THE INFLUENCE OF DIFFERENTIAL PRODUCTION AND DISSOLUTION ON THE STABLE ISOTOPE COMPOSITION OF PLANKTONIC FORAMINIFERA

by

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ABSTRACT

Planktonic foraminifera from plankton tows, sediment traps and sediments from the central North Atlantic were studied in order to understand how they acquire their oxygen and carbon isotope compositions. Shallow dwelling planktonic foraminifera (mostly spinose species), collected in plankton tows in the photic zone, show light isotopic compositions possibly in slight negative deviation from oxygen isotopic equilibrium.

Radioactive tracer experiments using ^{14}C and ^{45}Ca were conducted on shallow dwelling benthonic foraminifera and hermatypic corals. They show that photosynthesis of symbiotic algae within these organisms increases the amount of metabolic CO_2 incorporated into the skeleton which consequently becomes isotopically lighter. Because shallow dwelling planktonic foraminifera contain symbiotic algae it is suggested that their light isotopic compositions are also caused by photosynthetically enhanced incorporation of metabolic CO_2 in the skeleton.

Planktonic foraminifera collected in sediment traps and sediments show heavier oxygen isotope compositions that are in equilibrium for $CaCO_3$ deposited in the photic zone. At the same time the weight/individual for these foraminifera

is almost doubled compared to those from plankton tows. I suggest that these apparent equilibrium compositions are achieved by a combination of light, perhaps non-equilibrium skeletons deposited in the photic zone and isotopically heavier calcite deposited below the photic zone. The latter being isotopically heavy because temperatures are lower, metabolic activity is reduced, and photosynthesis by the symbiotic algae stops.

Dissolution of planktonic foraminifera on the ocean floor removes first the light-weight thin shelled individuals of a species population. Because these individuals are isotopically lighter, the isotopic composition of the surviving population is heavier.

The scheme described above is applied to explain the effect of dissolution on the glacial-interglaical amplitude of the Pleistocene isotopic record in the Atlantic and the Pacific Oceans. The timing of dissolution cycles in the two oceans is out of phase. Dissolution during the glacial in the Atlantic and during the interglacial in the Pacific makes the isotopic composition heavier. Preservation in the Atlantic during interglacials and in the Pacific during the glacials makes the isotopic composition lighter. The net effect is amplification of glacial-interglacial amplitude in the Atlantic and reduction of the amplitude in the Pacific.

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INTRODUCTION

The main purpose of this thesis is to follow the evolution and changes of the isotopic composition of planktonic foraminifera in their pathway from living populations in the water column to sediment populations on the ocean floor. In addition to thermodynamic fractionation, vital effects, skeleton deposition at depth and dissolution on the ocean floor are considered here to be major factors that modify the isotopic composition of planktonic foraminifera.

Chapter I describes experimental work that sheds light on the effects of symbiotic algae on the isotopic composition of benthonic foraminifera and hermatypic corals. In Chapter II observations on the isotopic composition of planktonic foraminifera from net tows and a vertical array of sediment traps demonstrate that skeleton deposition occurs at depth and affects the overall isotopic composition. Effects of dissolution were studied by comparison of shallow and deep samples and by in situ dissolution experiments. The results of these studies are found in various sections of Chapter II.

In Chapter III a scheme that describes the evolution of the isotopic composition of planktonic foraminifera is suggested, and this is used to explain the effects of dissolution cycles on the glacial-interglacial amplitude in deep sea cores during the upper Pleistocene.

CHAPTER I

THE "VITAL EFFECT" ON THE STABLE-ISOTOPE COMPOSITION OF FORAMINIFERA AND CORAL SKELETONS STUDIED BY RADIOACTIVE TRACERS

1. Introduction

Stable isotopes of oxygen and carbon in foraminiferal skeletons are widely used in paleoenvironmental studies. A basic assumption in these studies is that skeletal $CaCO_3$ is deposited at or close to isotopic equilibrium with sea water; any biogenic fractionation that may occur is assumed to be constant, and this can be accounted for. Under this assumption, depth habitat for various planktonic species, paleotemperatures, ice-volume estimates and paleoclimatological trends for the Pleistocene have been calculated (Deuser et al., 1976; Duplessy et al., 1975; Emiliani, 1954, 1955, 1966; Emiliani and Shackleton, 1974; Shackleton, 1977c; Shackleton and Opdyke, 1973, 1976). Foraminifera used in these studies usually are taken from sediment samples. However, when foraminifera from plankton tows were analyzed, the $^{16}O/^{18}O$ ratios (measured as permille deviation from a standard $-\delta^{18}O$) were often lighter (i.e. enriched in ^{16}O) than the expected equilibrium values by 0.5 to 1.5 °/₀₀. Light non-equilibrium values of up to 2-3 °/₀₀ for oxygen and carbon isotopes were also observed in shallow benthonic foraminifera (Vinot-Bertuille and Duplessy, 1973). It was pointed out (Grazzini, 1976; Shackleton et al., 1973;

Van Donk, 1970, 1977; Vinot-Bertuille and Duplessy, 1973) that different species of foraminifera as well as individuals within a species population can have different isotopic compositions although they live in the same water and are exposed to the same physico-chemical conditions. These observations conflict with the isotopic equilibrium assumption of the stable-isotope paleoceanographic method, and can impose severe limitation on the interpretation of stable isotope data. Parker (1958) suggested that symbiotic algae, existing in many species of planktonic and benthonic foraminifera (Boltovskoy and Wright, 1976) may influence the isotopic composition of the skeleton. Recently, Grazzini (1976) noted that planktonic foraminifera collected in the photic zone (possibly containing symbionts as indicated by their colored protoplasm) deviate from equilibrium more than foraminifera with 'milky' protoplasm collected below the photic zone. However, the effect of algal symbiosis on the stable-isotope composition of foraminiferal skeleton was never directly studied.

The relationship between $CaCO_3$ depositing organisms and their symbiotic algae are best demonstrated by hermatypic corals. Radioactive tracer experiments show that algal photosynthesis enhances coral calcification (Goreau, 1959, 1961, 1963) and that some metabolic carbon is incorporated into the skeleton (Goreau, 1959, 1961, 1963; Pearse, 1970). Hermatypic corals show specific and individual fractionation

of stable isotopes as well as light non-equilibrium isotopic compositions (Craig, 1957; Weber and Woodhead, 1970, 1972), again indicating incorporation of isotopically light metabolic CO_2 into the skeleton (Craig, 1957; Goreau, 1959, 1961, 1963; Pearse, 1970; Weber and Woodhead, 1970, 1972). Weber and Woodhead (1970, 1972) suggested a model emphasizing the importance of the symbiotic algae in determining the amount of light metabolic CO_2 incorporated in the skeleton of hermatypic corals. This model, however, was not tested rigorously, and does not seem compatible with some recent observations (Land et al., 1977).

In order to study more directly the effect of symbiotic algae on the stable-isotope composition of foraminifera and corals, a field experiment was conducted in the Gulf of Eilat, Israel, during the summer of 1975. Detailed description of field and laboratory procedures is reported elsewhere (Erez, 1977). Briefly, hermatypic corals and benthonic foraminifera were double labeled *in situ* using ^{45}Ca and ^{14}C as tracers to measure calcification and photosynthesis. Dark and dead controls were run with the illuminated experiments in order to evaluate the effect of photosynthesis on calcification and the non-biogenic incorporation of radionuclides. Experiments used different species at the same depth and the same species at different depths. Radioactive analysis of a sample yielded 3 fractions: (1) ^{45}Ca incorporated in the skeleton; (2) ^{14}C incorporated

skeleton (both measure calcification) and (3) ^{14}C incorporated into the soft tissue. This fraction was obtained by dissolving the skeleton and retaining the insoluble residue on a filter and represents mostly photosynthetically fixed carbon.

Radionuclides were measured by liquid scintillation. Counting efficiencies were 80% and 90% for ^{14}C and ^{45}Ca respectively. Photosynthetic rate is reported in μgC/mgIR/hr, (IR stands for Insoluble Residue), and calcification rate in μgCa/gCaCO3/hr. Precision of the analytical method (determined on 5 replicates) was ± 3.6% (1σ) for photosynthesis and ± 17% of the reported value (1σ) for calcification. Stable isotope analysis of oxygen and carbon in skeletal subsamples was carried out using commonly accepted techniques (McCrea, 1950). Prior to acid reaction, samples were combusted in a low-temperature oxygen-plasma furnace. Results are reported in °/₀₀ relative to the PDB standard.

$$\text{e.g. } \delta^{18}O = \left[\frac{(^{18}O/^{16}O)\ \text{sample}}{(^{18}O/^{16}O)\ \text{standard}} - 1\right] \times 1000$$

Precision for both isotopes is better than 0.1 °/₀₀, obtained by continuous runs of standards, as well as replicates of many samples.

2. Rates of Photosynthesis and Calcification

Rates of photosynthesis by coral symbionts (Table 1)

TABLE 1

Summary of *in-situ* radio-tracer experiments and stable isotope composition for corals and foraminifera in the Gulf of Eilat, Israel. Note the significant and consistent differences between experiments carried out in the light and the dark and dead controls.

TABLE 1

Species	Location	Depth (m)	PHOTOSYNTHESIS (µgC/mgIR/hr)			CALCIFICATION ($\mu gCa/gCaCO_3/hr$)			C^{14}/Ca^{45} ACCRETION RATE RATIO		δO^{18} (0/00)	δC^{13} (0/00)
			Light	Dark	Dead	Light	Dark	Dead	Light	Dark		
S. pistillata 1	Taba	4.5	.826	.085	.002	327	8	4	.172	1.192	-1.82	-.91
S. pistillata 2	Taba	4.5	.909	.028	.008	208	23	6	.321	.450	-1.68	-1.22
M. dichotoma	Taba	4.5	.546	.006	.002	129	19	16	.555	.945	+.69	+1.58
P. danae	Taba	4.5	1.707	.011	.0002	118	12	4	.081	.185	-2.32	-3.49
A. eurystoma	Taba	4.5	.986	.013	.004	70	9	5	.397	.538	-2.20	-1.81
S. pistillata	MBL	4	1.93	.003		508	4		.128	.281	-2.18	-1.45
S. pistillata	MBL	10	1.60	.001	.005	124	4	2	.229	.786	-1.81	-1.12
S. pistillata	MBL	15	2.04		.0004	1216		3	.073		-1.84	-1.68
S. pistillata	MBL	22	1.16	.005	.002	146	6	3	.243	1.014	-1.82	-1.57
S. pistillata	Ras Burka	20	1.48	.010		85	17	5	.471	1.523	-1.57	-.71
S. pistillata	Ras Burka	30	.90	.008		242	21		.104	1.290	-2.10	-1.55
A. variabilis	MBL	4	1.42			25			.100		-1.17	-1.69
A. variabilis	MBL	10	1.65			51		2	.344		-1.46	-.11
A. variabilis	MBL	15	2.16	.012	0	62	6	7	.351	.777	-2.13	-2.38
A. variabilis	MBL	22	2.32		.006	170			.256		-2.38	-2.18
A. lobifera	Taba	5	1.466	.036		1375	33		.162	.625	-.77	+1.12
A. lobifera	Taba	10	1.799	.045		1737	36		.187	.222	-.70	+1.35
A. lobifera	Taba	15	1.597	.052		1743	62		.323	.200	-.63	+1.27
A. lobifera	Taba	20	1.997	.048		2539	32		.192	.375	-.54	+1.24
A. lobifera	Taba	25	1.738	.052		3481					-.56	+1.36
A. lobifera	Taba	30	1.058			938					-.54	+1.13

range from 0.546 to 2.320 µgC/mgIR/hr, and are similar to those reported by Goreau (1959, 1961, 1963) for Jamaican corals. These rates average 60 times higher than the dark controls and 300 times higher than the dead controls. Photosynthetic rates in the foraminifer *Amphistegina lobifera* range from 1.058 to 1.997 µgC/mgIR/hr (Table 1); on the average these rates are 35 times higher than the dark controls.

Rates of calcification of corals based on ^{45}Ca uptake, range from 25 to 1212 µgCa/gCa/CO_3/hr. These values correspond to $CaCO_3$ accretion rates of 0.006% and 0.304%/hr, respectively (average 0.097%/hr), and are similar to those reported by Goreau (1959, 1961, 1963) for Jamaican corals. Drew (1973) reported rates of calcification for corals in the Gulf of Eilat when exposed to adverse conditions, that are an order of magnitude lower (see Erez, 1977, for detailed discussion). Calcification in the foraminifer *A. lobifera* range from 938 to 3481 µgCa/g$CaCO_3$/hr. The average calcification rate of this species which is a major constituent in the reef sediments (Erez and Gill, 1977) is 5 times higher than the average calcification rate of the fastest growing coral measured (*Stilophora pistillata*). For both corals and foraminifera there is a strong enhancement of calcification in illuminated experiments compared to dark controls. Light/dark calcification ratio average ∿ 11 for corals and ∿ 50 for foraminifera. Calcification rates in

dead controls are only half of those in dark controls. The consistent differences between illuminated experiments and the controls demonstrate the accuracy of the radiotracer method used in this study. The enhancement of skeleton deposition for corals and foraminifera in the light indicates that the symbiotic algae play an important role in this process. This confirms earlier *in-situ* observations for corals (Goreau, 1959, 1961, 1963), and laboratory studies on benthonic foraminifera (Röttger, 1972; Lee and Zucker, 1969).

In the depth profiles (Fig. 1) corals and foraminifera show mid-water photosynthesis and calcification maxima. Over-illumination in surface water may cause inhibition of photosynthesis as found in marine phytoplankton (Yentsch, 1974; Ryther, 1962). It seems that the symbiotic algae within the corals and foraminifera behave in a similar way. Because photosynthesis enhances calcification, mid-water calcification maxima are also observed (Fig. 1). It should be noted, however, that the positive correlation between photosynthesis and calcification observed here within a single species at different depths (Fig. 1) does not exist for different species at the same depth. This may indicate that different species have different calcification/photo-synthesis ratios.

3. Stable Isotope Composition

The stable-isotope results are shown in Table 1 and

Figure 1. Calcification and photosynthesis depth profiles for corals (a,b) and foraminifera (c), based on in-situ radiotracer experiments in the Gulf of Eilat, Israel. Note the reduction of photosynthesis in shallow water, possibly caused by photoinhibition, and the resulting mid-water photosynthetic maxima. Because calcification is strongly enhanced by photosynthesis, mid-water calcification maxima are also observed.

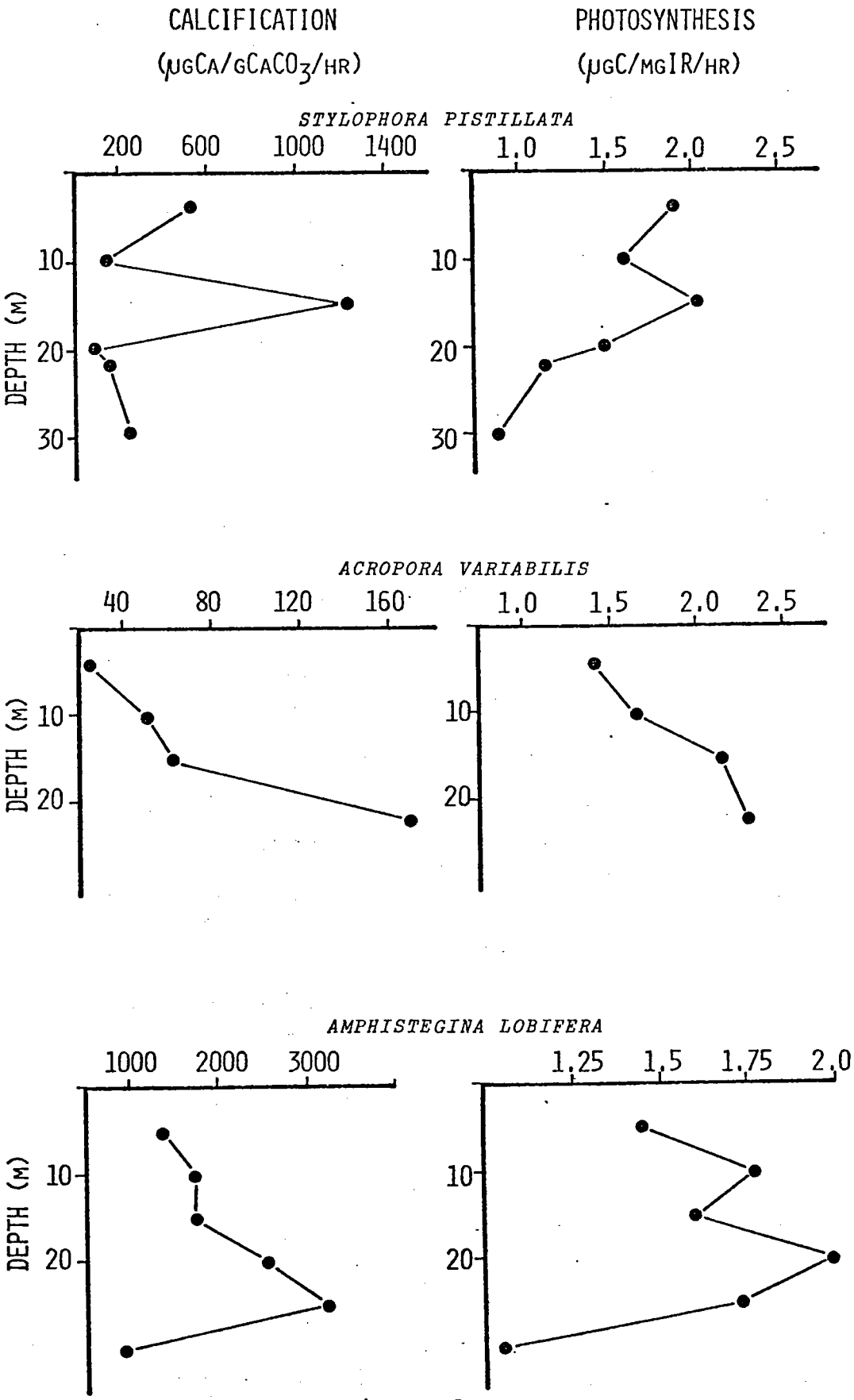

CALCIFICATION
(μgCa/gCaCO3/hr)
PHOTOSYNTHESIS
(μgC/mgIR/hr)
STYLOPHORA PISTILLATA
200
600
1000
1400
1.0
1.5
2.0
2.5
DEPTH (M)
10
20
30
ACROPORA VARIABILIS
40
80
120
160
1.0
1.5
2.0
2.5
DEPTH (M)
10
20
AMPHISTEGINA LOBIFERA
1000
2000
3000
1.25
1.5
1.75
2.0
DEPTH (M)
10
20

Figure 1.

Fig. 2. $\delta^{18}O$ range from 2.38 °/₀₀ to +0.69 °/₀₀ and $\delta^{13}C$ from -3.49 °/₀₀ to +4.65 °/₀₀. Equilibrium value for carbonate oxygen in the Gulf of Eilat is about +0.6 °/₀₀, based on the Epstein et al. (1953) equation, water $\delta^{18}O$ of +2.0 °/₀₀ (Nissenbaum, unpublished data) and average temperature of 22.9°C (Friedman, 1968). Almost all samples are much lighter than the expected equilibrium value for oxygen. The large range of values of $\delta^{13}C$ indicates that many of the corals and the foraminifera are out of equilibrium also with respect to carbon isotopes. Seasonal variations and different depth habitats cannot explain the large variation between and within species. Thus, it seems that most of the variability is ~~cuased~~ caused by a 'vital effect' (Epstein et al., 1951) involving metabolic activity in skeleton deposition. A clear trend can be observed (Fig. 2): the isotopic composition becomes lighter when rate of photosynthesis increases. This is observed for both elements (oxygen and carbon), for different species at the same depth (corals and foraminifera) and for one species (*Acropora variabilis*) at different depths. One should note that although there is considerable change in photosynthetic rates of *A. lobifera* in the depth range studied, there is relatively very small change in the isotopic composition of this species. The reasons for this are yet to be studied; however, the average $\delta^{18}O$ for this species is 1.2 °/₀₀ lighter than the expected equilibrium value. Isotopic

Figure 2. Stable-isotope composition of the skeletons vs. photosynthesis of symbiotic algae in corals and foraminifera in the Gulf of Eilat, Israel. (a) Different coral species at the same depth (4.5 m). M - *Millepora dichotoma*; S - *Stylophora pistillata*; A - *Acropora variabilis*; P - *Pocillopora danae*. (b) One coral species (*Acropora variabilis*) at different depths. (c) Different foraminifera species at one depth (∿ 10 m). A - *Amphistegina lobifera*; Ar - *Amphisorus hemprichii*; H - *Heterostigina sp.*; B - *Borelis sp.* Note that when photosynthesis increases, the isotopic composition becomes lighter. This indicates that as a result of photosynthesis by symbionts, more light metabolic CO_2 is incorporated in the skeleton of these organisms.

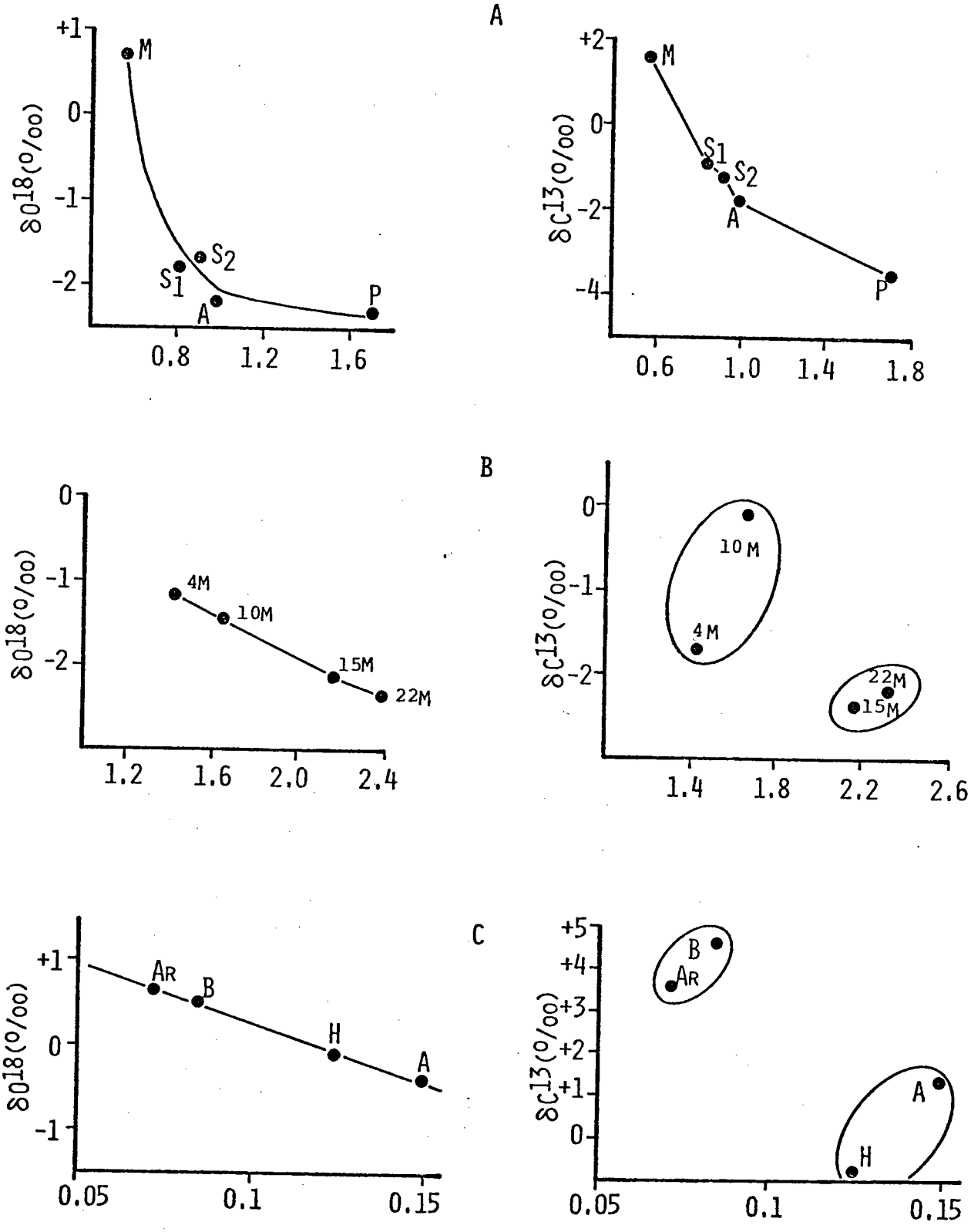

A
B
C
δO18(o/oo)
δC13(o/oo)
M
S1
S2
A
P
4M
10M
15M
22M
AR
B
H
A
PHOTOSYNTHESIS μgC/mgIR/hr

Figure 2.

