Planktonic foraminiferal depth habitat and $\delta^{18}O$ calibrations: Plankton tow results from the Atlantic sector of the Southern Ocean

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Plankton tows conducted in the Atlantic sector of the Southern Ocean allow analysis of the influence of water column structure on planktonic foraminiferal abundance and $\delta^{18}O$ composition. Foraminiferal abundance varies by several orders of magnitude across a large gradient in sea surface temperature and other hydrographic features, demonstrating high sensitivity of foraminiferal populations to regional differences in water properties. The depth of maximum abundance for key species such as *Globigerina bulloides* and *Neogloboquadrina pachyderma* is not constant from station to station. The pattern suggests that their abundance and shell chemistry are tied to density horizons or other conditions (such as food availability) that become more sharply defined with depth in the northern subantarctic. The consistent observation of *Globorotalia inflata* and *G. truncatulinoides* as relatively deep-dwelling species confirms their utility as indicators of upper thermocline processes, demonstrating high sensitivity of foraminiferal populations to regional differences in water properties. In studies, despite evidence that this does not hold for most homogeneous and is acquired at one time and place in the sedimentary populations of planktonic foraminifera is greatly simplified if we assume that the chemistry of dimensional sediment core record. Of course, this problem is among the more direct means of determining how these processes. The advantage of a plankton tow system, such as the multiple opening closing net and environmental sampling system (MOCNESS) [Wiebe et al., 1976, 1985], is that it collects material on a depth-discrete basis at a fixed sampling time and location. This approach allows for real-time observation of the depth-distribution of foraminiferal abundances and isotopic compositions for comparison to simultaneously collected data such as sea surface temperature (SST), salinity, and fluorescence (a proxy for chlorophyll a content) [Schreiber et al., 1998].

A variety of studies have documented the vertical distribution of planktonic foraminifera in the tropics [e.g., Fairbanks et al., 1980, 1982; Ravelo et al., 1990; Ravelo and Fairbanks, 1992; Watkins et al., 1996, 1998], yet there is an equal need to calibrate the paleoceanographic proxies derived from planktonic foraminifera in higher-latitude regions. This is especially true given that foraminifera represent one of the few means for retrospective monitoring of the processes of deep-water formation and heat exchange between the high-latitude ocean and the atmosphere. Core-top calibration approaches reveal a few basic properties of modern high-latitude foraminiferal assemblages, but generally, they cannot discriminate the mechanisms through which sedimentary signatures are acquired. For example, *Globigerina bulloides*, one of the principal species studied at middle-to-high latitudes, follows the predicted $\delta^{18}O$ of calcite for North Atlantic surface waters only over a narrow range of about 1‰ (from 1 to 2‰); outside of this range, *G. bulloides* core-top $\delta^{18}O$ does not reflect mean surface conditions [Bard et al., 1989]. The implication is that either seasonal growth or calcification at depth confounds the signal. By contrast, similar data from another planktonic species, *Neogloboquadrina pachyderma*, suggest a very different pattern in South Atlantic core tops: the $\delta^{18}O$ of this species follows the predicted $\delta^{18}O$ of calcite (and therefore SST variability) over a wide range of latitudes and hydrographic conditions [Charles and Fairbanks, 1990]. This pattern for *N. pachyderma* $\delta^{18}O$ is in marked contrast to that for the North Atlantic region [Kohfeld et al., 1996; Bauch et al., 1997]. Sorting
out such differences between two widely used planktonic species at high latitudes, including the differences between Northern and Southern Hemisphere patterns, is an essential prerequisite for refining reconstructions of past surface ocean conditions.

[5] Here we present a description of modern middle-to-high—latitude planktonic foraminifera caught in MOCNESS tows taken in the Southern Ocean sector of the Southern Ocean over a 1-month period during the austral summer of 1996. Specifically, we concentrate on the paleoceanographically important species that span a variety of depth habitats and latitudes. Previous work in the same study region made use of underlying core-top samples [Niebler et al., 1999] to decipher water column stratification. Our work here complements this and other work with core tops in the Southern Ocean [Labeyrie et al., 1996] because it allows a direct correlation between foraminiferal geochemistry and surface ocean properties. Our analysis does not consider all aspects of foraminiferal ecology, but we demonstrate that the unique intersection of depth habitat and South Atlantic water column structure might explain some of the main patterns of foraminiferal abundance and chemistry in the modern sediments (and by extension the paleoceanographic record).

