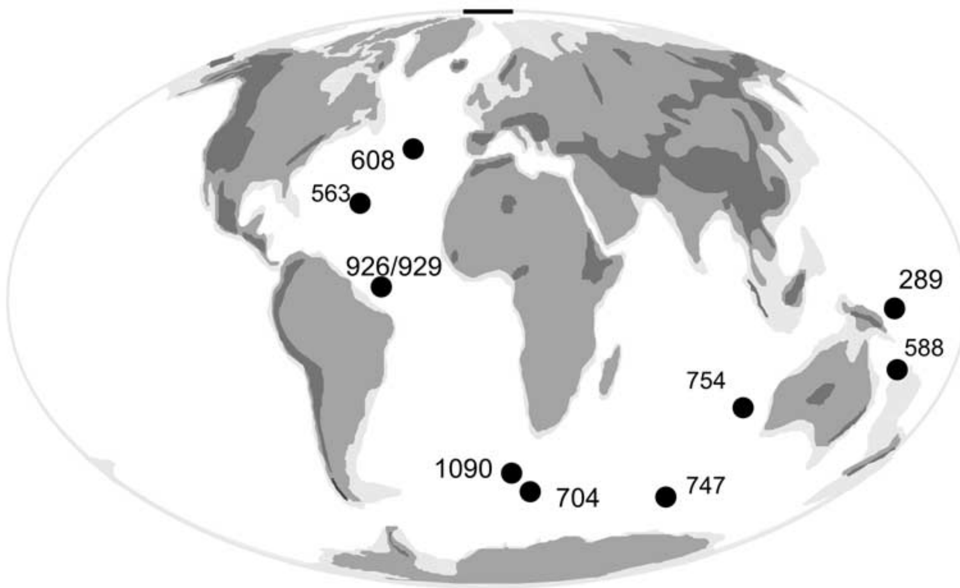

Figure 2. Locations of sites discussed in this study (Table 1).

salis δ^{18}O and δ^{13}C values to *Cibicidoides* (δ^{18}O correction, -0.4 ± 0.27‰; δ^{13}C correction, $+1.3 \pm 0.37$‰) (Figure 4).

3. Correlation of Oxygen Isotope Records to the Polarity Timescale (GPTS)

[10] At Site 1090 the recovery of an excellent paleomagnetic signal together with a high-resolution stable isotope record (Figure 5) allows us to refine the GPTS calibration of established oxygen isotope zones Mi1 through Mi2 [*Miller et al.*, 1991] (Table 2). In stratigraphic succession, Mi1 at 143.06 mcd correlates with C6Cn.2r, which places it just below the Oligocene/Miocene boundary. A large-amplitude δ^{18}O maximum at 117.24 mcd can be correlated to Mi1a and occurs within a polarity zone correlative to C6Ar. Mi1b is more difficult to identify in the Site 1090 record because of relatively high frequency variability in this interval. At 83.58 mcd, however, a δ^{18}O maximum within a polarity zone correlative to C5Dr marks a trend toward decreasing δ^{18}O values,

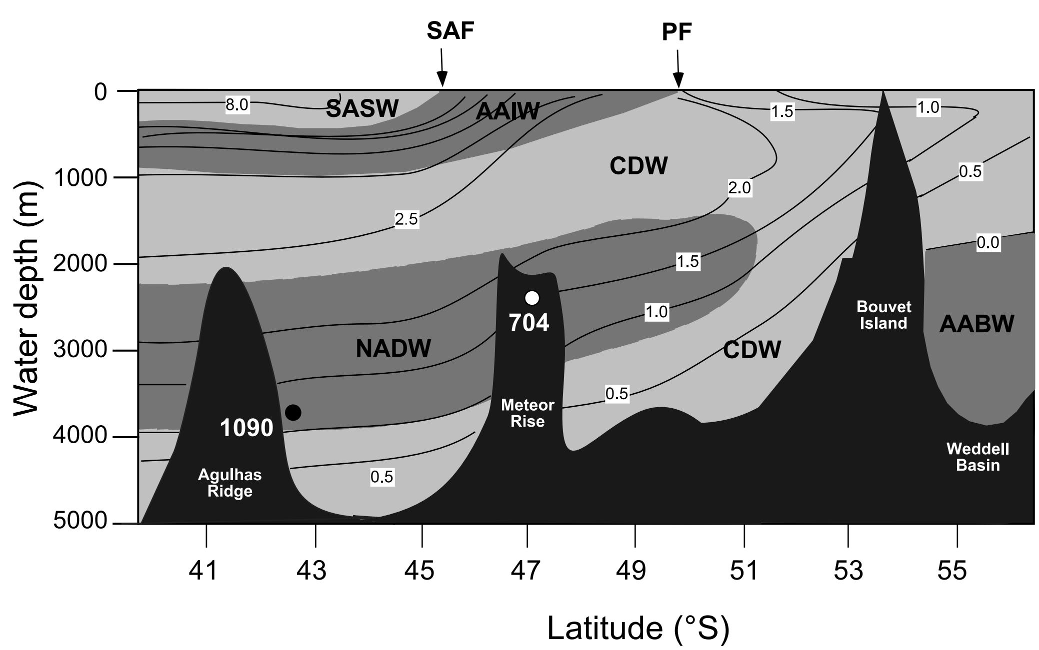

Figure 3. Bathymetric cross section through the Atlantic sector of the Southern Ocean showing the major surface, intermediate, and deep water masses [*Shipboard Scientific Party*, 1999]. Contours are potential temperatures (°C). NADW, North Atlantic Deep Water; CDW, Circumpolar Deep Water; AABW, Antarctic Bottom Water; AAIW, Antarctic Intermediate Water; SASW, Subantarctic Surface Water; SAF, Subantarctic Front; PF, Polar Front.

