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Stable isotopes in deep-sea carbonates: Box Core ERDC-92, West Equatorial Pacific

Oxygen isotopes
Carbon isotopes
Foraminifera
Deep sea sediments
Carbonates
Pacific ocean

Isotopes de l'oxygène
Isotopes du carbone
Foraminifères
Sédiments pélagiques
Carbonates
Pacifique

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ABSTRACT

Stable isotopes of oxygen and carbon have been determined for fifteen species of planktonic foraminifera and for various size fractions of carbonate material from a box-core raised in the western equatorial Pacific. There is a change in $\delta^{18}O$ of between 1.1 and 1.5‰ from the last glacial to the Holocene. Generally, the isotopically heavy foraminiferal species show the smaller ranges. Oxygen isotopes of coccoliths and small foraminifera tend to be somewhat heavier than average annual surface temperatures would indicate, perhaps due to increased production during times of increased upward mixing of cold subsurface waters. The ranking of species in terms of $\delta^{18}O$ resembles a depth habitat ranking (correlation coefficient 0.9). Oxygen isotopes of small foram tests are generally lighter than those of larger ones, within the same species. There is evidence of an upward migration of deep-living foram species during glacial times.
There is little change in $\delta^{13}C$ values of foram shell carbonate from glacial to postglacial. A $\delta^{13}C$ minimum occurs during deglaciation. Comparison of $\delta^{13}C$ values with reconstructed $\delta^{13}C$ depth profiles suggests that none of the shell carbonate is precipitated in thermodynamic equilibrium with respect to ^{13}C. The deviation is especially strong in small foraminifera. The degree of deviation from ^{13}C equilibrium is suggested to be a measure of growth rate, that is, the intensity of metabolism of the foraminifera. Shallow water species show a higher variability of $\delta^{13}C$ values than deep water foraminifera.

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RÉSUMÉ

Isotopes stables dans les sédiments calcaires profonds : Carotte boîte ERDC-92, Pacifique équatorial occidental

La teneur en isotopes stables de l'oxygène et du carbone a été déterminée dans les sédiments calcaires du Pacifique équatorial occidental prélevés au carottier boîte. Les mesures ont porté sur différentes classes granulométriques du sédiment total et sur quinze espèces de foraminifères planctoniques. On observe un changement du $\delta^{18}O$ entre le dernier glaciaire et l'Holocène. La différence est entre 1,1 et 1,5 ‰, les changements les plus faibles étant généralement observés pour les tests de foraminifères lourds sur le plan isotopique. Les isotopes de l'oxygène des coccolithes et des foraminifères de petite taille ont tendance à être plus lourds que l'indiquerait la moyenne annuelle de température des eaux de surface; ceci résultant peut-être d'une productivité accrue durant les périodes où les remontées d'eau froide de subsurface sont plus intenses. Les espèces se classent par rapport au $\delta^{18}O$ d'une façon semblable à leur ordre de classement en fonction de la profondeur d'habitat (coefficient de corrélation 0,9). La composition isotopique en

oxygène des tests de foraminifères de petite taille est généralement plus légère que celle des foraminifères de grande taille au sein d'une même espèce. Les espèces de foraminifères vivant en profondeur semblent remonter durant l'intervalle glaciaire.
La différence observée dans les valeurs du $\delta^{13}C$ du carbonate des tests de foraminifères entre le glaciaire et le postglaciaire est de faible amplitude. La période de déglaciation correspond à un minimum du $\delta^{13}C$. La comparaison entre les $\delta^{13}C$ mesurés et la reconstruction des variations du $\delta^{13}C$ en fonction de la profondeur d'eau montre que le carbonate des coquilles n'est en aucun cas précipité en équilibre thermodynamique par rapport au ^{13}C. La déviation est particulièrement forte pour les foraminifères de petite taille. L'amplitude de la déviation à partir de l'équilibre est apparamment fonction du taux de croissance, c'est-à-dire de l'intensité du métabolisme du foraminifère. Les espèces d'eaux peu profondes montrent une plus grande variabilité dans les valeurs de $\delta^{13}C$ que les foraminifères d'eaux profondes.

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INTRODUCTION

Following the work of Emiliani (1955, 1966) the oxygen isotope stratigraphy of planktonic and benthic foraminifera has become one of the major tools in the study of ocean history (Imbrie *et al.*, 1973; Shackleton, Opdyke, 1973, 1976; Emiliani, Shackleton, 1974). Recently, carbon isotope stratigraphy has been added to the pool of paleoceanographic techniques (Savin *et al.*, 1975; Shackleton, Kennett, 1975*a*, *b*; Berger *et al.*, 1978; Shackleton, Vincent, 1978), and coccoliths also are now studied with these methods (Margolis *et al.*, 1975). While the general shape of the isotope curves clearly displays the oscillations of ocean climate, the exact meaning of the range of isotope values has been the subject of considerable discussion. The variables presumably influencing the isotopic values include temperature fluctuations, exchange of water of the ocean with continental glaciers (glacial effect), changes in the patterns of evaporation and precipitation (water mass effect), changes in the deviation from thermodynamic equilibrium during bio-mineralization (vital effect), changes in the seasonality and in the mean depth level of shell production (differential production), and changes in the amount of selective removal of shells (differential dissolution) (see Table 1). Considerable progress has been made on the assessment of the amplitudes of isotopic signals since Emiliani's (1954) original work (see Lidz *et al.*, 1968; Hecht, Savin, 1970, 1972; Emiliani, 1971; Savin, Douglas, 1973; Vergnaud-Grazzini, 1976; Williams *et al.*, 1977). More can be learned from following the changes in isotopic composition of various kinds of particles through the deglaciation event. We present here such a detailed stratigraphy, the first for a deep sea core. In addition, we provide a matrix of compositions for planktonic foraminifera, for comparison between size classes within species, and between species at various characteristic times (present, climatic optimum, and near the glacial maximum). Data for one size fraction of two of the fifteen species considered here have been presented previously (Berger, Killingley, 1977).

MATERIALS AND METHODS

Among eighteen 50 x 50 cm box cores from Ontong-Java Plateau in the western Pacific (Sio Eurydice Expedition, Leg 9, April-May 1975), we chose a shallow core, ERDC-92, for analysis (2°13.5'S; 156°59.9'E; 1 598 m; 34 cm deep). The core consists of undisturbed, firm sediment (vane shear values near 130 g/cm^2), rich in carbonate (82 to 86%) and sand-sized particles (near 50%), and with a porosity near 71% and a bulk saturated density near 1.52 g/cm^3 (see Johnson *et al.*, 1977).

Wide-diameter (8.24 cm) subcores were taken by hand on board ship. One of these subcores (ERDC-92-2) was subsampled at closely spaced intervals (~ 1.5 cm) for isotopic analysis.

Analysis of the carbonate by mass spectrometer followed standard procedures. Samples were subjected to a treatment with "Calgon" solution and ultrasonic vibration to disaggregate the particles and provide clean specimens. The samples were washed through a series of sieves and individual foraminifera were hand-picked from the size fraction of interest. Approximately 0.5 mg amounts of prepared samples were reacted with 100% phosphoric acid at 50°C in an "in-line" vacuum system connected to VG Micromass 602 C mass spectrometer. After measuring the isotopic composition of the released CO_2 against that of a known CO_2 reference gas, the oxygen and carbon isotopic values for the samples were calculated with respect to PDB (as δ ‰) by the usual procedure (Craig, 1957). The reference gas was calibrated against standard NBS-20 limestone treated in the same way as the samples. The analytical precision expressed as σ for NBS-20 standard carbonate was 0.06‰ and the average difference between 60 duplicate foraminifera samples was 0.12‰ for $\delta^{18}O$ and 0.11‰ for $\delta^{13}C$.