2. Methods and Materials

[5] During January–March 1996 we collected MOCNESS plankton tow samples at or near piston coring stations that later became coring stations for Ocean Drilling Program (ODP) leg 177. The primary goals of this operation, on cruise TNO57 aboard the R/V Thomas Thompson, were twofold: (1) to document the vertical distribution of microfossil-producing plankton (foraminifera) and, by comparison with water column properties, to understand the ecological controls on their distribution and (2) to understand how and where in the water column the foraminifera acquire their chemical and isotopic signatures.

[6] Samples were collected from six stations spanning a latitudinal and longitudinal range between 41° and 53°S and 5° and 12°E, respectively (Figure 1). These subantarctic South Atlantic station locations were dictated by sediment coring interest, because they are areas with high sedimentation rates and important sedimentary sequences. However, these stations also happen to span surface waters that today exhibit, among other properties, an SST range of greater than 12°C in only 12° of latitude, making for one of the largest SST gradients in the global ocean [Levitus and Boyer, 1994]. The MOCNESS stations are therefore appropriately located in areas of interest for microfossil calibration.

[7] Of concern in any plankton tow study is the question of how representative such a “snapshot” view may be of seasonal or longer-term conditions. We assume that our snapshot is as representative of sedimentary conditions as is possible for a 1-month period. Because of ship time constraints, our sampling strategy could not explicitly resolve possible population migrations over diurnal or lunar-modulated [e.g., Bijma et al., 1990] reproductive cycles.

[8] Vertically stratified plankton tow samples (150-μm mesh net within a 333-μm net used for structural support) were collected using a MOCNESS device described in detail by Wiebe et al. [1976, 1985]. The MOCNESS system was also equipped with sensors to measure in situ properties from the water column, including temperature, salinity (and σt), and fluorescence. The temperature and salinity data are presented in Figure 2, while the σt data are shown with fluorescence in Figure 3. The individual nets were sequentially opened and closed over discrete depth intervals between 0 and 800 m, and temperature, salinity, and fluorescence were measured in situ with attached conductivity-temperature-depth (CTD) and fluorometer probes. The nine sampled intervals and their vertical spacing varied from station to station and were chosen using hydrographic information obtained from a separate CTD/Rosette cast just prior to MOCNESS deployment. MOCNESS samples were stored in a 100% ethyl alcohol (ethanol) solution to prevent dissolution of foraminiferal shells and to minimize artifacts to isotopic measurements that can arise with other preservatives such as buffered formalin [Ganssen, 1981]. Splits of these samples were density-separated in a hypersaline solution according to a modification of a method described by Bé [1959].

[9] Depending on station-specific foraminiferal abundances, sample splits were counted with the aim of totaling ~300 specimens per sample. In some cases the total abundance was low enough that the total counts came from the whole sample at less than 300 specimens. The abundance of key species from these sites are presented (Figure 4) as number of shells per unit volume of seawater, a scale made possible because the volume of filtered seawater was measured with a flowmeter attached to the frame of the MOCNESS system. The flowmeter data should be considered only an approximate representation of the effective flow through the nets (especially if biomass concentrations are high enough to cause net clogging, lower flow estimation, and hence overestimation of foraminiferal standing stock). For most stations, however, the relationship between the flowmeter data and effective flow should be relatively uniform for all the nets on a given cast. Fluorescence (mg/m³) is also presented with the absolute foraminiferal abundance data (Figure 4), after it was corrected by uniformly subtracting fluorescence contents below 500 m (which we assume to represent the background) for a given station. Planktonic foraminifera captured with the MOCNESS plankton tow system are not necessarily alive; however, their isotopic composition (discussed below) helps discern live from dead populations.
In order to minimize possible size-related differences, $\delta^{18}O$ analyses were carried out on the 150–250-μm fraction of *G. bulloides*, *N. pachyderma*, *Globorotalia truncatulinoides*, *Globorotalia inflata*, and *Orbulina universa*, where sufficient quantities were available. All samples were roasted under vacuum at 375°C for 1 hour prior to isotopic analysis. Measurements were made at the Scripps Institution of Oceanography (SIO) using a Carrousel-48 automatic carbonate preparation device coupled to a Finnigan MAT 252 mass spectrometer. The $\delta^{18}O$ precision (1σ) of the NBS-19 standard was better than 0.09‰ for 190 standards run along with the samples over an 18-month period during which the samples were analyzed. We will not deal with the $\delta^{13}C$ for the foraminifera (though the data exist) because measurements of seawater $\delta^{13}C$ (total dissolved CO$_2$) were probably compromised by sample storage effects.

