

17. OXYGEN AND CARBON ISOTOPE STRATIGRAPHY OF THE MIDDLE MIocene, HOLES 805B AND 806B¹

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ABSTRACT

Bulk carbon isotope records are an effective chemostratigraphic tool for the middle Miocene because of the large and systematic variation in first-order $\delta^{13}\text{C}$ signals. Bulk $\delta^{13}\text{C}$ measurements support the presence of a hiatus at 305 mbsf in Hole 805B (latest middle Miocene), provisionally located while on board ship using biostratigraphic and magnetostratigraphic events. Records at Holes 805B and 806B show the middle Miocene Monterey carbon isotope excursion although the record at Hole 806B is apparently more stratigraphically continuous. Detailed analysis of multispecies foraminiferal carbon isotope records during the middle Miocene ("Monterey excursion") segment at Hole 806B support the assertion that this carbon isotope excursion comprises mainly between-reservoir effects. The benthic $\delta^{18}\text{O}$ data increase after 15.3 Ma, which we suggest corresponds to the mid-Miocene cooling step/ice volume increase of other authors. Planktonic foraminiferal $\delta^{18}\text{O}$ evidence exists for steepening of the thermocline at 17.4 Ma. A second-order $\delta^{13}\text{C}$ excursion superimposed at 13.8 Ma on the first-order Monterey excursion is associated with a second-order negative $\delta^{18}\text{O}$ excursion.

INTRODUCTION

One of the most important achievements of Ocean Drilling Program (ODP) Leg 130 was the recovery of thick middle Miocene carbonate oozes and chalks drilled using the advanced hydraulic piston corer (APC), especially at Holes 805B and 806B (Fig. 1). Sedimentation rates at these sites were relatively high (12–13 m/m.y. at Hole 805B and between 20 and 30 m/m.y. at Hole 806B, as documented by the Shipboard Scientific Party), and this permits the generation of high-resolution $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records from the western equatorial Pacific through the middle Miocene. From the perspective of $\delta^{18}\text{O}$ reconstruction of marine paleoclimates, this is an important region because the thermal stability of the surface waters (Matthews and Poore, 1980; Prentice and Matthews, 1988; Ravelo et al., 1990) may permit further constraints on the relative contribution of ice-volume and temperature change to the marine carbonate $\delta^{18}\text{O}$ signal (Miller et al., 1991).

Several previous studies have focused on this important interval at other sites. Shackleton and Kennett (1975) and Savin et al. (1975) were the first to identify the profound $\delta^{18}\text{O}$ increase in benthic foraminifers that has been associated with the onset of Antarctic glaciation (Shackleton and Kennett, 1975; Savin et al., 1975, 1981; Woodruff et al., 1981), although the significance of this change in terms of ice-volume vs. temperature contribution to the $\delta^{18}\text{O}$ signal remains obscure (Matthews and Poore, 1980; see discussion in Miller et al., 1991). Woodruff et al. (1981) analyzed the benthic $\delta^{18}\text{O}$ record in detail at Deep Sea Drilling Project (DSDP) Site 289 (close to Hole 806B; Fig. 1) and concluded that the onset of major Antarctic glaciation occurred between 14.8 and 14.0 Ma. Shackleton (1982) examined the planktonic record at Site 289 and concluded that the middle Miocene stable isotope record offered the potential to examine climatic variability on time scales similar to that at present possible in the Pleistocene.

In a more recent series of studies, several authors (Savin et al., 1985; Barrera et al., 1985; Vincent et al., 1985; Miller et al., 1989, 1991; Wright et al., 1992) have identified this mid-Miocene $\delta^{18}\text{O}$ enrichment in deep waters at a number of sites in the Atlantic, Indian,

and Pacific oceans. However, estimates of the precise timing of this cooling episode vary. Woodruff and Savin (1989) developed a model that identifies a profound change in the nature of deep-water circulation during this interval. They hypothesize that the early Miocene was a time when deep-water circulation was dominated by Tethyan outflow but that following the mid-Miocene $\delta^{18}\text{O}$ enrichment the emphasis shifted to the production of cool, dense waters in the Antarctic region. Wright et al. (1992) used $\delta^{13}\text{C}$ variations in benthic foraminifers to monitor deep-water movements during the early and middle Miocene and concluded that two modes of deep-water formation were present during this interval. Between 24–20 Ma and 16–12.5 Ma, the deep oceans were ventilated by Southern Component Water (SCW), which is analogous to Antarctic Bottom Water (AABW); however, during other intervals, deep-water formation was from multiple sources with relatively warm Northern Component Water (NCW; analogous to North Atlantic Deep Water [NADW]) and Tethyan water masses (as suggested by Woodruff and Savin, 1989), supplementing SCW production between 20 and 16 Ma and NCW and SCW production dominating between 12.5 and 10 Ma. The combination of NCW and Tethyan water mass production between 20 and 16 Ma warmed the deep ocean by 3°–4°; after 16 Ma, deep-water temperatures cooled and have remained relatively cold ever since.

Stratigraphic reference sections for the middle Miocene interval based on stable isotopic changes in foraminifers have also been established. Loutit et al. (1983) subdivided the middle Miocene on the basis of $\delta^{13}\text{C}$ inflections and claimed that this improved the precision of stratigraphic correlation of events to ± 105 yr. Miller et al. (1991) and Wright and Miller (1992) defined eight oxygen isotope zones (their "Mi" zones), based on $\delta^{18}\text{O}$ inflections in benthic foraminifers that they assert correlate with glacioeustatic sea-level changes.