RESULTS AND DISCUSSION

Oxygen isotope stratigraphy

The oxygen isotope values of the various size fractions and species analyzed in stratigraphic sequence show the general trend from heavy to light and the separation of shallow and deep living species which has been established previously by Emiliani (1955) and by subsequent work (Fig. 1). The total change from glacial

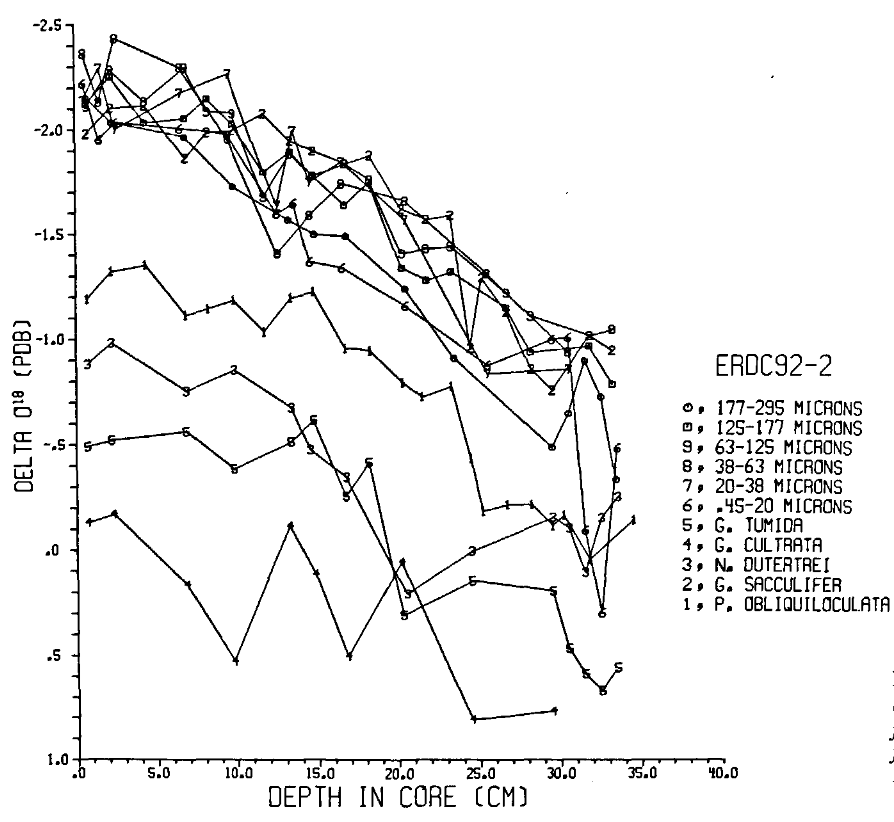

Figure 1

Oxygen isotope composition of various size fractions and of selected foram species, as a function of depth in Box Core ERDC-92. Note that the spacing is not exactly the same for all components.

to postglacial values is remarkably similar for the various types of particles, about $1.1°/_{oo}$, with the isotopically heavier species showing the smaller range. Beyond this trend and separation, there are remarkable differences in the sequences of the isotopic signals.

The separation of the shallow-water species *Globigerinoides sacculifer* from the deeper living species *Pulleniatina obliquitoculata*, *Neogloboquadrina dutertrei*, *Globorotalia tumida* and *Globoratalia cultrata* (= *G. menardii*) is very striking. *P. obliquiloculata* is about $0.8°/_{oo}$ heavier than *G. sacculifer*, corresponding to an apparent temperature difference of 3.3°C. This difference is constant from glacial to postglacial time in contrast to results in the Caribbean (Emiliani, 1955; Lidz *et al.*, 1968). There the shallow-living species have a much larger isotopic range ($\sim 1.7°/_{oo}$) than the deeper living ones (~ 0.9 to $1.0°/_{oo}$). This was interpreted by Lidz *et al.* (1968) as a contrast between a combined temperature plus glacial effect ($1.7°/_{oo}$) and a dominating glacial effect in the deep living species. As is evident from Table 1, other effects also enter, including any change-of-depth effect (stressed by Shackleton, 1968).

In the present data, *G. tumida* and *N. dutertrei* do show a trend of increased difference to *G. sacculifer*, from glacial to postglacial. However, it is much less pronounced than in the Caribbean. *G. sacculifer* is almost $0.9°/_{oo}$ lighter than *N. dutertrei* in the glacial and $1.2°/_{oo}$ lighter in the postglacial. Similarly, the *G. sacculifer-G. tumida* difference changes from 1.2 to $1.4°/_{oo}$.

The contributions from the various effects bearing on isotopic range (Table 1) are extremely difficult to identify. For example, a simple hypothesis consistent with our data is that the change in isotopic composition of the water from glacial to postglacial time is near $1°/_{oo}$ (Savin, Stehli, 1974), and a temperature change of near 2°C (Climap, 1976) accounts for a change of $0.4°/_{oo}$ in surface waters which brings the total range of *G. sacculifer* to $1.4°/_{oo}$. The same temperature effect would penetrate into the uppermost thermocline, the site of output of *P. obliquiloculata*. Below this level the temperature effect would be much reduced. Alternatively, the glacial effect could be $1.2°/_{oo}$ (following Shackleton, Opdyke, 1973) and the overall temperature change in mixed layer and uppermost thermocline would be only 1°C. The ranges of the deep living species would reflect entirely the glacial effect under this assumption.

In the present context, "glacial effect" strictly refers to that part of the amplitude of change in our data which is

Table 1

Variables influencing the glacial/interglacial range of isotopic composition of foraminifera ($\Delta\delta_f = \Delta\delta_T + \Delta\delta_G + \Delta\delta_E + \Delta\delta_V + \Delta\delta_D + \Delta\delta_R$) (from Berger, Gardner, 1975).

	Symbols	Magnitude ($°/_{oo}$)	Équivalent température (°C)
Temperature	$\Delta\delta_T$	~1	4
Glacial effect	$\Delta\delta_G$	~1	4
Water mass effect	$\Delta\delta_E$	~1	4
Vital effect	$\Delta\delta_V$	~0.5	2
Selective production	$\Delta\delta_D$	~0.5	2
Differential dissolution	$\Delta\delta_R$	~0.3	1

attributable to the addition (or subtraction) of glacial water, after thorough mixing of the ocean. The "true" glacial effect (usually implicitly defined as a maximum range in the literature) is greater, because of the effects of bioturbation and of incomplete oceanic mixing (Berger, Johnson, Killingley, 1977).

In either case, slight changes in the seasonality or habitat of shell output of *N. dutertrei* with respect to *G. tumida* are indicated. *G. cultrata* is seen to produce an entirely erratic signal, for reasons which will be discussed below.

The isotopic signals for the various undifferentiated size fractions (0.45-20 μm, coccolith fraction; 20-38, 38-63; 63-125; 125-177 and 177-295 μm) in essence behave much like *G. sacculifer*, that is, they are produced in waters of similar temperature and salinity. Because of the large number of specimens involved, results from the fine fraction actually are less erratic than those of *G. sacculifer*. Note for example that the very fine sand (63-125 μm) is consistently lighter than the fine sand (125-177 μm) by a small amount. The results are in agreement with field observations showing that small foraminifera are highly concentrated in surface waters (Berger, 1971).

The coccolith signal is of special interest, since there is relatively little information on coccolith isotopes.

Working on Pleistocene material from the Caribbean and the eastern tropical Pacific, Anderson and Cole (1975) reported a covariance relationship between the isotopic signals of coccolith-rich material (<44 μm) and planktonic foraminifera. They found that the $\delta^{18}O$ values for mixed foram samples were systematically heavier than the corresponding coccolith fraction in the eastern Pacific samples. Differential dissolution was invoked to explain the differences because the core was from below the lysocline. They assumed no differential dissolution effect in the coccoliths on the assumption that they all grow in surface water. In Figure 1 it is apparent that, for our well preserved core ERDC-92, the $\delta^{18}O$ signal for *G. sacculifer* is equal to or lighter than that for the coccolith fraction. Also, there is no obvious difference between the $\delta^{18}O$ of coccoliths and that of small foraminifera. However, the more resistant species, *P. obliquiloculata*, *N. dutertrei*, and *G. tumida* are indeed heavier, isotopically. Therefore, the explanation of Anderson and Cole that differential dissolution was the primary cause of the observed difference between forams and coccoliths is not contradicted, although the other effects mentioned (Table 1) also enter.