In order to compare equilibrium calcite $\delta^{18}O$ ($\delta^{18}O_{ec}$) values against measured foraminiferal values we estimated the $\delta^{18}O$ of calcite precipitated in isotopic equilibrium with seawater. We first predicted the $\delta^{18}O$ composition of the seawater using our own salinity measurements obtained from the MOCNESS downcasts and a South Atlantic salinity-$\delta^{18}O$ (water) relationship from Geochemical Ocean Sections Study (GEOSECS) [1987] data shown by Charles and Fairbanks [1990]. The applicability of this relationship is assessed by representative $\delta^{18}O$ (water) (D. Hodell, personal communication, 1996) and salinity analyses from our own downcasts of the CTD (Figure 5). We then used the predicted $\delta^{18}O$ (water) and the MOCNESS temperature data to estimate the $\delta^{18}O_{ec}$, according to a standard mean ocean water (SMOW) to Peedee belemnite (PDB) unit conversion [Hut, 1987] and a carbonate-water isotopic temperature scale from O’Neil et al. [1969]. These predicted $\delta^{18}O_{ec}$ values against measured foraminiferal values were estimated from the MOCNESS downcasts and a South Atlantic salinity-$\delta^{18}O$ (water) relationship from Geochemical Ocean Sections Study (GEOSECS) [1987] data shown by Charles and Fairbanks [1990]. The applicability of this relationship is assessed by representative $\delta^{18}O$ (water) (D. Hodell, personal communication, 1996) and salinity analyses from our own downcasts of the CTD (Figure 5). We then used the predicted $\delta^{18}O$ (water) and the MOCNESS temperature data to estimate the $\delta^{18}O_{ec}$, according to a standard mean ocean water (SMOW) to Peedee belemnite (PDB) unit conversion [Hut, 1987] and a carbonate-water isotopic temperature scale from O’Neil et al. [1969]. These predicted $\delta^{18}O_{ec}$ values against measured foraminiferal values were estimated from the MOCNESS downcasts and a South Atlantic salinity-$\delta^{18}O$ (water) relationship from Geochemical Ocean Sections Study (GEOSECS) [1987] data shown by Charles and Fairbanks [1990]. The applicability of this relationship is assessed by representative $\delta^{18}O$ (water) (D. Hodell, personal communication, 1996) and salinity analyses from our own downcasts of the CTD (Figure 5). We then used the predicted $\delta^{18}O$ (water) and the MOCNESS temperature data to estimate the $\delta^{18}O_{ec}$, according to a standard mean ocean water (SMOW) to Peedee belemnite (PDB) unit conversion [Hut, 1987] and a carbonate-water isotopic temperature scale from O’Neil et al. [1969]. These predicted $\delta^{18}O_{ec}$ values against measured foraminiferal values were estimated from the MOCNESS downcasts and a South Atlantic salinity-$\delta^{18}O$ (water) relationship from Geochemical Ocean Sections Study (GEOSECS) [1987] data shown by Charles and Fairbanks [1990]. The applicability of this relationship is assessed by representative $\delta^{18}O$ (water) (D. Hodell, personal communication, 1996) and salinity analyses from our own downcasts of the CTD (Figure 5). We then used the predicted $\delta^{18}O$ (water) and the MOCNESS temperature data to estimate the $\delta^{18}O_{ec}$, according to a standard mean ocean water (SMOW) to Peedee belemnite (PDB) unit conversion [Hut, 1987] and a carbonate-water isotopic temperature scale from O’Neil et al. [1969]. These predicted $\delta^{18}O_{ec}$ values against measured foraminiferal values were estimated from the MOCNESS downcasts and a South Atlantic salinity-$\delta^{18}O$ (water) relationship from Geochemical Ocean Sections Study (GEOSECS) [1987] data shown by Charles and Fairbanks [1990]. The applicability of this relationship is assessed by representative $\delta^{18}O$ (water) (D. Hodell, personal communication, 1996) and salinity analyses from our own downcasts of the CTD (Figure 5). We then used the predicted $\delta^{18}O$ (water) and the MOCNESS temperature data to estimate the $\delta^{18}O_{ec}$, according to a standard mean ocean water (SMOW) to Peedee belemnite (PDB) unit conversion [Hut, 1987] and a carbonate-water isotopic temperature scale from O’Neil et al. [1969]. These predicted $\delta^{18}O_{ec}$ values against measured foraminiferal values were estimated from the MOCNESS downcasts and a South Atlantic salinity-$\delta^{18}O$ (water) relationship from Geochemical Ocean Sections Study (GEOSECS) [1987] data shown by Charles and Fairbanks [1990]. The applicability of this relationship is assessed by representative $\delta^{18}O$ (water) (D. Hodell, personal communication, 1996) and salinity analyses from our own downcasts of the CTD (Figure 5). We then used the predicted $\delta^{18}O$ (water) and the MOCNESS temperature data to estimate the $\delta^{18}O_{ec}$, according to a standard mean ocean water (SMOW) to Peedee belemnite (PDB) unit conversion [Hut, 1987] and a carbonate-water isotopic temperature scale from O’Neil et al. [1969]. These predicted $\delta^{18}O_{ec}$
data are then compared against $\delta^{18}O$ measurements from foraminifera collected in the water column in Figure 6, while also shown in latitudinal section form in Figure 2.