Because ice-volume fluctuations, temperature effects, and salinity-correlated effects influence oxygen isotope fractionation, $\delta^{18}\text{O}$ records from foraminifers cannot be uniquely interpreted in terms of a single variable (see discussion in Miller et al., 1991). As a consequence of this, a considerable body of work has focused on quantifying the different contributions of these three variables to Cenozoic (particularly Pleistocene) $\delta^{18}\text{O}$ records. The usual approach to estimate the ice volume/temperature effect contribution is to compare planktonic foraminiferal records from tropical or subtropical regions and benthic foraminiferal $\delta^{18}\text{O}$ records. This strategy is based on the assumption that ice-volume growth or decay will affect both benthic

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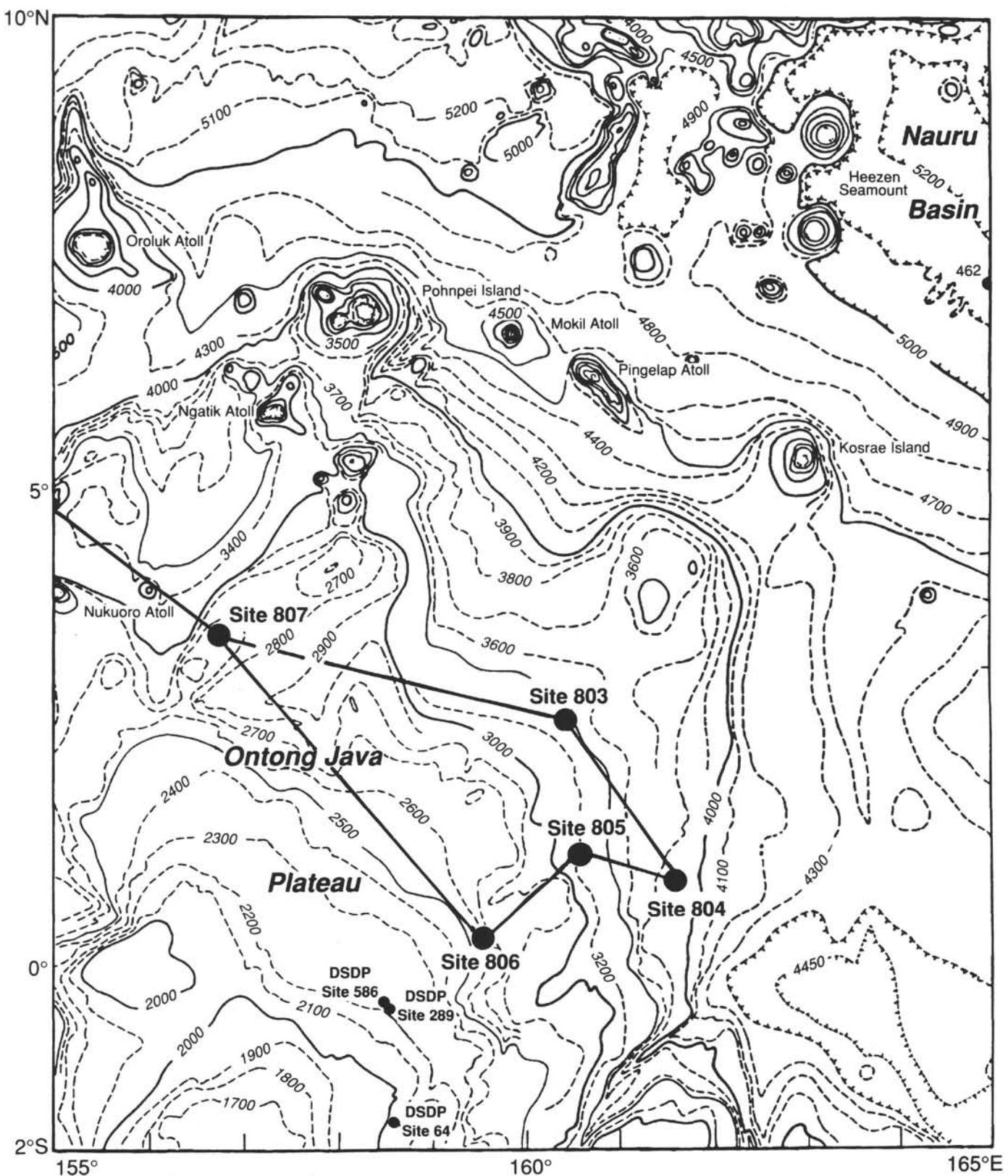


Figure 1. Location map of Leg 130 sites, including Holes 805B and 806B.

and planktonic records whereas temperature change will register only in planktonic records (e.g., Shackleton, 1967; Shackleton and Opdyke, 1973, 1976; Fairbanks and Matthews, 1978). Miller et al. (1991) advocate a similar approach further back in the Cenozoic, but they point out that the western equatorial regions offer the opportunity to obtain $\delta^{18}\text{O}$ records from planktonic foraminifers that are less likely to be affected by temperature variations than other regions because of their thermal stability (Ravelo et al., 1990).

Vincent and Berger (1985) developed a significant model that linked the middle Miocene $\delta^{18}\text{O}$ enrichment in benthic $\delta^{18}\text{O}$ records to the isotopically positive $\delta^{13}\text{C}$ values that typify middle Miocene foraminifers. They named this positive $\delta^{13}\text{C}$ interval the "Monterey excursion."

Our aim in this contribution is to develop an oxygen and carbon isotope stratigraphy for Holes 805B and 806B in the interval between 12 and 20 Ma. We wish to investigate the isotopically positive $\delta^{13}\text{C}$ values that characterize the early to middle Miocene interval and also to investigate their possible climatic consequences in the western equatorial Pacific. In conducting this preliminary contribution, our data were collected at a sampling resolution of one per core section (Appendices A and B). This work continues at a higher sampling resolution.

METHODS

The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ stratigraphies generated by the analysis of bulk samples were used to identify discontinuities that might be caused by the presence of hiatuses. In addition, we picked up to five species of planktonic foraminifers and four species of benthic foraminifers for study. However, because our focus is on the changing vertical gradients of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ across this time interval, we restrict the present discussion to data derived from a surface-water-dwelling planktonic foraminiferal species (*Dentoglobigerina altispira*), a deeper dwelling planktonic foraminiferal species (*Globoquadrina venezuelana*), and benthic foraminifers (usually *Cibicidoides* spp. and *Planulina wuellerstorfi*). The benthic foraminifer data were corrected for isotopic disequilibrium effects using the correction factors of Shackleton et al. (1984).