An important question that bears on the paleoclimatic use of calcareous plankton pertains to the magnitude of any "vital effect", that is, non-equilibrium precipitation. There has been considerable speculation with respect to this question. Duplessy *et al.* (1970), Vinot-Bertouille and Duplessy (1972) have presented evidence for non-equilibrium shell formation in benthic forams. Of course, this may not bear directly on the problem at hand. Vergnaud-Grazzini (1976), Kahn (1977) and Williams *et al.* (1977) have examined the same question for planktonic foraminifera. Shackleton *et al.* (1973) have shown that planktonic foraminifera taken in horizontal tows from a depth of 50 m can differ in isotopic composition. From this they conclude (p. 177), that "a substantial portion of the variation in isotopic composition between one species and another in foraminiferal death assemblages is due to different fractionation factors rather than to different life habitats." This may be true. However, without information on the gradients of the temperature and

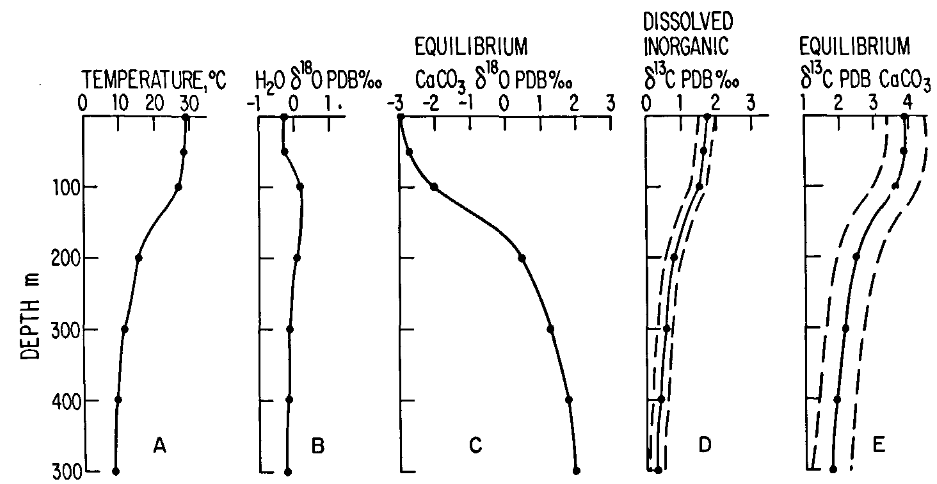

Figure 2

Temperature and reconstructed isotopic profiles for the eastern equatorial Pacific 2.1°S, 170°E. A, estimated average temperature depth profile from data given in Rotschi et al. *(1972); B, estimated ocean water $\delta^{18}O$ profile with depth reconstructed by using salinity data of Rotschi* et al. *(1972) and the oxygen 18, salinity relationship of Craig* et al. *(1965); C, equilibrium $CaCO_3$-$\delta^{18}O$ variation with depth calculated from the paleotemperature equation modified by Craig (1965) and profiles A and B; D, equilibrium $CaCO_3$-$\delta^{13}C$ calculated by relating apparent oxygen utilization (AOU) to $\delta^{13}C$. The AOU-$\delta^{13}C$ relationship was obtained by plotting treated data from Kroopnick (1974), Chung and Craig (1973), and Kroopnick* et al. *(1972). AOU for the Eastern Equatorial Pacific was computed by taking the difference between measured dissolved oxygen (Rotschi* et al.*, 1972) and the O_2 saturation values from Sverdrup, Johnson and Fleming, 1942, p. 188; using the salinity/temperature data of Rotschi* et al.*, 1972. The dashed profiles indicate the limits of the estimated error range; E, equilibrium $CaCO_2$-$\delta^{13}C$ profile obtained from profiles A and D using the carbon isotope fractionation relationships of Emrich* et al. *(1970). The dashed profiles indicate the limits of the estimated error range.*

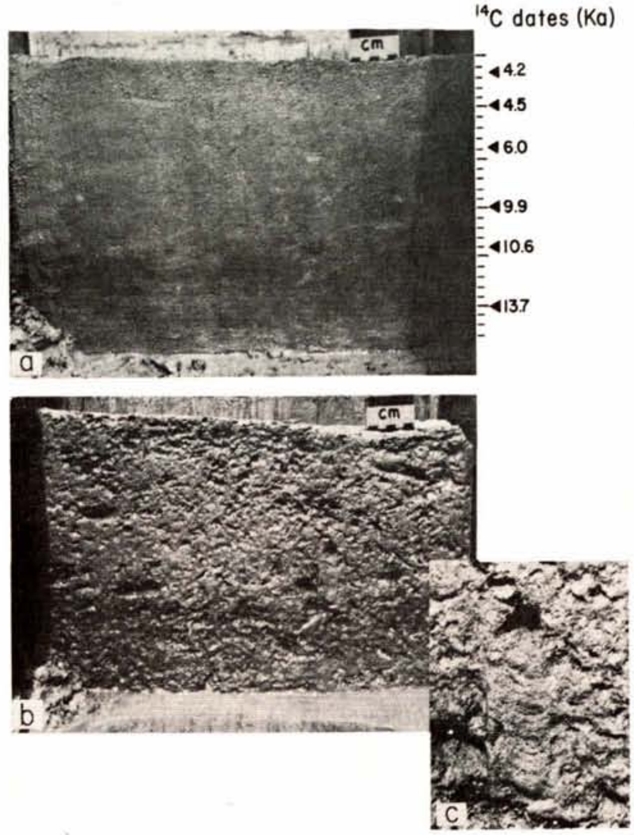

Figure 3

Freshly cut vertical face of Box Core ERDC-92. a, *smoothed with spatula and cleaned with wet paper towel* (^{14}C *dates from Peng* et al., 1978); b, *gently washed with a spray of water to bring out structures;* c, *close-up of* Teichichnum *spreite.*

isotopic composition of the water, or on the depth migrations of the foraminifera during shell growth, this suggestion cannot be evaluated.

Here we merely note that the value of $-2.1°/_{oo}$ for the $\delta^{18}O$ of the fine fractions in the surface sediment corresponds to 27°C under the assumption of equilibrium. This is slightly lower than the temperatures measured in surface waters (Fig. 2). This observation agrees with the suggestion that tropical shell output may be enhanced during times of upwelling and mixing (Berger, Gardner, 1975).

Apart from the differences between the mean values and the amplitudes of the various signals, there also are differences between the shapes of their sequences. For example, comparing the curves for *P. obliquiloculata* and *G. tumida*, we note that the maximum change in the signal from the heavy glacial to postglacial light values is near 24 cm (~ 14 ky ago) in one and near 19 cm (~ 11 ky) in the other. Considering the relatively short mixing times of the upper ocean waters, one cannot assume that the deeper living species experienced any changes in water temperature or chemistry thousands of years later than the shallower one.

Before any such apparent "leads and lags" (Pisias *et al.*, 1975; Luz, Shackleton, 1975; Moore *et al.*, 1977) can be discussed in oceanographic terms, the effects of vertical mixing on the deep sea floor must be considered. Evidence for benthic mixing was given by Suess (1956), who found that ^{14}C dates of core-top deep sea sediments were surprisingly old. He stated (p. 2) that "an admixture of old carbonate can be seen from the fact that the total carbonate in the fine fraction has an apparent age sometimes greater by almost 2 000 years than the coarse fraction, which consists mainly of foraminifera tests.". Thus, he not only recognized mixing *per se*, but also its differential effects on the various size fractions. He also realized that the admixture of young carbonate into older sediments will make subsurface ^{14}C ages appear too young. Much thought has been given to these problems since (Berger, Heath, 1968; Guinasso, Schink, 1975; Peng *et al.*, 1978).

The smooth freshly cut face of the core (Fig. 3 *a*) belies the true mixed-up nature of the record, which becomes visible after gentle washing (Fig. 3 *b*). The close-up (Fig. 3 *c*) shows a large burrow at 13 cm depth which is associated with a distinct "spreiten" pattern (*Teichichnus*; see Ekdale, Berger, 1977). Obviously, any samples taken from such burrow stuffings would be uninterpretable, especially since the feature would not normally be recognized during sampling. No amount of mathematical manipulation can eliminate such errors from "lumpy" mixing. Furthermore any averaging of replicate subsamples, or values from different fractions of the same sample, also would hardly decrease this particular type of noise (although it might decrease other types).

Effects of lumpy mixing can be seen in the oxygen isotope stratigraphies of ERDC-92. For example, the sharp gradients of the signal of *G. sacculifer* between 23 and 25 cm, of *P. obliquiloculata* at 24 cm, of *G. tumida* and *N. dutertrei* at 19 cm are rather unreasonable considering the expected effects of homogeneous mixing. The extremely steep gradient of the coccolith signal between 31 and 32 cm is another striking example of "impossible" contrast. The very heavy value of $+0.3°/_{oo}$ is unexplained, perhaps it represents a non-mixed parcel of sediment from a glacial maximum.