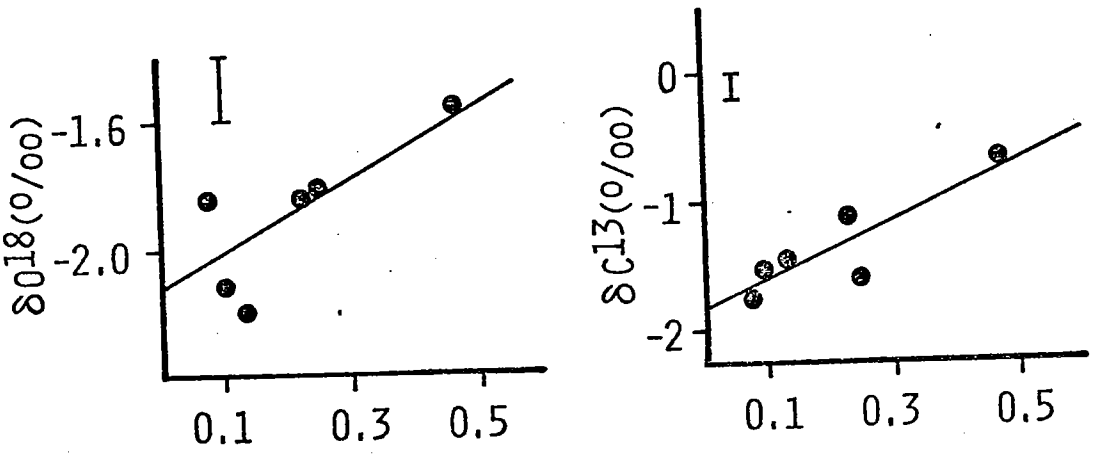
composition and rates of calcification do not correlate well; therefore, it seems that skeleton growth rate per se is not the prime factor controlling the isotopic composition, as was concluded also by Weber et al. (1976).

4. Metabolic CO_2 in the skeleton and the effect of symbiotic algae

Based on the light isotopic compositions of coral skeletons, several authors suggested that some metabolic CO_2 is incorporated in the skeleton (Craig, 1957; Goreau, 1959, 1961, 1963; Pearse, 1970; Weber and Woodhead, 1970, 1971). The present study supplies independent evidence to support this idea based on calcification rates: because double labeling technique was used, calcification rates can be measured by ^{45}Ca or ^{14}C uptake into the skeleton. Ideally these rates should be equal. However, accretion rates based on ^{45}Ca uptake always are 5 to 10 times higher than those based on ^{14}C uptake (Table 1). As a result, the $^{14}C/^{45}Ca$ accretion rate ratio is usually between 0.1 to 0.5. This phenomenon was first observed by Goreau (1959, 1961, 1963) who interpreted it as a dilution of the ^{14}C tracer by metabolic carbon pool, which is available in the coral tissue for skeletogenesis. My observations strongly support Goreau's idea. The correlations with the isotopic compositions (Fig. 3) further suggest that the dilution is indeed by isotopically light metabolic CO_2. In dark experiments, the $^{14}C/^{45}Ca$ accretion rate ratio is almost

Figure 3. The ratio between calcification rates based on ^{14}C and ^{45}Ca incorporation in the skeleton vs. the stable isotope composition in the coral *S. pistillata*. Ideally this ratio should be 1.0; the ratio is lower than expected indicating dilution of the ^{14}C tracer by metabolic CO_2. Note that when dilution by metabolic CO_2 increases the isotopic composition becomes lighter. The error bars represent 2σ of the stable-isotope analysis.

STYLOPHORA PISTILLATA



C^{14}/Ca^{45} ACCRETION RATE RATIO

Figure 3.

always higher than in the illuminated experiments, and in many experiments the ratio approaches one (Table 1) which indicates no relative dilution of the tracer. This suggests that photosynthesis increases the amount of metabolic CO_2 in the skeleton, and independently supports the observation that when photosynthesis increases, the isotopic composition of the skeleton becomes lighter (Fig. 2).

Weber and Woodhead (1970, 1971) proposed that the isotopic composition of coral skeletons is a mixture of isotopically heavy sea-water bicarbonate and isotopically light metabolic (or respiratory) CO_2. They assumed that symbiotic algae utilize metabolic CO_2 for photosynthesis and therefore predicted that when photosynthesis increases, less metabolic CO_2 is available for incorporation in the skeleton which consequently will become heavier. This prediction does not agree with the observations presented here (Fig. 2) - that higher rates of photosynthesis are associated with isotopically lighter skeletons. Weber and Woodhead's model is based on numerous observations that corals have lighter isotopic composition at 18 m than at the surface, and on the addumption that photosynthesis is maximal at the surface and decreases with depth. The depth profiles presented here (Fig. 1) as well as the profile of Barnes and Taylor (1973) show that photosynthesis is maximal at some mid-depth (15-25 m) rather than at the surface. Thus it is possible that the lighter isotopic composition observed by Weber and

Woodhead (1970, 1971) around 18 m is associated with higher photosynthetic rates, in agreement with the relationship between photosynthesis and isotopic composition presented here (Fig. 2). Recently, Weber et al. (1976) measured the isotopic composition of the coral *Montastrea annularis* in the Caribbean at different depths. They found a minimum of $\delta^{13}C$ at 18.3 m and interpreted this as support for their model (Fig. 9). In view of the present results this profile can be reinterpreted to represent a mid-water photosynthetic maximum, that is supported by other investigators (Barnes and Taylor, 1973; Land et al., 1977) (see also Erez, 1977).

5. Conclusions and Implications for Paleoceanographic Studies

It is concluded that skeletal carbonate of hermatypic corals and foraminifera is composed of a mixture of two components: (1) sea water bicarbonate that may be incorporated into the skeleton by an isotopic equilibrium process; (2) metabolic CO_2 that is enriched in ^{12}C and ^{16}O. The symbiotic algae, when active, enhance the overall metabolic activity of the host-algae complex, and increase the amount of metabolic CO_2 in the organisms internal CO_2 pool. This will increase the share of metabolic CO_2 component in the skeleton which consequently will become isotopically lighter. This model can explain the non-equilibrium compositions and the individual and specific fractionations that were reported for corals and shallow benthonic

foraminifera (Vinot-Bertuille and Duplessy, 1973; Weber and Woodhead, 1970, 1971). Because many shallow planktonic foraminifera contain symbiotic algae (Boltovskoy and Wright, 1976; Anderson and Bé, 1976), the same model may be applied for these organisms, to explain the non-equilibrium compositions and the specific fractionations they exhibit (Grazzini, 1976; Shackleton et al., 1973; Van Donk, 1970, 1977) (see Chapter II).

Paleoenvironmental interpretations (especially paleo-temperatures) based on the stable isotope composition of planktonic foraminifera that contain symbiotic algae should be made with great caution. Isotopically light skeletons, for example, can represent change in isotopic composition of sea-water or elevation of temperature (as was interpreted before), but also can represent an increase in the photo-synthetic rate of the symbiotic algae.

This is well demonstrated by the depth profile for the coral *A. variabilis* (Table 1, Fig. 2). A temperature gradient for this profile can be calculated, assuming that $\delta^{18}O$ vs. temperature curve for this species is parallel to the paleotemperature curve of Epstein et al. (1953) as suggested by Weber and Woodhead (1970, 1971). The "isotopic" temperature thus calculated at 22 m is 5°C warmer than at 4 m. Such inverted thermocline is obviously ~~erroneous~~ erroneous in the high energy reef environment of the Gulf of Eilat, where temperature is known to be constant or slightly lower with

depth (Friedman, 1968; Klinker et al., 1976). This eroneous result is caused by increase of photosynthesis in this depth interval (Fig. 2). The role of metabolic CO_2 from sources other than photosynthesis, may be of equal importance in producing a "vital effect" on skeletal isotopic composition of organisms, as suggested by Goreau (1977). This may also apply for organisms that do not have symbiotic algae (e.g. deep planktonic and benthonic foraminifera). In order to accurately calculate paleotemperatures, depth habitats, ice volume estimates, or other environmental parameters based on stable isotope composition of biogenic carbonates, the "vital effect" should first be studied and accounted for.

CHAPTER II

PATHWAY OF THE ISOTOPIC COMPOSITION OF PLANKTONIC FORAMINIFERA FROM PLANKTON TOWS TO SEDIMENT TRAPS AND SEDIMENTS

1. Introduction

Despite the fact that planktonic foraminifera from plankton tows exhibit non-equilibrium isotopic compositions (Shackleton et al., 1973; Grazzini, 1976; Kahn, 1977; Van Donk, 1970, 1977) foraminifera in surface sediments seem to be in equilibrium with respect to oxygen isotopes (e.g. Emiliani, 1954, 1955; Shackleton and Vincent, 1978; Berger et al., 1978). There is no direct proof that these are indeed equilibrium values. However, the isotopic temperatures calculated from $\delta^{18}O$ of these skeletons are within the yearly temperatures ranges for the sample locations. Furthermore, depth stratifications of foraminifera in the water column deduced from these isotopic temperatures are compatible with information from plankton tows (see Van Donk, 1977; Bé, 1977, for comprehensive reviews). The important question then becomes: what is the mechanism that converts the light non-equilibrium isotopic compositions of living planktonic foraminifera to equilibrium compositions in sediment populations. To shed some light on this problem and other related problems, I have compared the isotopic composition of living, descending and sediment populations of planktonic foraminifera at one spot in the Sargasso Sea

(Central North Atlantic). Foraminiferal populations descending in the water column were collected by an array of sediment traps at different depths. The traps have significant advantages compared to the traditional method of plankton net tows: 1) Plankton nets collect only the material that is larger than its mesh size. Usually a large mesh size (200-300 μ) is used because net clogging is a severe problem in small sizes. The sediment traps, on the other hand, do not discriminate against size fractions; 2) The traps collected material for periods of a few months, thus averaging out part of the time-variability that is observed when nets are used; 3) The traps were attached to a taut line at specific depths, while the depth at which plankton nets are towed varies (especially in the lower part of the water column); 4) Plankton tows supply information on standing stocks in the water column while sediment traps supply information on fluxes of planktonic foraminifera to the ocean floor. The composition of sediment population is determined by the input flux rather than by the standing-stock of the populations in the overlying waters. Therefore sediment traps are an advantage. Plankton tows were used in this study only to determine the isotopic composition of living populations in the photic zone. This material is bound to all the limitations mentioned above. Sediment samples were collected by box coring and only the uppermost material from the undisturbed surface was considered for this study.

2. Material and Experimental Procedures

A. Field Methods

Sediment traps: Five sediment traps with 1.5 m^2 collecting area designed and described by Honjo (1978) were deployed in the central Sargasso Sea (PARFLUX S station) at 31°34'N, 55°03'W, in 5581 m of water (see fig. 4). The first deployment (marked S1,5300) lasted for 75 days from October 19, 1976 to January 5, 1977 and consisted of two identical traps at 5367 m. The second deployment (marked S2,400; S2,1000; and S2,400) lasted for 110 days from July 20, 1977 to November 7, 1977. It consisted of three sediment traps at one mooring at depths of 372 m, 976 m and 3694 m (see table 2). Pertinent data on the fluxes, chemistry and biology of these sediment traps as well as water column chemistry in PARFLUX S station are available in Honjo (1978) and Spencer et al. (1978). After collection samples were kept at 2°C in the dark.

Plankton tows: Plankton tows were collected during the recovery of the S1 mooring and the deployment and recovery of the S2 mooring (table 2). Data on depth, time, location and mesh size of these tows is summarized in table 2. Samples were frozen until analyzed.

Box cores: Two box cores were collected by F. Grassle (W.H.O.I.) slightly northwest of the S station (table 2,

fig. 4). Visual observation of these box cores indicated that the sediment-water interface was successfully retrieved (Grassle, pers. comm.). Core OC28, 395 was raised from 4500 m. It is a calcareous ooze and the foraminifera are well preserved. Core OC28, 396 is from 4950 m and contains ∿ 50% clay. The foraminifera are moderately preserved and show signs of dissolution, chalky appearance and enrichment in thick calcified individuals.

B. Laboratory procedures

Sample preparation: The sediment trap and the plankton samples were placed in petri dishes and combusted in a low temperature plasma furnace to eliminate most of the organic matter. Samples were washed in distilled water, wet sieved through 250 µ, and dried at 50°C. The foraminifera above 250 µ were hand picked, counted and weighed. Most samples contained at least 30 individuals and in many cases over 100 individuals. *Globigerinoides ruber* was split into two morphotypes:

1) *G. ruber* larger, with wide apertures, thin shells and round last chamber. These were often red individuals.

2) *G. elongatus* smaller, narrow apertures, thicker wall and flattened or somewhat deformed last chambers.

Globigerinoides sacculifer contained many morphotypes of *G. trilobus* but they were not separated. Box core samples were wet sieved through 63 µm, dry sieved through 250 µm, and analyzed like all other samples.

TABLE 2.
SAMPLES USED FOR ISOTOPIC ANALYSIS IN CHAPTER II

PLANKTON SAMPLES

Sample	Cruise	Day Collected	Time	Mesh Size	Depth Range	Location
						31°34'N
PLANKTON 1	OC30	7/22/77	1540-1640	150 μ	0-200 m	55°03'W
PLANKTON 2+4	OC30	7/22/77	1200-1250	150 μ, 333 μ	0-20 m	"
PLANKTON 7	OC30	7/22/77	1300-1400	150 μ	0-100 m	"
PLANKTON 3	KNR63	1/10/77	night	240 μ	0-5 m	"
PLANKTON 5	KNR63	1/11/77	1100-1200	240 μ	0-200 m	"
PLANKTON 8	KNR63	1/10/77	0355-0425	240 μ	0-200 m	"
PLANKTON 6	OC34	11/13/77	2125-2225	150 μ	0-5 m	"
PLANKTON 9	OC34	11/13/77	2255-2355	333 μ	0-5 m	"

SEDIMENT TRAPS

Sample	Cruise Collected	Deployment Dates	Duration	Depth	Location
					31°34'N
S2, 400	OC34	July 20-November 7, 1977	110 days	372 m	55°03'W
S2, 1000	OC34	July 20-November 7, 1977	110 days	976 m	"
S2, 4000	OC34	July 20-November 7, 1977	110 days	3694 m	"
S2, 5300	KNR63	October 20, 1976-January 5, 1977	75 days	5367 m	"

SEDIMENT SAMPLES (Box Cores)

Sample	Cruise	Depth	Location	Interval Sampled
Box Core 395	OC28	4500 m	33°42.5'N, 57°27.0'W	Surface
Box Core 396	OC28	4950 m	34°18.0'N, 57°12.0'W	Surface

Figure 4. Location map for samples discussed in Chapter II

PARFLUX S - Plankton tows and sediment traps

OC28, 395 - Box core at 4500 m

OC28, 396 - Box core at 4950 m

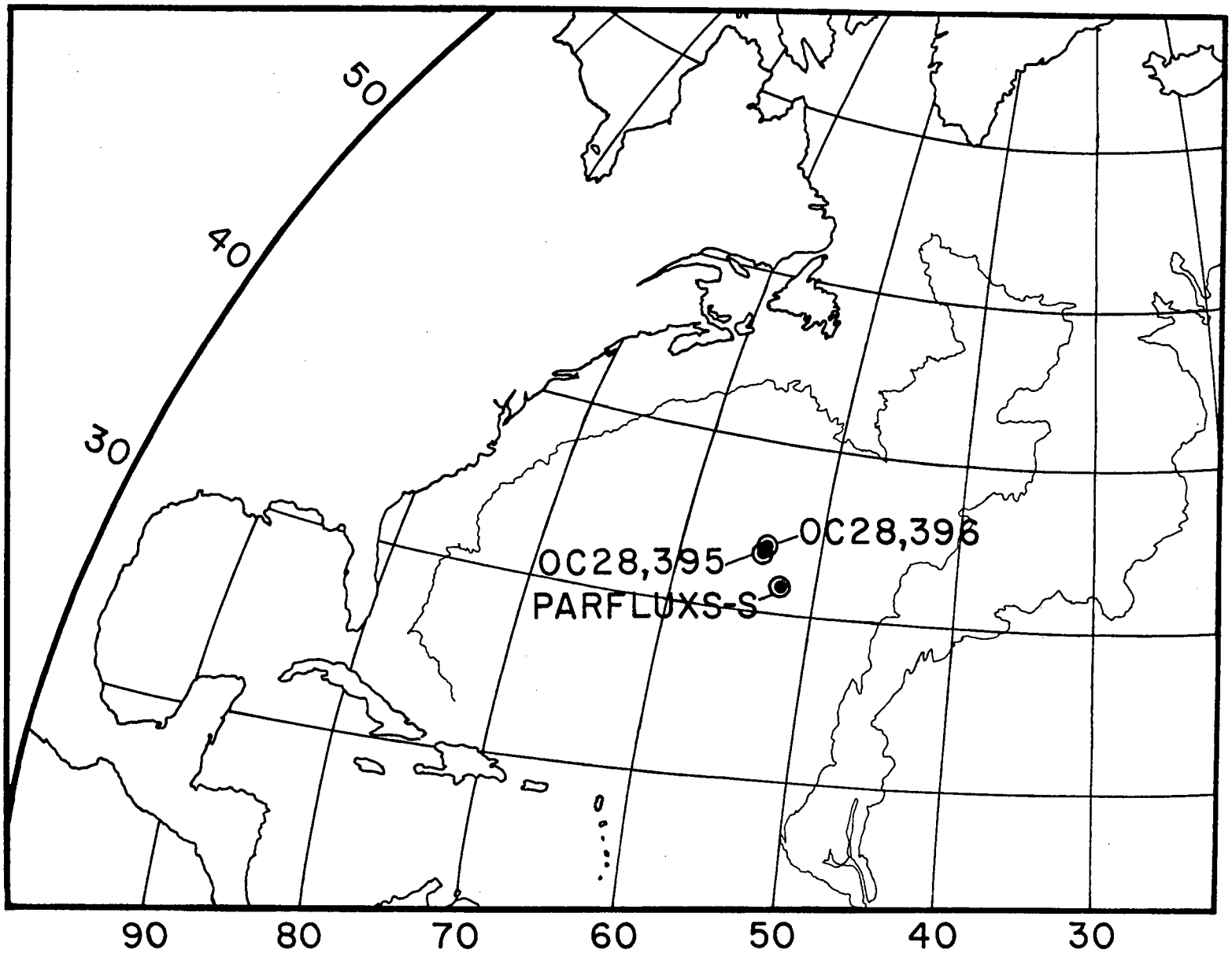


Figure 4.

Combustion: Before samples were reacted with phosphoric acid, they were crushed in pure ethanol, ultrasonicated and combusted in a vacuum furnace at 450°C for half an hour. Thus the samples discussed below were combusted twice: One combustion in a plasma furnace, which probably removed most of the organic matter on the outer surface area. Second, combustion in a vacuum furnace after they were crushed. In general, combustion tends to make the isotopic composition lighter (Epstein et al., 1953; Emiliani, 1966; Kahn, 1977). We attempted to test the effect of combustion on a few species by comparing unroasted samples to samples roasted in a vacuum furnace, plasma furnace and both. Four species were analyzed. For each, a few hundred individuals were collected from a sediment trap in the Equatorial Atlantic at 400 m. They were crushed and the fragments mixed in a glass vial. Subsamples were combusted in plasma furnace for 30 minutes and for 60 minutes, and in vacuum furnace for 30 minutes and 120 minutes (at 450°C). Portions from the plasma roasted samples were roasted additionally for 30 minutes in vacuum. The results are confusing and cannot be regarded as conclusive (table 12). It seems that different species show different trends with different treatments. In general, plasma roasting seems to make the isotopic composition lighter as it proceeds. Vacuum roasting for 30 minutes makes the isotopic composition lighter but additional roasting does not change it significantly. Samples

TABLE 12

EFFECTS OF DIFFERENT COMBUSTION METHODS ON THE ISOTOPIC COMPOSITION OF FORAMINIFERA FROM SEDIMENT TRAPS

E 389	$\delta^{18}O_{PD8}$	$\delta^{13}C_{PD8}$
G. ruber		
not roasted	-1.87	+1.06
30 min. vac. roast	-2.34	+0.90
120 min. vac. roast	-2.25	+0.96
plasma 30 min.	-2.82	+0.38
plasma 60 min.	-2.44	+0.81
plasma 60 min. + 30 min. vac.	-2.30	+0.86
G. sacculifer		
not roasted	-1.73	+1.64
30 min. vac. roast	-1.79	+1.63
120 min. vac. roast	-1.72	+1.45
plasma 30 min.	-1.77	+1.43
plasma 30 min. + 30 min. vac.	-1.90	+1.58
plasma 60 min.	-2.35	+1.19
plasma 60 min. + 30 min. vac.	-1.73	+0.76
G. dutertrei		
not roasted	-0.68	+1.52
30 min. vac. roast	-0.82	+1.43
120 min. vac. roast	-0.91	+1.46
plasma 30 min. + 30 min. vac.	-0.96	+1.32
plasma 60 min.	-1.38	+0.91
plasma 60 min. + 30 min. vac.	-1.95	-0.16
G. tumida		
not roasted	inpure gas	
30 min. vac. roast	+0.04	+1.76
120 min. vac. roast	0.00	+1.61
plasma 30 min.	-0.29	+1.44
plasma 30 min. + 30 min. vac.	-0.57	+1.13
plasma 60 min.	-1.00	+0.54
plasma 60 min. + 30 min. vac.	-0.99	+0.20

that were double roasted (30 minutes plasma + 30 minutes vacuum) seem to be very similar to the vacuum roasted samples. Roasting by either method or combination of methods produces lighter isotopic composition than found in unroasted samples. In what follows I assume that the roasting used is sufficient to remove the organic contaminants that interfere with the isotopic measurements. In absence of more conclusive data, this problem is not considered resolved. However, the internal consistancy shown in the data discussed below and its agreement with results of other studies allows the above assumption.

Mass-spectrometry: Isotopic analysis was carried out in a V.G. Micromass 602C mass-spectrometer in N. Shackleton's laboratory, Cambridge, England. Samples were reacted by 100% orthophosphoric acid at 50°C. The analytical precision reported for the measurement is $\pm$ 0.05 °/₀₀ (1 σ) (Shackleton and Opdyke, 1973). But it is more reasonable to regard 1 σ of 0.1 °/₀₀ for different pickings from the same sample. Results are expressed in δ values relative to the PDB standard, e.g. $\delta^{18}O = 1000 \left(\frac{R \text{ sample} - R \text{ std}}{R \text{ std}}\right)$ where R is the 46 to 44 mass ratio.