Table 1. Site Summary

Site	Water Depth	Latitude	Longitude	Reference
608	3526	42°50′N	23°50′W	*Woodruff and Savin* [1989][a]
563	3796	33°39′N	43°46′W	*Woodruff and Savin* [1989]
926	3598	3°43′N	42°54′W	*Flower et al.* [1995]
929	4361	5°58′N	43°44′W	*Zachos et al.* [1997]
1090	3699	43°S	20°W	this study
704	2532	46°53′S	7°25′E	*Wright et al.* [1992][a]
747	1695	54°48′S	76°47′E	*Wright and Miller* [1992][a]
754	1086	30°54′S	93°35′E	*Rea et al.* [1991][a]
588	1533	26°02′S	161°13′E	*Kennett* [1986][a]
289	2206	0°29′S	158°30′E	*Woodruff and Savin* [1989]

[a] Data are from the *Zachos et al.* [2001b] compilation.

which is consistent with definitions of Mi1b at other sites (e.g., Sites 608 and 563 [*Miller et al.*, 1991] and Site 747 [*Wright and Miller*, 1992]). Identification of Mi2 at Site 1090 is hampered by the presence of a hiatus at ~70 mcd (top of lower Miocene [*Shipboard Scientific Party*, 1999]). There is a double peak of similar magnitude at 72.99 and 72.69 mcd within C5Cn.1r and C5Cn.1n, respectively. Furthermore, a second maximum occurs just below the hiatus at 71.59 mcd, also within C5Cn.1n. These maxima have amplitudes of >0.5‰, which is characteristic for the $\delta^{18}O$ excursion that defines Mi2 [*Miller et al.*, 1991; *Wright and Miller*, 1992].

[11] In addition to these previously established events, high-resolution oxygen isotope data exhibit a number of smaller amplitude $\delta^{18}O$ cycles (Figure 5). In stratigraphic sequence these events include discrete maxima at 151.09 mcd just below Mi1 and at 133.84 mcd between Mi1 and Mi1a. These maxima occur within C6Cr and C6Br, respectively, and could prove useful for global-scale oxygen isotope correlations. To be consistent with the numbering scheme of *Miller et al.* [1991], the $\delta^{18}O$ maximum within C6Cr (151.09 mcd), which is the last maximum during the late Oligocene, could be referred to as Oi2b.1, and the $\delta^{18}O$ maximum within C6Br (at 133.84 mcd) could be referred to as Mi1.1. In addition to these secondary correlation points, there are several distinct maxima between 125 mcd and 117 mcd leading up to Mi1a and numerous smaller-amplitude but high-frequency $\delta^{18}O$ maxima between Mi1a and Mi1b.

[12] As more high-resolution records become available, the numbering scheme for the early Miocene isotope zones will undergo revision. Eventually, this oxygen isotope chronostratigraphy should follow a nomenclature that incorporates the GPTS correlation such as proposed by *Shackleton et al.* [1995] for the Pliocene, where numbers are reinitiated at each polarity chron boundary from youngest to oldest. In this scheme, Mi2 would then become 5Cn*x*, Mi1b would be 5Dr*x*, Mi1a would be 6Ar*x*, Mi1.1 would be 6Br*x*, Mi1 would be 6Cn*x*, and Oi7 would be 6Cr*x*, where *x* stands for the *x*th $\delta^{18}O$ maximum within a chron.

[13] Similarly, the high-resolution carbon isotope record displays a series of maxima and minima that are characteristic of the late Oligocene and early Miocene and could serve as global correlation points (Figure 5 and Table 2). In stratigraphic order, there are three distinct $\delta^{13}C$ events during the late Oligocene. A minimum occurs at 154.69 mcd within C6Cr, a maximum at 149.30 mcd also occurs within C6Cr, and another minimum occurs at 146.71 mcd, which correlates with C6Cn.3n. At 141.86 mcd a $\delta^{13}C$ maximum is coincident with the C6Cn.2n/r chron reversal at 142.10 mcd to within the limits of our current stratigraphic resolution. Because this $\delta^{13}C$ maximum falls only 24 cm above the polarity chron boundary C6Cn.2n/r, it can be used to identify the O/M boundary, which is placed at the C6Cn.2n/r boundary in recent timescales [*Berggren et al.*, 1995; *Cande and Kent*, 1992, 1995] and at the O/M boundary stratotype section in northern Italy [*Steininger et al.*, 1997]. Our work in progress (K. Billups et al., work in progress, 2001) will resolve in more detail the stratigraphic position of this $\delta^{13}C$ maximum with respect to the C6Cn.2n/r reversal. Additional high-frequency, high-amplitude $\delta^{13}C$ fluctuations are common above 141.86 mcd (Figure 5). At 85 mcd, values begin to increase toward a maximum at 72.59 mcd within C5Cn.1n. With $\delta^{13}C$ values of 1.9‰ this maximum is comparable to the amplitude of Mi2 displayed in other early Miocene records (e.g., Figure 1).

[14] Using the age model of *Cande and Kent* [1995], it is possible to refine the relative timing of late Oligocene through