Among the various $\delta^{18}O$ paleotemperature equations proposed in the literature [e.g., Bemis et al., 1998, and references therein] we chose the inorganic calcite equation of O’Neil et al. [1969] as our reference point because it was the only one calibrated to temperatures as low as $0^\circ C$. However, we recognize that this definition of equilibrium calcite is arbitrary and perhaps not even directly applicable to living planktonic foraminifera. Over the temperature range of interest here ($0–15^\circ C$) the absolute offset between measured foraminiferal $\delta^{18}O$ and “predicted equilibrium” varies significantly if other equations are extrapolated to the low temperature range [e.g., Kim and O’Neil, 1997; Erez and Luz, 1982]. Because there is as yet no agreement on how best to define equilibrium for biologically precipitated calcite, we cannot interpret the average apparent disequilibrium. On the other hand, the choice of paleotemperature equation does not affect the relative interspecies and intraspecies offsets from equilibrium, an important point that we discuss below.

**Figure 2.** Latitudinal sections of temperature, salinity, and the predicted $\delta^{18}O$ of equilibrium calcite. The temperature and salinity data are downcast measurements collected with the CTD system, while the predicted $\delta^{18}O$ (calcite) data are calculated using $\delta^{18}O$ (standard mean ocean water (SMOW)) = 0.5(salinity) - 17‰ [Charles and Fairbanks, 1990], the Hut [1987] SMOW to Peedee belemnite (PDB) correction factor of $-0.27$‰, and the O’Neil et al. [1969] paleotemperature equation (see Figure 6 caption for further details).
In order to compare the foraminiferal $\delta^{18}O$ from the water column to that from material at the modern sediment-water interface, we also measured "core-top" samples beneath three of the six stations where material permitted. Core-top material is not presented from TNO57-22, TNO57-9, and TNO57-11 since we have reason to doubt the presence of modern or even late Holocene sediments at these locations. The representative data from the other stations are shown along the lines beneath individual panels of Figure 6. These samples were provided by either multicore or trigger core from each individual station for the depth interval of either 0–1 cm or 0–2 cm into the sediment. Table 1 presents a summary of the core samples used for the core-top data. Species-specific foraminiferal samples were generally picked from the 150–250-$\mu$m size fraction, although in one case (TNO57-21) it was necessary to...
supplement this with material from the >250-μm size fraction.

3. Results
3.1. Oceanographic Setting

[14] Figure 2 shows that the surface temperature ranges from about 1 to 13°C across the 12° latitudinal span of our study area. The density structure becomes much better defined to the north, with a relatively shallow thermocline and a clear surface mixed layer. The salinity section (Figure 2) shows generally more salty surface waters in the north, while also illustrating southern middepth anomalies that are likely the result of interleaving layers of water with different origin. Thus the local structure affecting the foraminiferal population is certainly partly a product of eddy mixing, which must be a perennial process associated with the Antarctic Circumpolar Current [Savchenko et al., 1978; Gille et al., 2000].

[15] Phytoplankton abundance, as in many other areas of the world’s oceans, is sensitive to the hydrographic properties associated with the pycnocline. Figure 3 shows that fluorescence maxima all occur within the mixed layer, but usually at relatively deep depths near the top of the pycnocline. The origin of the so-called “deep chlorophyll maximum” in other regions has been a source of long standing debate, and the possible mechanisms for its occurrence include the intersection of high nutrients and high light, the effect of density gradients on settling, or preferential grazing patterns. It is also possible that fluorescence peaks are partly the result of phytoplankton photoadaptation, involving changes in the carbon to chlorophyll or carbon to fluorescence ratios. Without ancillary evidence our data cannot test these various possibilities. Accordingly, we only note the relationship between fluorescence and foraminifera on a descriptive (nonmechanistic) level.