Foraminiferal samples were disaggregated in distilled water. Between 5 and 15 planktonic specimens were picked from the 300–355- μm size fraction to minimize ontogenetic effects. The number of benthic specimens picked from this size fraction was generally lower as a result of their scarcity within the samples. Samples were cleaned using H_2O_2 and $(\text{CH}_3)_2\text{CO}$ and then dried for 30 min at 60°C. Samples were analyzed isotopically using a VG Isotech PRISM mass spectrometer in the Oxford Laboratory calibrated to NBS 19 and Cambridge Carrara marble.

The lower portion of the interval studied at Hole 806B in particular was relatively deeply buried (300–600 mbsf). The interval studied at Hole 805B was shallower (290–390 mbsf). In sections more deeply buried than about 400 m, diagenetic alteration of stable isotope ratios becomes an important consideration (Miller and Curry, 1982; Barrera et al., 1987; Wright et al., 1992). Clearly, therefore, the possibility of diagenetic overprinting must be investigated in the deeper portions of Hole 806B. This is especially important because the adjacent DSDP Site 289 has a similar burial depth and has been used as a reference section for this interval in the western equatorial Pacific (Shackleton, 1982; Wright et al., 1991, 1992). Other authors have suggested that the $\delta^{18}\text{O}$ signal may be degraded because of diagenetic overprinting (Elderfield et al., 1982; Miller et al., 1991). We have followed the approach used by Wright et al. (1992) and adopted two criteria for the identification of diagenesis: (1) optical examination for calcite overgrowths and (2) similarity to other, shallower records. With respect to the first criterion, sediments disaggregated easily, suggesting only minimal cementation and significant calcite overgrowth was not observed. Comparison of the $\delta^{18}\text{O}$ record from Hole 806B with other records is discussed below. In general, however, the major features of

middle Miocene $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ history are well represented in Holes 805B and 806B, and we conclude therefore that diagenetic overprinting is not a major problem.

RESULTS AND DISCUSSION

Bulk Isotope Stratigraphy

Figures 2 and 3 illustrate the middle Miocene bulk isotope stratigraphy at Holes 805B and 806B. Both Figures 2 and 3 exhibit $\delta^{13}\text{C}$ values that reach broad maxima in the middle of the sections studied. At Hole 805B, these values are most positive between Cores 130–805B-35X (325 mbsf) and -38X (352 mbsf). At Hole 806B, a particularly positive second-order $\delta^{13}\text{C}$ excursion (marked "A" on Fig. 3) is discernible between Cores 130–806B-50X (452 mbsf) and -52X (470 mbsf), whereas values in general remain high between Cores 130–806B-53X (490 mbsf) and -55X (530 mbsf). We interpret this positive first-order feature to be the local expression of the middle Miocene $\delta^{13}\text{C}$ excursion documented by Vincent and Berger (1985). The second-order excursion may correlate with one of the benthic $\delta^{13}\text{C}$ events of Loutit et al. (1983). Pending further examination, we have tentatively identified this as the "Zone N11 subevent."

Detailed analysis of Hole 805B suggests that the middle Miocene portion of Hole 805B is less continuous than the equivalent section at Hole 806B. In particular, we note steps in the $\delta^{13}\text{C}$ signal within Core 130–805B-33X (between Samples 130–805B-33X-3, 139–141 cm, and -33X-4, 9–11 cm; marked "A" on Fig. 2) that correspond to a hiatus as identified by the shipboard biostratigraphers. Breaks in sedimentation may exist between Samples 130–805B-35X-3, 99–101 cm, and -35X-4, 9–11 cm (marked "B" on Fig. 2) and between Samples 130–805B-37X-4, 69–71 cm, and -38X-1, 59–61 cm (marked "C" on Fig. 2). Judging from the positive $\delta^{13}\text{C}$ values immediately above our suggested stratigraphic break at point "B," we suggest that the second-order $\delta^{13}\text{C}$ excursion, which we have termed the "Zone N11 subevent" (marked "A" on Fig. 3), is probably partially missing in Hole 805B.

It is clear that the bulk carbonate records of the first-order middle Miocene $\delta^{13}\text{C}$ excursion provide a sufficiently systematic signal of high signal-to-noise ratio to serve as a high-resolution stratigraphic tool. In this it is comparable to the large amplitude of the $\delta^{13}\text{C}$ signal in the Paleocene, which similarly provides high-resolution chemostratigraphic control as discussed by Shackleton et al. (1985).

In contrast, the bulk $\delta^{18}\text{O}$ records at Holes 805B and 806B do not show systematic change over the whole of the middle Miocene interval. However, systematic changes on shorter time scales are visible. Hole 805B shows a trend toward positive $\delta^{18}\text{O}$ values between Cores 130–805B-34X and -33X (marked "D" on Fig. 2), and Hole 806B shows the same trend between Core 130–805B-48X and -45X (marked "B" on Fig. 3). This probably corresponds to the ^{18}O enrichment associated with the cooling/ice-volume increase effect discussed in the introduction and well known from the benthic records at several sites (e.g., Miller et al., 1991). In general, however, the greater amplitude of the high-frequency variation in the bulk carbonate $\delta^{18}\text{O}$ signal in the middle Miocene sequence at Holes 805B and 806B probably precludes the same stratigraphic usefulness as the $\delta^{13}\text{C}$ signal.

A possible exception to this is the pronounced positive excursion between Cores 130–806B-50X (452 mbsf) and -52X (470 mbsf) (marked "C" in Fig. 3), which is contemporaneous with the maximum noted in the d13C record. These positive $\delta^{18}\text{O}$ values are missing from Hole 805B, which supports our assertion that this portion of the succession is missing in Hole 805B.