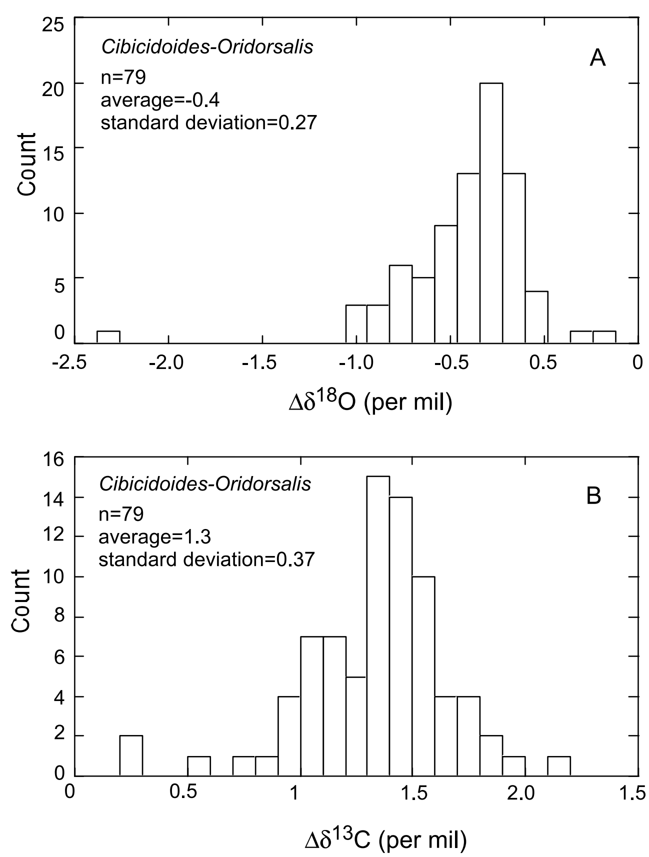

Figure 4. Distribution of the differences between the values of (a) $\delta^{18}O$ and (b) $\delta^{13}C$ for specimens of *Cibicidoides* and *Oridorsalis* from the same stratigraphic levels ($\Delta\delta^{18}O$ and $\Delta\delta^{13}C$, respectively). On the basis of 79 duplicates the average $\Delta\delta^{18}O$ is -0.4‰, and the average $\Delta\delta^{13}C$ is 1.3‰. We use these values to correct *Oridorsalis* $\delta^{18}O$ and $\delta^{13}C$ values to *Cibicidoides* $\delta^{18}O$ and $\delta^{13}C$ values.

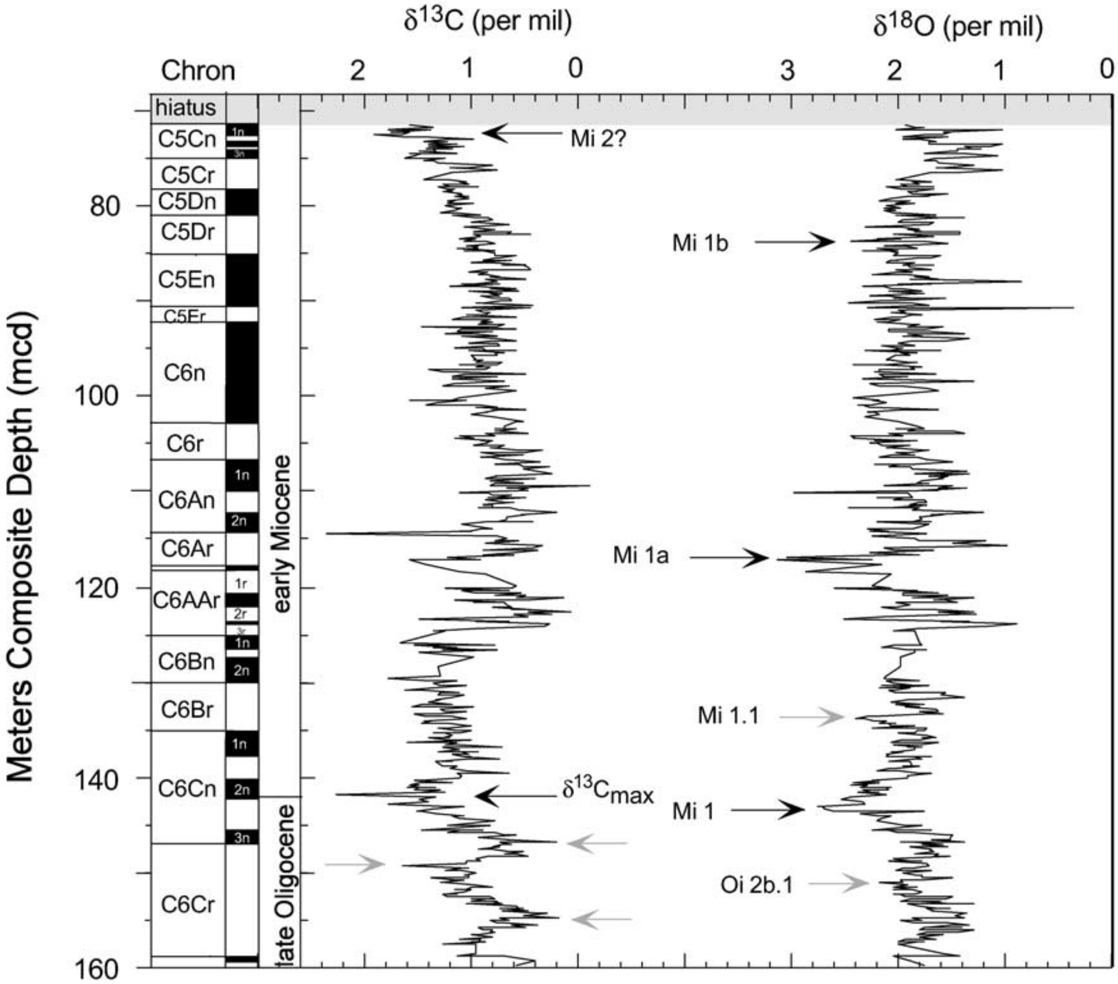

Figure 5. Site 1090 stable isotope records and correlation to geomagnetic polarity chrons plotted versus meters composite depth (mcd). Oxygen isotope zones Mi1 through Mi2 are labeled after *Miller et al.* [1991] and marked by solid arrows. Two additional correlation points based on the $\delta^{18}O$ record and four additional correlation points based on the $\delta^{13}C$ record are highlighted by shaded arrows. For specific stratigraphic designations, see Table 2.

early Miocene stable isotope events (Figure 6). The temporal resolution of ~10 kyr is an order of magnitude higher than in previous studies [e.g., *Miller et al.*, 1991; *Wright and Miller*, 1992]. From oldest to youngest, the $\delta^{18}O$ maximum ascribed to Oi7 occurs at 24.34 Ma, Mi1 occurs at 23.86 Ma, Mi1.1 occurs at 23.29 Ma, Mi1a occurs at 21.69 Ma, Mi1b occurs at 18.0 Ma, and Mi2, on the basis of the $\delta^{13}C$ maximum, occurs at 16.2 Ma (Table 2).