3.2. Planktonic Foraminiferal Abundance

[16] Several important observations can be drawn from the abundance data presented in Figure 4. One is that the overall abundances vary considerably from site to site, by as much as 4–5 orders of magnitude between TNO57-16 and TNO57-13. The TNO57-9 site shows the highest surface foraminiferal abundance, about 235/m³ of seawater, in apparent response to a phytoplankton bloom suggested by the coincident peak in fluorescence (>0.7 mg/m³). The TNO57-16 site shows the highest middepth (75–100 m water depth) abundance, about 2484/m³ of seawater, also in association with a fluorescence peak of about 1.4 mg/m³. The highest deep (>100 m water depth) abundances are shown at TNO57-16, TNO57-9, and TNO57-11.

Figure 4. (opposite) Water column abundances of various species of planktonic foraminifera presented from each of our six stations, arranged as in Figure 3, along with in situ fluorescence data from the MOCNESS system. In each stacked histogram figure the varying column thickness corresponds to the towed depth interval in the water column, while the foraminiferal species absolute and relative abundances are shown along the bottom x axis according to the legend color scheme. The fluorescence data are shown in each panel along the upper x axis in concentration units of mg/m³ (as in Figure 3).

[17] The association of total foraminiferal abundance peaks with fluorescence maxima is not always straightforward. For instance, surface peaks in fluorescence at sites TNO57-21 and TNO57-11 are not coincident with abundance peaks at those stations. Also, fluorescence peak depths vary from site to site. For example, the peaks occur well below the surface at the TNO57-16 and TNO57-13 sites, while they generally occur closer to the surface at the other stations. Despite these differences, however, there is reasonable correspondence between foraminiferal abundance and fluorescence maxima. In fact, TNO57-22 shows a double fluorescence peak in the upper 100 m, with high foraminiferal abundances closely associated with each.

[18] It is possible to separate the abundance patterns by species. Despite the fact that G. bulloides is encountered over a wide depth range the maximum of this species is consistently associated with a fluorescence peak (the only exception being TNO57-21), suggesting that G. bulloides closely tracks phytoplankton blooms. The next most abundant species is N. pachyderma (left-coiling), which shows a wide range of occurrence for abundance maxima (anywhere from the surface at TNO57-13 to the 75–175 m depth at TNO57-9. N. pachyderma (right-coiling) is observed mostly at middepths, with one exception (TNO57-22). Turborotalita quinqueloba also generally shows its highest abundances below 100 m, although the surface maximum at TNO57-22 is also an exception. G. inflata is consistently most abundant at subsurface depths (between 50 and 300 m), while the maximum abundances of G. truncatulinoides are always deeper than 100 m. The maxima of this species are even deeper at TNO57-21 and TNO57-9, roughly 500 and 200 m, respectively. O. universa is the least abundant planktonic species in these taws, showing measurable middle-to-deep quantities at only TNO57-21 and TNO57-22.

[19] In addition to this close association with a probable food source in phytoplankton (suggested by the fluorescence peaks), maximum foraminiferal abundances generally occur within the mixed layer and just above the pycnocline (the only exception being TNO57-21). In all cases, G. bulloides maxima coincide with overall foraminiferal maxima because it is everywhere the most dominant species.