3. Results and Discussion

A. $\delta^{18}O$ and $\delta^{13}C$ relationship

The isotopic data for all samples analyzed are shown in fig. 5 and table 3. $\delta^{18}O$ values range from -1.63 °/₀₀ to

+ 1.23 °/₀₀ and $\delta^{13}C$ ranges from -2.47 to + 2.35 °/₀₀. The $\delta^{18}O$ range can be accommodated within the expected range for isotopic equilibrium of $CaCO_3$ in the upper 1000 m of water given the yearly range of temperatures. However, the $\delta^{13}C$ range is 5 times larger than the $\delta^{13}C$ range of ΣCO_2 in upper 1000 m (Kroopnick et al., 1972; Kroopnick, in press). No particular trend is shown by the data as a whole (fig. 5), but the fields occupied by different genera and even different species within a genus are well separated from each other. This is quite apparent in fig. 6 where only sediment trap data are displayed. In fig. 7 the average isotopic compositions and their standard deviations for each species are shown. Here again the separation between genera and species is quite clear. This suggests that biological controls are involved in determining the isotopic composition of genera as well as species within a genus. All species grouped as Globigerinoides in figures 5, 6 and 7 are spinose, contain symbiotic algae and are known to occupy the photic zone (Bé et al., 1977; Van Donk, 1977). The Globorotaliids, on the other hand, are smooth, do not contain symbiotic algae and live well below the photic zone (op. cit). In Chapter I, I have shown that symbiotic algae influence the isotopic composition of hermatypic corals and associated benthic foraminifera. It is possible that similar effects can also be seen in planktonic foraminifera, and cause this difference between the two groups. When individual species

TABLE 3

$\delta^{13}C$(°/oo) RELATIVE TO PDB

	PLANKTON - JANUARY			PLANKTON-NOVEMBER		PLANKTON - JULY			PARFLUX S SEDIMENT TRAPS				OC28 SEDIMENT		
Species	2-5 m	200 m	200 m	2-5 m	2-5 m	2-5 m	200 m	100 m	S2 400	S2 1000	S2 4000	S1 5300	Bx 395 4500 m	Bx 396 4950 m	Av. ± (σ)
G. ruber	+0.49	+0.28		+1.33	+1.16				+1.27	+1.15	+1.48	+0.72	+0.86	+0.95	+1.02(.37)
G. truncatulinoides	+0.33	+0.20	+0.25			+0.09	+0.56		+0.77	+0.93	+0.81		+0.81	+1.28	+0.60(.38)
G. inflata									+0.36	+0.36	+0.47		+0.40	+0.65	+0.45(.12)
H. pelagica			-2.47						+2.45	-1.83	-2.35	-1.71		-	-2.16(.36)
G. hirsuta							+0.46		+0.53	+0.82			+0.73	+1.04	+0.72(.23)
G. conglobatus	+1.80	+1.53	+1.49	+2.35 +1.74	+2.30 +1.75				+1.71	+1.63		+1.87	+2.11	+2.09	+1.86(.29)
G. elongatus	+0.43		+0.55						+0.80	+0.84	+0.93	+0.74		-	+0.72(.19)
G. sacculifer				+.151	+1.26				+1.73	+1.40			+1.11	+1.65	+1.44(.23)
G. siphonifera	-0.35	-0.50	-0.35					-0.06	-0.05	-0.24	-0.32	-0.07	+0.15	+0.93	+0.01(.50)
G. calida									-0.12	-0.81	-0.31			-	-0.41(.36)
O. universa			-.151	+2.10	+2.25			+1.95	+1.94	+2.10	+0.73	+1.57	+1.51	+1.69	+1.73(.42)
G. crassaformis									+0.64						+0.64
P. obliquiloculata	+0.56	+0.43	+0.69										+0.57	+0.69	+0.59(.11)
G. dutertrei			+1.17										+0.75		+0.96(.30)

TABLE 3

$\delta^{18}O$(°/₀₀) RELATIVE TO PDB

Species	PLANKTON - JANUARY			PLANKTON-NOVEMBER		PLANKTON - JULY			PARFLUX S SEDIMENT TRAPS				OC28 SEDIMENT		Av. ± (σ)
	2-5 m	200 m	200 m	2-5 m	2-5 m	2-5 m	200 m	100 m	S2 400	S2 1000	S2 4000	S1 5300	4500 m	4950 m	
G. ruber	-0.66	-0.60		-1.17	-1.31				-1.32	-1.41	-1.63	-1.41	-0.71	-0.43	-1.03(.42)
G. truncatulinoides	-0.21	-0.17	-0.28			-0.28	+0.31		+0.56	+0.85	+0.53		+0.55	+0.98	+0.28(.48)
G. inflata									+0.19	+0.68	+0.41		+0.74	+1.01	+0.61(.32)
H. pelagica			-0.78						-1.32	-1.41	-0.91	-1.07			-1.10(.27
G. hirsuta							+0.33		+0.50	+0.97			+0.89	+1.23	+0.78(.36)
G. conglobatus	-0.32	-0.35	-0.37	-0.97 -0.98	-1.05 -1.19				-0.14	+0.02		-0.27	-0.20	+0.13	-0.43(.47)
G. elongatus	-0.71		-0.67						-1.20	-1.21	-1.29	-0.96			-1.01(.27)
G. sacculifer				-1.14	-1.14				-1.43	-1.02			-0.78	-0.55	-1.01(.31)
G. sifonifera	-0.68	-0.47	-0.76					-0.41	-0.01	-0.08	-0.03	-0.63	-0.06	+0.15	-0.27(.34)
G. calida									-0.35	-0.06	+0.50				+0.03(.43)
O. universa			-0.46	-1.13	-1.14			-0.47	-0.36	-0.49	-0.81	-0.79	-0.20	-0.15	-0.57(.36)
G. crassaformis									+0.56						+0.56
P. obliquiloculatus	-0.25	-0.23	-0.24										+0.06		+0.16(.13)
G. dutertrei			-0.28										+0.17		+0.06(.32)

Figure 5. $\delta^{18}O$ vs. $\delta^{13}C$ for all planktonic foraminifera analyzed.

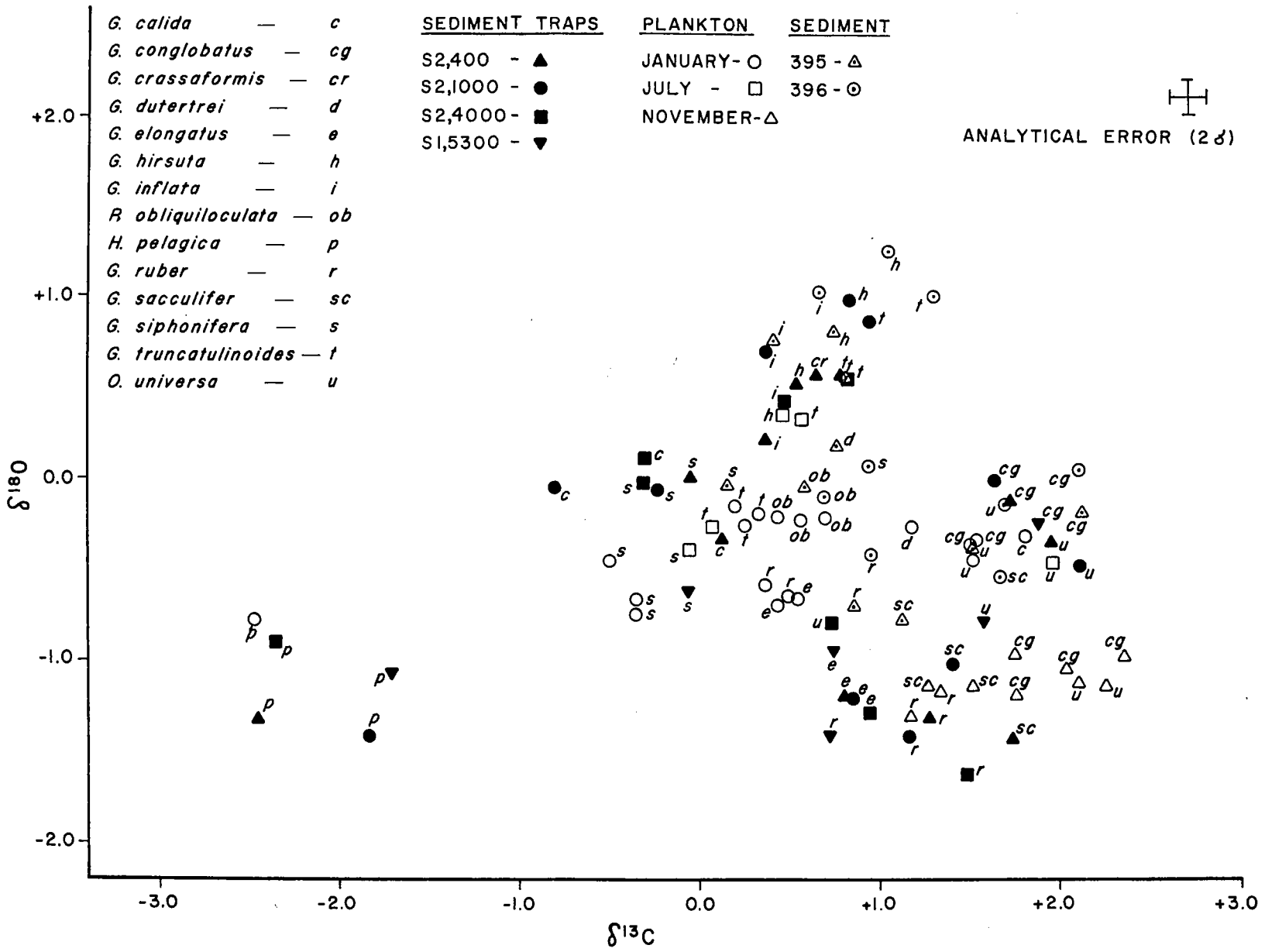

G. calida — c
G. conglobatus — cg
G. crassaformis — cr
G. dutertrei — d
G. elongatus — e
G. hirsuta — h
G. inflata — i
R obliquiloculata — ob
H. pelagica — p
G. ruber — r
G. sacculifer — sc
G. siphonifera — s
G. truncatulinoides — t
O. universa — u
SEDIMENT TRAPS
S2,400 - ▲
S2,1000 - ●
S2,4000 - ■
S1,5300 - ▼
PLANKTON
JANUARY - ○
JULY - □
NOVEMBER - △
SEDIMENT
395 - △
396 - ⊙
ANALYTICAL ERROR (2 δ)
+2.0
+1.0
0.0
-1.0
-2.0
δ18O
-3.0
-2.0
-1.0
0.0
+1.0
+2.0
+3.0
δ13C

Figure 5.

Figure 6. $\delta^{18}O$ vs. $\delta^{13}C$ for sediment traps. Note the separate fields occupied by different genera. *O. universa* was grouped together with other *Globigerinoides* species in accordance with the taxonomic scheme suggested by Bé (1977).

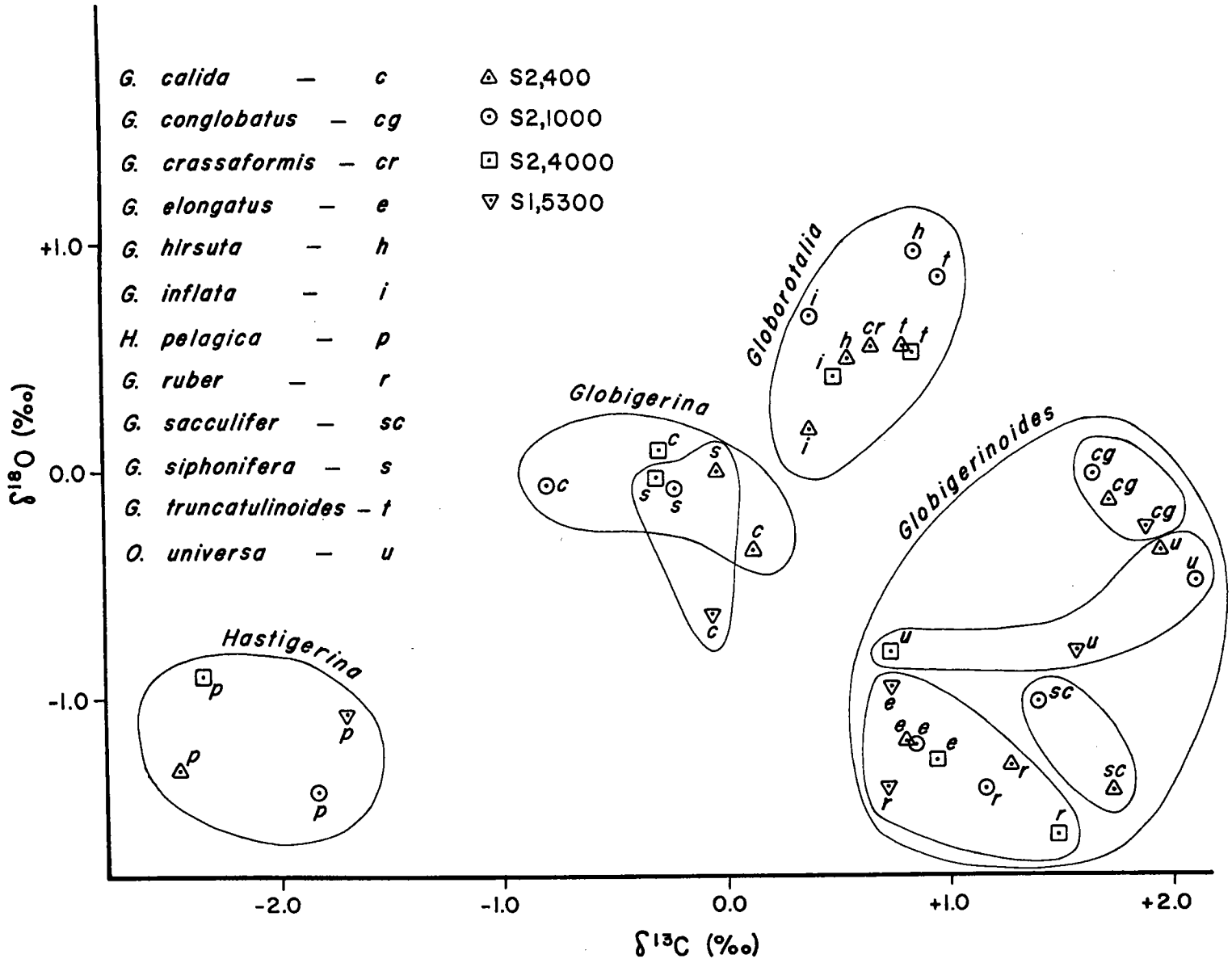

G. calida — c
G. conglobatus — cg
G. crassaformis — cr
G. elongatus — e
G. hirsuta — h
G. inflata — i
H. pelagica — p
G. ruber — r
G. sacculifer — sc
G. siphonifera — s
G. truncatulinoides — t
O. universa — u
S2,400
S2,1000
S2,4000
S1,5300
Globorotalia
Globigerina
Globigerinoides
Hastigerina
+1.0
0.0
-1.0
δ18O (‰)
-2.0
-1.0
0.0
+1.0
+2.0
δ13C (‰)

Figure 6.

Figure 7. $\delta^{18}O$ vs. $\delta^{13}C$ averages for planktonic foraminifera from plankton tows, sediment traps and sediments. The error bars represent 2 σ about the mean. *O. universa* is grouped with other Globigerinoides species and *P. obliquiloculata* is grouped with Globorotalia species. In both cases this fits the taxonomic scheme suggested by Bé (1977).

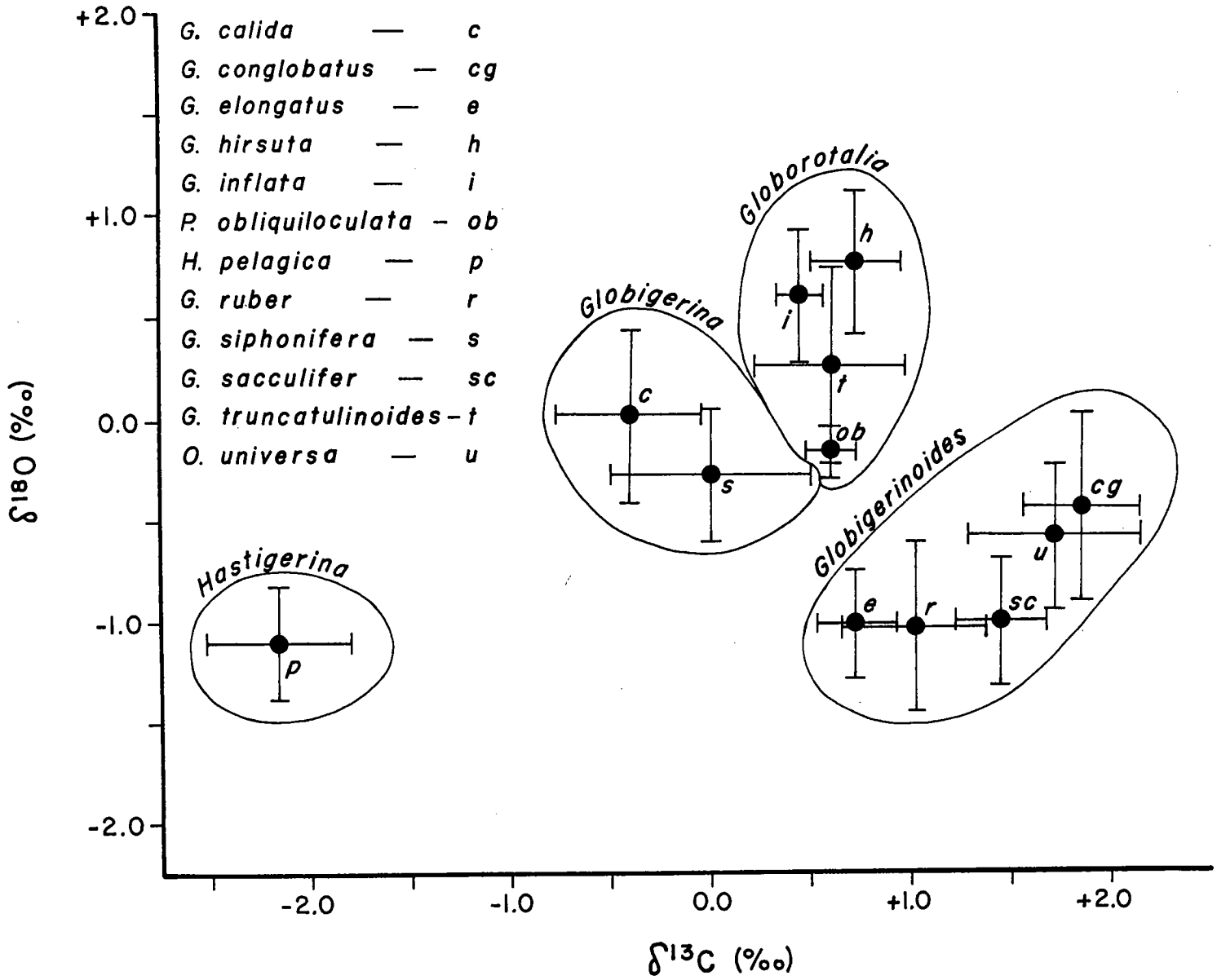

G. calida — c
G. conglobatus — cg
G. elongatus — e
G. hirsuta — h
G. inflata — i
P. obliquiloculata — ob
H. pelagica — p
G. ruber — r
G. siphonifera — s
G. sacculifer — sc
G. truncatulinoides — t
O. universa — u
Globorotalia
Globigerina
Globigerinoides
Hastigerina
+2.0
+1.0
0.0
-1.0
-2.0
δ18O (‰)
δ13C (‰)

Figure 7.

are examined (fig. 8), some show significant trends. *G. truncatulinoides, G. hirsuta* and *G. inflata* (all belong to the genus Globorotalia), clearly show positive correlation between $\delta^{18}O$ and $\delta^{13}C$. In addition, both oxygen and carbon isotopes become heavier as these species proceed from plankton samples to sediment trap and sediment samples (fig. 8, table 3). The increase in $\delta^{18}O$ between the plankton samples and the sediment trap samples can be explained by additional skeleton deposition below the photic zone, at lower temperatures. If carbon isotopes in the skeleton were fractionated by a constant specific factor from the dissolved total CO_2 of seawater (as suggested by some authors, e.g. Williams et al., 1977; Shackleton and Vincent, 1978), one would expect negative correlation between $\delta^{18}O$ and $\delta^{13}C$. This is because $\delta^{13}C$ of ΣCO_2 decreases with depth due to oxidation of organic matter that was fixed in the photic zone (Deuser and Hunt, 1968; Kroopnick et al., 1972; Kroopnick, in press). Instead, $\delta^{13}C$ in the Globorotallids increases with depth and thus shows positive correlation with $\delta^{18}O$ (table 3). It is possible that carbon isotopes in the skeleton are controlled by metabolic activity of the organism, as suggested by Kahn (1977), Berger et al. (1978), and my data in Chapter I. When metabolic activity is high, more isotopically light metabolic CO_2 is incorporated in the skeleton. Decrease in temperature can significantly lower the metabolic rate of the

Globorotaliids as they sink in the water column and thus account for the heavier carbon isotope compositions.

The other species shown in fig. 8 do not show a definite trend. They are *Globigerinoides* species, *Orbulina universa* and *Globigerina siphonifera*. All these species are known to contain symbiotic algae (Bé, 1977; Bé et al., 1977). If the $\delta^{18}O$ axis in fig. 8 is taken to represent depth scale, it seems that some of these species show a mid-water minimum in $\delta^{13}C$ values. This pattern is quite clear for *G. sacculifer* and *G. ruber* and is possible for *O. universa*, *G. conglobatus* and *G. siphonifera*. The dashed line in fig. 8 is a schematic representation of this idea. A mid-water $\delta^{13}C$ minimum was found by Weber et al. (1976) for the hermatypic coral *M. annularis* in the Caribbean. This was interpreted in Chapter I to represent mid-water photosynthetic maximum by the symbiotic algae. This interpretation is displayed in fig. 9 and is based on the experimental data and discussion shown in Chapter I. Similar interpretation may be applied to the data in fig. 8. It implies that these species that show mid-water $\delta^{13}C$ minima have symbionts that show mid-water photosynthetic maxima, much like the hermatypic corals and shallow benthonic foraminifera shown in fig. 1. Primary productivity experiment was carried out at station S during July 21, 1977, using the ^{14}C tracer method (fig. 10). It shows that indeed phytoplankton in the Sargasso Sea has photosynthetic maximum at roughly 50 m, and that photoinhibition occurs at shallower depths.

Figure 8. $\delta^{18}O$ vs. $\delta^{13}C$ for individual species from plankton tows, sediment traps and sediments. Note the linear correlation between oxygen and carbon isotopes shown by the Globorotallids. The dashed line shown for species that contain symbiotic algae represents a schematic photosynthesis vs. depth relationship.

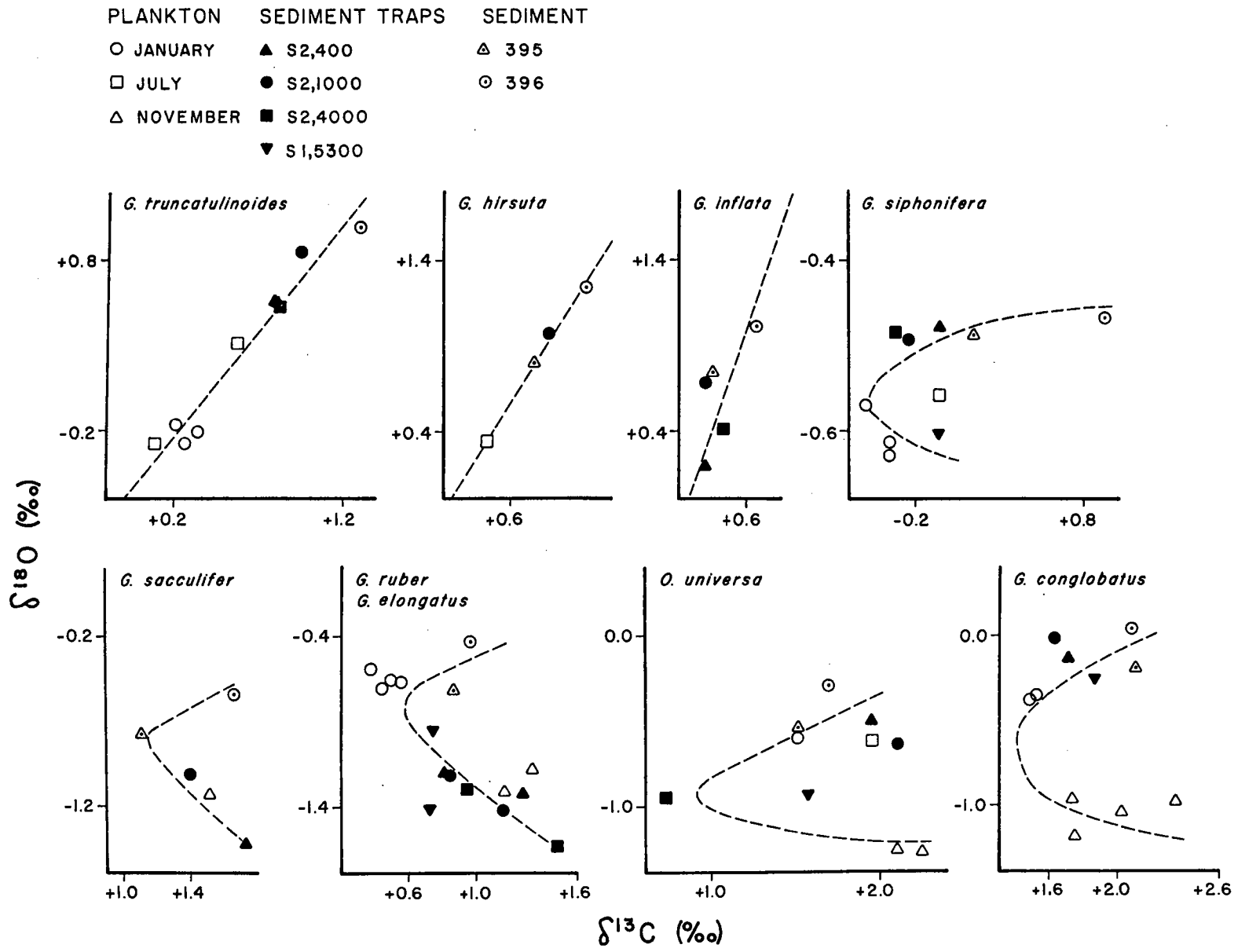

PLANKTON
JANUARY
JULY
NOVEMBER
SEDIMENT TRAPS
S2,400
S2,1000
S2,4000
S1,5300
SEDIMENT
395
396
G. truncatulinoides
G. hirsuta
G. inflata
G. siphonifera
G. sacculifer
G. ruber
G. elongatus
O. universa
G. conglobatus
+0.8
-0.2
+0.2
+1.2
+1.4
+0.4
+0.6
+1.4
+0.4
+0.6
-0.4
-0.6
-0.2
+0.8
-0.2
-1.2
+1.0
+1.4
-0.4
-1.4
+0.6
+1.0
+1.6
0.0
-1.0
+1.0
+2.0
0.0
-1.0
+1.6
+2.0
+2.6
δ18O (‰)
δ13C (‰)

Figure 8.