[15] These ages are likely to change as orbital-scale resolution stable isotope records are generated (and as the orbital computations are refined). For example, the Site 929 astronomically tuned age model for the time interval between 21 and 25 Ma suggests that the age of the O/M boundary (C6Cn.2n/r) is ~0.9 Ma younger than previously thought (22.9 versus 23.8 Ma [*Shackleton et al.*, 2000]). To illustrate, we compare the 1090 record with the precessional-

Table 2. Site 1090 Depth-Age Control Points

Polarity Chron[a]	Depth, mcd[b]	Age,[c] Ma	Event	Description
C5Cn.1n (0.15)	72.59	16.2	~Mi 2	$\delta^{13}C$ maximum
C5Dr (0.40)	83.58	18.0	Mi 1b	$\delta^{18}O$ maximum
C6Ar (0.56)	117.24	21.69	Mi 1a	$\delta^{18}O$ maximum
C6Br (0.18)	133.84	23.29	Mi 1.1	$\delta^{18}O$ maximum
C6Cn.2n (0.09)	141.86	23.79	O/M	$\delta^{13}C$ maximum
C6Cn.2r (0.71)	143.06	23.86	Mi 1	$\delta^{18}O$ maximum
C6Cn.3n (0.24)	146.71	24.10		$\delta^{13}C$ minimum
C6Cr (0.80)	149.30	24.24		$\delta^{13}C$ maximum
C6Cr (0.65)	151.09	24.34	Oi2b.1	$\delta^{18}O$ maximum
C6Cr (0.34)	154.69	24.52		$\delta^{13}C$ maximum

[a] The number in brackets indicates the position within of polarity chron(i.e., 0.25 denotes 25% from onset of chron).
[b] Here mcd stands for meters composite depth.
[c] See *Cande and Kent* [1995].

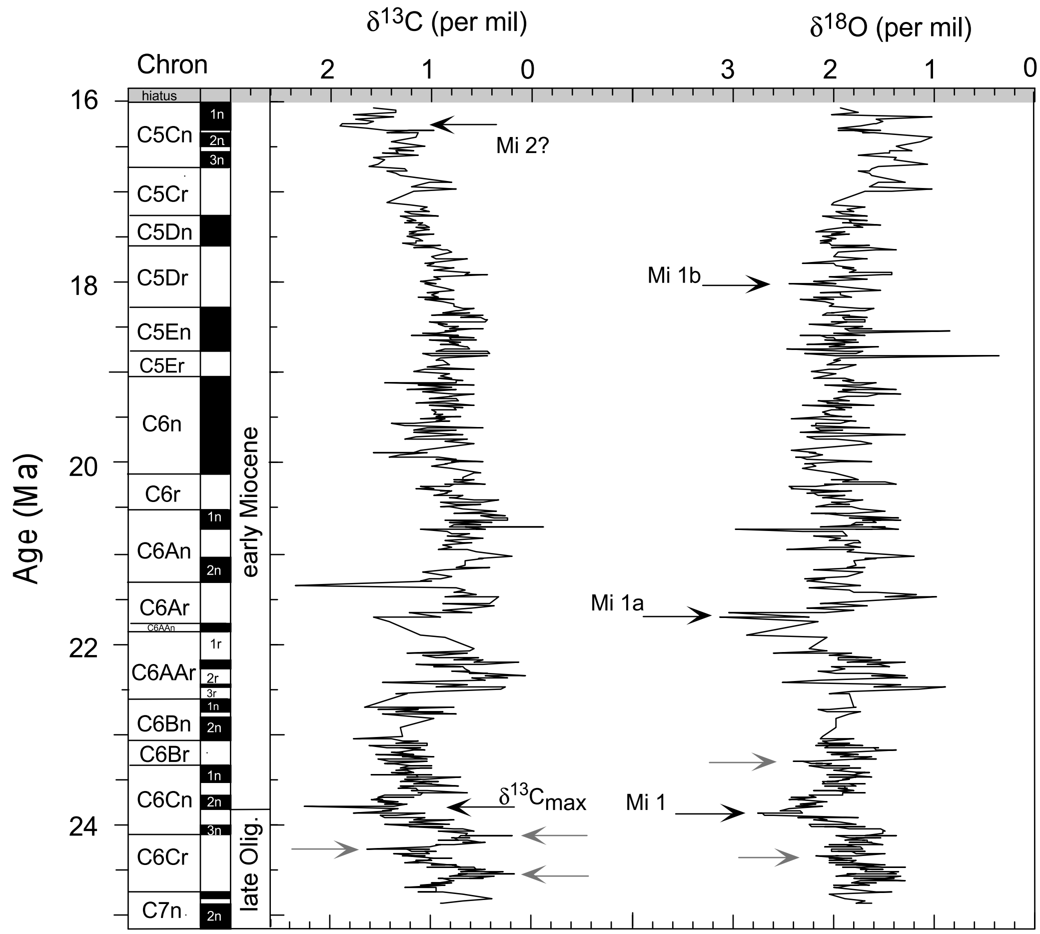

Figure 6. Site 1090 stable isotope records and geomagnetic polarity chrons plotted versus age (Ma) using the *Cande and Kent* [1995] timescale. Oxygen isotope zones Mi1 through Mi2 are labeled after *Miller et al.* [1991]. Arrows are as in Figure 5.

scale stable isotope record from Ceara Rise in the western equatorial Atlantic (ODP Leg 154 Site 929 [*Zachos et al.*, 1997; *Paul et al.*, 2000]) (Figure 7). For the purpose of this exercise, we shift the ages of the Site 1090 record by a constant 0.9 Ma. Superimposing the two records in this manner allows us to determine how the relative timing of the primary correlation points (i.e., Mi1, Mi1a, and the δ^{13}C events) compare with those from an astronomically tuned record (Figure 7). After accounting for the age model offset, excellent agreement exists between the Site 1090 and Site 929 higher-frequency stable isotope variability at and below the O/M boundary even though the resolution of the Site 1090 record is only about half that of the Site 929 record. The ages of Mi1 and the δ^{13}C events are therefore consistent for the two sites. The agreement between the records becomes questionable for ages younger than ~23 Ma. We conclude that the polarity chron ages are internally consistent with a 0.9 Myr offset in astrochronology only at and below the O/M boundary. Above the boundary the mismatch between Sites 1090 and 929 cannot be accounted for by a constant difference between the GPTS and astronomical timescales.