3.3. Oxygen Isotopic Signatures and Water Column δ18O Gradients

[20] Oxygen isotopic data (δ18O in %PDB) from the water column and core-top samples are presented in Figure 6. In the water column at all stations, predicted δ18O curves are more positive in δ18O relative to foraminiferal measurements, implying that all species are well outside our
calculations of isotopic equilibrium (using the O’Neil et al. [1969] equation) in this region. However, at the depth of maximum abundance for any given species, the species-specific offset from expected equilibrium is fairly constant. We specifically calculated this $\delta^{18}O$ departure for *G. bulloides*, *N. pachyderma*, *G. inflata*, and *G. truncatulinoides*, using all six stations and all appropriate depths (where peak abundances coincided with measurable $\delta^{18}O$). This collective analysis yielded an average $\delta^{18}O$ departure from equilibrium of $1.0 \pm 0.4\%$, with a high degree of similarity for each species (*G. bulloides* = 0.95, *N. pachyderma* = 1.00, *G. inflata* = 0.94, and *G. truncatulinoides* = 1.10).
Figure 5. (opposite) Comparison of the predicted equilibrium $\delta^{18}O$ (in %PDB) of calcite estimated with the same method but with two different starting points. The first involves the prediction of the $\delta^{18}O$ of seawater based on salinity (light circles), while the second involves actual measurements of the $\delta^{18}O$ of seawater (dark squares, D. Hodell, unpublished data, 1996) where data are available from four of the six stations. The results of this comparison justify the use of a South Atlantic salinity versus $\delta^{18}O$ (water) relationship derived from GEosecs [1987] data (see text for details). With either starting point the $\delta^{18}O$ (water) information is necessary to take the prediction to the next step. The $\delta^{18}O$ (water) prediction is converted to PDB units [Hut, 1987] and then used to estimate the $\delta^{18}O$ of equilibrium calcite according to the paleotemperature equation of O’Neil et al. [1969]. Panel arrangement is consistent with previous figures, and locations for which this comparison was not possible (TNO57-11 and TNO57-16) are not shown.
there have been various attempts to determine the controls on the modern depth habitat of this species (e.g., Ortiz et al. [1995] for the North Pacific and Kohfeld et al. [1996] for the North Atlantic). Our South Atlantic results suggest this species is most abundant at pycnocline depths (strongly dictated by thermocline depths) and chlorophyll maxima in the sub-Antarctic (Figures 3 and 4), an observation that is generally consistent with the other previous
studies. In our study area the *N. pachyderma* abundance peak south of the modern polar front constitutes a case where the maximum does not coincide with a fluorescence peak below the pycnocline (Figures 3 and 4). This observation suggests that at higher latitudes, *N. pachyderma* may indeed be more surface-restricted. Other northern peak below the pycnocline (Figures 3 and 4). This observation suggests that at higher latitudes, *N. pachyderma* may indeed be more surface-restricted. Other northern high-latitude MOCNESS studies [Carstens et al., 1997; Bauch et al., 1997] found the modern distribution of *N. pachyderma* to be dictated more by water mass differences (i.e., eddy mixing or advection) potentially influence *N. pachyderma* abundances in the water column. In our case here the slope of the isopycnal surfaces might facilitate communication between surface dwelling populations south of the Antarctic Polar Front and thermocline dwelling populations to the north [Mortyn et al., 2002].

[27] Despite the fact that there was not one truly surface-dwelling species that spanned our South Atlantic study region there were species of planktonic foraminifera that were at least restricted to deeper depths. Our results from higher latitudes in the South Atlantic corroborate earlier findings [Fairbanks et al., 1980; Hemleben et al., 1989] and support the idea that *G. inflata* is a relatively deep dweller. In our tows, *G. inflata* is consistently most abundant at depths between 50 and 300 m. Similarly, our South Atlantic observations corroborate previous findings [Fairbanks et al., 1980; Hemleben et al., 1985, 1989] that *G. truncatulinoides* is a deep dweller, consistently shown beneath 100 m and well below both the fluorescence maxima and pycnocline depths at each of our stations. Our South Atlantic abundances are much lower than in other studies, on the order of 1 shell/m³. This difference might result from the fact that we sampled during the austral summer at a time of year when this species does not typically reproduce [Lohmann and Schweitzer, 1990]. In any case it is clear that at least in the summer months, two of the dominant species of foraminifera are not observed in the surface mixed layer of the high-latitude South Atlantic, regardless of thermal structure. This observation suggests that the shell chemistry of these species should reflect thermocline, or perhaps even subthermocline, conditions.

[28] If various planktonic foraminiferal species are stratified with depth, then it is appropriate to consider whether their respective (or combined) δ¹⁸O composition reflects this depth distribution. Synthesizing a large body of previous work [e.g., Emiliani, 1955, 1966; Bé and van Donk, 1971; Hecht and Savin, 1972; Shackleton et al., 1973; Vergnaud-Graizzi, 1976; Williams et al., 1977; Shackleton and Vincent, 1978; Kahn, 1977, 1979; Fairbanks et al., 1980; Curry and Matthews, 1981; Kahn and Williams, 1981], one might conclude that there is no systematic picture of how the δ¹⁸O of any particular species deviates from equilibrium and therefore that the exercise of resolving vertical water column structure from foraminiferal δ¹⁸O might be inherently problematic. However, many of the apparent previous interpretations of disequilibrium may have been the result of an evolving range of calibrations and equations used for estimating equilibrium calcite δ¹⁸O values [e.g., Bemis et al., 1998, and references therein]. We observe offsets from arbitrarily defined equilibrium calcite values that are consistent across a range of stations. Our field-based data also tend to support extrapolation of culture-based δ¹⁸O calibration of Bemis et al. [1998], though our data are not sufficient to conclude that subantarctic foraminifera precipitate their skeletons in equilibrium with seawater according to a universal linear trend. In any event the observation of constant offsets from calculated equilibrium would at least seem to minimize the strictly biogeochemical complications of foraminiferal δ¹⁸O.