Based on our assessment of the stratigraphic continuity of Holes 805B and 806B, we have concentrated this initial investigation on the apparently more complete record of Hole 806B. We intend to combine records from Holes 805B and 806B as our contribution to the high-resolution Neogene stratigraphy project of the Leg 130 Shipboard Scientific Party.

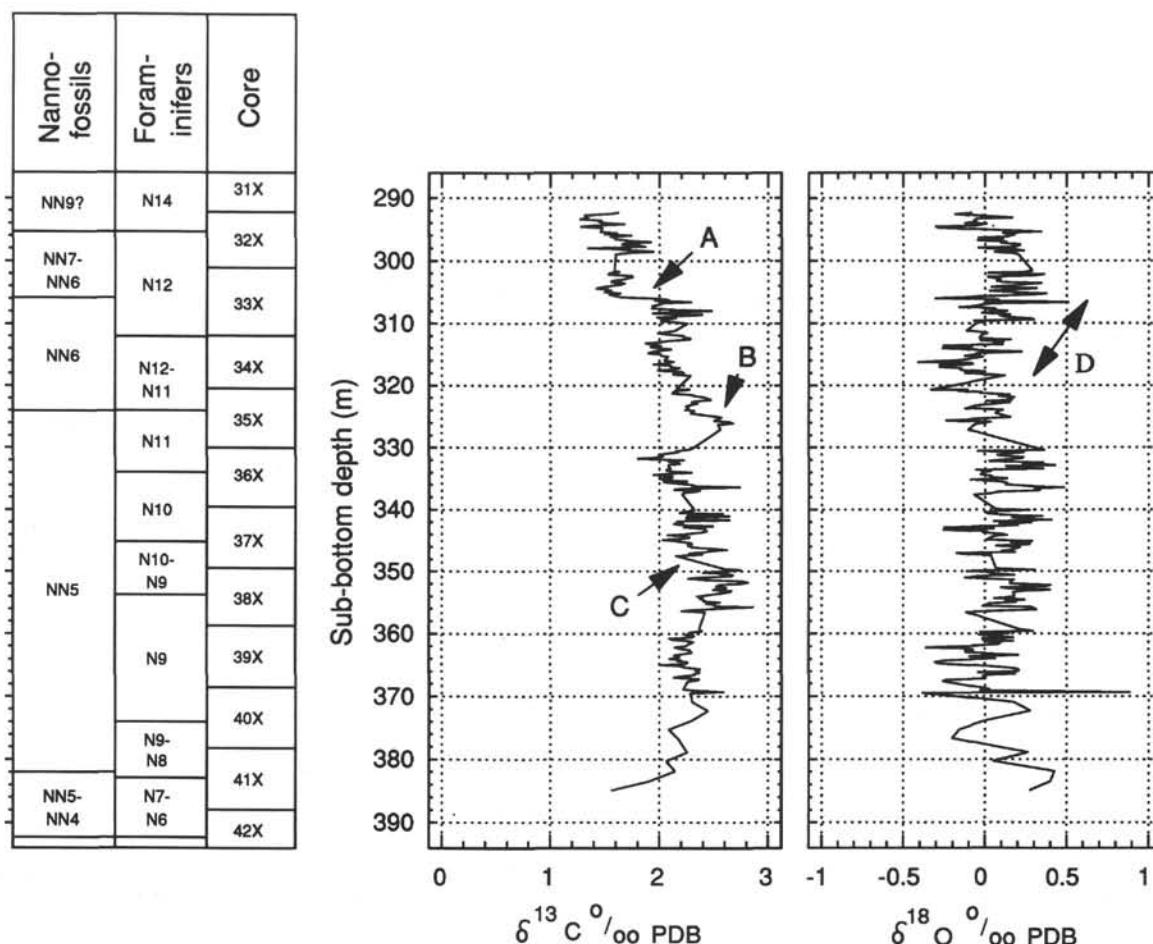


Figure 2. Bulk sample oxygen and carbon isotope stratigraphy, Hole 805B. A = probable hiatus in succession also noted by shipboard bio- and magnetostratigraphers, B and C = possible hiatuses within Monterey carbon isotope excursion, D = positive $\delta^{18}\text{O}$ excursion.

Foraminiferal Isotope Stratigraphy

Figure 4 illustrates the data set for Hole 806B. The Monterey $\delta^{13}\text{C}$ excursion is clearly expressed in all three groups of foraminifers. In particular, the second-order positive excursion between Cores 130-806B-50X (452 mbsf) and -52X is clearly depicted in *D. altispira*, *G. venezuelana*, and the benthic foraminifers. Examination of the $\delta^{18}\text{O}$ data shows that this second-order event is contemporaneous with a marked positive excursion in both sub-surface dwellers (*G. venezuelana*) and benthic foraminifers.

The $\delta^{18}\text{O}$ data for all three groups of foraminifers trend toward negative values between 600 and 550 mbsf, which we interpret as the early Miocene warming trend noted by Vincent and Berger (1985). Within Core 130-806B-58X (550 mbsf), separation of the *D. altispira* and *G. venezuelana* $\delta^{18}\text{O}$ signals occurs. The $\delta^{13}\text{C}$ signals for the same species separate at about the same level, and we interpret this as the steepening of the thermocline and probably the sub-surface oxygen minimum as suggested by Vincent and Berger (1985). Within Core 130-806B-54X, benthic $\delta^{18}\text{O}$ values start to trend toward more positive values, which we interpret as the cooling/ice-volume increase associated with the buildup of Antarctic ice.

We developed a chronology for our samples from Hole 806B using the biostratigraphic datums identified by Takayama (this volume). We used the time scale of Berggren et al. (1985) and calculated sample ages by linear interpolation between age control points. To reduce scatter in the data over this relatively long interval, we stacked the data into 0.1-m.y. increments of modeled time. Figure 5 shows the bulk data when processed by this method, and Figure 6 illustrates the

foraminiferal data. A problem with Hole 806B is the lack of detailed magnetostratigraphy. Hence, our chronology, which is based solely at present on the nannofossil zonation of Takayama (this volume), must be regarded as preliminary.