There is an additional complication. Mixing not only changes the gradients in the isotope signal, it also changes the gradients in the proportion of the signal carriers themselves. This effect can only be neglected if the proportions of foram species or other fractions do not change much. Clearly, this is not the case in several of the

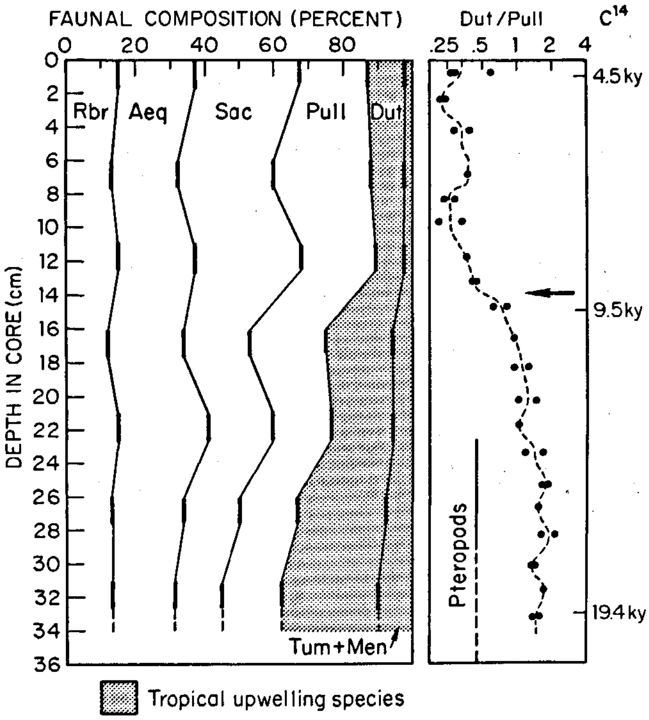

Figure 4
Faunal stratigraphy of Box Core ERDC-92 (particles > 295 µm) (from Berger, 1977). Rbr, Globigerinoides ruber, *Aeq,* Globigerinella aequilateralis (= G. siphonifera)*; Sac,* Globigerinoides sacculifer*; Pull,* Pulleniatina obliquiloculata*; Dut,* Neogloboquadrina dutertrei*; Tum,* Globorotalia tumida*; Men,* Globorotalia menardii (= G. cultrata)*. Dut/Pull: upwelling index. Arrows shows maximum change.* ^{14}*C dates from Peng* et al. (1978).

components (see Fig. 4). *G. tumida* and *G. cultrata* for example, decrease in proportion from glacial to postglacial time, as does *N. dutertrei*. The maximum change is between 14 and 15 cm (corresponding to an apparent ^{14}C age of about 9 000 years). The effect of this change in proportion is to keep the isotopic signal at relatively heavier values, in going from glacial to postglacial. This occurs because a relatively small admixture of old sediment will carry a relatively high proportion of isotopically heavy signal carriers. In addition, it must be remembered that the percentage of forams refers to the size fraction greater than 295 µm. If the proportion of this size fraction changes (as it does, see Fig. 5), additional corrections are necessary when attempting to recover the original signal. Finally, the proportion of carbonate changes also (Fig. 5). Since changes are largely confined to the silt and clay fractions, their effects in distorting isotopic signals through benthic mixing will be restricted to those fractions. Again, opportunities arise for creating leads and lags in various signals due to mixing. We will not elaborate further on this phenomenon of what may be called "mictic phase shifts". One way to deal with this complex of problems is to investigate a series of cores, correlate and average the signals, much as Emiliani (1955) did in deriving his generalized curve, except in greater detail. Only when a standard curve is available for the "true" signal will it be possible to interpret the deviations correctly. An attempt to produce such a "true" signal for the last 18 000 years has been published, including data from ERDC-92 (Berger *et al.*, 1977).

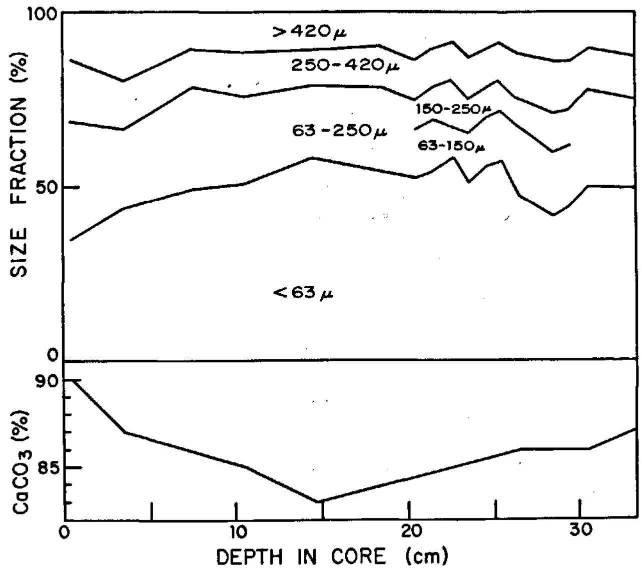

Figure 5
Stratigraphy of particle sizes and of carbonate percentage in Box Core ERDC-92.

The thickness of the mixed layer is a crucial factor in the interpretation of a bioturbated record. In core ERDC-92 the ^{14}C age of surface sediment (4 500 years) and the overall sedimentation rate (1.7 cm/ky) indicates a mixed layer thickness near 8 cm (the product of age and rate). This estimate is high because sedimentation rates are expected to drop in the postglacial, due to a drop in fertility (see Fig. 4, Pull/Dut signal), and because it refers to bulk carbonate, rather than to sand-sized particles. Additional ^{14}C determinations (Peng *et al.*, 1978) are in accordance with a mixed layer thickness of about 8 cm, for bulk carbonate.

It is interesting that the δ^{18}O isotopes of the two fine fractions (38-63, and 20-38 *µm*) are clearly distinct. However, they fall within the range of values for the foraminiferal size fractions, and the species *G. sacculifer*, on the whole.

If the Holocene coccolith fraction precipitated carbonate in thermodynamic equilibrium with surface water (29°C and salinity 34.5‰ from Reid, 1969) then using the paleotemperature equation as modified by Craig (1965) and assuming a water-oxygen value equal to −0.3‰ PDB at 34.5‰ salinity (Craig, Gordon, 1965), the equilibrium isotopic composition of calcium carbonate should be −3.0‰ PDB. The fact that the measured values average around −2.0‰ could indicate an average precipitation temperature for the coccoliths, in the Holocene, of about 27°C which corresponds to an estimated depth range of 50 to 150 m (Rotschi *et al.*, 1972). Alternatively the 1.0‰ difference could be due to non-equilibrium precipitation of coccolith test material or to seasonal effects or to some combination of these.

Oxygen isotope ranges in foraminiferal species

The oxygen isotopic composition of tests within any one foram species varies with test size and morphology

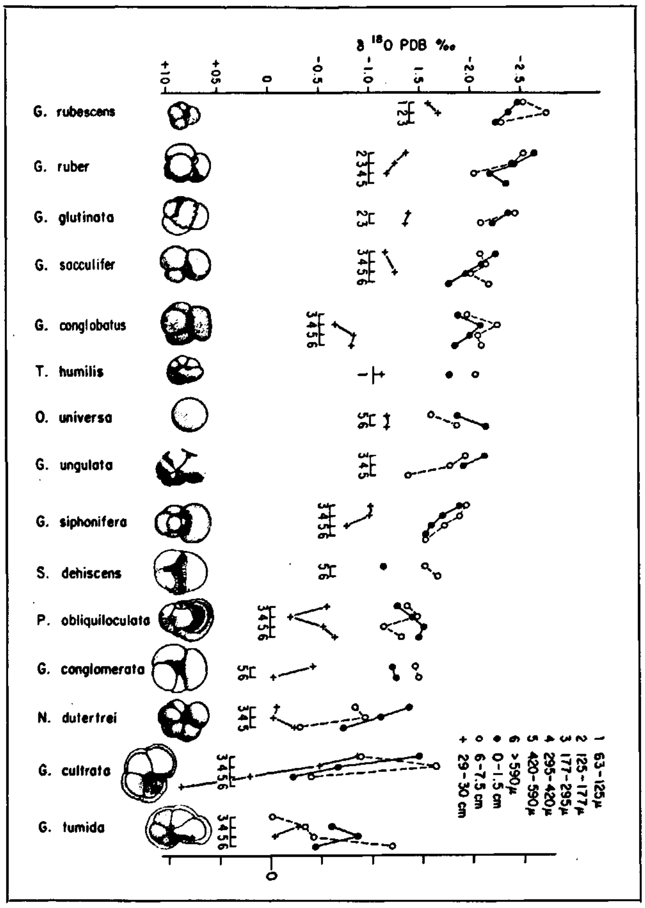

Figure 6

Oxygen isotope composition of selected foraminiferal species, as a function of size, at three stratigraphic levels in Box Core ERDC-92 (0-1.5; 6-7.5 *and* 29-30 cm; *corresponding to recent, hypsithermal, and glacial maximum, respectively). Foraminifera illustrations are from Parker* (1962) *except those of* Orbulina universa *and* Globorotalia ungulata. *The illustration of* G. ungulata *is from Blow* (1969).