Figure 9. $\delta^{13}C$ vs. depth for the coral *Montastrea annularis* in the Caribbean (after Weber et al., 1976). The schematic photosynthetic curve on the right is suggested to explain this profile.

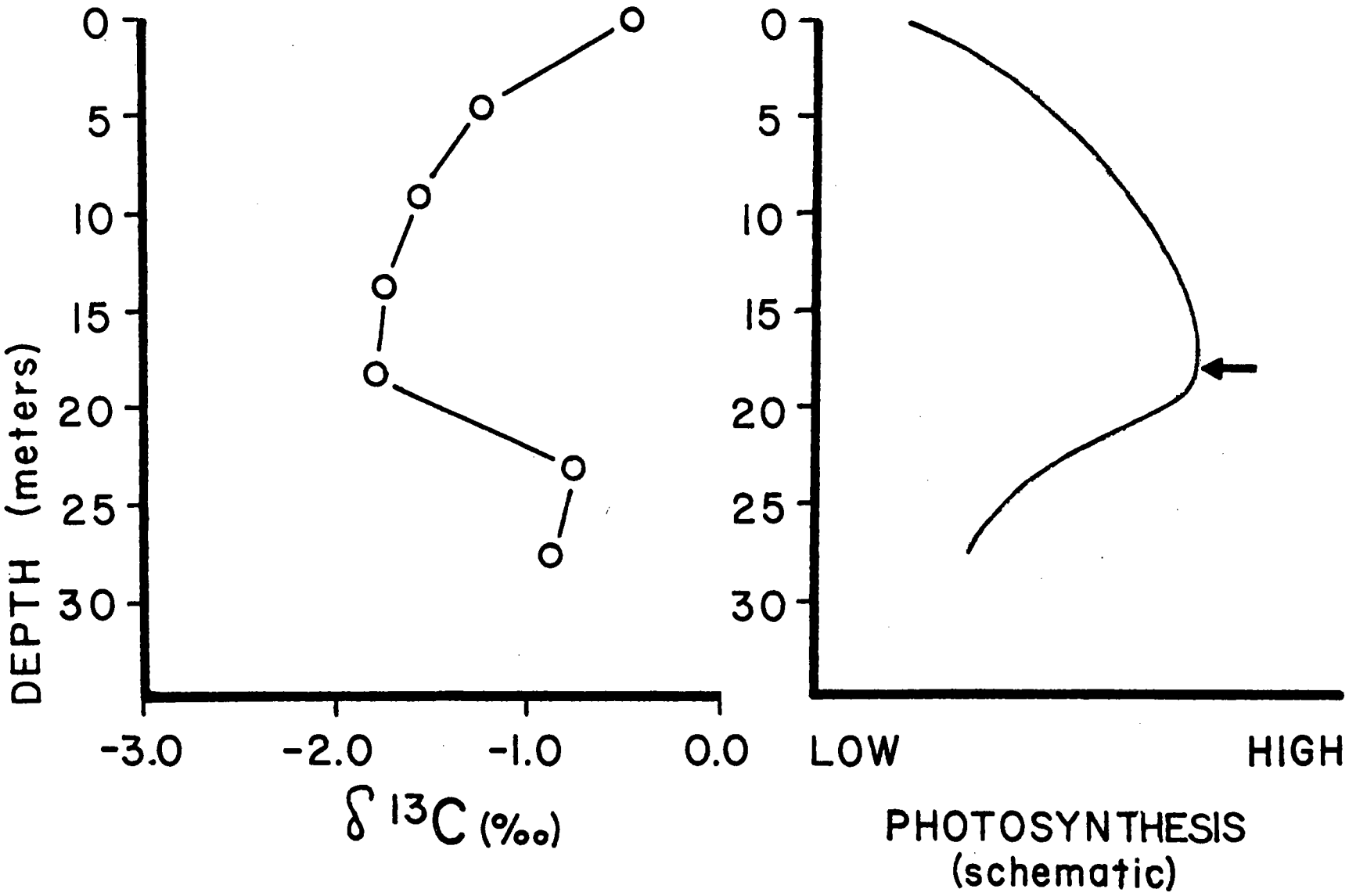

0
5
10
15
20
25
30
DEPTH (meters)
-3.0
-2.0
-1.0
0.0
δ 13C (‰)
LOW
HIGH
PHOTOSYNTHESIS
(schematic)

Figure 9.

Figure 10. In-situ primary productivity profile for Station S during July 1977. Note the mid-water photosynthetic maximum.

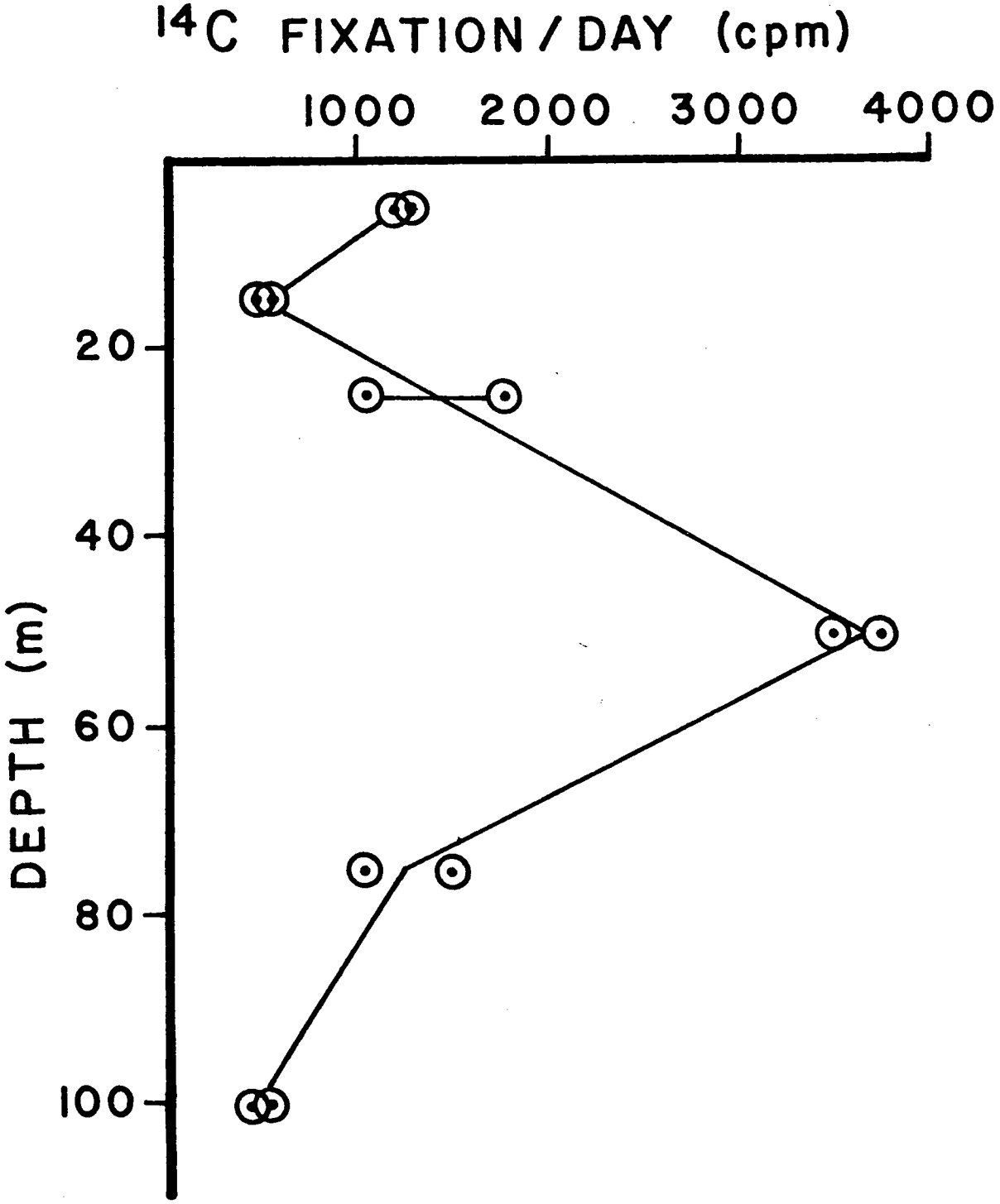

^{14}C FIXATION / DAY (cpm)
1000
2000
3000
4000
20
40
60
80
100
DEPTH (m)

Figure 10.

B. Seasonality and depth stratification

The isotopic method to derive paleotemperatures was originally suggested by Urey (1947). The method became operational after Epstein et al. (1951, 1953) derived an empirical paleotemperature scale (or equation) based on calcareous molluscs that were grown at known temperatures. Emiliani (1954) applied this scale to planktonic foraminifera in surface sediments and showed that the "isotopic temperatures" he calculated were reasonable for the overlying water. He further showed that depth stratification of various species inferred from the isotopic temperatures was in agreement with information gathered by plankton tows. However, direct attempts to test whether Epstein's equation really holds for planktonic foraminifera were not so successful. In these studies foraminifera from plankton tows were analyzed and the isotopic temperatures derived were compared to the actual temperatures measured in the field (Van Donk, 1970, 1977; Shackleton et al., 1973; Kahn, 1977). In most samples and species the isotopic temperatures were higher than expected, indicating light non-equilibrium $\delta^{18}O$ values (op. cit.). In contrast, studies following Emiliani's (1954) using surface sediment were almost all in agreement with his conclusions that foraminifera deposit their skeleton in equilibrium with sea water (Lidz et al., 1968; Shackleton and Vincent, 1978; Berger et al., 1978).

In what follows I attempted to test the isotopic paleo-temperature method using plankton samples as well as sediment traps and surface sediments in one spot in the Sargasso Sea (Station S). Temperature and salinity data for each month were averaged for a 4 degree square surrounding Station S. Data comes from NODC file and our own measurements (see appendix for the data and the averages). Isotopic temperatures were calculated using the equation of Epstein and Mayeda (1953):

$$t(°c) = 16.5-4.3(\delta^{18}O_c-\delta^{18}O_w)+0.14(\delta^{18}O_c-\delta^{18}O_w)^2$$

where $\delta^{18}O_c$ and $\delta^{18}O_w$ are the deviation in °/oo from the PDB standard for calcium carbonate and sea-water respectively. $\delta^{18}O_w$ was calculated using linear regression between salinity and $\delta^{18}O_w$ based on data for the North Atlantic from Epstein and Mayeda (1953) (see fig. 11)

$$\delta^{18}O_w = 0.48 \cdot S - 17.02 \qquad S = \text{salinity in °/oo}$$

for which $r^2 = 0.95$.

The isotopic temperatures were compared with the actual temperature profile observed. Plankton samples were compared with the average of the month in which the samples were collected. Sediment trap samples were compared with the average temperature for the duration of the deployment, and sediment samples were compared with the yearly average. The results are depth-temperature profiles on which the foraminifera are positioned according to their isotopic

temperature (figs. 13-20). (This includes corrections to account for the salinity change with depth.)

Before considering the results, the question of seasonality must be addressed. Bé (1959, 1960) and Tolderlund and Bé (1971) have shown that abundance of foraminifera in the western North Atlantic varies seasonally. It was suggested by Bé (1960) and Berger (1971) that seasonality may be reflected in the isotopic compositions of planktonic foraminifera. From fig. 12 it is clear that surface plankton samples collected in November are well separated from surface plankton samples from January, the latter samples showing $\delta^{18}O$ heavier by roughly 0.7 °/₀₀. This corresponds to 3°C which is close to the difference in the averages between the two months. The isotopic temperatures for the common species in these two samples is summarized in table 4, and compared to the monthly averages. It shows that these species recorded the seasonal temperature change.

TABLE 4

COMPARISON OF ISOTOPIC TEMPERATURES FOR PLANKTON SAMPLES WITH ACTUAL TEMPERATURES

	G. ruber	*O. universa*	*G. conglobatus*	surface monthly average
November	24.4	23.5	23.9	23.2
January	22.0	21.0	20.5	19.9
Difference	2.4	2.5	3.4	3.3

Figure 11. $\delta^{18}O$ vs. salinity for the North Atlantic after Epstein and Mayeda (1953).

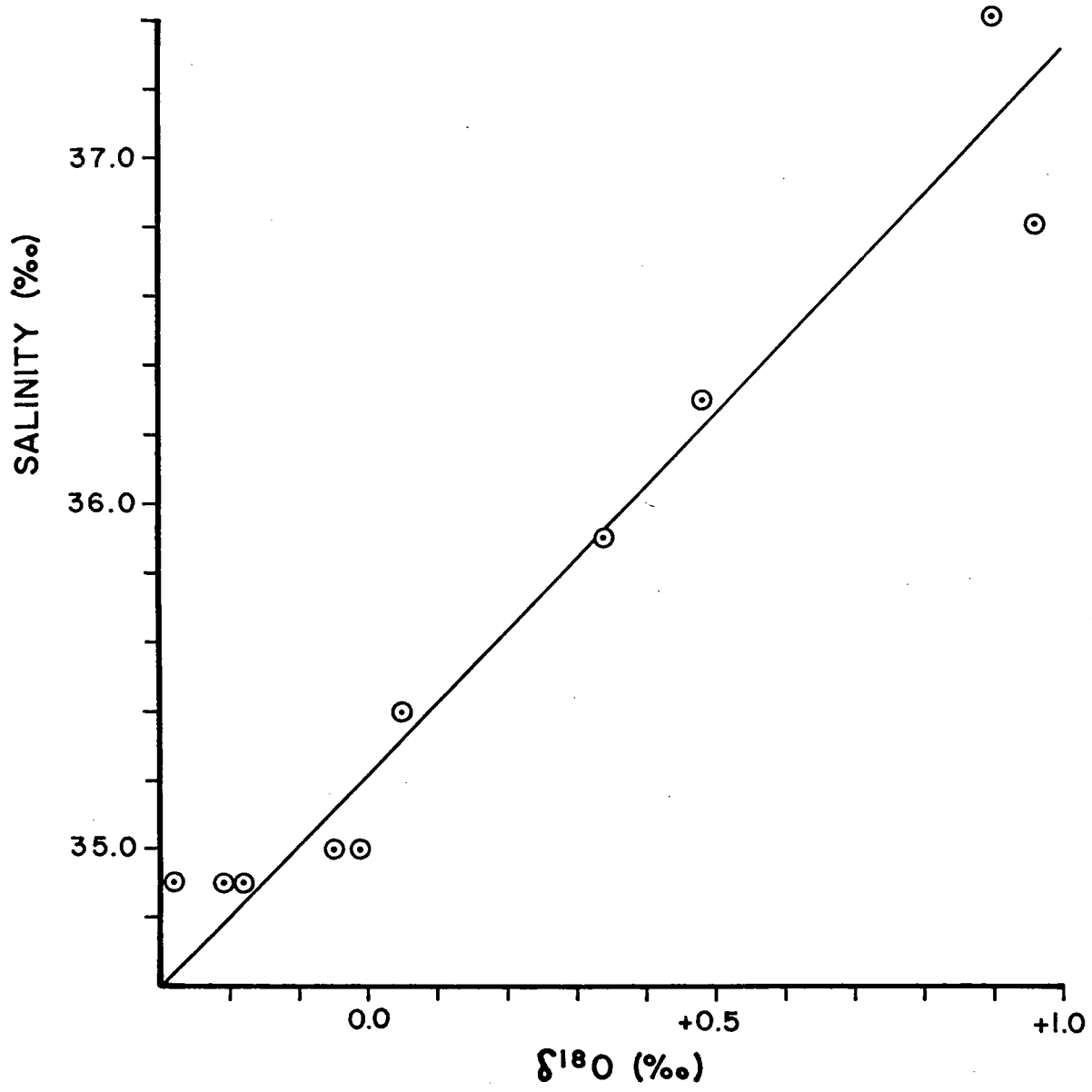


Figure 11.

Figure 12. $\delta^{18}O$ vs. $\delta^{13}C$ for plankton tow samples. Note the higher $\delta^{18}O$ values shown by January tows compared to November tows. July samples are all from 100 m or 200 m, and this may account for the high $\delta^{18}O$ values they have.

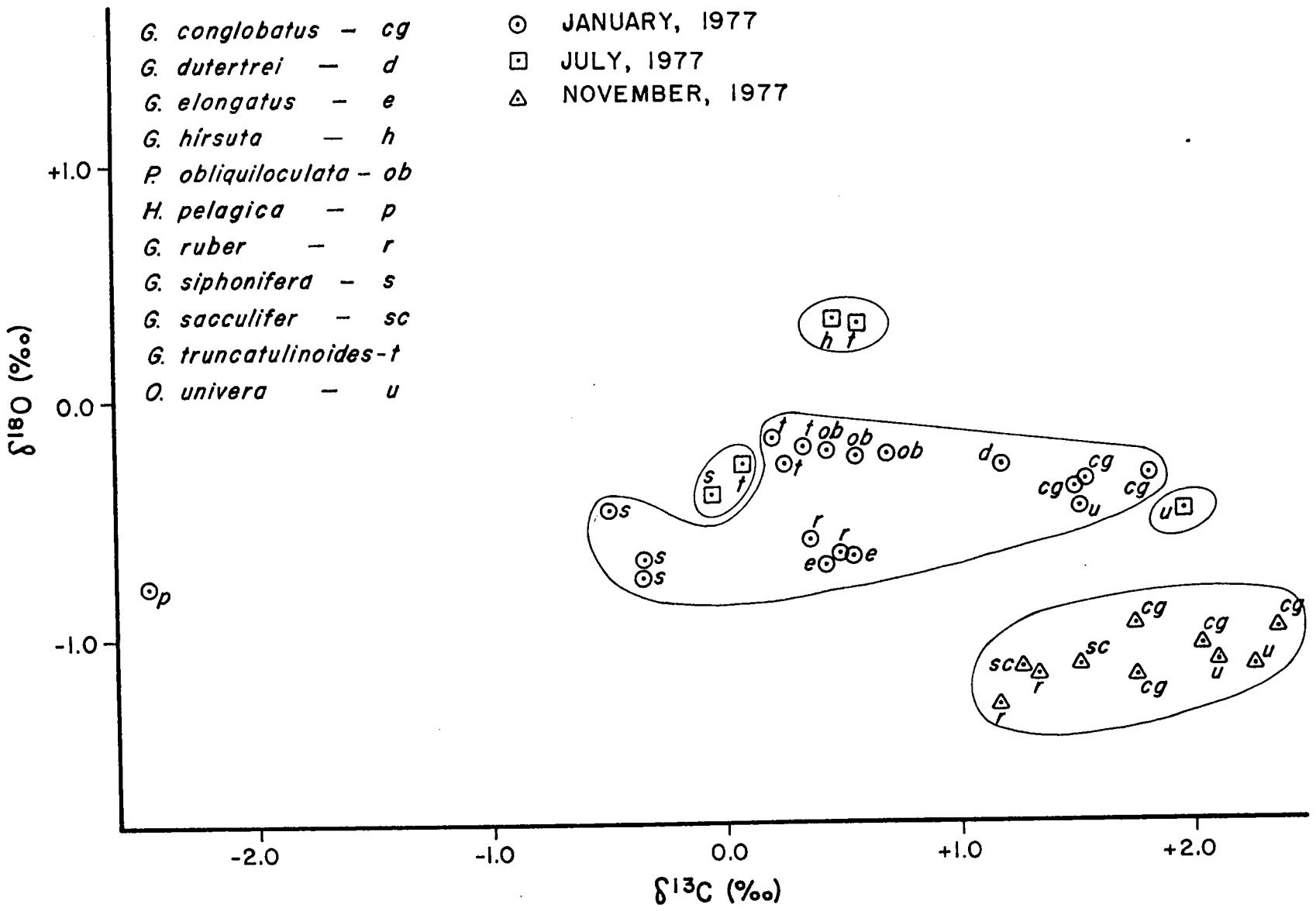

G. conglobatus – cg
G. dutertrei – d
G. elongatus – e
G. hirsuta – h
P. obliquiloculata – ob
H. pelagica – p
G. ruber – r
G. siphonifera – s
G. sacculifer – sc
G. truncatulinoides – t
O. univera – u
JANUARY, 1977
JULY, 1977
NOVEMBER, 1977
+1.0
0.0
-1.0
δ18O (‰)
-2.0
-1.0
0.0
+1.0
+2.0
δ13C (‰)

Figure 12.

TABLE 5
DEPTH ASSIGNMENT FOR PLANKTONIC FORAMINIFERA (in meters)
BASED ON THEIR ISOTOPIC COMPOSITION

Species	PLANKTON			SEDIMENT TRAPS				SEDIMENT		
	JANUARY PLANKTON	JULY PLANKTON	NOVEMBER PLANKTON	S2 400	S2 1000	S2 4000	S2 5300	028 395	OC28 396	Av.
G. ruber	0*		0*	30	25	0*	0*	25	50	15
G. elongatus	0*			35	35	30	0*			25
G. sacculifer			0*	25	40			12.5	37.5	30
G. conglobatus			0*	100	130		110	85	140	105
H. pelagica	0*			30	25	45	0*			25
G. siphonifera	0*	55		125	110	120	60	150	200	115
G. calida				75	115	480				225
O. universa	0*	50	0*	75	70	50	0	85	85	55
G. truncatulinoides	105†	70† 270		500†	605	485		475	620	515
G. inflata				245	560	450		550	630	450
G. crassaformis				500†						500
G. hirsuta		290†		475†	735			580	665	600
P. obliquiloculata	100†							150	100	150
G. dutertrei	0*							225		225

* light non-equilibrium

† heavy non-equilibrium

The plankton samples from July (fig. 12), when surface temperatures are high, shows unexpectedly higher $\delta^{18}O$ values than samples collected in November and January. However, all July samples except one were collected at 100 m and 200 m. At these depths the temperature during July is the same or lower than in November and January. Therefore, seasonal changes in temperature indeed seem to be recorded in the isotopic composition of planktonic foraminifera, and this will be considered in the following discussion.

Depth assignments for different species based on the profiles shown in figs. 13-20 is summarized in table 5. The average depth for a species represents the depth at which most of shell production occurs. In general, the results of this study support the accepted scheme of depth stratification (compare Van Donk, 1977; Bé, 1977; Hecht, 1974; and table 6).

TABLE 6

DEPTH GROUPS OF PLANKTONIC FORAMINIFERA BASED ON THEIR ISOTOPIC COMPOSITION

0-50 m SHALLOW	50-255 m INTERMEDIATE	225-600 m DEEP
G. ruber	*G. conglobatus*	*G. truncatulinoides*
G. sacculifer	*G. siphonifera*	*G. inflata*
H. pelagica	*G. calida*	*G. hirsuta*
	O. universa	*G. crassaformis*
	P. obliquiloculata	
	G. dutertrei	

The differences are: 1) *H. pelagica* is grouped here with the shallow species whereas other workers grouped it with intermediate or deep species; 2) The intermediate and deep species show a range roughly twice that indicated from earlier $\delta^{18}O$ stratification studies (Van Donk, 1977). It seems like the foraminifera are stretched over a longer water column. Perhaps this can be related to the longer primary productivity profile in the central oceanic Sargasso Sea compared to coastal and equatorial regions (Ryther, 1963).

The plankton profiles for November and January suggest that most shallow and intermediate species are out of equilibrium (figs. 14, 15); i.e., they show a higher temperature than the average surface temperature. Earlier studies by Van Donk (1970, 1977), Shackleton et al. (1973), Grazzini (1976), and Kahn (1977) have also shown light non-equilibrium $\delta^{18}O$ values in planktonic foraminifera from plankton tows and sediments. These deviations from equilibrium are small and may possibly be caused by improper combustion of organic matter associated with the skeleton or by inaccurate calibration of standards between different laboratories. Another possibility is that the foraminifera collected in January deposited their skeleton earlier and thus recorded warmer temperatures. Bé et al. (1977) found that generation time for most surface dwelling species is less than 20 days. Similar estimates were obtained by Berger and Soutar (1968). If we assume that the foraminifera

Figure 13. Temperature depth profile for Station S in July. The foraminifera are located according to their isotopic temperature.

Figure 14. Temperature depth profile for Station S in November. The foraminifera are located according to their isotopic temperature.

Figure 15. Temperature depth profile for Station S in January. The foraminifera are located according to their isotopic temperature.