[29] The core-top foraminiferal δ¹⁸O values, however, are consistently higher than those from the water column, in some cases by amounts that exceed the 0.2–0.3‰ that might be attributable to the warming of the 20th century alone. This generalization includes the deeper-dwelling *G. inflata* as well as *G. bulloides*. Since there were no reasons to question the integrity of the representative core-top data that we report here, such as physical evidence for bioturbation or a systematic trend with carbonate saturation state (core tops were distributed both above and below the present carbonate lysocline [Hodell et al., 2001]), other processes must be involved. For example, significant calcification in colder seasons than we sampled by the plankton tows (peak of the summer) is one obvious possibility. Another plausible explanation for these observa-

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**Table 1. Summary of Core-Top Sample Material Used in This Study**

<table>
<thead>
<tr>
<th>Site</th>
<th>Core-Type</th>
<th>Water Depth, m</th>
<th>Available Planktonic Species</th>
<th>Size Fraction, μm</th>
<th>Sample Interval, cm</th>
</tr>
</thead>
<tbody>
<tr>
<td>TNO57-21</td>
<td>multicore</td>
<td>4978.5</td>
<td><em>G. inflata</em></td>
<td>&gt;250</td>
<td>0–1</td>
</tr>
<tr>
<td>TNO57-16</td>
<td>trigger core</td>
<td>3665</td>
<td><em>G. bulloides, N. pachyderma</em></td>
<td>150–250</td>
<td>0–2</td>
</tr>
<tr>
<td>TNO57-13</td>
<td>multicore</td>
<td>2851</td>
<td><em>N. pachyderma</em></td>
<td>150–250</td>
<td>0–1</td>
</tr>
</tbody>
</table>

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**Figure 6.** (opposite) Oxygen isotopic data (δ¹⁸O in ‰PDB) for each site, similar to the arrangement in previous figures. Each y axis corresponds to depth in the water column to 800 m, and each x axis corresponds to δ¹⁸O from 0 to 4.5‰. In each panel the line with the most data points and the heaviest isotopic values corresponds to that of predicted δ¹⁸O of calcite (using predicted δ¹⁸O (water) rather than measured δ¹⁸O (water), see Figure 5), while the other lines correspond to species-specific δ¹⁸O (calcite) measurements according to the legend in the lower right panel. Isotopic analyses were made on any sample that yielded enough foraminiferal specimens for measurement. Core-top δ¹⁸O data are shown along the bottom line of the panels where representative data were available, at TNO57-21, TNO57-16, and TNO57-13 (see text for details).
tions is that all species acquire a calcite crust at a depth below that of our plankton tows, i.e., somewhere below 800 m in the water column while enroute to the sediment-water interface (at lower temperatures and therefore with higher $\delta^{18}O$). This process has at least been documented for *G. truncatulinoides* and seems plausible for *G. inflata* but is less well established for species such as *G. bulloides* [Hemleben et al., 1989]. If only the deeper-dwelling species acquire secondary calcite preferentially, the sedimentary $\delta^{18}O$ differences between them and the shallower species should be accentuated. Our data do not show evidence for such differential secondary calcification between shallow and deep-dwelling species (Figure 6). We therefore have a stronger basis for interpreting their $\delta^{18}O$ differences from the sedimentary record as a reflection of changes in the upper surface ocean or foraminiferal habit.

[30] Does the water column depth habitat at least explain the geographic pattern, if not the absolute value, of $\delta^{18}O$ observed in sediment core-top samples? Prior compilations of Southern Ocean (Indian and Atlantic sectors) core-top *G. bulloides* and *N. pachyderma* $\delta^{18}O$ data can be used to address this problem (Figure 7). From these data it is evident that core-top *G. bulloides* $\delta^{18}O$ values deviate from the trend of predicted $\delta^{18}O$ values at higher temperatures (northern sub-antarctic). In contrast, the core-top *N. pachyderma* $\delta^{18}O$ pattern follows the slope of predicted $\delta^{18}O$ over the full range of the data set (compare solid squares against open triangles in Figure 7).