(Emiliani, 1954, 1971; Oba, 1969; Hecht, Savin, 1970, 1972; Weiner, 1975; Vergnaud-Grazzini, 1976; Kahn, 1977). This relationship was first investigated by Emiliani (1954). He divided the populations of various species into two size groups (250 to 500 µm, and 500 to 1 000 µm) and determined the oxygen isotope composition for each class. Only the samples of *Orbulina universa* showed an appreciable difference between the two groups, the larger one yielding a lower $\delta^{18}O$. From this he concluded that this organism migrates upward as it grows. (No such conclusion is in fact warranted, since spherical *Orbulina* are in all likelihood a terminal stage of growth, that is, the difference between little ones and big ones does not reflect differences in growth stages, but differences in terminal size attained).

Subsequently, Emiliani (1971) published the data of this initial study, together with additional determinations on three size classes. He found (p. 1122) that the following species "appear not to change the ^{18}O concentration in their shells during growth": *Globigerinoides ruber, G. sacculifer, N. dutertrei, P. obliquiloculata, Sphaeroidinella dehiscens.* Other species showed "a marked increase in $\delta^{18}O$ with specimen age in the Caribbean": *Globorotalia menardii* (*G. cultrata*), *Globorotalia truncatulinoides.* Also "in marked contrast to the Caribbean, *G. menardii* does not appear to change appreciably its depth habitat in the Pacific and Indian Oceans," whereas *G. tumida* does. Emiliani (1971) also considered the difference in size dependence of the $^{18}O/^{16}O$ ratio between glacial and interglacial conditions. He finds a smaller range during glacials. This to him suggests that depth habitat changes are more important than seasonal effects, the latter being presumably increased during glacials.

We have repeated Emiliani's experiments in somewhat greater detail analyzing several size fractions in the fourteen most common species in Core ERDC-92. Analyses were made for three levels: surface sediment (0-1.5 cm), climatic optimum (6-7.5 cm) and glacial condition (29-30 cm). Our results (Fig. 6) show that, in general, the larger shells within each species tend to be enriched in ^{18}O. We will call this trend "normal". We have no evidence that this trend can be interpreted solely as a migration during growth, since we do not know the relationships between size, age, and shell growth. In the following species, the trend is clearly "normal": *Globigerinoides ruber, Globigerinita glutinata, Globorotalia ungulata, Globigerinella siphonifera.* In *Orbulina universa*, the trend is "reversed" in agreement with Emiliani's (1954) results. In the following, trends are not clear or they are "mixed," that is, they differ between levels: *Globigerina rubescens, Globigerinoides conglobatus, G. sacculifer, Pulleniatina obliquiloculata, Globoquadrina conglomerata, Neogloboquadrina dutertrei, Globorotalia cultrata, Globorotalia tumida.* Note that four of these species (*G. rubescens, G. sacculifer, N. dutertrei* and *G. cultrata*) would appear in the "normal" category if we consider surface sediment only.

Several possible explanations can be given for the trends observed: 1) Growth is linked to a systematic change of depth habitat, in some cases. This is the hypothesis of Emiliani (1954, 1971). It implies that small specimens as found in the sediments are juveniles who were somehow prematurely terminated. There is some evidence for greater concentration of juveniles in surface waters, and a greater proportion (not necessarily concentration) of larger individuals deeper in the mixed layer (Berger, 1971). However, these distributional patterns may largely result from the loss of spines and sinking which accompanies the reproduction process (Bé, Anderson, 1976). Shell material would not necessarily be added during such sinking. 2) Small adults are more abundant than large ones under certain conditions (related to temperature) and *vice versa*. For example the "reverse" trend of *O. universa* can be explained by a correlation of temperature with size of adults (Bé, Duplessy, 1976). 3) Adults which fail to reproduce for some reason may sink and keep on building shells, contributing to a "normal" trend. There is evidence for such a process (Bé, Lott, 1964; Bé *et al.*, 1966; Berger, 1971). 4) Shells are deposited out of equilibrium (Vergnaud-Grazzini, 1976), and the degree of disequilibrium changes during growth.

Clearly, the problem of the relationship between oxygen isotopic composition and shell size is closely linked to the questions of foram life cycles and the stages of the life cycles at which empty shells are produced. These questions are unresolved.

In addition to the $\delta^{18}O$ versus size relationships the results of the size analyses can serve to "fingerprint" the various species isotopically. Some characteristics which are readily extracted are given in Table 2.

The rank sequence of the species, with respect to postglacial isotopic composition, from a $\delta^{18}O$ value of -2.43 ± 0.2 (*G. rubescens*) to one of -0.55 ± 0.38 (*G. tumida*) agrees closely with the depth habitat ranking of Berger (1971) (rank difference correlation: $r_d = 0.9$). Thus, about 80% (square of correlation coefficient) of the variation in the average isotopic composition can be formally "explained" by a depth habitat effect, in agreement with suggestions by Emiliani (1954, 1971), Savin and Douglas (1973) and other workers. *G. cultrata* which has been listed as having a "variable" depth habitat (Berger, 1971), has the largest standard deviation of the mean isotopic composition (0.6‰) in "postglacial" sediments, and the largest range (1.4‰). Generally, the ^{18}O rich species (colder, deeper) have a greater spread of values than the ^{18}O poor species. This can be reconciled with one type of habitat straddling the uppermost part of the thermocline, versus another type strictly within the mixed layer.

The rank sequence of average isotopic composition for the glacial level is not exactly the same as for the postglacial, but is very close ($r_d > 0.9$). The one species that distinctly changes its position is *G. conglobatus*, which becomes unusually ^{18}O-rich (cold) within the glacial level. Thus, its glacial/postglacial differential (1.4‰) is the largest one found. The others are generally between 0.8‰ and 1.1‰ (roughly corresponding to the magnitude of the "glacial effect"). Exceptions are the ^{18}O-rich forams *N. dutertrei*, *G. tumida*, and *G. cultrata* whose differential is distinctly less. This effect also has been previously noted by Emiliani (1954) and was discussed at some length by Lidz *et al.* (1968) and by Shackleton (1968).

If the 1‰ differential in the shallow habitat species records the glacial effect, and if any vital effect (disequilibrium) stays constant through time, then the smaller differential of the deep habitats forms would indicate that these shells were grown in warmer water during glacials, presumably indicating an upward migration. One possible explanation is contained in Emiliani's (1971) suggestion that depth habitats are ultimately tied to food supply and hence penetration of light. Glacial time is characterized by increased production, hence reduced light transmission. Levels of equal photosynthetic activity, therefore, will be relatively shallow during glacial time and relatively deep during postglacial time. Hypotheses having to do with gradients of density and of osmotic pressure also are conceivable (see Savin, Douglas, 1973). Such mechanistic explanations are inherently less attractive for finding the motives of behavior of organisms, which have to be primarily worried about food and reproductive success. This is not to exclude the possibility that such gradients serve as signals for the guidance of any depth adjustment, as long as the gradients are correlated with environmental factors that bear on food, survival, or reproduction.