Figure 5 shows clearly the start of the Monterey excursion at 18.75 Ma. Values increase to 16.8 Ma and then stabilize at about 2.3‰ before declining again at 14.6 Ma. Subsequently, $\delta^{13}\text{C}$ values peak to values of 2.7‰ at 13.8 Ma. This is the expression in the bulk sediment of our foraminiferal "Zone N11 subevent." The $\delta^{18}\text{O}$ values remain rather constant throughout the succession with the exception of a positive inflection between 14.3 and 13.4 Ma, which is contemporaneous with the "Zone N11 subevent."

Figure 6 illustrates the foraminiferal data chronologically. The possible steepening of the thermocline discussed previously occurs at 17.4 Ma. Our data suggest that the onset of the increasing $\delta^{13}\text{C}$ values characteristic of the Monterey excursion occur in planktonic foraminifers at about 18.8 Ma. This is effectively contemporaneous with the onset of the increase in bulk sediment $\delta^{13}\text{C}$ values. However, our benthic $\delta^{13}\text{C}$ data do not start to increase until about 17.9 Ma; this lag is likely to be an artifact of our low benthic sampling density.

Vincent and Berger (1985) inferred on the basis of covarying foraminiferal $\delta^{13}\text{C}$ data from DSDP Site 216 that the middle Miocene carbon isotope maximum was composed largely of between-reservoir carbon isotope fractionation effects. In contrast, Corfield and Cartlidge (1992) noted that the Paleocene carbon isotope maximum was comprised of both between-reservoir (burial of organic carbon) and within-reservoir (surface-water productivity) change. We have assessed the intensity of the vertical carbon isotope gradient during the

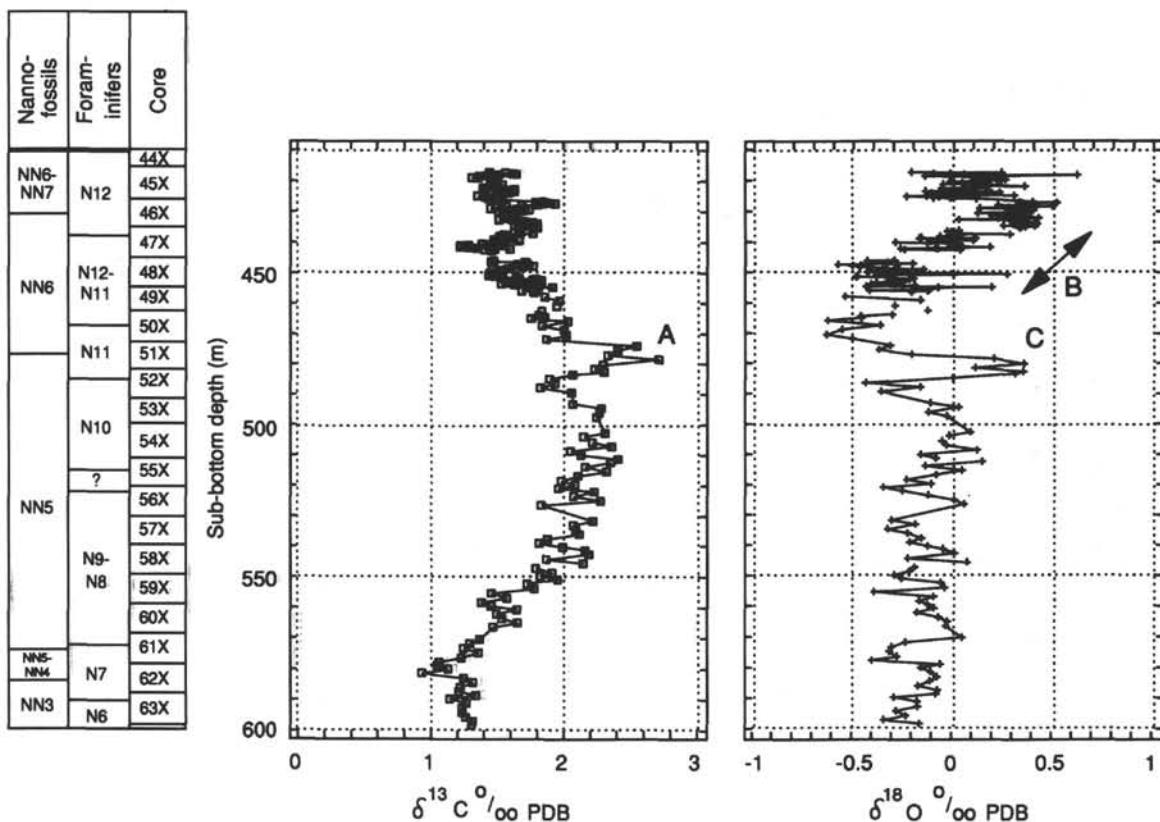


Figure 3. Bulk sample oxygen and carbon isotope stratigraphy, Hole 806B. A = "Zone N11 $\delta^{13}\text{C}$ subevent" discussed in the text, B = positive $\delta^{18}\text{O}$ excursion, C = "Zone N11 $\delta^{18}\text{O}$ subevent" discussed in the text.

middle Miocene interval at Hole 806B ($\Delta\delta^{13}\text{C}$) by subtracting benthic $\delta^{13}\text{C}$ values from *D. altispira* values ($\delta^{13}\text{C}_{\text{alt-ben}}$) as well as subtracting *G. venezuelana* values from *D. altispira* ($\delta^{13}\text{C}_{\text{alt-ven}}$) values. The results are plotted as a function of bulk $\delta^{13}\text{C}$ (Fig. 7). This plot suggests a weak increase in $\Delta\delta^{13}\text{C}_{\text{alt-ven}}$, whereas no relationship exists between bulk $\delta^{13}\text{C}$ values and $\Delta\delta^{13}\text{C}_{\text{alt-ben}}$. This suggests that a minor increase in surface-water productivity may have been associated with the Monterey excursion but that the bulk of the increase in $\delta^{13}\text{C}$ values was caused by the lithospheric sequestration of isotopically light carbon. This finding is consistent with the hypothesis of Vincent and Berger (1985) that the Monterey carbon isotope excursion was caused by an increase in the rate of burial of organic carbon.