[31] The tow and hydrographic data suggest a possible interpretation for this difference. The northern hydrographic conditions enhance the temperature contrast between water at the base of the mixed layer and water just below the mixed layer, such that any diapycnal mixing and entrainment across this boundary would stir up relatively cold water and allow proliferation of *G. bulloides* populations with a relatively positive $\delta^{18}O$ signature. In the warmer regions of the sub-antarctic, and therefore where predicted $\delta^{18}O$ is relatively low, the *G. bulloides* $\delta^{18}O$ signal is more effectively “smeared” up from below the mixed layer, and measured $\delta^{18}O$ values are consequently heavier than predicted for the surface mixed layer. This process is best illustrated by our results that show shallow *G. bulloides* $\delta^{18}O$ values much closer to predicted values in our more northern stations (Figure 6).

[32] Despite this potential bias it appears that *G. bulloides* $\delta^{18}O$ does track the predicted $\delta^{18}O$ of calcite throughout the upper 100 m at several stations, while *G. inflata* and *G. truncatulinoides* $\delta^{18}O$ reliably trace the profile of predicted $\delta^{18}O$ of calcite below this depth (Figure 6, stations TNO57-22, TNO57-9, and TNO57-11 especially). Thus, if these deeper *G. inflata* or *G. truncatulinoides* data are combined with the shallow (<100 m) *N. pachyderma* or *G. bulloides* data, the structure of the predicted equilibrium $\delta^{18}O$ of calcite profile is reproduced well in the depth-discrete tow samples. Because $\delta^{18}O$ is a strong reflection of the ocean’s thermal and density structure (Figure 2), this shallow-deep $\delta^{18}O$ differencing signature can, in effect, be used as a recorder of water column structure conditions in regions where the abundance profiles of the different species are sharply resolved peaks. Sediment cores from

![Figure 7. Plot of measured $\delta^{18}O$ (%PDB) versus predicted $\delta^{18}O$ (%PDB) for Southern Ocean core-top data compiled from both Labeyrie et al. [1996] and this study. Predicted values are all derived from austral summer temperature and salinity data in order to minimize seasonality effects (see Labeyrie et al. [1996] and text of this study for details). Foraminiferal species shown from both the south Indian and South Atlantic sectors are *G. bulloides* and *N. pachyderma* (left-coiling). The dashed line corresponds to a 1:1 slope where measured values follow those that are predicted. *N. pachyderma* data are represented by solid squares; *G. bulloides* (where predicted values are >2.6%) data are represented by open circles; and *G. bulloides* (predicted values <2.6%) data are represented by open triangles. Note that the *N. pachyderma* and the *G. bulloides* (predicted >2.6%) data follow the 1:1 slope, while the *G. bulloides* (predicted <2.6%) data do not, indicating that *G. bulloides* $\delta^{18}O$ from warmer regions farther north are typically heavier than predicted, a pattern that does not occur farther south (see text for details). One of the *G. bulloides* data points (predicted value of 0.87%) comes from a core location close to TNO57-21 and is not presented elsewhere in this study.](image)

5. Conclusion

[33] The combined ecological and geochemical approach of the MOCNESS plankton tow sampling suggests that the $\delta^{18}O$ of both deep- and shallow-dwelling species reflects the
three-dimensional thermal structure of the upper ocean throughout much of the subtropical region. Though the results clearly imply that any one planktonic foraminiferal record from the sediment may be affected by the complexities of upper ocean structure, the conclusion is relevant for paleoceanographic work in a number of specific ways. For example, it provides an explanation for why the amplitude of planktonic foraminiferal \( {\delta}^{18}O \) signals in the northern sub-Antarctic may differ from their counterparts in other regions. Second, it suggests that changes in the difference between shallow- and deep-dwelling species is a legitimate strategy for monitoring surface water stratification, provided that the most appropriate shallow indicator species is chosen (G. bulloides in the central sub-Antarctic and N. pachyderma to the south). Third, it provides constraints on the possible interpretations of high-latitude carbon cycling from various existing foraminiferal records of the Southern Ocean [e.g., Rickaby and Elderfield, 1999]. Thus the hydrographic and ecological complexities may offer opportunities as well as challenges.

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**References**