One more note of caution on the values labeled "glacial/postglacial differences" in Table 2. They are not strictly comparable from one species to the next, because in some species the full size spectrum was not available for both the postglacial levels and the glacial one. Since the differential depends on which size fractions are chosen, it is clear that a switch in size spectrum from one level to the other, however slight, will affect results. This

Table 2

$\delta^{18}O$ characteristics of planktonic foraminifera for "postglacial" (0-1.5 *and* 6-7.5 *cm*) *and "glacial"* (29-30 *cm*) *sampling intervals. Data given in Figure* 7. *(Per mil with respect to PDB.)*

Species	Trend	Average "postglacial"	Range "postglacial"	Average "glacial"	Range "glacial"	Difference of averages "glacial"/ "postglacial"
1. *Globigerina rubesens*	Mixed	−2.43±0.2	0.5	−1.67	0.2+	0.8
2. *Globigerinoides ruber*	Normal	−2.36±0.2	0.6	−1.26	0.2+	1.1
3. *Globigerinita glutinata*	Normal	−2.28±0.2	0.3	−1.37	0.1	0.9
4. *Globigerinoides sacculifer*	Mixed	−2.06±0.15	0.5	−1.20	0.1+	0.9
5. *Globigerinoides conglobatus*	Mixed	−2.02±0.2	0.4	−0.77	0.2+	1.4
6. *Turborotalita humilis*	—	−1.91±0.2	0.3	−1.12	—	0.8
7. *Orbulina universa*	Reversed	−1.86±0.2	0.5	−1.17	0	0.7
8. *Globorotalia ungulata*	Normal	−1.78±0.3	0.7	—		—
9. *Globigerinella siphonifera*	Normal	−1.73±0.16	0.4	−0.93	0.25+	0.8
10. *Sphaeroidinella dehiscens*	—	−1.44±0.3	0.6	—	—	—
11. *Pulleniatina obliquiloculata*	Mixed	−1.36±0.13	0.4	−0.49±0.2	0.4	0.9
12. *Globoquadrina conglomerata*	Mixed	−1.34±0.1	0.25	−0.24	0.4	1.1
13. *Neogloboquadrina dutertrei*	Mixed	−0.87±0.36	1.0	−0.12	0.3	0.75
14. *Globorotalia cultrata*	Normal	−0.87±0.6	1.4	−0.7	1.1	0.8
15. *Globorotalia tumida*	Mixed	−0.55±0.38	1.2	−0.16	0.2+	0.4

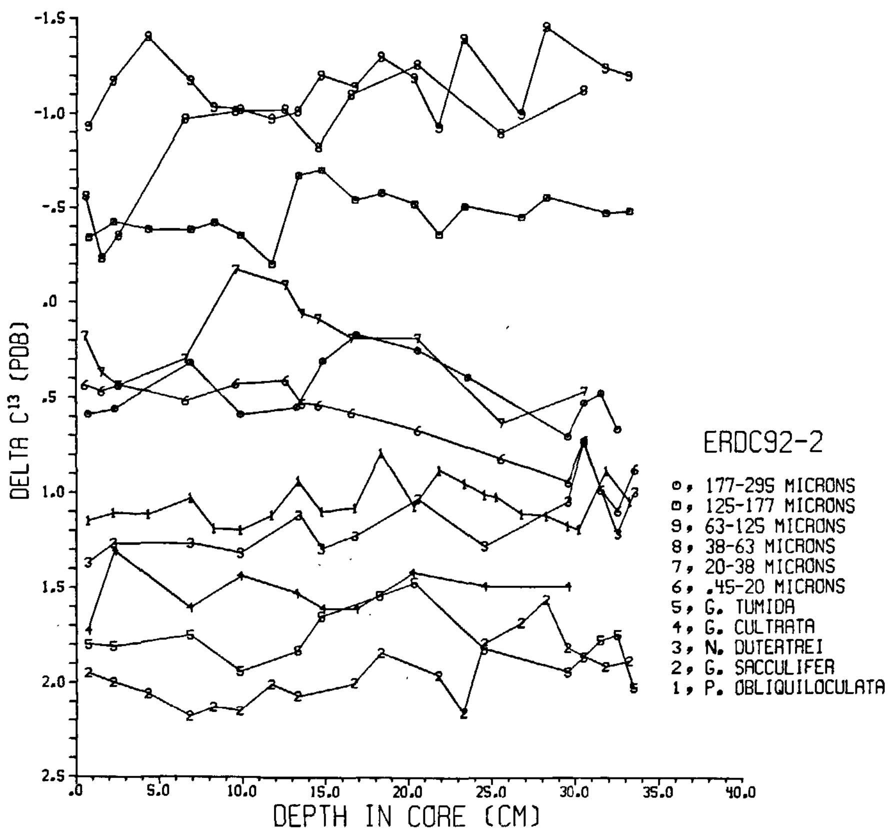

Figure 7
Stratigraphy of carbon isotopes for various size fractions and selected foram species in Box Core ERDC-92.

is quite a general problem with oxygen isotope stratigraphies which apparently has received little attention.

Carbon isotope stratigraphy

Another dimension is added to the study of carbonate sedimentation by considering the stable carbon composition (Broecker, 1973; Savin, Douglas, 1973; Vinot-Bertouille, Duplessy, 1973; Saito, van Donk, 1974; Anderson, Cole, 1975; Margolis *et al.*, 1975; Weiner, 1975; Shackleton, Kennett, 1975*a*, *b*; Vergnaud-Grazzini, 1976; Kahn, 1977; Kroopnick *et al.*, 1977; Shackleton, 1977; Williams *et al.*, 1977; Berger *et al.*, 1978; Shackleton, Vincent, 1978).

In Core ERDC-92, *G. sacculifer* is C-isotopically heavier than *P. obliquiloculata* by about 0.8‰ (Fig. 7), presumably indicating a closer association of *G. sacculifer* with the ^{12}C-depleted shallow water layer (Duplessy, 1972; Kroopnick, 1974). However, the elegance of this simple interpretation suffers somewhat when it is noted that *N. dutertrei*, *G. cultrata*, and *G. tumida* do not fit the same trend. Since their habitat is thought to be deeper than that of *P. obliquiloculata* (see Fig. 1), the increase in the $^{12}C/^{13}C$ ratio that goes with decreasing oxygen values (Deuser, Hunt, 1969; Craig, 1970; Duplessy, 1972; Kroopnick, 1974) should result in a further decrease of ^{13}C, beyond that of *P. obliquiloculata*. Clearly, this is not the case.

Before considering these unexplained differences between the various species (and also size fractions) we discuss the stratigraphic trends as such.

Assuming that upwelling decreased from the glacial to the Holocene (see Fig. 4), one might expect a trend from light $\delta^{13}C$ to heavy $\delta^{13}C$ values. Presumably, ^{12}C (light $\delta^{13}C$) goes parallel with nutrients, because like these it is freed upon oxidation of organic matter, which is ^{12}C enriched (Sackett *et al.*, 1965). The expected trend is

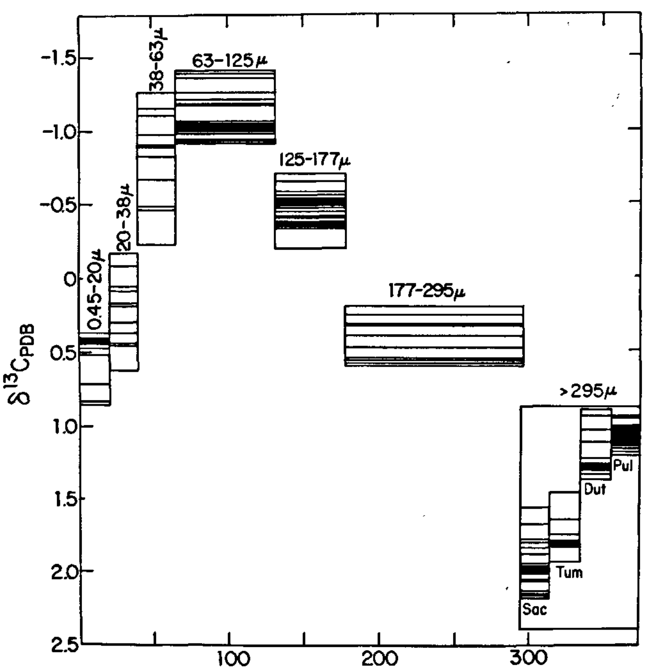

Figure 8
Carbon isotope composition of calcareous particles in Box Core ERDC-92 as a function of size. Sac, Globigerinoides sacculifer; *Tum,* Globorotalia tumida; *Dut,* Neogloboquadrina dutertrei; *Pul,* Pulleniatina obliquiloculata. *Each horizontal line represents one determination.*

indeed visible in the $\delta^{13}C$ stratigraphy of *G. sacculifer* (Fig. 8). It is also seen in the curves for 177-295 μm (which is rich in *G. sacculifer*) and in the coarse silt fraction, 38-63 μm. The other signals do not clearly show the expected trend, or show the reverse (20-38 μm).