4. Comparison to Other Records

4.1. Basin-Scale Stable Isotope Gradients

[16] As sedimentary archives from the southern ocean are rare, the Site 1090 isotope record provides critical constraints on late Oligocene and early Miocene deep water circulation. We compare the stable isotope record from Site 1090 to published records from other ocean basins (Figure 2 and Table 1). With the exception of Sites 926, 289, and 563, all data were taken from the compilation of *Zachos et al.* [2001b], who have updated the individual age models to be consistent with the GPTS of *Cande and Kent* [1995]. For Sites 926, 289, and 563 we match the large-scale features (e.g., Mi events) of each stable isotope record to the Site 1090 record to bring these age models in line with the *Cande and Kent* [1995] timescale.

[17] Carbon isotope gradients between the North Atlantic and Pacific Oceans are small or nonexistent during much of the late Oligocene through early Miocene (25–16 Ma; Figure 8 and Table 3). Only δ^{13}C values at Indian Ocean Site 754 are consistently higher than those in the deep Southern Ocean, on average by ~0.5‰ (Figure 8c and Table 3). Between ~22 and 17 Ma, intermittent δ^{13}C gradients develop between Site 1090 and North Atlantic Sites 563 and 608 (Figure 8a), between Site 1090 and Southern Ocean Site 704 (Figure 8c), and between Site 1090 and Pacific Sites 289 and 588 (Figure 8d). No δ^{13}C gradient exists between the deep Southern Ocean and the tropical Atlantic (Figure 8b and Table 3).

[18] In contrast to the δ^{13}C records the δ^{18}O records exhibit persistent interbasin and intrabasin differences during the late Oligocene and early Miocene, with Site 1090 consistently showing the highest δ^{18}O values (Figure 9 and Table 3). With a ~0.7‰ difference the largest δ^{18}O gradient exists between Site 1090 and Indian Ocean Site 754 (Figure 9c and Table 3 for average δ^{18}O values). Interpreting the foraminiferal δ^{18}O values at face value and ignoring possible interlaboratory δ^{18}O offsets, which can be as large as ~0.2‰ [*Ostermann and Curry*, 2000], this observation suggests

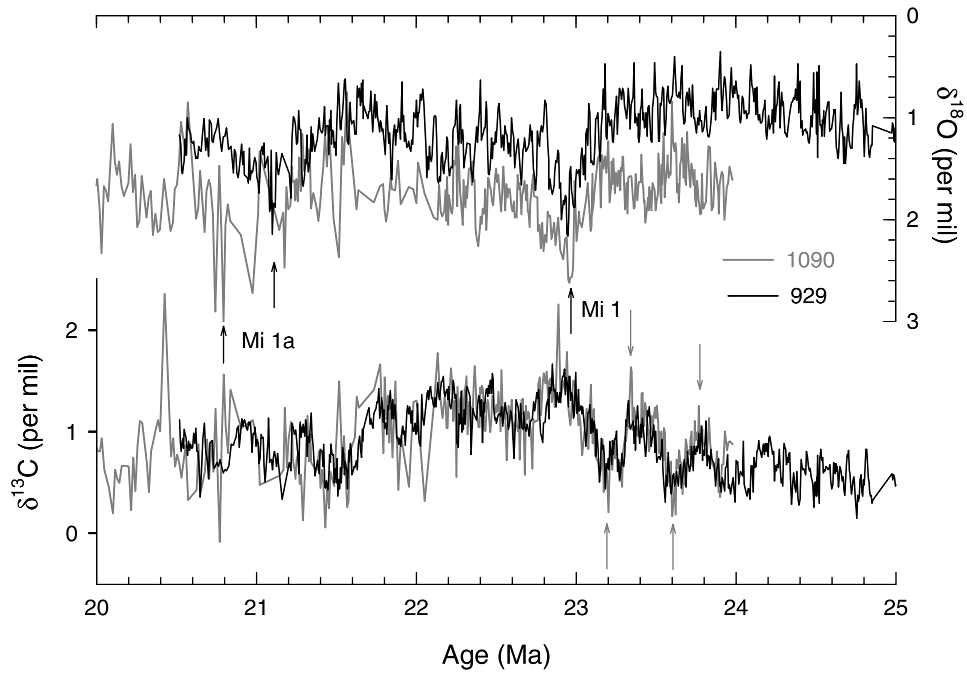

Figure 7. Comparison of the Site 1090 (a) δ^{18}O and (b) δ^{13}C records to those from western equatorial Atlantic Site 929 [*Zachos et al.*, 1997] for the late Oligocene to early Miocene climate transition (25–20 Ma). Site 1090 ages are based on the magnetostratigraphy at Site 1090 correlated to the GPTS of *Cande and Kent* [1995]. For purposes of comparison of the two sites we have made a constant adjustment of −0.9 Myr to the GPTS ages of Site 1090 in order to bring them in line with the orbitally tuned ages of the Site 929 record. Shaded arrows indicate additional δ^{13}C correlation points (Table 2).

that the warmest waters were present in the tropical Indian Ocean, consistent with the modern ocean (Table 3). On average, the Southern Ocean δ^{18}O values are higher by ~0.5‰ than those recorded in the tropical and North Atlantic (Figures 9a and 9b). With respect to the two tropical Atlantic sites (Sites 929 and 926) the magnitude of the δ^{18}O gradient certainly reflects environmental differences because in this case, we are able to discount measurement artifacts (see Section 2). Sites 1090 and 929 are both deep water sites (3700 and 4400 m, respectively, Table 1). The ~0.6‰ difference in δ^{18}O values provides evidence for warmer deep water mass in the north (~3°C warmer according to a 1°C for every 0.23‰ change [e.g., *Erez and Luz*, 1982]). A relatively small (~0.3‰) δ^{18}O gradient exists between Site 1090 and the other shallower Southern Ocean Sites 704 and 747 (Figure 9c and Table 3). The corresponding 1.3°C temperature gradient is larger than in the modern ocean (Table 3). Between Sites 704 and 747 the δ^{18}O difference is negligible (Table 3), which is not inconsistent with the small difference in water depth between the sites (~800 m, Table 1). Between the Pacific and the deep Southern Ocean, the δ^{18}O gradient is relatively large in comparison to the modern oceanic temperature gradient.