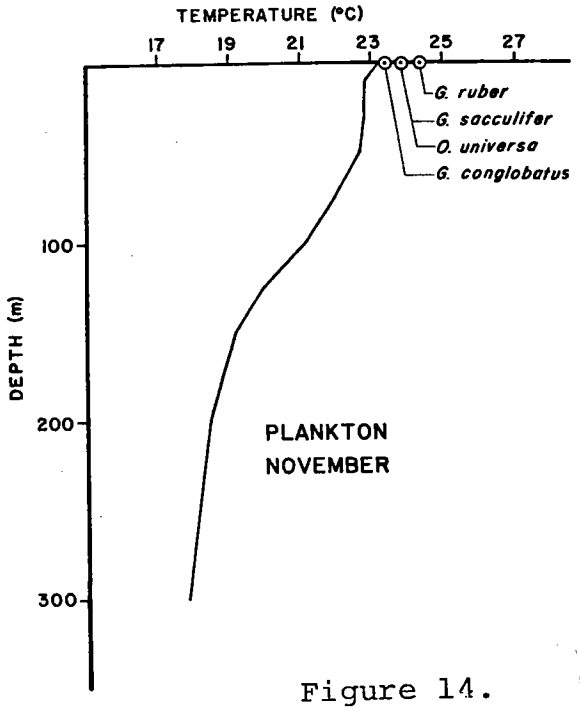


Figure 14.

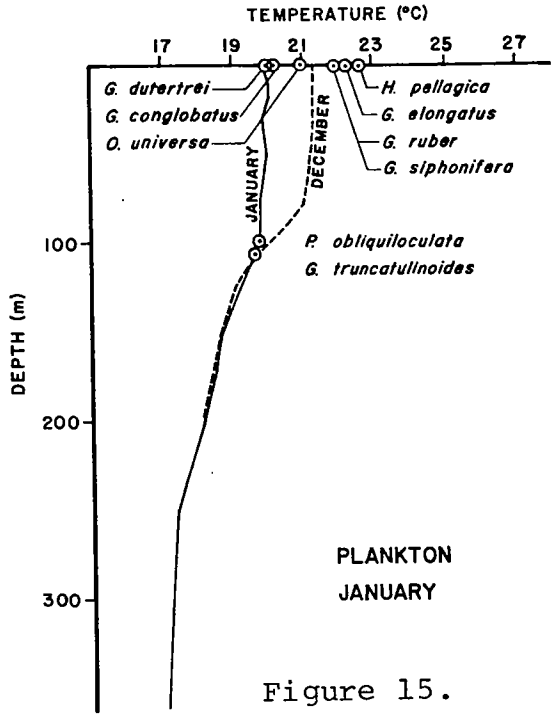


Figure 15.

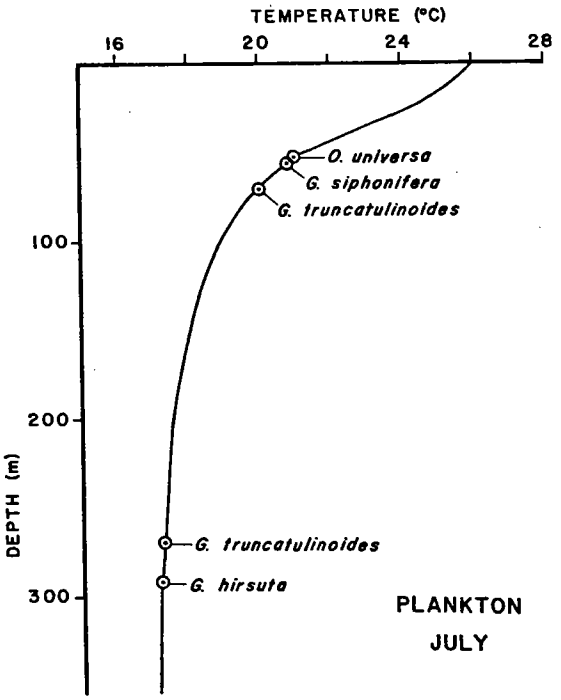


Figure 13.

Figure 16. Average temperature depth profile for the duration of S2 deployment (July 20-November 7, 1977). The foraminifera collected at a sediment trap at 400 m are located according to their isotopic temperature.

Figure 17. Average temperature depth profile for the duration of S2 deployment (July 20-November 7, 1977). The foraminifera collected at a sediment trap at 1000 m are located according to their isotopic temperature.

Figure 18. Average temperature depth profile for the duration of S2 deployment (July 20-November 7, 1977). The foraminifera collected at a sediment trap at 4000 m are located according to their isotopic temperature.

Figure 19. Average temperature depth profile for the duration of S1 deployment (October 20, 1976-January 5, 1977). The foraminifera are located according to their isotopic temperature (the dashed line is average temperature for October). Note that the disequilibrium values displayed by surface dwelling species disppears if one assumes that their major production was during October.

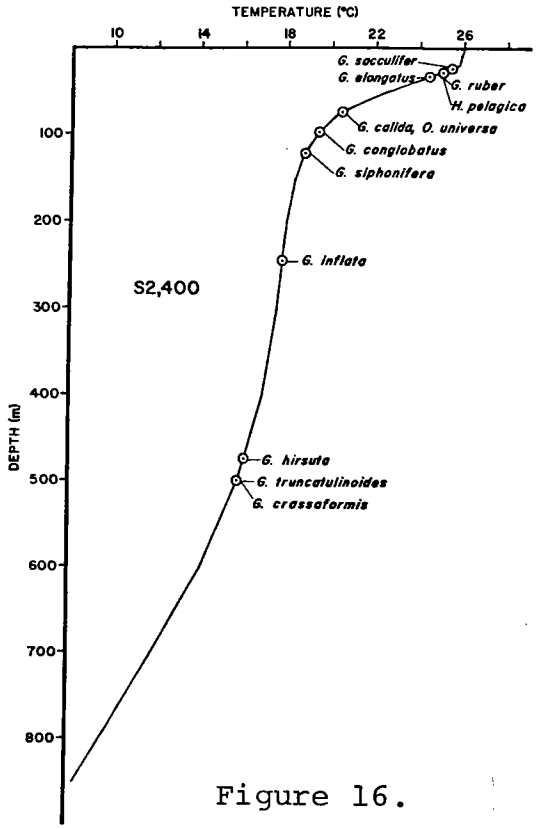


Figure 16.

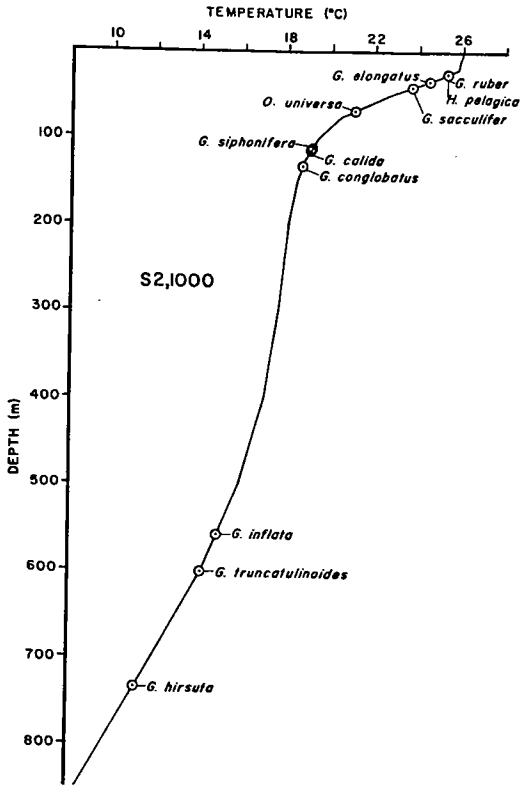


Figure 17.

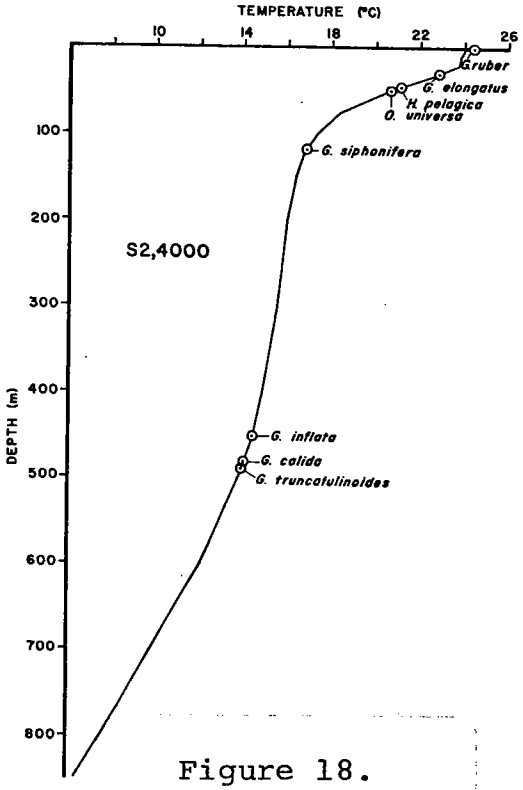


Figure 18.

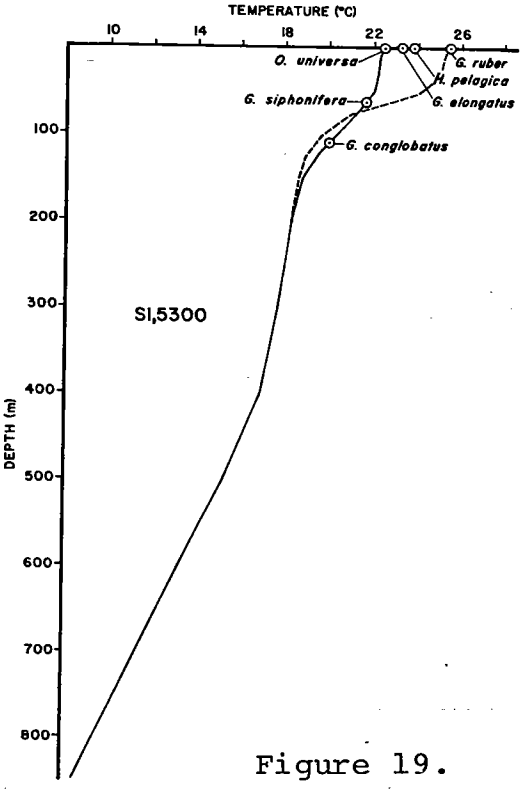


Figure 19.

collected in January deposited their skeleton in December (fig. 15), *G. dutertrei*, *G. conglobatus*, and *O. universa* will shift to equilibrium. However, *H. pelagica*, *G. elongatus*, *G. ruber* and *G. siphonifera* will still be out of equilibrium. The sediment trap at 5367 m (S1,5300) also seem to have collected few species that are out of equilibrium (fig. 19). However, it is possible that in this case seasonality can account for these values. If the species that are out of equilibrium deposited their skeletons in early October and were already settling down while the sediment trap was deployed (October 20), their isotopic composition can be accommodated within the equilibrium deposition (fig. 19). This will imply a settling time of roughly 10-20 days for 5367 m well within the experimental data of Berger and Piper (1972), for foraminifera larger than 250 μm.

In table 7 the deviations from equilibrium for oxygen isotopes are summarized. Most of the negative deviations are from plankton data. All but two are shown by shallow dwelling, spinose species that contain symbiotic algae. It is interesting that all the positive deviations are shown by the Globorotaliids that are deep dwelling, smooth and do not contain symbiotic algae. It is hard to explain the positive deviations by seasonality, because temperature fluctuations below 200 m do not exceed 1°C (fig. 21). Planktonic foraminifera exhibit diurnal vertical migration (e.g. Berger,

1971; Bé, 1977). Perhaps the positive isotopic deviation can be explained by upward migration of population that deposited most of its skeleton in deeper and colder water. The Globorotaliids from plankton tows may thus represent upward migration of ∿ 100 m. The isotopic composition of the Globorotaliids collected in sediment traps at 400 m (S2,400) show isotopic temperature of 500 m. However, these species were collected also in shallow plankton tows and seem to deposit a good part of their shell in shallower water (see also Bé and Ericson, 1963; Bé and Lott, 1964). To counter-balance this isotopically lighter part of their skeleton and show isotopic temperature of 500 m, they must deposit some skeleton deeper than 500 m. The average weight per individual for *G. truncatulinoides* collected in plankton tows is 23.3 µg and $\delta^{18}O$ is -0.22(°/₀₀). The same species at the 400 m sediment trap (S2,400) weighs 42 µg per individual and $\delta^{18}O$ is +0.56 °/₀₀. The average isotopic composition of the material added between the plankton sample and the trap sample ($\delta^{18}Ox$) can be obtained from material balance:

$$(-0.22)\ \frac{23.3}{42} + \delta^{18}Ox\ \frac{(42-23.3)}{42} = (+0.56)$$

$$\delta^{18}Ox = +1.53(°/₀₀)$$

Such a skeleton must have been deposited between 800 m and 1000 m (table 9). *G. truncatulinoides* from the trap at 1000 m (S2,1000) have $\delta^{18}O$ value of 0.85 (°/₀₀) which indicates that skeleton deposition indeed occurs between 400

TABLE 7

DEVIATION FROM OXYGEN ISOTOPIC COMPOSITION -
NEGATIVE DEVIATIONS

sample	species	$\delta^{18}O$(°/oo)	$\Delta\delta^{18}O$(°/oo)
Plankton-January	*G. ruber*	-0.63	-0.44
"	*G. elongatus*	-0.69	-0.50
"	*G. conglobatus*	-0.35	-0.16
"	*H. pelagica*	-0.78	-0.71*
"	*G. siphonifera*	-0.64	-0.43
"	*G. dutertrei*	-0.28	-0.21*
"	*G. universa*	-0.46	-0.39*
"	*G. truncatulinoides*	-0.22	-0.15*
"	*P. obliquiloculata*	-0.24	-0.05
Plankton-November	*G. ruber*	-1.24	-0.25
"	*G. sacculifer*	-1.14	-0.15
"	*G. conglobatus*	-1.05	-0.06
"	*O. universa*	-1.13	-0.14
Plankton-July	*G. siphonifera*	-0.41	-0.08†
S2,4000	*G. ruber*	-1.63	-0.10
S1,5300	*G. ruber*	-1.41	-0.64
"	*G. elongatus*	-0.96	-0.19
"	*H. pelagica*	-1.07	-0.30
"	*O. universa*	-0.79	-0.02

* collected in tows of 200 m only assumed to live in 125 m
† collected in tows of 100 m only assumed to live in 50 m

TABLE 7 (continued)

DEVIATION FROM OXYGEN ISOTOPIC COMPOSITION-
POSITIVE DEVIATIONS

sample	species	$\delta^{18}O$	$\Delta\delta^{18}O$
Plankton - July	*G. truncatulinoides*	-0.28	+1.28
"	*G. truncatulinoides*	-0.31	+0.24*
"	*G. hirsuta*	+0.33	+0.26*
S2,400	*G. hirsuta*	+0.50	+0.18
"	*G. truncatulinoides*	+0.56	+0.24
"	*G. crassaformis*	+0.56	+0.24

* collected in tows of 200 m only assumed to live in 125 m

and 1000 meters. These observations are in good agreement with Bé and Ericson (1963); and Bé and Lott (1964) who suggested that skeleton deposition for this species starts in the euphotic zone and continues to 1000 m or below, based on plankton tow data. This discussion implies that the population of *G. truncatulinoides* must have migrated at least 500 m upwards in order to be collected by the sediment trap at 400 m. This is not unreasonable because other zooplankton are known to show similarly large vertical migrations (Banse, 1964).

Depth stratifications of the species collected in the traps at 400 m, 1000 m, 4000 m and the box core samples are summarized in figs. 16-18 and table 5. The internal consistency between these samples and their agreement with other schemes of stratification seem to indicate that the skeletons are deposited in equilibrium with respect to oxygen isotopes as suggested by Emiliani (1954), Shackleton and Vincent (1978), and Berger et al. (1978). Thus, the discrepancy between non-equilibrium values shown by surface dwelling species in plankton tows and equilibrium values shown by these species in sediment samples disappears in the upper 400 m of the water column. The best explanation I can offer for this observation is that additional skeletal deposition takes place below the photic zone at least down to a few hundred meters below the surface. This skeleton is deposited in colder water and thus will record higher $\delta^{18}O$

values. In addition, the effects of symbiotic algae must disappear below the photic zone, and this also will make the skeleton isotopically heavier. In table 8 the increase in the average weight per individual between plankton material and sediment trap material is summarized for the surface dwelling species. The weight/individual was averaged for the S2,400, S2,1000 and S2,4000 traps because no depth trend was observed for these species below 400 m. The increase in average weight/individual between the plankton and trap samples ranges from 36% to 233% and averages roughly 80%. In table 9 the equilibrium isotopic composition for $CaCO_3$ is calculated for the deployment period of S2-traps. If one assumes that plankton samples deviate by 0.5 (°/₀₀) from equilibrium surface values, and if all additional skeleton deposition is at 200 m, the overall isotopic composition should be (see table 9):

$$(-2.0) \times \frac{100}{180} + (+ 0.16 \times \frac{80}{180}) = -0.76$$

This will indicate average skeleton deposition at roughly 50 m which is in good agreement with the average depth assigned to the surface dwelling species.

A plot of average weight/individual vs. $\delta^{18}O$ for all samples (except *G. conglobatus*) shows a general trend of increase in $\delta^{18}O$ as weight/individual increases. This further supports the idea that as foraminifera sink down the water column they continue to calcify and add isotopically heavier skeleton. Sediment samples show heavier weight and

Figure 20. Yearly average temperature depth profile for S site. Foraminifera from box cores are located according to their isotopic temperature. Note that almost all species, individuals from box core 395 (taken at 4500 m) show higher isotopic temperature than individuals from box 396 (taken at 4950 m). This difference is attributed to dissolution.

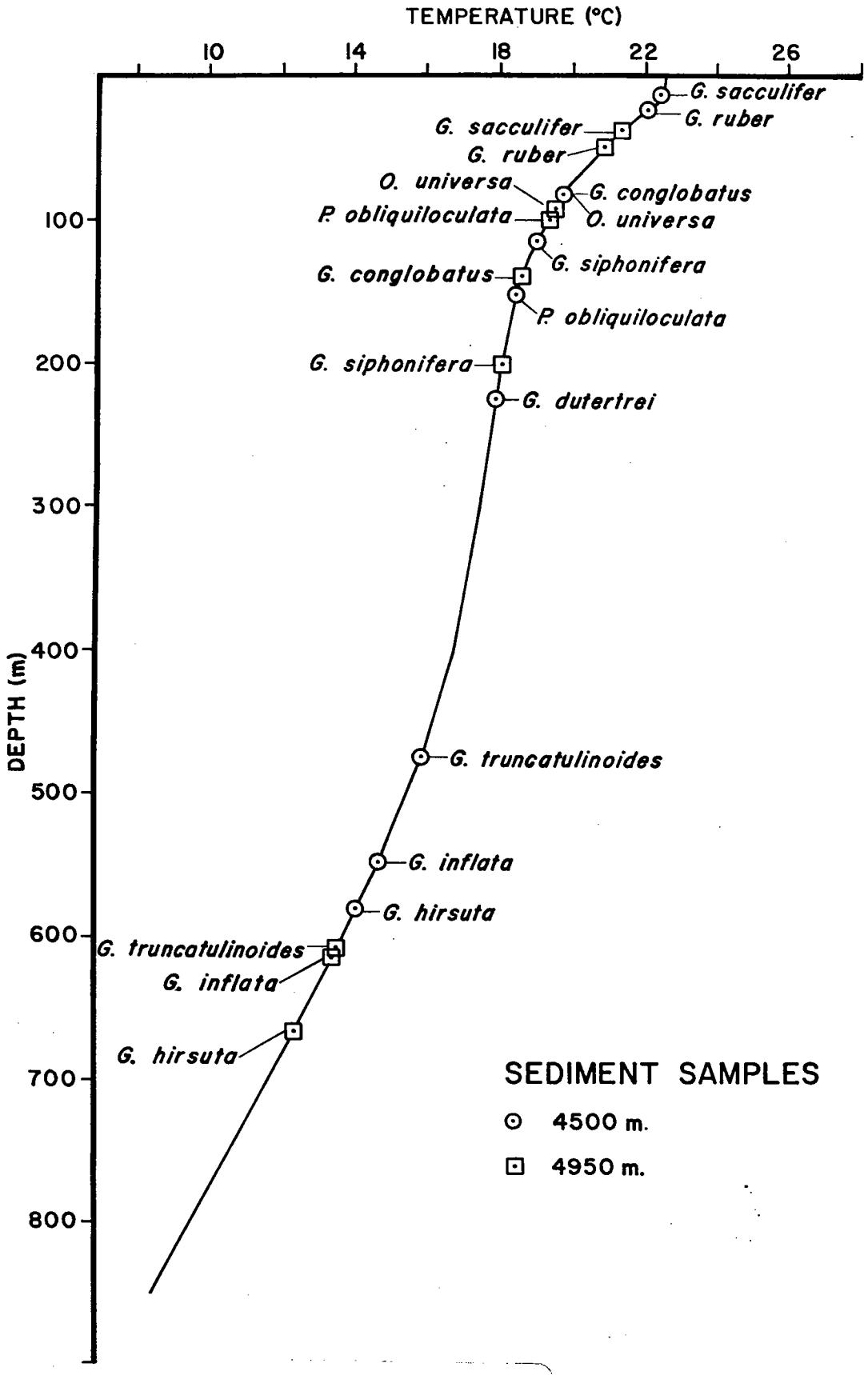

TEMPERATURE (°C)
10
14
18
22
26
DEPTH (m)
100
200
300
400
500
600
700
800
G. sacculifer
G. ruber
G. sacculifer
G. ruber
O. universa
G. conglobatus
P. obliquiloculata
O. universa
G. siphonifera
G. conglobatus
P. obliquiloculata
G. siphonifera
G. dutertrei
G. truncatulinoides
G. inflata
G. hirsuta
G. truncatulinoides
G. inflata
G. hirsuta
SEDIMENT SAMPLES
4500 m.
4950 m.

Figure 20.

Figure 21. Monthly average temperature depth profile for S site for the duration of S1 and S2 deployments. Note that seasonal variability is important only in the upper 100 m.

Figure 22. Yearly average temperature profile for the warmest and coolest months averages for S site.

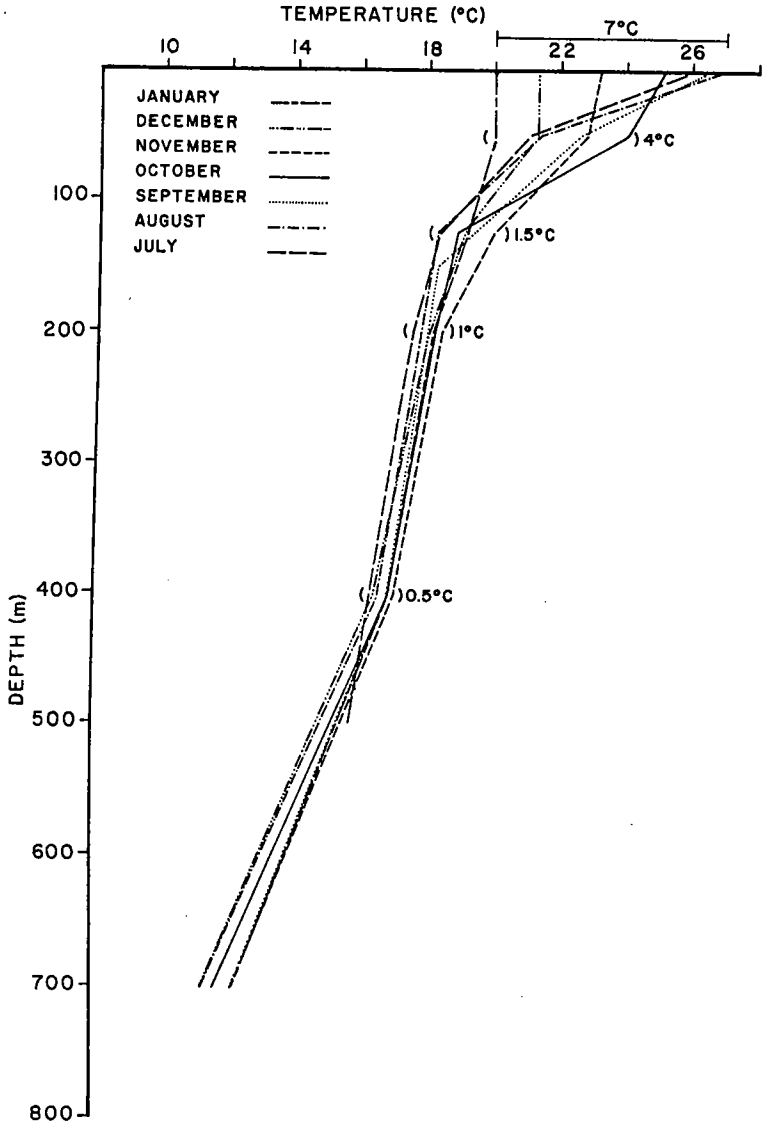


Figure 21.