It is evident that the mean values of $\delta^{13}C$ vary greatly between the various fractions, much more than between glacial and postglacial (Fig. 8).

Perhaps the most striking facet of the large spread in $\delta^{13}C$ values is the great enrichment of the foram fine fractions with ^{12}C. This is entirely unexpected because the oxygen values are so similar to those of *G. sacculifer* (Fig. 1). Thus, while the small tests apparently are made at the same temperature and salinity as the shells of *G. sacculifer*, they are produced at entirely different $\delta^{13}C$ concentrations or show entirely different deviation from thermodynamic equilibrium. A partial explanation may be that the smallest tests are made during times of upwelling, a hypothesis consistent with earlier evidence relating small tests to high fertility (Berger, 1969). However, the differences between the size fractions are decidedly too great to explain the observed spread entirely by a correlation between fertility and ambient $\delta^{13}C$.

The sea surface temperature in the Ontong-Java Plateau regions is about 29°C year-round (Reid, 1969). Measurements of $\delta^{13}C$ in dissolved inorganic carbon (mostly bicarbonate) in the Pacific show a gradation from about $+2$‰ at the surface to a minimum value of around 0‰ at a depth of a few hundred meters with occasional minima as low as -0.5‰ (Craig, 1970; Kroopnick *et al.*, 1970; Kroopnick, 1974). Deep water has a $\delta^{13}C$ value close to 0‰ in the equatorial Pacific.

Figure 2 shows reconstructed isotope profiles for the Ontong-Java area from surface to 500 m. The equilibrium ^{13}C curve for $CaCO_3$ was plotted assuming that the fractionation relationship of Emrich *et al.* (1970) are applicable.

The observed $\delta^{18}O$ values of recent-foraminifera (Fig. 6) when superimposed on the calculated equilibrium oxygen $CaCO_3$ curve (Fig. 2*c*) show a reasonable correspondence to temperature-depth habitats. For example *G. sacculifer* values indicate a temperature range of 25-28°C whereas those for *G. tumida* show a range of 19-23°C. These temperatures are compatible with the estimated temperature ranges of Savin and Douglas (1973) for the same species (20-30 and 15-22°C respectively).

The compatibility of the $\delta^{18}O$ values with equilibrium precipitation is in sharp contrast to the incompatibility of $\delta^{13}C$ values (Fig. 9). Clearly all of the carbon signals are too light with respect to carbon equilibrium precipitation. Similar conclusions were drawn by Kahn (1977), Williams *et al.* (1977) and by Shackleton and Vincent (1978) from isotopic measurements on various planktonic foraminiferal species.

It appears virtually certain that none of the finer fractions we measured for $\delta^{13}C$ were precipitated at carbon equilibrium. The coccolith fraction was probably deposited between about -3 and -4‰ from equilibrium and small forams in the 63-125 μm fraction possibly precipitated their calcium carbonate carbon as much as -6‰ from equilibrium.

The considerable enrichment in ^{12}C of small foram species compared with the larger ones is interesting. Some possible explanations are: 1) a partial utilization of dissolved CO_2 (-8‰) in their shell-building processes; 2) recovery of carbon for shell-building from the degradation of organic molecules (-20‰); 3) rapid $CaCO_3$ production of the initial structure of the test so that there is, in effect, kinetic fractionation operating with consequent ^{12}C enrichment. Such an initial nucleation followed by a slower, near-equilibrium thickening and enlarging phase would presumably result in small forams showing much lighter carbon than larger, more massive forms.

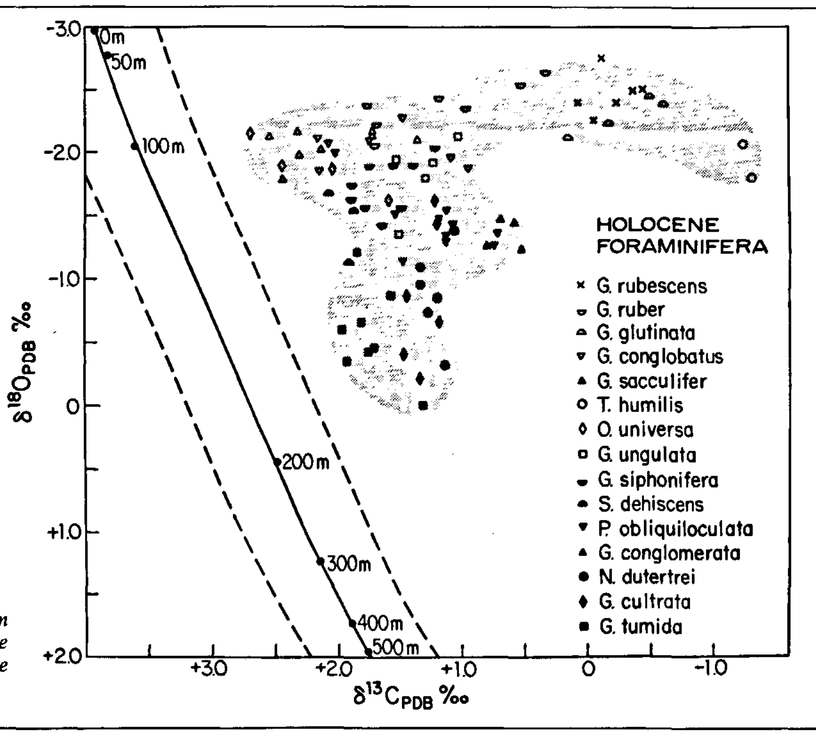

Figure 9

$\delta^{18}O_{PDB}$ *versus* $\delta^{13}C_{PDB}$ *for 0-1.5 cm and 6-7.5 cm intervals of Box Core ERDC-92. Also shown is the estimated* ^{18}O-^{13}C *equilibrium curve. (Dashed lines are limits of estimated error in equilibrium curve).*

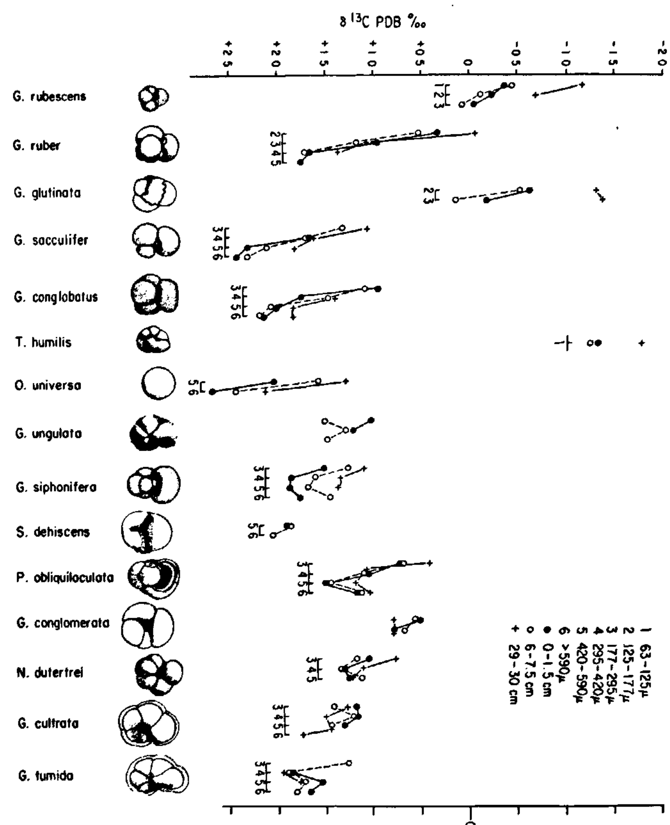

Figure 10
Carbon isotope composition of selected species as a function of size in Box Core ERDC-92. Sizes and levels as in Figure 6.