Vincent and Berger (1985) and Berger and Vincent (1986) related the middle Miocene increase in benthic $\delta^{18}\text{O}$ values found globally to the increase in $\delta^{13}\text{C}$ values in surface, deeper water, and benthic foraminifers. The mechanism they propose is that an increase in the rate of burial of organic carbon in the sediments rimming the Pacific Basin (the Monterey Shale and time-equivalent organic-rich strata) cause re-equilibration of CO_2 between the ocean and the atmosphere, with the result that atmospheric pCO_2 decreased and triggered polar cooling. Our data broadly support their speculation although we note that explicitly quantitative modeling experiments are required to verify their hypothesis. Our estimates of the chronology (subject to the reservations outlined above) of these events are as follows:

1. The $\delta^{13}\text{C}$ values in bulk sediment, as well as surface and deeper dwelling planktonic foraminifers, begin to increase at 18.8 Ma. The $\delta^{13}\text{C}$ increase in benthic foraminifers begins at 17.9 Ma; however, this lag may be a function of our low benthic sampling density. The

increase in $\delta^{13}\text{C}$ in these three groups of foraminifers may indicate the onset of organic carbon burial in Pacific Basin sediments.

2. The $\delta^{18}\text{O}$ records of surface and deeper dwelling planktonic foraminifers begin to diverge at 17.4 Ma, suggesting a steepening in the thermocline.

3. The benthic $\delta^{18}\text{O}$ record begins to increase at 15.3 Ma.

4. By the early/middle Miocene boundary (16.4 Ma), the thermocline is at its steepest in our record and the broad maximum of the Monterey carbon excursion has been reached.

5. The "Zone N11 subevent" occurs between 14.1 and 13.2 Ma. The $\delta^{13}\text{C}$ values peak to their most positive Miocene values in all three foraminiferal groups. Contemporaneous deeper dwelling planktonic and benthic $\delta^{18}\text{O}$ values increase and then recover. This positive $\delta^{18}\text{O}$ inflection in the benthic curve may be the expression in the record from Hole 806B of oxygen isotope Zone Mi3 of Miller et al. (1991). The $\delta^{13}\text{C}$ excursion with which it is apparently associated has been identified in other sites from benthic $\delta^{13}\text{C}$ data (e.g., DSDP Sites 289 and 563; see fig. 14 in Wright et al., 1992), although this $\delta^{13}\text{C}$ event has apparently not been explicitly named. Our data from Hole 806B suggests that our "Zone N11 subevent" affects all foraminiferal groups examined in this study and therefore all levels in the water column. If, as Miller et al. (1991) assert, oxygen isotope Zone Mi3 is associated with glacioeustatic lowering, this is not surprising. However, the reasons behind the apparently synchronous $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ change during this event need further examination.

Finally, our planktonic and benthic foraminiferal data from this western equatorial site may contribute some additional constraints to the discussion concerning the relative contributions of ice-volume