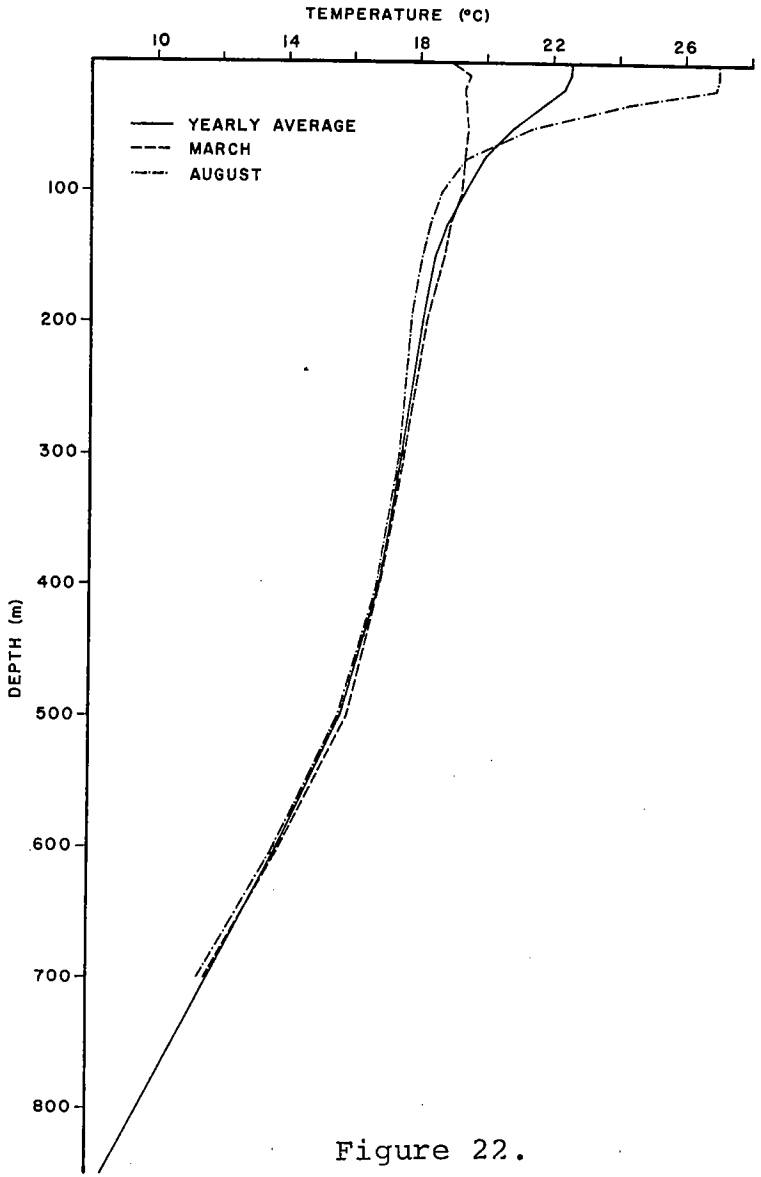


Figure 22.

TABLE 8

INCREASE IN WEIGHT PER INDIVIDUAL BETWEEN PLANKTON TOWS AND SEDIMENT TRAPS FOR SHALLOW DWELLING SPECIES

SPECIES	weight per individual (µg) PLANKTON TOWS	SEDIMENT TRAPS	DIFFERENCE	wt/ind. increase in %	av. depth (m)
G. ruber	10.4 (5)	14.7 (3)	4.3	41	15
G. sacculifer	13.7 (3)	21.0 (3)	7.3	53	30
G. conglobatus	50.0 (4)	83.0 (3)	33.0	66	105
O. universa	27.0 (5)	36.7 (3)	9.7	36	55
G. siphonifera	4.5 (4)	15.0 (3)	10.5	233	115
H. pelagica	7.7 (4)	10.7 (3)	3.0	39	25

TABLE 9

EQUILIBRIUM OXYGEN ISOTOPIC COMPOSITION FOR $CaCO_3$ BASED ON AVERAGE TEMPERATURE AND SALINITY FOR S2-DEPLOYMENT PERIOD (July 20 to November 7)

depth	(σ) av. t°C	av. δO^{18} eq.	av. δO^{18}_w
0	25.92 (1.16)	-1.53	+0.52
50	22.54 (1.10	-0.74	+0.56
125	18.63 (0.53)	+0.03	+0.51
200	17.99 (0.31)	+0.16	+0.48
400	16.93 (0.18)	+0.32	+0.40
700	11.90 (0.46)	+1.15	+0.04
1250	5.42 (0.12)	+2.77	-0.18

heavier isotopic composition while plankton samples are lighter in weight and in isotopes. Sediment trap samples have a wider spread, occupy the upper and the lower end of the general trend, and look like a mixture of sediment and plankton samples. The trend shown in fig. 23 can also be seen for individual species (fig. 24); however, the data is quite noisy and quantitative treatment cannot be applied.

Carbon isotopes also become heavier when weight/individual increases (fig. 25). However, the trend here is less clear than for the oxygen isotopes. In fig. 26 $\delta^{13}C$ vs. weight/individual is shown for different species. Despite the spread of the data points, the same trend of heavier isotopic composition when weight/individual increases can be observed. This trend is especially clear for the Globorotaliids, who also showed positive linear correlation between $\delta^{18}O$ and $\delta^{13}C$ (fig. 8).

The progression of oxygen isotopic composition for individual species from plankton tows to sediment traps and sediments is shown in fig. 27. The vertical axis represents schematic depth so that the steepness of the gradient in $\delta^{18}O$ should be considered qualitatively. Three types of behavior are observed: 1) increase in $\delta^{18}O$ with depth (referred as normal), 2) decrease in $\delta^{18}O$ with depth (reverse), and 3) no change with depth. When considering the trends in fig. 27 (see also table 3) it must be remembered that plankton and sediment trap samples are affected by seasonality,

Figure 23. Weight per individual vs. $\delta^{18}O$ for planktonic foraminifera. Note the general positive trend, and the separation between plankton and sediment samples. (*G. conglobatus* data are excluded.)

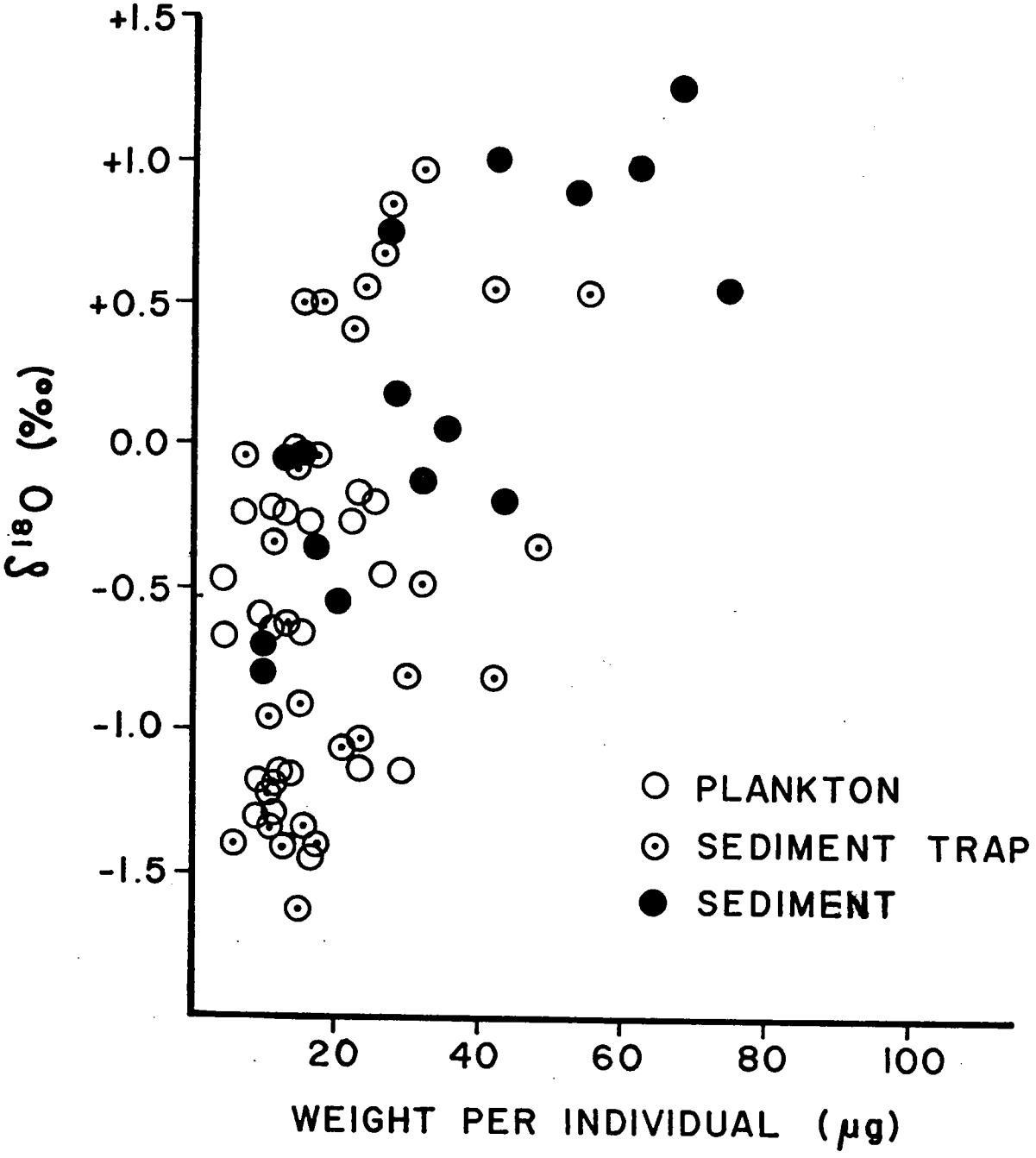

+1.5
+1.0
+0.5
0.0
-0.5
-1.0
-1.5
δ18O (‰)
20
40
60
80
100
WEIGHT PER INDIVIDUAL (μg)
PLANKTON
SEDIMENT TRAP
SEDIMENT

Figure 23.

Figure 24. Trends shown by individual species for weight per individual vs. $\delta^{18}O$ plot. Note the positive trend shown by most species.

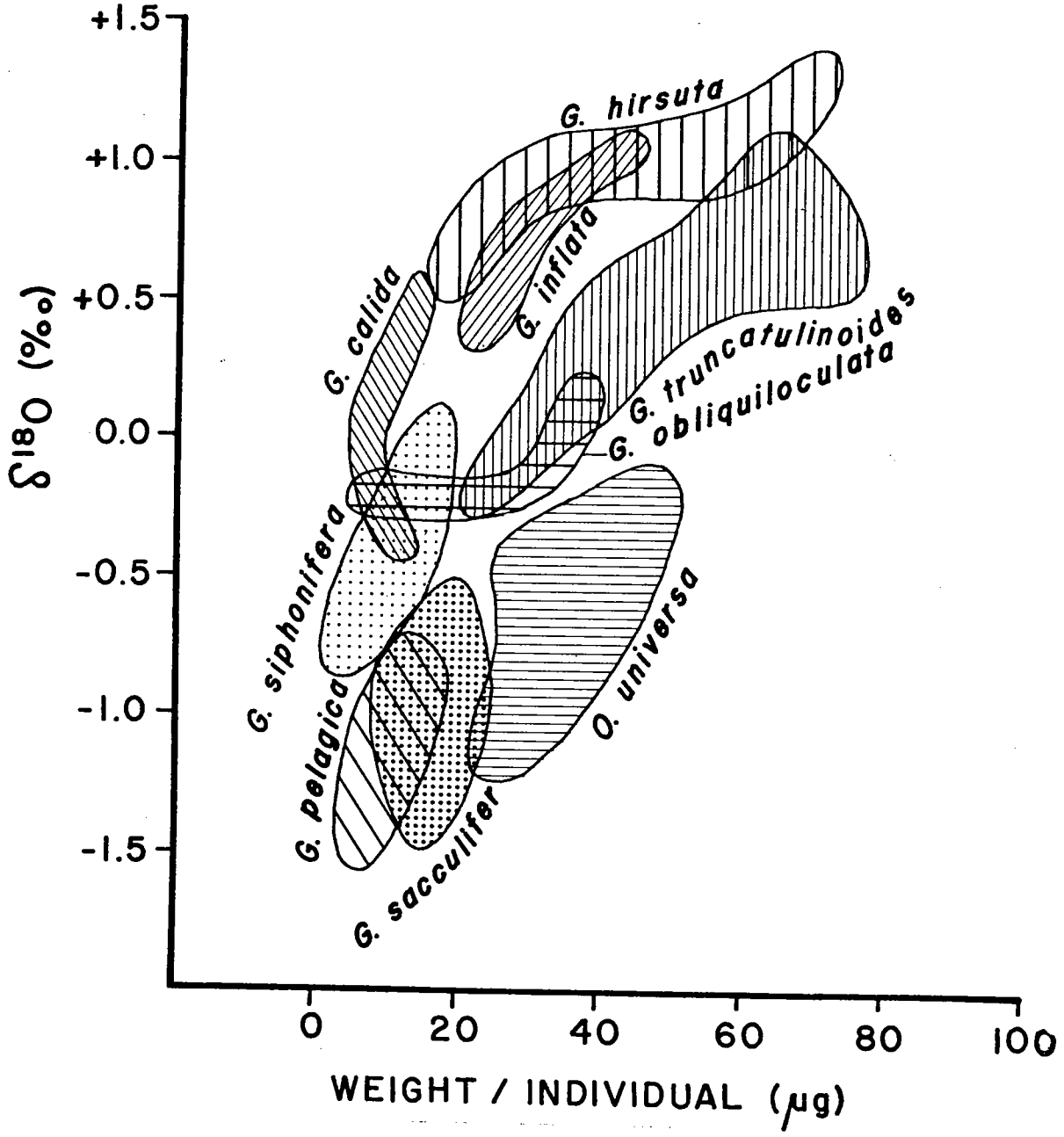

+1.5
+1.0
+0.5
0.0
-0.5
-1.0
-1.5
δ18O (‰)
G. hirsuta
G. calida
G. inflata
G. truncatulinoides
G. obliquiloculata
G. siphonifera
G. pelagica
O. universa
G. sacculifer
0
20
40
60
80
100
WEIGHT / INDIVIDUAL (μg)

Figure 24.

Figure 25. Weight per individual vs. $\delta^{13}C$ for planktonic foraminifera. Note the general positive correlation.

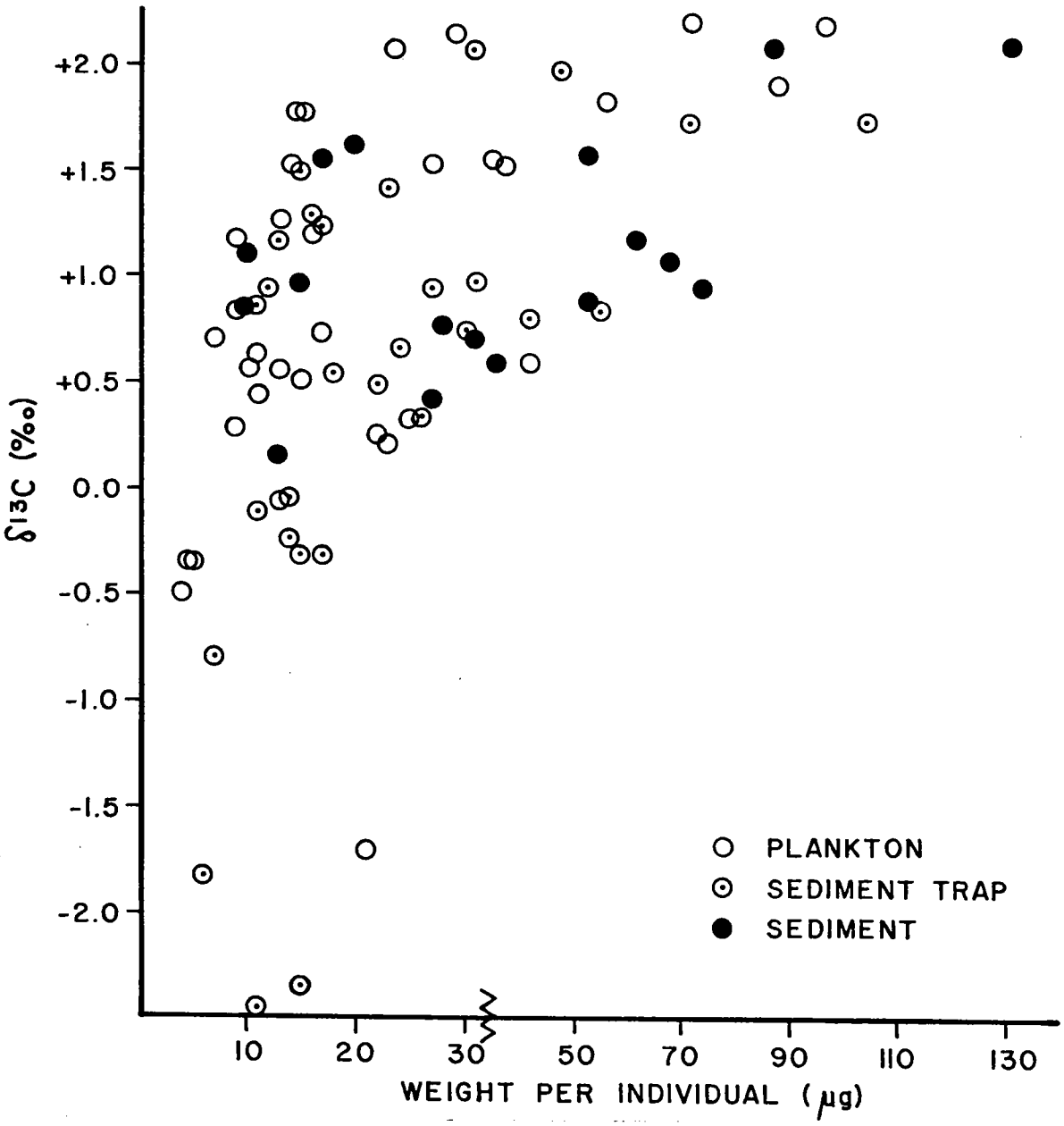

+2.0
+1.5
+1.0
+0.5
0.0
-0.5
-1.0
-1.5
-2.0
δ13C (‰)
10
20
30
50
70
90
110
130
WEIGHT PER INDIVIDUAL (μg)
PLANKTON
SEDIMENT TRAP
SEDIMENT

Figure 25.

Figure 26. Weight/individual vs. $\delta^{13}C$ for separate species. Note the general trend of increasing $\delta^{13}C$ with increase in weight per individual.

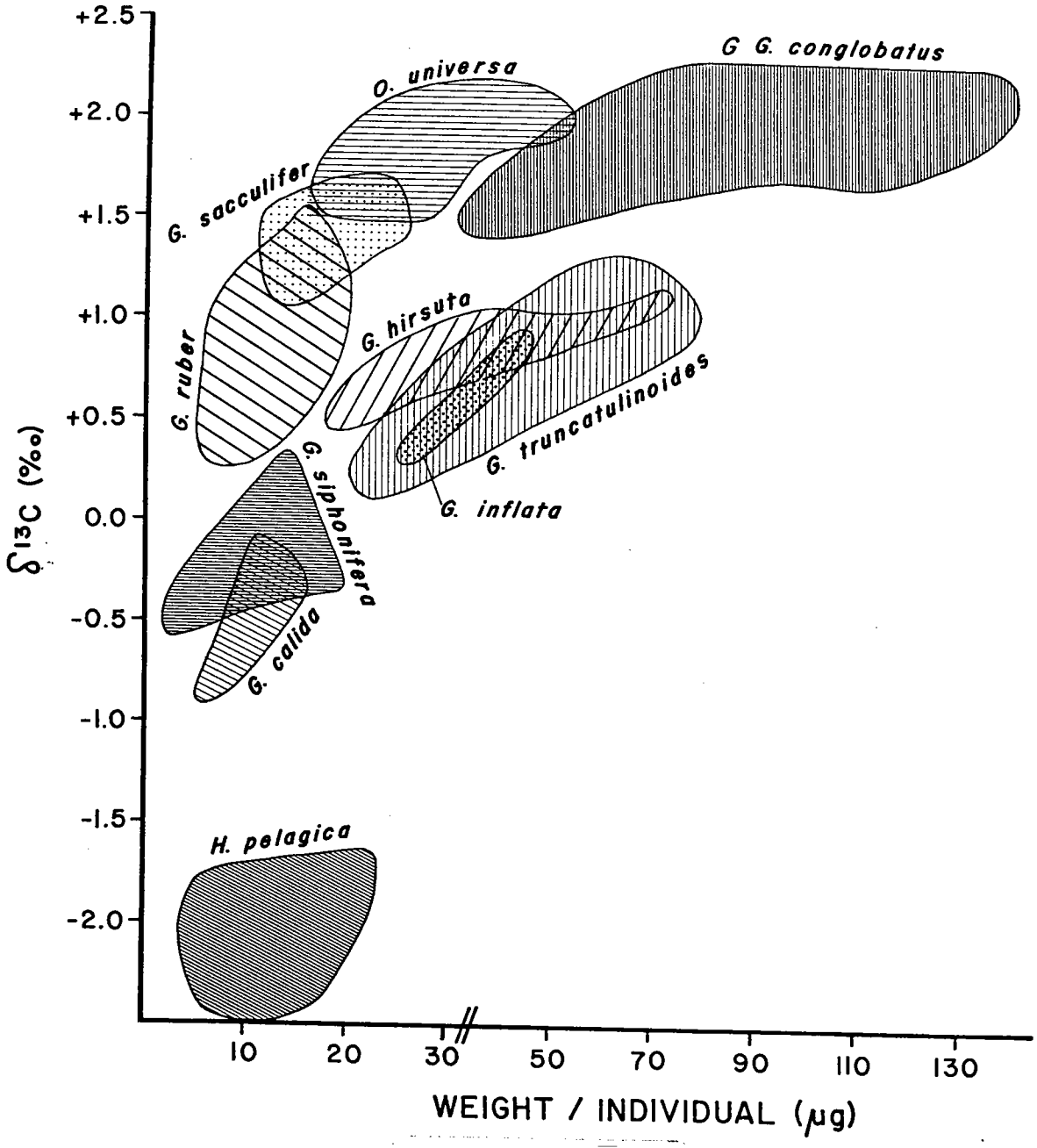

+2.5
+2.0
+1.5
+1.0
+0.5
0.0
-0.5
-1.0
-1.5
-2.0
δ13C (‰)
G. conglobatus
O. universa
G. sacculifer
G. ruber
G. hirsuta
G. truncatulinoides
G. inflata
G. siphonifera
G. calida
H. pelagica
10
20
30
50
70
90
110
130
WEIGHT / INDIVIDUAL (μg)

Figure 26.

Figure 27. Comparison of $\delta^{18}O$ for individual species in plankton tows, sediment traps and bottom sediments.