[19] Between 17 and 16 Ma, Site 1090 δ^{18}O values decrease. While decreasing δ^{18}O values are also observed at the other sites, it appears that the thermal gradient between the three Southern Ocean sites and the tropical Indian Ocean is getting smaller at this time (Figure 9c). Decreasing δ^{18}O values are consistent with a decrease in ice volume and increase in deep water temperatures associated with the early Miocene climatic optimum [e.g., *Shackleton and Kennett*, 1975, *Savin et al.*, 1975].

4.2. Implications on Deep Water Circulation

[20] Deep-sea carbon isotope gradients as represented by benthic foraminiferal δ^{13}C values are commonly used to reconstruct deep water circulation patterns [*Duplessy et al.*, 1984; *Mix and Fairbanks*, 1985; *Oppo and Fairbanks*, 1987; *Raymo et al.*, 1990; *Charles and Fairbanks*, 1990; *Wright et al.*, 1991, 1992]. This approach has been applied successfully to the late Plio-Pleistocene when North Atlantic to Pacific deep water nutrient gradients and hence foraminiferal δ^{13}C differences were large, similar to the modern ocean differences in the δ^{13}C of ΣCO_2 (Table 3). This approach has also demonstrated that modern-type deep water δ^{13}C gradients evolved during the late Miocene between 8 and 6 Ma [*Wright et al.*, 1991].

[21] Previous compilations of late Oligocene through early Miocene benthic foraminiferal δ^{13}C values illustrate that basin to basin δ^{13}C gradients were small [*Woodruff and Savin*, 1989; *Wright et al.*, 1992; *Wright and Miller*, 1992], perhaps related to low mean ocean nutrient levels [*Delaney and Boyle*, 1987; *Wright et al.*, 1991]. On the basis of foraminiferal δ^{13}C values, *Woodruff and Savin* [1989] suggested that early Miocene deep water circulation was similar to present in that a cold Southern Component Deep Water mass filled the majority of the deep ocean but differed from today in that the low-latitude Tethys was a source of warm and salty waters to intermediate water depths. Using the same methodology, *Wright et al.* [1992] showed that the North Atlantic was an additional source of deep water during the early Miocene at least between 20 and 16 Ma.

[22] The Site 1090 stable isotope record adds another dimension to these reconstructions because at 3700 m water depth it provides a means of characterizing the distribution of deep Southern Ocean water masses and their geographic extent. However, there is lack of site to site carbon isotope gradients during the late Oligocene through early Miocene that would allow an unequivocal reconstruction of the direction of deep water flow. Only the persistent δ^{13}C gradient between Indian Ocean Site 754 and the Southern Ocean sites indicates that there were two different water masses. The presence of a nutrient-depleted water