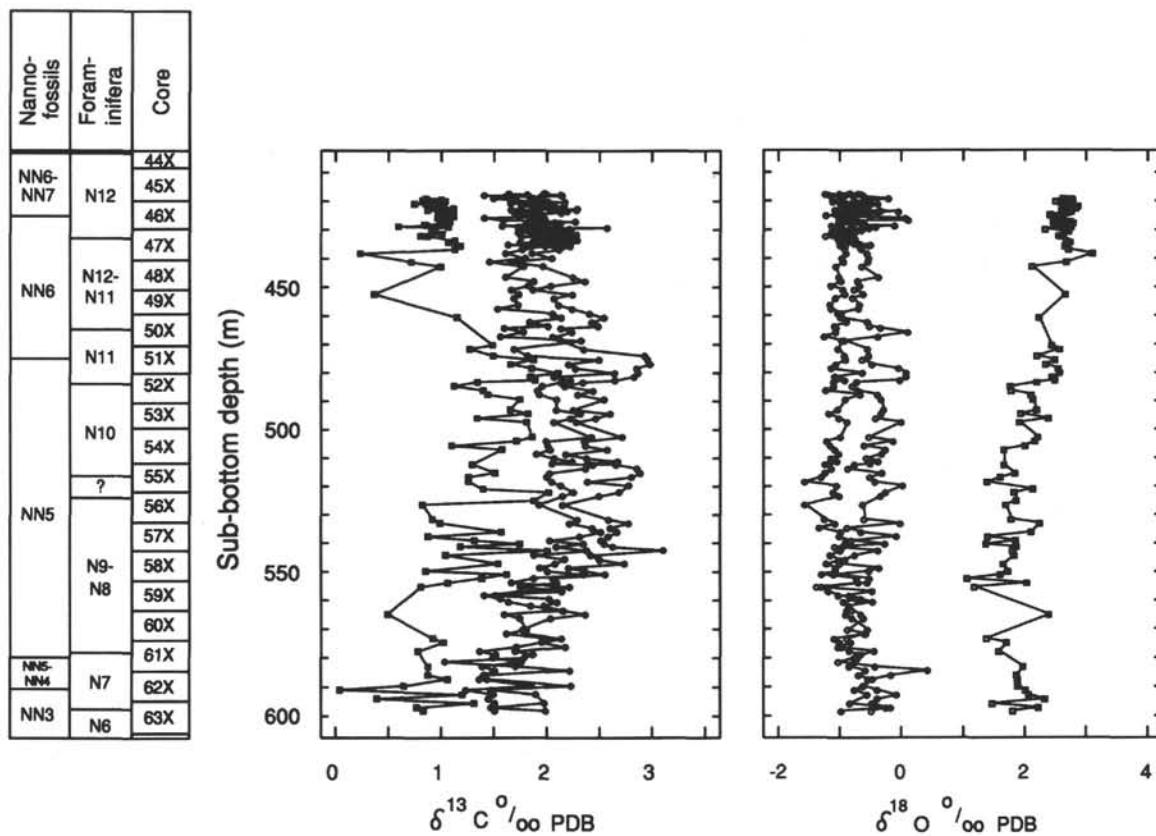


Figure 4. Foraminiferal $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ stratigraphy, Hole 806B. The Monterey carbon isotope excursion is clearly expressed. Note also the steepening of the thermocline indicated by both $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ data near 550 mbsf. The benthic $\delta^{18}\text{O}$ signal begins to decline after about 510 mbsf. The "Zone N11 subevent" is clearly shown in all $\delta^{13}\text{C}$ water mass monitors and is well expressed in both the deeper dwelling planktonic and benthic $\delta^{18}\text{O}$ record. Solid circles = *Dentoglobigerina altispira*, open circles = *Globoquadrina venezuelana*, and squares = benthic data.

increase and temperature decrease to the middle Miocene ^{18}O enrichment. As Figure 6 shows, over the interval between 15.3 Ma (our estimate of the onset of the middle Miocene ^{18}O enrichment) and the top of the measured section at 12.4 Ma, mean benthic $\delta^{18}\text{O}$ data increase by 0.8‰ whereas surface planktonic foraminiferal data increase by 0.4‰. Miller et al. (in press) have compared data from Holes 563 (North Atlantic) and 707 (western equatorial Indian Ocean) and also noted that the planktonic $\delta^{18}\text{O}$ change is smaller than the benthic $\delta^{18}\text{O}$ change. Their estimates are that benthic $\delta^{18}\text{O}$ increased by 0.7‰ and planktonic $\delta^{18}\text{O}$ increased by 0.4‰ over the middle Miocene ^{18}O "shift." Their estimates, therefore, are similar to ours, although ours are made from the same site in a western equatorial location and thus avoid problems of comparing distant records.

Miller et al. (in press) point out that this well-known difference in amplitude has been attributed either (1) to warming of equatorial surface waters during the growth of the Antarctic Ice Sheet (Shackleton and Kennett, 1975; Savin et al., 1975, 1985) or (2) to cooling of deep waters (Matthews and Poore, 1980; Prentice and Matthews, 1988) in addition to the ^{18}O enrichment from the growth of the Antarctic Ice Sheet.

Miller et al. (in press) suggest that the surface-water warming hypothesis is unreasonable because of planktonic and benthic foraminiferal covariance subsequent to 12.5 Ma that implies larger ice volumes on Antarctica than the scarcity of glaciomarine sediments supports (Miller et al., 1991). They therefore conclude that the change in benthic foraminiferal values is attributable to a combination of ice-volume increase and a cooling of deep waters by 1°–4°C. Our data from Hole 806B do not contradict this suggestion.

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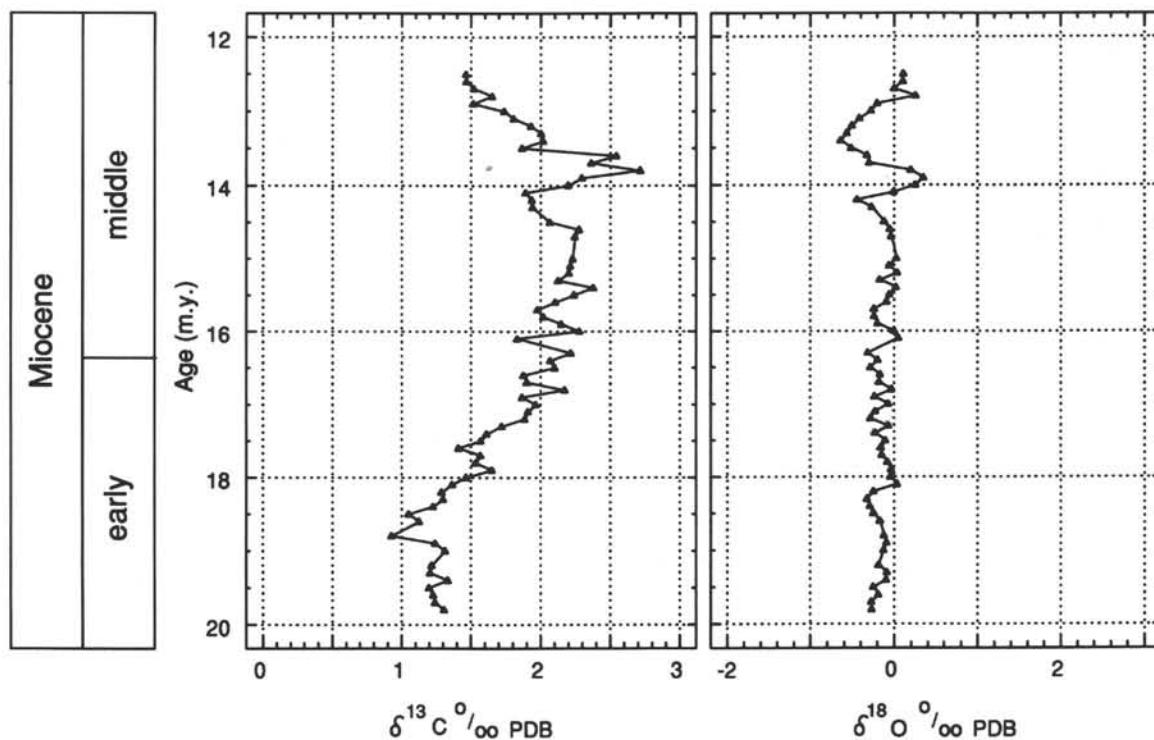


Figure 5. Chronostratigraphy of bulk sample $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ measurements, Hole 806B. These $\delta^{13}\text{C}$ data show the onset of the middle Miocene Monterey excursion at 18.8 Ma and its decline following the "Zone N11 subevent" at 14.5 Ma. The $\delta^{18}\text{O}$ data are more monotonous except for the increase in $\delta^{18}\text{O}$ values during the "Zone N11 subevent."