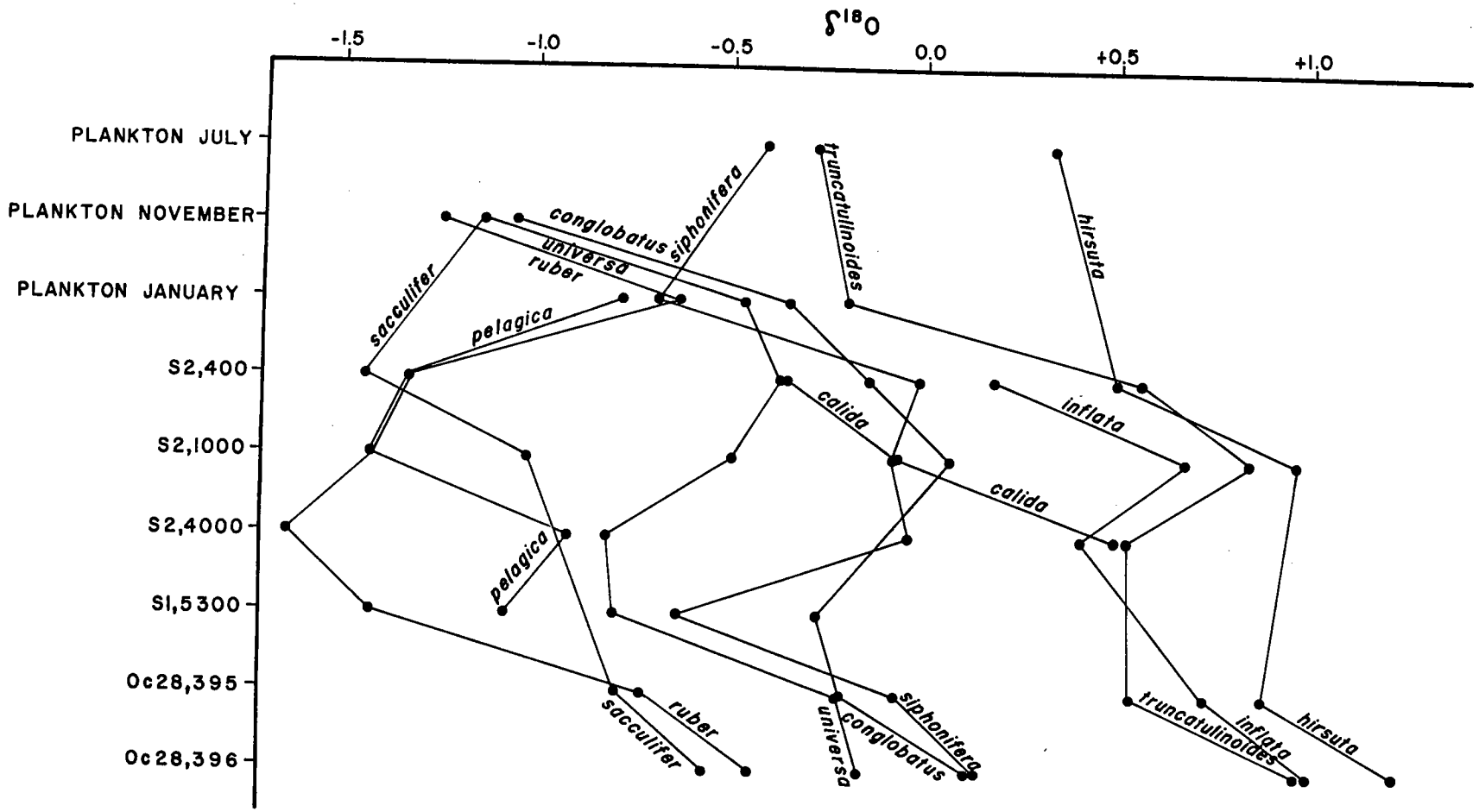

δ18O
-1.5
-1.0
-0.5
0.0
+0.5
+1.0
PLANKTON JULY
PLANKTON NOVEMBER
PLANKTON JANUARY
S2,400
S2,1000
S2,4000
S1,5300
Oc28,395
Oc28,396
conglobatus siphonifera
universa
ruber
sacculifer
pelagica
truncatulinoides
hirsuta
calida
inflata
calida
pelagica
sacculifer
ruber
universa
conglobatus
siphonifera
truncatulinoides
inflata
hirsuta

Figure 27.

especially for the shallow and intermediate species.

The Globorotaliids (*inflata, hirsuta* and *truncatulinoides*) basically show the expected normal trend from plankton to S2,1000. The reverse trend between S2,1000 and S2,4000 may be caused by seasonality. It suggests that a part of the population of these species was already sinking below 1000 m when the S2 array was deployed. Alternatively, a population of these species could have bloomed towards the end of the deployment period (20 November). This bloom reached the upper two traps but did not sink yet to reach the 4000 m trap. The isotopic composition in box-core 395 seem to represent the yearly production and is slightly ligher than S2,1000. The normal trend shown between the two box cores can be attributed to dissolution and will be discussed below. *G. calida* shows a normal trend between the three upper traps. This may suggest that this species calcifies even below 1000 m. *G. conglobatus* shows a normal trend down to 1000 m. The reverse trend below that cannot be explained easily. If it represents seasonality, and peak of productivity for this species started towards the end of November, it must have been collected in the S1,5300 that lasted from October 20 1975 to January 5, 1976. But *G. conglobatus* in S1,5300 is isotopically lighter than in S2,1000. Perhaps seasonal changes in depth habitat may explain this profile. The reverse trend shown by *O. universa, H. pelagica* and *G. ruber* between January plankton tows and the deep traps may well be

explained by seasonality. It seems that the lower traps did not have a chance to collect the peak winter production of these populations. *G. siphonifera* shows a very complex behavior somewhat like *G. conglobatus* that cannot be explained with the information at hand. *G. sacculifer* shows normal trend between 400 m and 1000 m that indicates additional skeleton deposition below 400 m. Perhaps this part of the skeleton is a crust (or "cortex" that eventually leads to the formation of *S. dehiscens* according to the scheme suggested by Bé (1965) and Bé and Hemleben (1970). The profile for *G. sacculifer* also suggests that the production of this species peaks in August and September.

My results are in good agreement with many micropaleontological studies that show additional skeleton deposition at depth and below the photic zone (Murray, 1897; Ericson, 1959; Bé and Ericson, 1963; Bé and Lott, 1964; Orr, 1969; Bé, 1965; Bé and Hemleben, 1973; Bé et al., 1973). Few isotopic studies that addressed this problem are also in agreement with my data and interpretation. Horibe et al. (1969) found a difference of 1.8 °/₀₀ and 4.5 °/₀₀ between encrusted and non-encrusted stages of *G. menardii* and *G. truncatulinoides*, respectively. Grazzini (1976) found that *G. truncatulinoides*, *G. inflata* and *G. ruber* with thick encrusted shells were isotopically heavier than thin shelled ones by 0.60 °/₀₀, 0.52 °/₀₀ and 1.06 °/₀₀, respectively.

In summary:

1) Surface dwelling planktonic foraminifera collected in plankton tows show light oxygen isotopic compositions which possibly can be out of equilibrium by roughly 0.5 °/₀₀.

2) Additional skeleton deposition below the photic zone in these species can account for 40-50% of their skeletal mass. The isotopic composition of this skeleton is heavier and counterbalances the light isotopic composition shown by the plankton tow populations. Thus the shallow dwelling foraminifera collected in sediment traps show equilibrium compositions for skeletons deposited within the photic zone.

3) Deep dwelling species do not show negative deviation from oxygen isotopic equilibrium but may have considerable vertical migrations in the water column. These species continue to deposit their skeleton well below the photic zone roughly to a depth of 800-1000 m.

4) The oxygen isotopic composition of surface dwelling and perhaps also deep dwelling species records seasonal temperature changes. This introduces considerable variability into the isotopic data.

5) The average weight/individual increases with depth and shows positive correlation with $\delta^{18}O$ and $\delta^{13}C$. This indicates additional skeleton deposition at depth especially for deep dwelling species.

C. Dissolution of planktonic foraminifera and its effect on the stable-isotope composition

Differential dissolution of planktonic foraminifera shells was first observed by Murray (1897) and was later described by many authors (Arrhenius, 1952; Phleger et al., 1953; Ruddiman and Heezen, 1967; Pytkowicz and Fowler, 1967; Berger, 1968, 1970a&b, 1971; Parker and Berger, 1971; and many others). It was realized that as a result of dissolution the assemblage composition changes and becomes enriched in the "resistant" foraminifera which represent a colder, deeper, or higher latitude character than the original one.

In these studies emphasis was given to the differences between species rather than within species. Skeleton deposition at depth was not studied directly, and the material used (foraminifera from plankton tows and from sediments) does not represent death populations. Nevertheless, the following tentative conclusions are summarized by Berger (1970a):

1) "Various phenotypes of the same species possess different resistances to solution. In general, the thin-shelled transparent forms with the last chamber larger than the previous ones are destroyed more rapidly.
2) The phenotypic composition of a foraminiferal death assemblage is changed by the effects of solution. The assemblage tends to become enriched with opaque (usually

thick shelled), zero and negative forms, i.e. specimens with small terminal chambers.

3) The pattern of the change is not necessarily identical for all species; layer-by-layer removal, chamber-by-chamber solution and overall etching may be of relatively different importance."

Berger (1971) pointed out that if additional skeleton deposition occurs in deeper and colder water, dissolution may have an important influence on the stable isotope composition of foraminiferal population because these heavily calcified forms are preferentially preserved. Savin and Douglas (1973) and Savin and Stehli (1974) discussed the effects of dissolution on the $\delta^{18}O$ of planktonic foraminifera shells. They showed that the actual observed surface temperature is higher than the isotopic temperature, and that this deviation (ΔT) between the two is increasing with depth (at least for *G. sacculifer, C. conglobatus* and *G. menardii*). This was attributed to dissolution which removes preferentially the more fragile individuals which happened to live in shallower and warmer water. Berger and Killingley (1977) have also shown that the isotopic composition of *G. sacculifer* becomes heavier due to dissolution. My data support Berger's (1971) interpretation that the shift towards heavier isotopic composition is caused by preferential removal of the thin shelled and isotopically light individuals. In table 10, comparison between the two box cores that were

collected at 4500 m and 4950 m respectively shows that for almost all species $\delta^{18}O$ and $\delta^{13}C$ were heavier in the deeper core. On the average, the difference in $\delta^{18}O$ is quite small (+ 0.22 °/₀₀). If shallower and deeper box cores were available, perhaps a larger deviation would be observed. The average isotopic temperature difference is 1.07°C and it implies an average depth difference of 46 m (see also fig. 20). In addition to the change in isotopic composition, five species showed an increase in the average weight/individuals in the deeper sample. This suggests that the fragile and light weight individuals were indeed removed by dissolution, and as a result $\delta^{18}O$ and $\delta^{13}C$ become heavier, in accord with the relationship between weight/individuals and isotopic composition shown in figs. 23-26.

An in-situ dissolution experiment was carried out during the S1 deployment, using an in-situ water circulator (ISWAC). The results are described by Honjo and Erez (1978). A foraminiferal assemblage from the Rio Grande Rise at 2190 m was used in this study and lost 30.4% of its weight during 79 days at 5518 m.

Isotopic composition and weight/individual were measured on selected species (larger than 250 μ) from this dissolved assemblage and compared to the undissolved standard samples. In addition, similar comparison was done for monospecific populations of *G. bulloides*, *G. pachyderma*

and *G. sacculifer* that were dissolved in separate chambers of the ISWAC (Honjo and Erez, 1978). The results are summarized in table 11. Essentially there is no difference between dissolved and undissolved samples. Some species become lighter, others somewhat heavier but the average difference is certainly insignificant. This unexpected result does not agree with all the other observations discussed above. However, it can be explained if the nature of the samples used here is considered. The Rio Grande Rise material is almost pure foraminiferal sand. Considerable deep currents exist in this area which winnow the sediment intensively (Johnson et al., 1977). As a result, this assemblage does not contain fragile, transparent and light weight individuals. It was suggested that the removal of these fragile individuals causes the shift in the isotopic composition of a species population. Winnowing and/or dissolution had already removed this fragile part of the population before our experiment was carried out. Therefore, additional dissolution did not change the isotopic composition. The scheme shown in fig. 28 may help to explain this subject. It suggests that as dissolution proceeds, the fragile individuals are dissolved. As a result, the weight/individual reaches a maximum. At this stage $\delta^{18}O$ and $\delta^{13}C$ also reach their maxima and additional dissolution may only make the foraminifera thinner but will not change the isotopic composition. In case of species that get encrusted

like *G. sacculifer* and the Globorotaliids, it is even possible to get lighter isotopic compositions as dissolution proceeds. This is because the crust which is isotopically heavier (Grazzini, 1976; Horibe et al., 1969) is removed and the leftover skeleton is isotopically lighter (fig. 28). Support for this scheme and its application to the ISWAC sample comes from table 11. It shows that there was no change in the species abundance as a result of dissolution. At the same time the % weight loss/individual was more or less the same for the species under consideration, and the average was close to the weight loss for the entire sample (which had 20 different species and ranged from 63 to 1000 μ). This suggests that the only change caused by dissolution was to make the shells thinner, and thus every individual lost some 30% of its weight, while the number of individuals remained the same. If dissolution would have increased the weight/individual (at least for some species), the isotopic composition would have become heavier, as indeed was found for the box core samples (table 10).

TABLE 10.

DIFFERENCE BETWEEN THE ISOTOPIC COMPOSITION OF FORAMINIFERA IN BOX CORES 396 AT 4950 m AND 395 AT 4500

species	$\Delta\delta^{18}O$ (‰) 4950-4500 m	Δt (°C) 4950-4500 m	$\Delta\delta^{13}C$ (‰) 4950-4500 m	Depth (m) 4950-4500 m	Δ Av. wt/ind. 4950-4500 m (µg)
G. ruber	+0.28	-1.24	+0.09	+25	0**
G. truncatulinoides	+0.43	-2.34	+0.47	+145	-12†
G. inflata	+0.27	-1.28	+0.20	+80	+15
G. hirsuta	+0.34	-1.71	+0.31	+85	+15
G. conglobatus	+0.33	-1.65	+0.02	+55	+44
G. sacculifer	+0.23	-1.07	+0.54	+25	+10
G. siphonifera	+0.21	-1.02	+0.78	+50	+2
G. universa	+0.05*	-0.27	+0.18	+0	-26
G. obliquiloculata	-0.18*	+0.93	+0.12	-50	-4
Average	+0.22	-1.07	+0.31	+46	+5

* Note that these species do not show increase in $\delta^{18}O$, and also do not show increase in the average weight/individuals.

** *G. ruber* did not show increase in weight/individual because dissolution removes the large transparent (*G. ruber* type) and retain the small thick shelled and isotopically heavier *G. elongatus* type.

† *G. truncatulinoides* shows slight decrease in weight/individual apparently because all fragile individuals were already dissolved and even the thick ones become thinner (see fig. 28).

TABLE 11.
Comparison between isotopic composition of dissolved and non-dissolved foraminifera in the ISWAC experiment

	species	$\Delta\delta^{18}O$ diss.- std.	$\Delta\delta^{13}C$ diss.- std.	% wt. loss/indiv.	abundance (dissolved)	% abundance (std.)
Rio Grande Assemblage	*G. ruber*	-0.09	-0.08	25	31.1	32.3
	G. sacculifer	-0.13	-0.24	21	8.7	7.9
	O. inflata	-0.13	-0.27	18	22.6	16.5
	G. pachyderma	+0.11	+0.12	18	5.0	3.3
	G. truncatulinoides	-0.09	-0.07	29	13.0	14.8
	G. menardii	+0.36	-0.01	47	0.9	0.4
	G. bulloides	+0.13	-0.24	23	2.6	3.1
monospecific	*G. bulloides*	+0.03	+0.10	23		
	G. pachyderma	-0.13	-0.14	31.1		
	G. sacculifer	-0.01	-0.06	36.3		
	Average	-0.02	-0.09	26.0		

Figure 28. Scheme for the influence of dissolution on the average weight per individual and $\delta^{18}O$ of planktonic foraminifera.

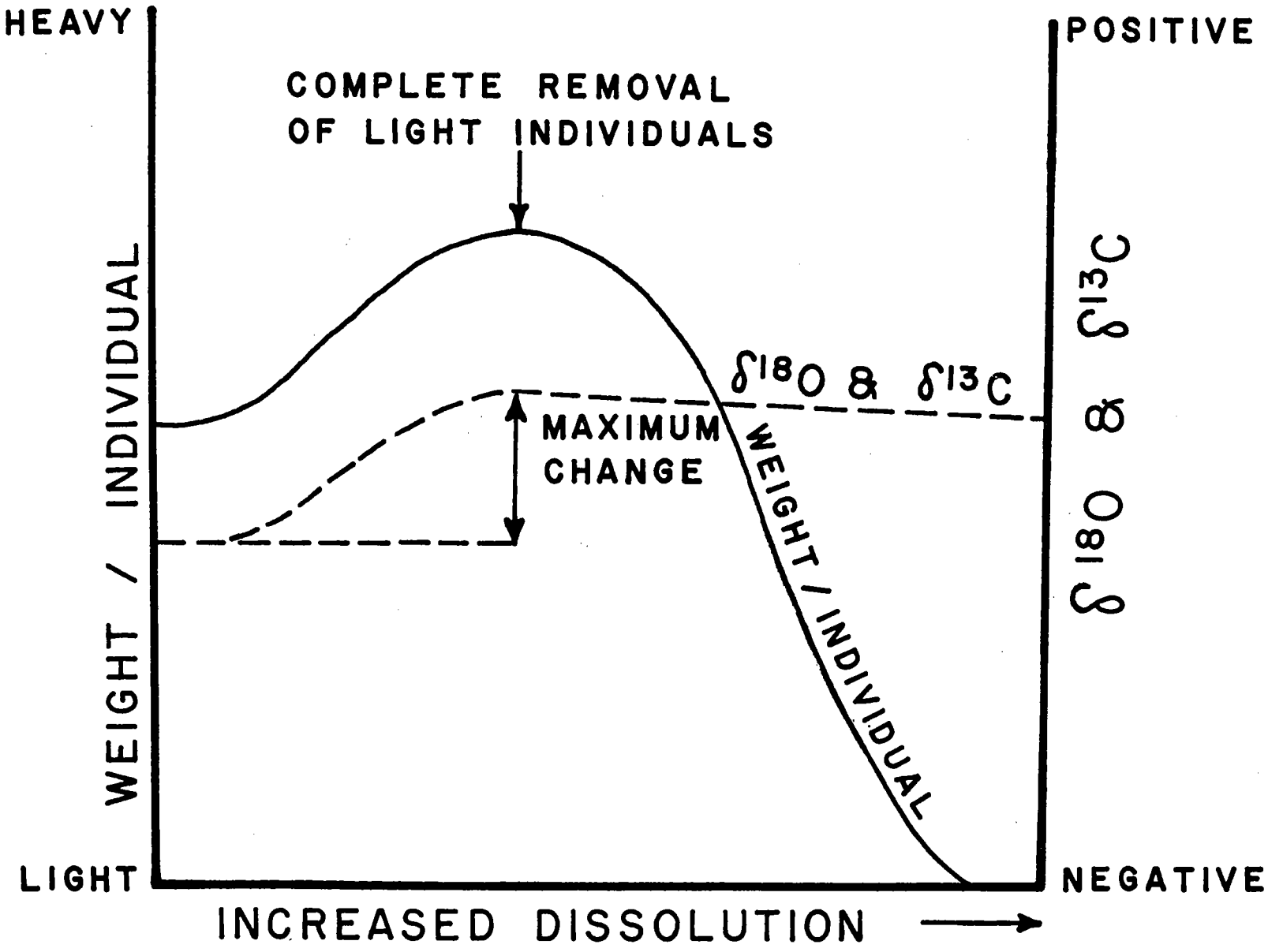

HEAVY
POSITIVE
COMPLETE REMOVAL
OF LIGHT INDIVIDUALS
WEIGHT / INDIVIDUAL
δ18O & δ13C
MAXIMUM
CHANGE
WEIGHT / INDIVIDUAL
δ18O & δ13C
LIGHT
NEGATIVE
INCREASED DISSOLUTION

Figure 28.

CHAPTER III

DISSOLUTION EFFECTS ON THE ISOTOPIC RECORD OF THE PLEISTOCENE

1. Introduction

Based on the earlier two chapters, the following scheme can be adapted to describe how planktonic foraminifera acquire their oxygen isotope composition. Living populations in the photic zone deposit isotopically light skeleton in slight negative deviation from isotopic equilibrium. It appears that deviation from equilibrium may be caused by incorporation of metabolic CO_2 which is enriched in ^{16}O and ^{12}C. Symbiotic algae, when present, tend to increase the amount of metabolic CO_2 in the skeleton, hence make it isotopically lighter. Part of the living population starts to sink below the photic zone, and into the thermocline. While sinking calcification continues, more chambers and layering are added, and some species start to develop thick $CaCO_3$ encrustations over their existing skeletons. The thickened outer parts of the skeleton are isotopically heavier because they are deposited in colder water and because the metabolic activity of the organisms is probably less intensive, especially for species that contain symbiotic algae. Thus, individuals that continue to calcify while sinking have thicker, larger and isotopically heavier skeletons than individuals that do not calcify below the photic zone. Foraminifera that reach the ocean floor contain many fragile, transparent and isotpically light individuals as indicated in deep sediment trap samples. However, foraminifera populations

in sediments, even in relatively shallow water, contain very few transparent and fragile individuals, and are significantly enriched in the thicker, more heavily calcified individuals. This fact has been described by many workers before and is documented again in Chapter II. Dissolution and (to a lesser degree) winnowing are the prime causes for this difference between foraminifera collected in the water column (thanatocoenosis) and foraminifera that accumulate in the sedimentary column (taphocoenosis).

Because thicker and larger individuals are isotopically heavier, sediment populations are isotopically heavier than water column populations. This is probably the reason why foraminifera from core tops and box cores show oxygen isotopic composition that is in equilibrium with the overlying water, while plankton tows and sediment traps may show slight, non-equilibrium isotopic compositions.

In summary, dissolution of planktonic foraminifera will make the isotopic composition of a species population heavier. Preservations (or no dissolution) will retain the original isotopic composition of a species population which is often somewhat lighter than equilibirum.

Dissolution and preservation cycles of planktonic foraminifera are well documented in the upper Pleistocene record in deep sea cores. The purpose of this chapter is to show how these cycles influenced the isotopic record of the upper

Pleistocene section in deep sea cores. Before doing so, however, the timing of dissolution cycles in the Atlantic and Pacific Oceans must be assessed.

2. Dissolution Cycles of $CaCO_3$ in the Atlantic and Pacific Oceans

Existence of cyclic variations in percent $CaCO_3$ in the upper Pleistocene section of deep sea cores was reported by many authors (e.g. Arrhenius, 1952; Hays et al., 1969; Phleger et al., 1963; Gardner, 1975; Damuth, 1975; and Johnson et al., 1977). There is a general agreement that these cycles have a similar frequency to the glacials and interglacials (op. cit.). However, in the Pacific $CaCO_3$ percent is high during the glacials and low during the interglacials. In the Atlantic the opposite is observed: $CaCO_3$ percent is higher during the interglacials and lower during the glacials.

3. Evidence for $CaCO_3$ cycles representing dissolution cycles

Much debate exists on the interpretation of the carbonate cycles. Perhaps the main reason for this debate is the above mentioned phase lag between the Atlantic and the Pacific with respect to the timing of the $CaCO_3$ cycles.

In the Pacific, Berger (1973) has shown that during the glacial, shells of planktonic foraminifera are better preserved, that benthonic/planktonic ratio is lower, and dissolution susceptible species are more abundant. These qualitative

dissolution indices coincide well with higher $CaCO_3$ percent during the glacials. During the interglacials poor preservation of planktonic foraminifera coincides with low $CaCO_3$ percent. Luz and Shackleton (1975) support Berger's interpretation for the Pacific and provide better stratigraphical control on the timing of dissolution/preservation cycles. In applying the same criteria to assess dissolution/preservation cycles in the Atlantic, Berger's data (1973) are not conclusive and can be interpreted either way. More recent work in the Atlantic, especially by CLIMAP has substantiated that interglacial high $CaCO_3$ percent in the Atlantic is associated with better foraminiferal preservation and that glacial low $CaCO_3$ percent is associated with poor foraminiferal preservation (Gardner, 1975; Damuth, 1975; Bé et al., 1976; Thunnell, 1976). Several authors, however, claimed that $CaCO_3$ cycles in the Atlantic are not caused by dissolution (e.g. Broecker, 1971; Berger, 1968a, 1973; Luz and Shcakleton, 1975; Ruddiman, 1971). Therefore, in the following section I summarize the evidence showing that $CaCO_3$ cycles in the Atlantic are caused by dissolution. Gardner (1975) showed that in the eastern equatorial Atlantic, cores from relatively shallow water ($\sim$3000 m) have high $CaCO_3$ percent ($\sim$80%) and that glacial/interglacial variations are small. Cores from increasing water depth show lower average $CaCO_3$ percent and progressively larger glacial/interglacial amplitudes (Gardner, 1975, fig. 3,4). This systematic variation in depth cannot be attributed to changes in productivity because these

cores are located close to each other. Carbonate dissolution is strongly controlled by depth (i.e. pressure, temperature) and thus can better explain this variation. Gardner (1975) has also shown that dilution alone cannot be responsible for the variation in $CaCO_3$ observed in this region. In the following, Gardner's calculation is expanded to calculate the minimum possible extent of dissolution during glacials in the eastern equatorial Atlantic. The following quantities will be considered (expressed in length units of a core section):

Ci - carbonate section in the interglacial

Cg - carbonate section in the glacial

Ni - non-carbonate section in interglacial

Ng - non-carbonate section in glacial

$$\text{interglacial } \%\ CaCO_3 = \frac{Ci}{Ci+Ni} \times 100 = Fi$$

$$\text{glacial } \%\ CaCO_3 = \frac{Cg}{Cg+Ng} \times 100 = Fg$$

Ni + Ci = total sediment section accumulated during interglacial

Ng + Cg = total sediment section accumulated during glacial

$R = \frac{Ng+Cg}{Ni+Ci}$ sediment accumulation ratio between glacial and interglacial

If sedimentation is uniform, R should be 1.23 because this is

the time ratio between the glacial and interglacial sections considered (Y/X time zones ratio in Gardner, 1975). If there is no dissolution of $CaCO_3$ (Ci = Cg), the changes in the carbonate fraction must be caused by dilution. Cores deeper than 4000 m show an average of 50% $CaCO_3$ in the interglacials and 10% $CaCO_3$ in the glacials. Thus, assuming Ci = Cg,

$$Fi = \frac{Ci}{Ci+Ni} \times 100 = 50\%$$

$$Fg = \frac{Ci}{Ci+Ng} \times 100 = 10\%$$

$$R = \frac{Fg}{Fi} = \frac{Ng+Ci}{Ni+Ci} = 5$$

This implies that the glacial section should be 5 times longer than the interglacial section. However, Gardner's cores show that R is approximately 2, which indicates some dilution by non-carbonate material but not nearly enough to make R equal 5. Therefore, there must be some dissolution during the glacial in order for Ci to be larger than Cg. Assuming R = 2, it is possible to calculate the ratio between Cg and Ci.