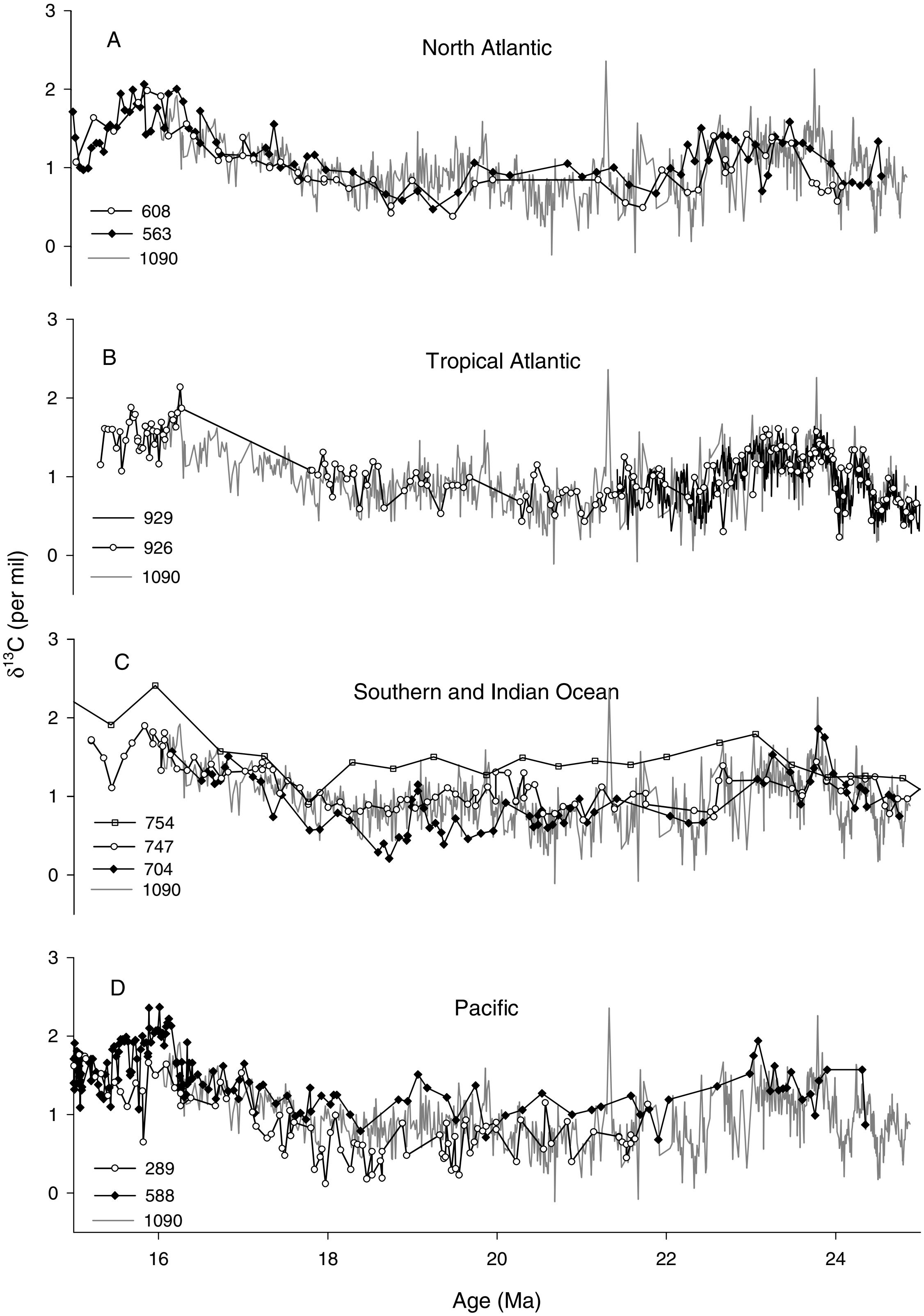

Figure 8. Carbon isotope gradients between the deep Southern Ocean (Site 1090) and (a) the North Atlantic (Sites 563 and 608), (b) western tropical Atlantic (Sites 929 and 926), (c) records from the Southern Ocean (Sites 704 and 747) and tropical Indian Ocean (site 754), and (d) two records from intermediate depths in the western equatorial Pacific (Sites 588 and 289) (Table 1). During the Oligocene/Miocene climate transition as well as during the late early Miocene, $\delta^{13}C$ gradients between the major ocean basins were small or nonexistent with the exception of a positive gradient between the Southern Ocean and the tropical Indian Ocean (Figure 8c).

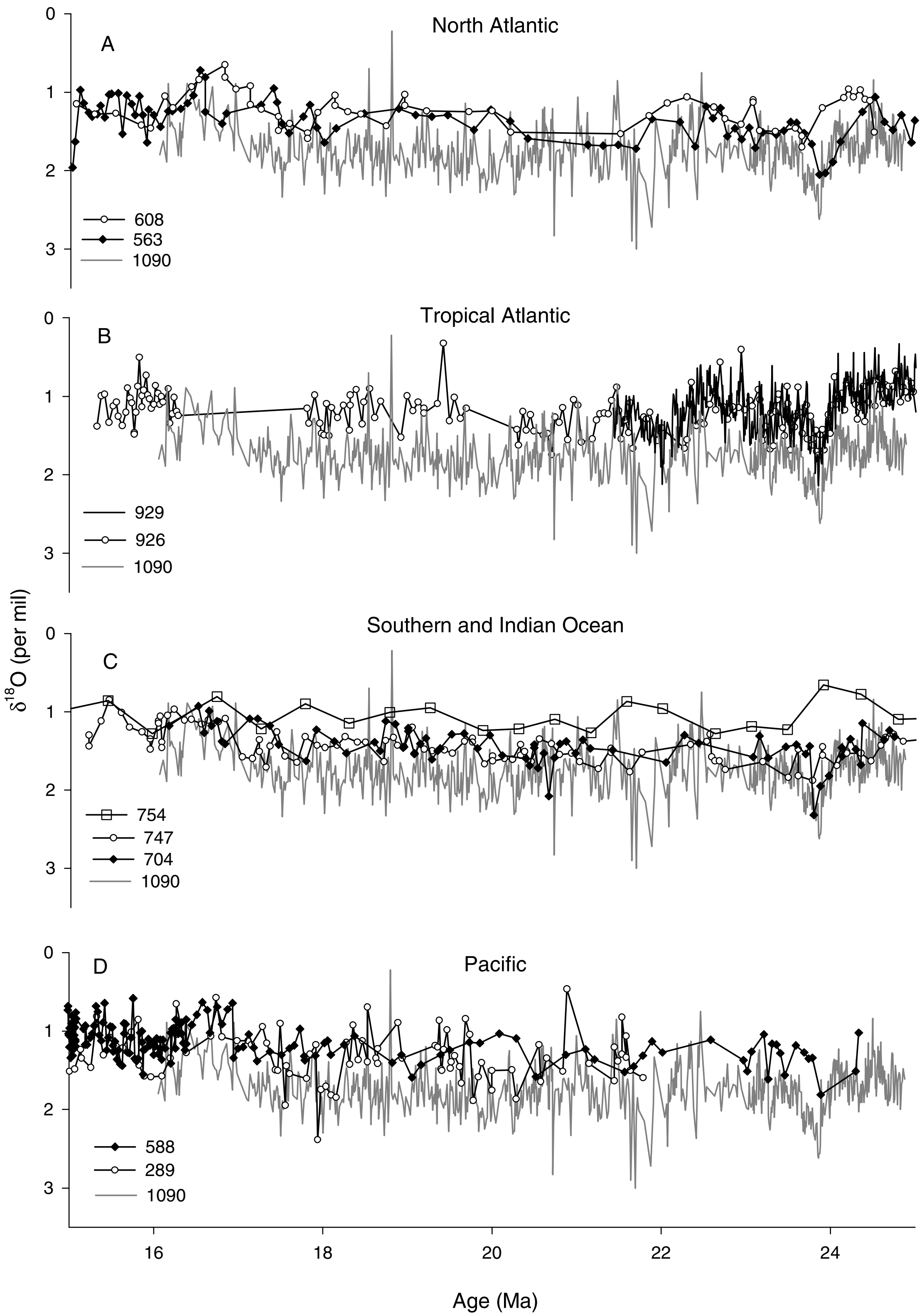

Figure 9. Oxygen isotope gradients between the deep Southern Ocean (Site 1090) and (a) the North Atlantic (Sites 563 and 608), (b) western tropical Atlantic (Sites 929 and 926), (c) records from the Southern Ocean (Sites 704 and 747) and tropical Indian Ocean, and (d) two records from intermediate depths in the western equatorial Pacific (Sites 588 and 289) (Table 1). Oxygen isotope records differ significantly from site to site, with Site 1090 displaying the highest $\delta^{18}O$ values. This suggests that the deep Southern Ocean was very cold in comparison to the deep North Atlantic (Figures 9a and 9b), the Indian, and the Pacific Oceans. The presence of $\delta^{18}O$ gradients between the North Atlantic, Indian, and Pacific suggests that warm water masses were present in these ocean basins.