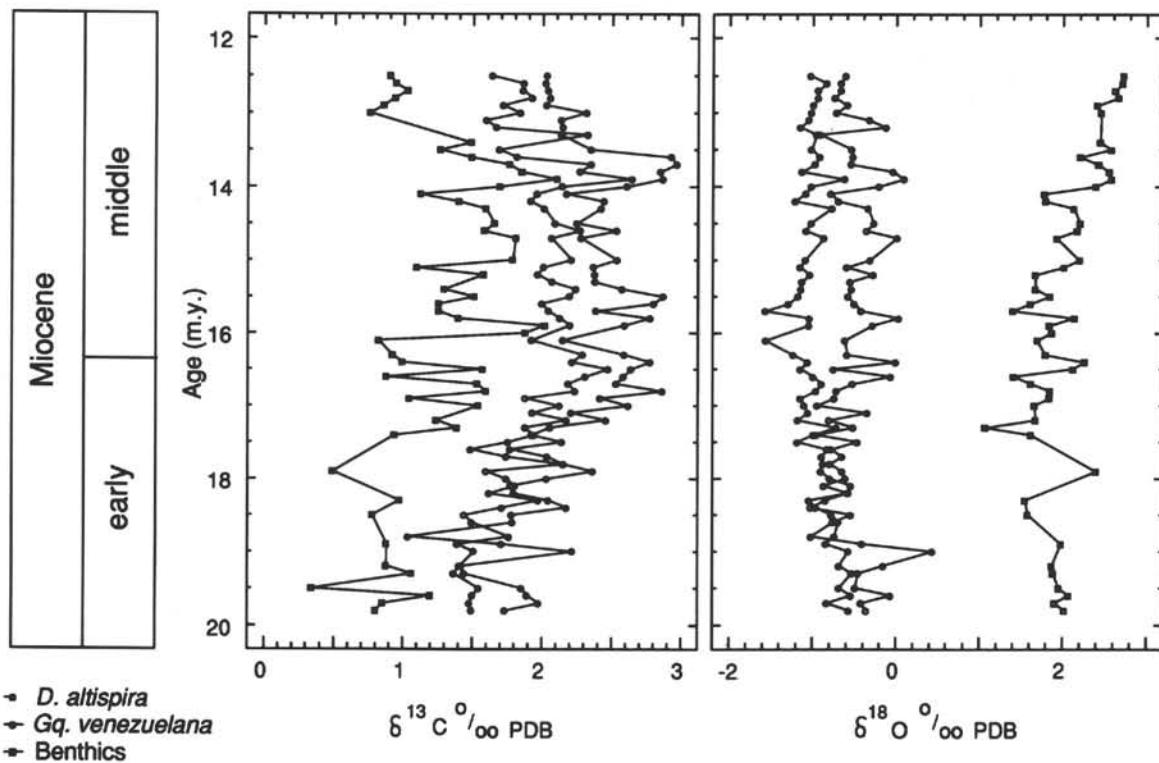


Figure 6. Chronostratigraphy of foraminiferal $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ measurements, Hole 806B. The chronology of events is discussed in the text. Note the increase in $\delta^{18}\text{O}$ values starting at 15.3 Ma. With the exception of the data set discussed by Shackleton et al. (1984), our estimate of the onset of the $\delta^{18}\text{O}$ increase is 1–2 m.y. older than other published estimates (e.g., Barrera et al., 1985), although our estimates are provisional (as discussed in the text) because they are based solely on nanofossil age control points. Symbols as for Figure 4.

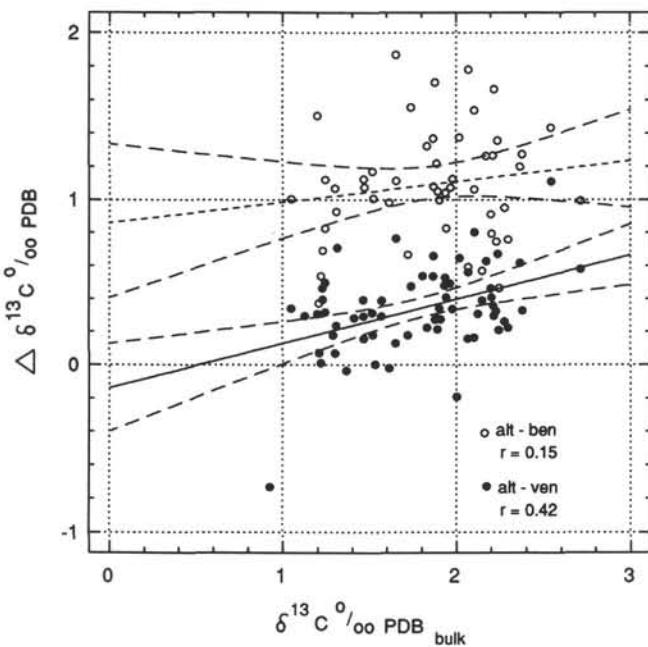


Figure 7. Two estimates of the vertical carbon isotope gradient in the middle Miocene Pacific Ocean. The increase in $\Delta\delta^{13}\text{C}_{\text{alt-ven}}$ suggests a minor contribution from surface-water productivity to the structure of the Monterey carbon isotope excursion. However, $\Delta\delta^{13}\text{C}_{\text{alt-ben}}$ are not even weakly correlated, suggesting that the major component of the Monterey carbon isotope excursion is caused by between-reservoir fractionation of carbon isotopes.