Whatever the causes, the close correlation of size and $\delta^{13}C$ in the carbonate fraction of calcareous ooze should prove to be a valuable tool in determining the transfer of carbonate from one size class to the next upon breakup of foram shells and coccoliths during dissolution.

Carbon isotope ranges within foram species

The carbon isotopic composition of shells within individual species of planktonic foraminifera varies considerably (Fig. 10). Rather than producing a large scatter, these variations are of a very regular kind: generally, the ^{13}C content increases with size, and generally, the glacial specimens are enriched with ^{12}C relative to the postglacial ones. We will call this the "normal" condition. There are only two clearly anomalous patterns: the one in *G. tumida* and the one in *G. cultrata.*

With respect to comparison between species, we note that there is no evidence that the average $\delta^{13}C$ content reproduces the ranking derived from $\delta^{18}O$, that is, there seems to be no obvious correlation of (inferred) depth habitat with the $\delta^{13}C$ average. However, regarding the *range* of values within each species, there does seem to be an indication that the oxygen isotopically light forms are characterized by large $\delta^{13}C$ differences between sizes, as well as between age levels. The $\delta^{13}C$ data given by Savin and Douglas (1973) also shows this trend of increased variability in the shallow water species (*G. ruber* and *G. sacculifer versus G. tumida* and *G. truncatulinoides*).

We have argued earlier that the very light $\delta^{13}C$ values seen in the fine foram fractions indicate disequilibrium precipitation, presumably due to the "vital" effects originally postulated by Urey (1947). If so, the increase of $\delta^{13}C$ with size suggests that this "vital" effect decreases with size, and hence presumably with age and rate of growth. One of us (Berger, 1969) suggested that shell building has an "automatic" quality which allows it to proceed in the correct proportion to the build-up of protoplasm during active growth, but manifests itself as excess thickening when growth of protoplasm slows or ceases. This idea was based on earlier observations of Bé and coworkers (Bé, Ericson, 1963; Bé, Lott, 1965) and on consideration of steady state buoyancy. The present data could be interpreted in the light of this hypothesis: during active growth (small size), the shell material precipitated reflects the metabolic processes within the cell, during slow or terminal growth (large size) metabolism all but ceases, but considerable shell material is still being added. This material tends to go toward equilibrium values.

Incidentally, the same mechanism could also explain at least part of the "normal" oxygen isotope effect: if protoplasm growth ceases while shell growth continues, the individual foraminifera will sink and hence build its shell in increasingly cooler water.

What might be the advantage in such sinking behavior? In order to reproduce, a foraminiferal cell has to clean itself from food residues, and prepare for meiosis or mitosis. A good place for this activity is a depth level where predation is low. However, the chance of the juveniles to reach the food-rich upper layers must not be sacrificed. The best depth level therefore is the uppermost thermocline. That this level is the site of production of empty shells has been proposed earlier, on the basis of net tow data (Berger, 1971).

If our conjecture about the significance of the "normal" $\delta^{13}C$ trend within planktonic foram species has merit, there should be interesting correlations of $\delta^{13}C$ with morphology: the presumably slowly growing kummerforms should have high $\delta^{13}C$ values. We have evidence from a detailed study of *N. dutertrei* that this is indeed so.

This evidence, however slight, and the internal consistency of the data at hand embolden us to present a general model of foraminiferal carbon isotope composition (see Fig. 11). It is a relationship between size and $\delta^{13}C$ composition which has the shape of a "V" with unequal limbs. The long limb, the "disequilibrium" part, corresponds to rapid growth and high metabolic activity. The short limb, the "approaching equilibrium" part, corresponds to slow growth and low metabolic activity.

The small, presumably fast growing species, are well out of equilibrium, due to a high contribution of metabolic carbon to shell material. The large shallow water species have fast growing early stages, but come closer to equilibrium at later growth stages. As they sink they record the relatively low $\delta^{13}C$ values associated with lowered dissolved O_2 values in the water. The life history of the deep living forms is analogous, except that their maximum $\delta^{13}C$ reflect the lighter dissolved carbon of the deeper water. Incidentally the $\delta^{13}C$ values of *S. dehiscens* can be considered the short limb of the "V" of *G. sacculifer*, if we follow the suggestion of Bé (1965) that these two forms are the same species.

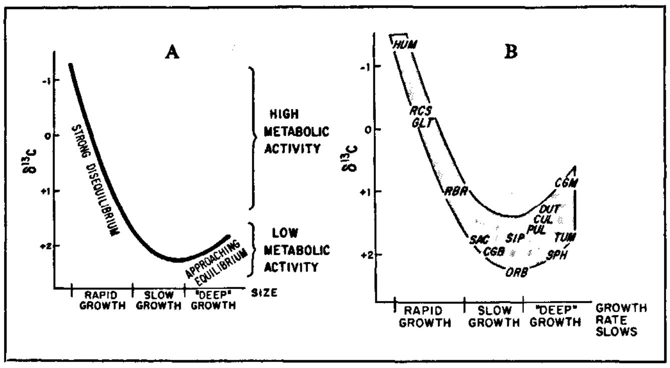

Figure 11

A, model of $\delta^{13}C$ *content of foram shells. B, approximate central position of selected species within the model: HUM,* Turborotalita humilis; *RCS,* Globigerina rubescens; *GLT,* Globigerinita glutinata; *RBR,* Globigerinoides ruber; *CGB,* Globigerinoides conglobatus; *SAC,* Globigerinoides sacculifer; *CUL,* Globorotalia cultrata; *SIP,* Globigerinella siphonifera; *ORB,* Orbulina universa; *DUT,* Neogloboquadrina dutertrei; *PUL,* Pulleniatina obliquiloculata; *CGM,* Globoquadrina conglomerata; *TUM,* Globorotalia tumida; *SPH,* Sphaeroidinella dehiscens.

SUMMARY AND CONCLUSIONS

The investigation of ERDC-92, a box core from the uppermost part of Ontong-Java plateau, taken close to the equator, yielded the first detailed isotopic stratigraphy of the Glacial-Holocene transition in deep sea sediments. The following conclusions were drawn, in part tentative:

1) The amplitude of the glacial/postglacial $\delta^{18}O$ signal is about 1.1 to $1.5°/_{oo}$ with the isotopically heavier shells showing the smaller range.

2) *P. obliquiloculata* is about $0.8°/_{oo}$ ($\hat{=}3.3°C$) heavier than *G. sacculifer*, throughout the entire time interval.

3) The "glacial effect" in our data (uncorrected for bioturbation) is about $1°/_{oo}$ or slightly larger (comparing favorably with the $1.2°/_{oo}$ sugested by Shackleton, Opdyke, 1973).

4) The fine sand fraction (small forams) has an oxygen isotope signal much like *G. sacculifer*, as does the coccolith fraction.

5) Benthic mixing introduces considerable noise into the isotopic sequences, by "lumpy mixing".

6) Coccolith oxygen isotopes (and also those of mixed small foraminifera) are somewhat heavier than expected for equilibrium with average temperatures of the surface water. Seasonal variation in shell output may contribute to this discrepancy (that is, increased output during increased upward mixing of subsurface waters), as well as "vital" effects.

7) Oxygen isotope values of small shells tend to be lighter than those of large shells, within species, except in *O. universa*, as previously noted by Emiliani (1954, 1971).

8) The ranking of foram species in terms of average $\delta^{18}O$ resembles a depth habitat ranking: $r_d = 0.9$ (as suggested by Emiliani, 1954, 1971; also Berger, 1969; Savin, Douglas, 1973; Shackleton, Vincent, 1978).

9) There is evidence for a change in depth habitat of deep-living species, that is, an upward migration during glacial time.

10) There is a $\delta^{13}C$ minimum during deglaciation.

11) All of the shell carbonate is precipitated out of equilibrium (too light) with sea water, with regard to ^{13}C. The deviation is especially strong in the smallest foraminifera, both between and within species.

12) The fact that size fractions are tagged with a typical $\delta^{13}C$ signal is useful in studying breakup of shells during diagenesis.

13) Shallow water species show a higher variability of $\delta^{13}C$ values than deep water species of foraminifera.

14) The degree of deviation from equilibrium is suggested to be a measure of rate of growth, that is, intensity of metabolism of the foraminifera.

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Note: The works of Vergnaud-Grazzini (1976), Kahn (1977) and Williams *et al.* (1977) became available after the present manuscript was completed. No attempt was made to reconcile their results with ours.

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