Table 3. Late Oligocene Through Early Miocene (25–16 Ma) Average Low-Resolution $\delta^{13}C$ and $\delta^{18}O$ Values

Site	$\delta^{13}C$ of ΣCO_2[a]	Temperature,[a] °C	$\delta^{13}C$, ‰	$\delta^{18}O$, ‰
608	1.0–1.1	2.2	1.01	1.23
563	1.0	2.2	1.15	1.40
926	1.0	2.1	1.01	1.17
929	0.6	0.7	0.85	1.10
1090	0.4	1.3	0.94	1.74
704	0.6	1.9	0.90	1.43
747	0.5	1.0	1.05	1.47
754	N/A	4.5	1.42	1.07
588	−0.3	1.7	1.39	1.18
289	−0.3	1.7	0.74	1.33

[a] Values are from Geochemical Ocean Sections Study (GEOSECS) [*Ostlund et al.*, 1987].

mass in the tropical Indian Ocean would be consistent with the low latitude Tethyan ocean being a source of warm intermediate water as proposed by *Woodruff and Savin* [1989]. High $\delta^{13}C$ values at a water depth of ~1000 m may, however, reflect ^{13}C enrichment due to more fully equilibrated isotope exchange with atmospheric CO_2 at the sea surface, analogous to the formation of ^{13}C-enriched modern Antarctic Intermediate Water [*Charles and Fairbanks*, 1990; *Wright et al.*, 1992]. Intermittent $\delta^{13}C$ gradients between ~22 and 17 Ma may reflect changes in ambient surface water productivity superimposed on a generally low nutrient ocean. On the basis of the $\delta^{13}C$ values it is not possible to be certain of the direction of intermediate or deep water flow between the Indian and Southern Oceans.

[23] Unlike previous studies, which have relied on benthic foraminiferal $\delta^{13}C$ records, we propose that it is the presence of persistent site to site $\delta^{18}O$ gradients that provides further insight into early Miocene deep water circulation. Benthic foraminiferal $\delta^{18}O$ values in general provide a record of ice volume and local- to regional-scale deep ocean temperature and salinity changes. When comparing $\delta^{18}O$ time series from different regions in the ocean, the gradients between sites reflect temperature and salinity differences associated with the presence of different water masses, water mass stratification, and mixing. Consistent variations in the amplitude of the $\delta^{18}O$ signal among sites reflect fluctuations in global ocean ^{18}O content as brought about by the growth and decay of polar ice sheets. Benthic foraminiferal $\delta^{18}O$ gradients between the North and South Atlantic provide evidence for a source of relatively warm deep waters forming in the North Atlantic that reached the deep tropical Atlantic. Whether this water mass formed due to favorable climatic conditions in the North Atlantic (e.g., a negative precipitation versus evaporation balance) or whether it reflects a significant contribution of waters exiting the low-latitude western Tethys is not possible to say. Regardless, this NCW did not reach the South Atlantic sector of the Southern Ocean. Southern Ocean $\delta^{18}O$ values may have been high at Site 1090 because of the absence of relatively warm NCW. The presence a relatively warm deep water mass in the modern Atlantic can be seen in the relatively small north to south temperature difference (Table 3). Dominance of cold Southern Component Deep Water (SCW) at Site 1090 suggests that relatively cold circumpolar climatic conditions prevailed between 25 and 17 Ma. Between 17 and 16 Ma the site 1090 $\delta^{18}O$ values would allow for some warming of the deep Southern Ocean. Warming at Site 1090 may reflect a decrease in the relative flux of SCW associated with generally warmer high latitudes or an increase in the relative flux of other warmer water deep waters such as those of North Atlantic or perhaps Tethyan origin. In light of recent evidence for the limited role of pCO_2 in early Miocene climate change, it is likely that ocean circulation played a key role in early Miocene warming [*Pagani et al.*, 1999]. In this regard, one important implication of the relatively higher values at 1090 is that a large part of the late Oligocene $\delta^{18}O$ decrease (26 Ma, Figure 1) may reflect on a warming of low- and middle-latitude deep waters of the Atlantic rather than a reduction in ice volume.

[24] Persistently cold deep waters at Site 1090 throughout the late Oligocene and early Miocene are reminiscent of a circulation regime similar to the modern Southern Ocean with Circumpolar Deep Water, Antarctic Intermediate Water, and a strong Antarctic Circumpolar Current. The Site 1090 $\delta^{18}O$ record supports therefore those tectonic reconstructions of Southern Ocean bathymetry that call for an opening of the Drake Passage by ~26 Ma to allow the thermal isolation of Antarctica from poleward flowing warm subtropical surface water [*Barker and Burell*, 1977, 1982; *Lawver et al.*, 1994].

5. Conclusions

[25] At ODP Site 1090 (Leg 177), a high-resolution stable isotope record can be correlated through magnetostratigraphy to the GPTS for the late Oligocene though early Miocene (25–16 Ma) interval. Established oxygen isotope zones Mi1 through Mi2 as well as several other isotope events can be assigned ages based on correlation to the GPTS. Our correlation suggests that the $\delta^{18}O$ maximum commonly associated with the O/M boundary is slightly older (23.86 Ma) and falls within C6Cn.2r. The $\delta^{13}C$ maximum, which slightly lags the $\delta^{18}O$ maximum, occurs at the C6Cn.2n/r boundary (23.79 Ma) and can therefore be used to define the O/M boundary. We present an initial comparison of the GPTS-calibrated stable isotope record from Site 1090 to the orbitally tuned stable isotope record from Site 929. This comparison illustrates that very good agreement can be achieved at and below the boundary, but above the boundary, higher-frequency stable isotope variability in the two records cannot be easily correlated. With respect to Southern Ocean paleoceanography the Site 1090 stable isotope records provide the first deep ocean end-member for reconstructions of deep ocean circulation patterns and late Oligocene to early Miocene climate change. A general lack of carbon isotope gradients between ocean basins during the majority of the latest Oligocene and early Miocene is inconclusive with respect to paleocirculation changes. Benthic foraminiferal $\delta^{18}O$ gradients between the North and South Atlantic provide evidence for a source of relatively warm deep waters forming in the North Atlantic that reached the deep tropical Atlantic. NCW did not reach the South Atlantic sector of the Southern Ocean. Dominance of cold Southern Component Deep Water at Site 1090 suggests that relatively cold circumpolar climatic conditions prevailed between 25 and 17 Ma and may reflect a circulation regime similar to the modern Southern Ocean with a well developed Antarctic Circumpolar Current. If correct, the Site 1090 $\delta^{18}O$ record supports those tectonic reconstructions of Southern Ocean bathymetry that call for unrestricted flow through the Drake Passage by ~26 Ma.

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