(1) $$R = \frac{Cg+Ng}{Ci+Ni} = 2$$

(2) $$Fg = \frac{Cg}{Cg+Ng} \times 100 = 10$$

(3) $$Fi = \frac{Ci}{Ci+Ni} \times 100 = 50$$

Divide (2) by (3)

$$\frac{Cg}{Ci} = \frac{Cg+Ng}{Ci+Ni} \cdot \frac{10}{50}$$

and from (1) $\frac{Cg}{Ci} = \frac{2}{5}$

This calculation shows that $CaCO_3$ accumulation during glacials is only $\frac{2}{5}$ (or 40%) of $CaCO_3$ accumulation during the interglacial and suggests that at least 60% of the $CaCO_3$ deposited during the glacial have dissolved. Sixty percent is only a lower limit on dissolution because the dilution observed is made by material that contains some $CaCO_3$ (up to 30%). In addition, Gardner (1975) showed that benthonic/planktonic ratio and shell fragmentation indicated dissolution in the low carbonate glacial sections and preservation in the high carbonate interglacial sections. The stratigraphic control on these cores is based on faunal zonation (Gardner, 1975) as well as $\delta^{18}O$ curves and ^{14}C dates (Damuth, 1975; Gardner and Hays, 1976), leaving little doubt on the timing of these dissolution cycles.

Bé et al. (1976) showed a similar pattern in the western equatorial Atlantic. Glacial sections have low carbonate percent, decrease of the coarse fraction, excessive fragmentation of planktonic foraminifera, absence of pteropods and increase in the benthonic/planktonic foraminifera ratio. Johnson et al. (1977), Prell and Hays (1976) and Thunnell (1976) found similar dissolution cycles in the southwestern Atlantic Columbia Basin and the Gulf of Mexico, respectively. In summary, $CaCO_3$ cycles in both oceans are caused by dissolution; they have similar frequencies as the glacials and interglacials; and the Atlantic and the Pacific are out of phase with respect to the timing of dissolution.

Today's oceans seem to support the scheme mentioned above. In the present interglacial the Atlantic is in a preservation mode, the CCD is roughly at 5 km and the saturation level (Ω=1) level is at 4 km (Takahashi, 1975). The Pacific is in a dissolution mode, its CCD is between 3 and 4 km and the saturation level (Ω=1) is at 1 to 2 km (Takahashi, 1975). This major difference between the oceans is attributed to the difference in the total CO_2 (ΣCO_2) to alkalinity ratio which determine the undersaturation with respect to $CaCO_3$ in seawater (Li et al., 1969; Broecker, 1974; Takahashi, 1975; Bender and Graham, 1978). Deep water in the Atlantic has a short residence time ($\sim$ 400 yrs) (Broecker and Li, 1970) due to the high rate of production of NADW in the Norwegian and Labrador Seas. The NADW has low values of ΣCO_2 becuase it has lost its excess CO_2 to the atmosphere before being downwelled, and does not have enough time to accumulate significant amount of CO_2. The Pacific on the other hand does not have a source of deep water in its northern end, and its deep water has a much longer residence time ($\sim$ 1500 yrs) (Broecker and Li, 1970). Oxidation of organic matter produces CO_2, which accumulates with time, resulting in a high ΣCO_2/alkalinity ratio.

Weyl (1968) in his theory of glaciation suggested that NADW were not produced during the glacials. There is now evidence that during the last glacial the Norwegian and Labrador Seas were covered by ice (Kellogg, 1976) and that

production of NADW was significantly reduced (Schnitker, 1974; Streeter, 1973; Duplessy et al., 1975). Such circumstances may have made the glacial Atlantic more similar to the present-day Pacific, i.e. longer residence time for deep water, higher ΣCO_2/alkalinity ratio, shallower Ω=1 level and thus, increased dissolution. This mechanism, when alternating between the glacials and interglacials, could produce dissolution cycles in the Atlantic. It is harder to speculate on the mechanism that retards dissolution in the Pacific during the glacial because of lack of relevant data. One possibility can be offered:

During the glacial when sea level was 130 to 150 m lower than today the coral reefs and the other $CaCO_3$ rich shelves in the Pacific were exposed and probably eroded. Reefs growing at lowered sea levels accumulated smaller mass of $CaCO_3$ because the area suitable for reef growth on the shelves is drastically reduced (see hypsographic curve in Sverdrup et al., 1942). Thus, former sinks for $CaCO_3$ (the Pacific shelves and the Atlantic pelgaic sediments) became sources for dissolved carbonate which could add excess alkalinity to the deep Pacific and make the deep Pacific water less corrosive. In addition, lowering of sea level by 130-150 m during glacials lowers the pressure on the deep water by 13 to 15 atmospheres. Although this will cause only a few percent increase in Ω, it can reduce dissolution rates by a significant amount. Recent dissolution kinetic studies

(Morse, 1978; Keir, 1979) have shown that $CaCO_3$ dissolution in the ocean is governed by high order kinetics, where the rate of dissolution is proportional to $(1-\Omega)^n$ and n = 3 to 5.

Regardless of the mechanism involved, the timing of dissolution cycles during the Pleistocene are out of phase in the Atlantic and the Pacific.

4. Effects of dissolution cycles on the glacial-interglacial amplitude shown by $\delta^{18}O$ curves in deep sea cores

Most authors agree that the major factor controlling the glacial-interglacial isotopic amplitude (GIA) is the ice volume at the northern hemisphere and its isotopic composition (Shackleton, 1967; Imbrie et al., 1973; Shackleton and Opdyke, 1973, 1976; Shackleton and Emiliani, 1974; Berger and Gardner, 1975). Temperature, assumed to be the main factor by Emiliani (1955, 1966, 1971) seems to be of secondary importance (see more recent reviews by Hecht, 1974; Van Donk, 1977; and Shackleton, 1977a,b). As a result, remarkable similarity exists in the shape and frequency content of $\delta^{18}O$ curves between the two oceans. However, a large difference exists in the GIA between the two oceans (Emiliani, 1955, 1966; Emiliani and Shackleton, 1974; Berger and Gardner, 1975). In Table 13 most of the pertinent data, available on the GIA exhibited by planktonic foraminifera, is summarized. Most Atlantic cores show amplitudes of 1.9 °/oo while most Pacific cores show GIA of 1.1 °/oo. Emiliani, 1955, 1971), Imbrie et al. (1973), Shackleton and

TABLE 13

Isotopic Glacial-Interglacial Amplitude (GIA) for Planktonic Foraminifera from the Atlantic Ocean

Core #	Depth (m)	Location*	Species	$\delta^{18}O$(°/oo) Glacial	$\delta^{18}O$(°/oo) Interglacial	GIA(°/oo)	Sedimentation Rate(cm/1000yr)
V28-14	1855	NA	*G. pachyderma*	+4.64	+2.11	2.35	9.4
RC11-86	2829	SA	*G. succulifer*	+0.79	-0.55	1.34	2.5
RC12-294	3308	SA	*G. bulloides*	+2.88	+0.96	1.92	2.8
V19-240	3103	SA	*G. inflata*	+2.37	+0.72	1.65	3.1
V19-248	3321	SA	*G. ruber*	+0.90	-0.13	1.03	1.2
V19-282	4356	EA	*G. dutertrei*	+1.60	+0.31	1.29	5.0
V22-38	3797	EA	*G. sacculifer*	+0.19	-0.95	1.14	1.8
V22-174	2630	EA	*G. ruber*	+0.32	-1.23	1.55	3.3
V22-174	2630	EA	*G. sacculifer*	+0.81	-0.92	1.73	3.3
A172-6	4160	CR	*G. sacculifer*	+0.09	-1.87	1.96	3.3
A179-4	2965	CR	*G. ruber*	-0.23	-2.04	1.81	1.8
A179-4	2965	CR	*G. sacculifer*	+0.16	-1.80	1.96	1.8
A179-4	2965	CR	*G. dubia*	+0.86	-1.04	1.90	1.8
A180-73	3749	EA	*G. sacculifer*	+0.51	-1.44	1.95	1.7
234A	3577	EA	*G. sacculifer*	+0.08	-1.76	1.84	1.2
234	3577	EA	*G. sacculifer*	+0.15	-1.80	1.95	2.4
235A	4560	EA	*G. sacculifer*	+0.24	-1.90	2.14	1.8
246	3210	EA	*G. sacculifer*	+0.04	-1.64	1.68	4.6
280A	4256	NA	*G. inflata*	+1.92	+0.63	1.29	1.6
P6304-8	3927	CR	*G. sacculifer*	+0.63	-1.29	1.92	3.9
P6304-9	4126	CR	*G. sacculifer*	+0.44	-1.35	1.79	3.3
A240-M1	4180	CR	*G. sacculifer ?*	+0.01	-1.71	1.72	2.2
A245 BR-C	2968	CR	*G. ruber*	-0.30	-1.57	1.27	2.8
V12-122	2800	CR	*G. sacculifer*	-0.01	-2.19	2.18	3.0
V12-122	2800	CR	*G. sacculifer*	-0.15	-2.47	2.32	3.0

* EA = Equatorial Atlantic
NA = North Atlantic
SA = South Atlantic
CR = Caribbean

TABLE 13 (continued)

Isotopic Glacial-Interglacial Amplitude (GIA) for Planktonic Foraminifera from the Pacific Ocean

Core #	Depth (m)	Location*	Species	$\delta^{18}O$(°/oo) Glacial	$\delta^{18}O$(°/oo) Interglacial	GIA(°/oo)	Sedimentation Rate(cm/1000yr)
BNFC43 PG3	2874	EP	*G. sacculifer*	-0.44	-1.80	1.36	1.1
RC8-94	3074	SP	*G. sacculifer*	+0.81	+0.05	0.76	1.8
RC10-114	2791	EP	*G. sacculifer*	-1.04	-1.66	0.62	0.9
RC11-210	4420	EP	*G. sacculifer*	+0.16	-1.30	1.46	2.2
RC11-210	4420	EP	*P. obliquiloculata*	+0.45	-0.92	1.37	2.2
RC11-230	3259	EP	*G. sacculifer*	-0.06	-1.08	1.02	1.7
V21-33	3726	EP	*G. sacculifer ?*	+1.48	+0.18	1.30	2.2
V21-59	2992	NP	*G. sacculifer*	-0.39	-1.19	0.80	0.8
V21-146	3968	NP	*G. inflata*	+2.32	+1.01	1.31	2.1
V24-109	2367	EP	*G. sacculifer*	-0.67	-2.08	1.41	2.7
V28-203	3243	EP	*G. sacculifer*	-0.45	-1.72	1.27	2.2
V28-235	1746	EP	*G. sacculifer*	-0.71	-1.80	1.09	1.9
V28-238	3120	EP	*G. sacculifer*	-0.96	-1.97	1.01	2.3
V28-239	3490	EP	*G. sacculifer*	-0.84	-1.72	0.88	1.6
WAH-8F2	4308	EP	*G. obliquiloculata*	+0.73	-0.63	1.36	1.7
Y69-106P	2870	EP	*G. sacculifer*	-0.20	-1.95	1.75	1.9
Y71-7-45P	3096	EP	*G. sacculifer*	+0.24	-1.12	1.36	0.5

* EP = Equatorial Pacific
NP = North Pacific
SP = South Pacific

Opdyke (1973), and Berger and Gardner (1975) tried to explain this difference by assuming that the Pacific responded only to the ice volume effect whereas in the Atlantic additional effects of temperature and evaporation/precipitation are superimposed on the ice volume effect.

I suggest that part of the difference observed in the GIA between the Atlantic and the Pacific can be caused by dissolution cycles. Dissolution effects on a species population are quite small, perhaps in the range of 0.2-0.3 °/oo (Savin and Douglas, 1973; Berger and Killingley, 1977; and Chapter II). However, the timing of dissolution-preservation cycles within and between the two oceans can amplify the effects of dissolution on the GIA up to 4 times its original effect on a species population. To understand this one should recall that dissolution will shift the isotopic composition of a species population to heavier value, while increased preservation will tend to make the isotopic composition lighter. In complete absence of dissolution, even light non-equilibrium isotopic compositions may be observed (Grazzini, 1976; Weiner, 1972, 1977; Emiliani, 1955, 1966; Douglas and Savin, 1973).

In the Atlantic relatively heavy isotopic composition during the glacial will become even heavier due to dissolution and light isotopic composition during interglacials will become even lighter because of preservation. Thus the total effect is to increase the GIA (fig. 31). In the Pacific the

light isotopic composition during interglacials will become slightly heavier due to dissolution while the heavy isotopic composition during glacials will become lighter due to preservation. This will reduce the GIA in the Pacific (fig. 31).

If we assume that dissolution or preservation shifts the isotopic composition only by 0.2 °/₀₀ the GIA in the Pacific will be reduced by 0.4 °/₀₀, and amplified in the Atlantic by 0.4 °/₀₀, thus the difference between the GIA in the two oceans will become 0.8 °/₀₀ which is quite close to the observed value (table 13, fig. 29). The following studies by other investigators support my interpretation:

Berger and Gardner (1975) have suggested that dissolution caused the extremely low GIA in the Pacific (0.2-0.8 °/₀₀) reported by Emiliani (1955). Shackleton and Opdyke (1976) also have demonstrated the effects of dissolution within the Pacific. They showed that Core V23-238 from 3120 m recorded a larger GIA than core V28-239 from 3450 m. The average difference in GIA between the two cores was 0.29 °/₀₀. In addition, Shackleton and Opdyke (1973) showed that the GIA recorded by benthonic foraminifera in core V28-238 is higher by 0.2 °/₀₀ than that recorded by *G . sacculifer*. The authors suggested that *G . sacculifer* migrated downwards during the interglacial to account for this observation. While this is possible, it is simpler to explain this difference by dissolution that reduced the GIA exhibited by *G . sacculifer* .

Figure 29. Glacial-interglacial amplitude for $\delta^{18}O$ curves of planktonic foraminifera in the Atlantic and the Pacific Oceans (sources of data in table 13). Note the separation between the Atlantic and the Pacific data points. The model of Peng et al. (1977) for the effect of bioturbation on the isotopic record cannot account for all the variance in the data. The difference between the two oceans is attributed to the influence of dissolution cycles on the isotopic record.

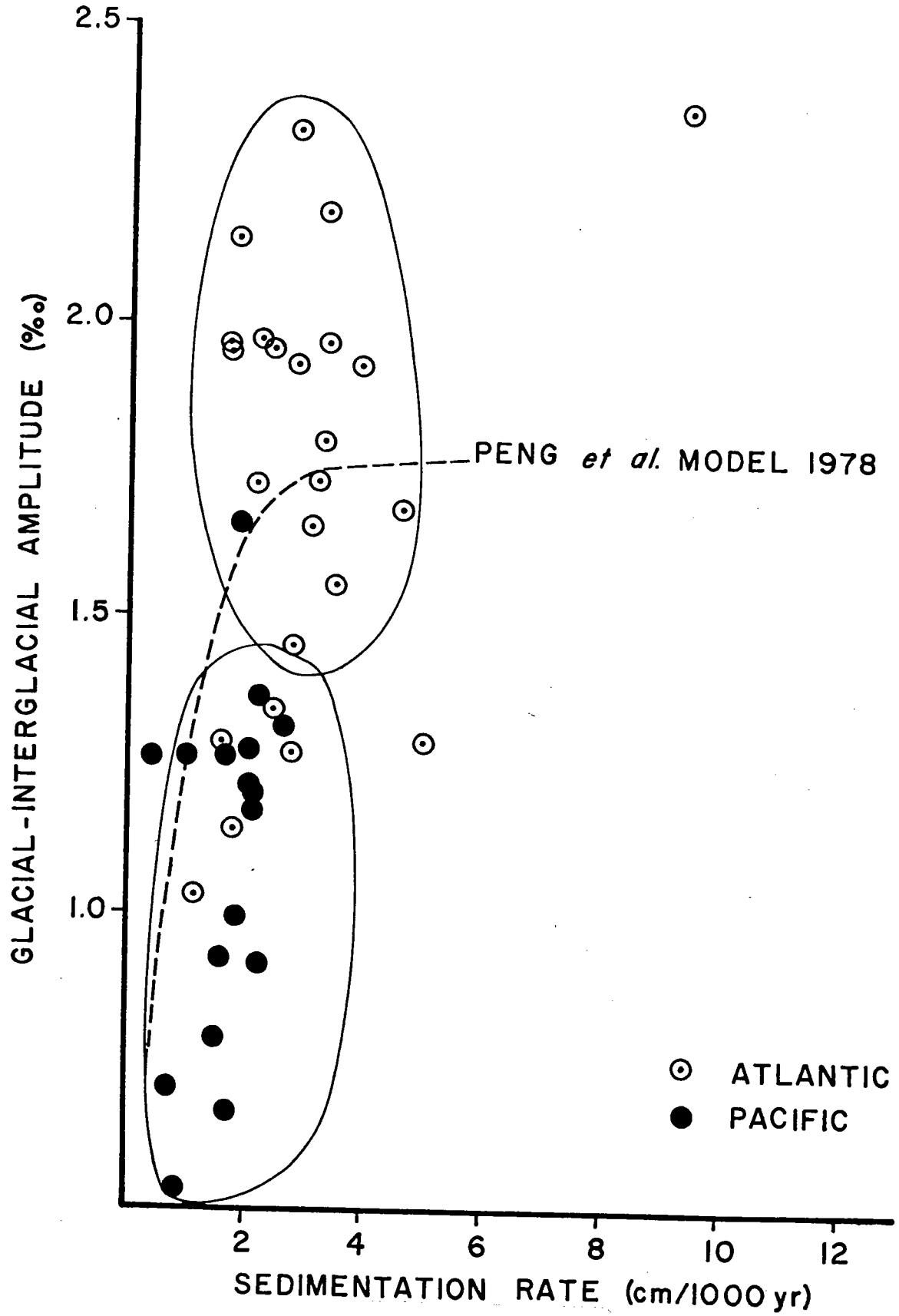

2.5
2.0
1.5
1.0
GLACIAL-INTERGLACIAL AMPLITUDE (‰)
PENG et al. MODEL 1978
ATLANTIC
PACIFIC
2
4
6
8
10
12
SEDIMENTATION RATE (cm/1000 yr)

Figure 29.

Berger and Killingley (1977) have recently compared the isotopic record of the last 19,000 years in two box cores from 1597 m and 3732 m in the western equatorial Pacific. They showed that the isotopic record of *P. obliquiloculata* (a dissolution resistant species) was unaffected by dissolution, while the isotopic record of *G. sacculifer* (which is dissolution susceptible) was significantly heavier in the deepr box core due to dissolution in the last 10,000 yrs. Thus the GIA recorded by *G. sacculifer* in the deeper core is lower by 0.3 °/₀₀ than the GIA recorded by *P. obliquiloculata*. In the Atlantic it is expected that dissolution resistant species will show a smaller GIA than dissolution susceptible species. In the Caribbean core A-179-4 (from 2965 m), Emiliani (1955), indeed found that while the dissolution susceptible *G. ruber* and *G. sacculifer* showed high GIA of 2.05 °/₀₀ and 1.94 °/₀₀ respectively, the dissolution resistant species, *G. dubia (dutertrei)* and *G. menardii* showed GIA of 1.37 °/₀₀ and 0.75 °/₀₀ respectively. This trend has been shown to persist through time by Lidz et al. (1968) who found that *G. tumida, G. truncatulinoides* and *G. crassaformis* exhibited lower GIA compared to dissolution susceptible species (*G. sacculifer, H. pelagica, O. aniversa* and *G. conglobatus*).

5. Discussion

Almost every observation discussed above can be interpreted to show that processes other than dissolution were in operation. Emiliani (1955, 1966, 1971), Shackleton and Opdyke (1973), Lidz et al. (1968), Berger and Gardner (1975) and many others suggested that higher GIA in the Atlantic is caused by temperature fluctuations between the glacial and the interglacial. Vertical migration of planktonic foraminifera combined with temperature changes between glacials and interglacials can account for the differences between the benthonics and the planktonics as well as the differences between shallow and deep dwelling planktonics. Comparison between isotopic temperatures and paleontologically determined temperatures (e.g. Emiliani, 1971; Imbrie et al., 1973; Van Donk, 1976; Gardner and Hays, 1976) seems to support the idea that some temperature fluctuations are superimposed on the isotopic record in the Atlantic. However, micropaleontological techniques have a large uncertainty in determining paleotemperatures (Berger and Gardner, 1975). This uncertainty, even when most sophisticated techniques are used, has the same magnitude as the estimated temperature changes between the glacials and interglacials (i.e., $\pm$ 2° to 3°C). In addition, the disagreement between different micropaleontological techniques is quite large (see table 6 in Hecht, 1974). Perhaps these uncertainties exist because

foraminifera are not so sensitive to temperature per se. Most planktonic species are horizontally distributed over quite a range of temperatures (see Bé, 1977, Table 3, for comprehensive review). Furthermore, many species migrate diurnally through a large temperature gradient which far exceeds the proposed glacial-interglacial temperature change. Thus, despite the fact that faunal composition changes were exhibited by foraminifera between the glacials and interglacials, their significance for accurate temperature determinations is questionable.

Recently Shackleton (1977a) has shown that the GIA recorded by benthonic foraminifera is at least 1.65 °/₀₀. This value is believed to represent the change in $\delta^{18}O$ of ocean water due to ice accumulation and melting. In comparison, *G. sacculifer* in the Pacific shows an average GIA of 1.13 °/₀₀ (based on 13 cores in Table 13). If this difference is interpreted in terms of temperature changes, it implies that *G. sacculifer* in the Pacific lived in warmer water during the glacial and colder water during the interglacial. While this is not impossible, it is more likely that this difference between *G. sacculifer* and benthonic foraminifera is caused by dissolution. If dissolution does reduce the GIA in the Pacific, it must also increase the GIA in the Atlantic. Indeed, *G. sacculifer* in the Atlantic shows an average GIA of 1.82 °/₀₀ (based on 14 cores, Table 13). Therefore, there is no need to evoke cooling and warming of

Atlantic surface water in order to explain the differences in the GIA between the two oceans.

Finally, effects of bioturbation on the isotopic record must be considered. It has been recognized since the pioneer work of Emiliani that bioturbation can blur the isotopic record and thus reduce the GIA (Emiliani, 1955, 1966; Shackleton, 1977a,b; Berger et al., 1977). Because biogenic mixing affects only the upper few centimeters, it is expected that cores with low accumulation rates will show lower GIA than those with high sedimentation rates. Shackleton (1977a,b) showed that this is indeed the case in a few selected cores from the Atlantic and the Pacific. However, in these comparisons no differentiation was made between planktonic and benthonic foraminifera. Peng et al. (1977) plotted the GIA vs. sedimentation rate for all cores analyzed by Shackleton (fig. 30). Their data show that cores with higher sedimentation rates indeed display higher GIA. A model of bioturbation suggested by these authors matches the data to a certain extent. However, the fit for benthonic foraminifera matches the model much better than for planktonic foraminifera. Most of the planktonic data points are from the Pacific and they display a lower GIA than expected if bioturbation was the main factor in operation. In fig. 31 I have shown the GIA vs. sedimentation rate for all the data available for planktonic foraminifera (mostly from Shackleton, Emiliani and Van Donk). In addition to the difference in GIA between the Atlantic and

the Pacific, it is also clear that most Pacific points fall below the predicted model of the GIA by Peng et al. (1977) and most Atlantic points fall above the expected GIA. Thus, one can conclude that while effects of bioturbation are certainly important, dissolution adds a significant modulation to the isotopic record that cannot be ignored. Perhaps when techniques like those suggested by Berger et al. (1977) to "unmix" the isotopic record are improved, better assessment of dissolution effects will be possible.

Figure 30. A model to account for effect of bioturbation on the glacial-interglacial amplitude of $\delta^{18}O$ (reproduced from Peng et al., 1977). Black circles represent planktonic foraminifera and open circles represent benthonic foraminifera. Note that while the model can account for the variability shown by benthonic foraminifera, it cannot fully explain the data shown by planktonic foraminifera.

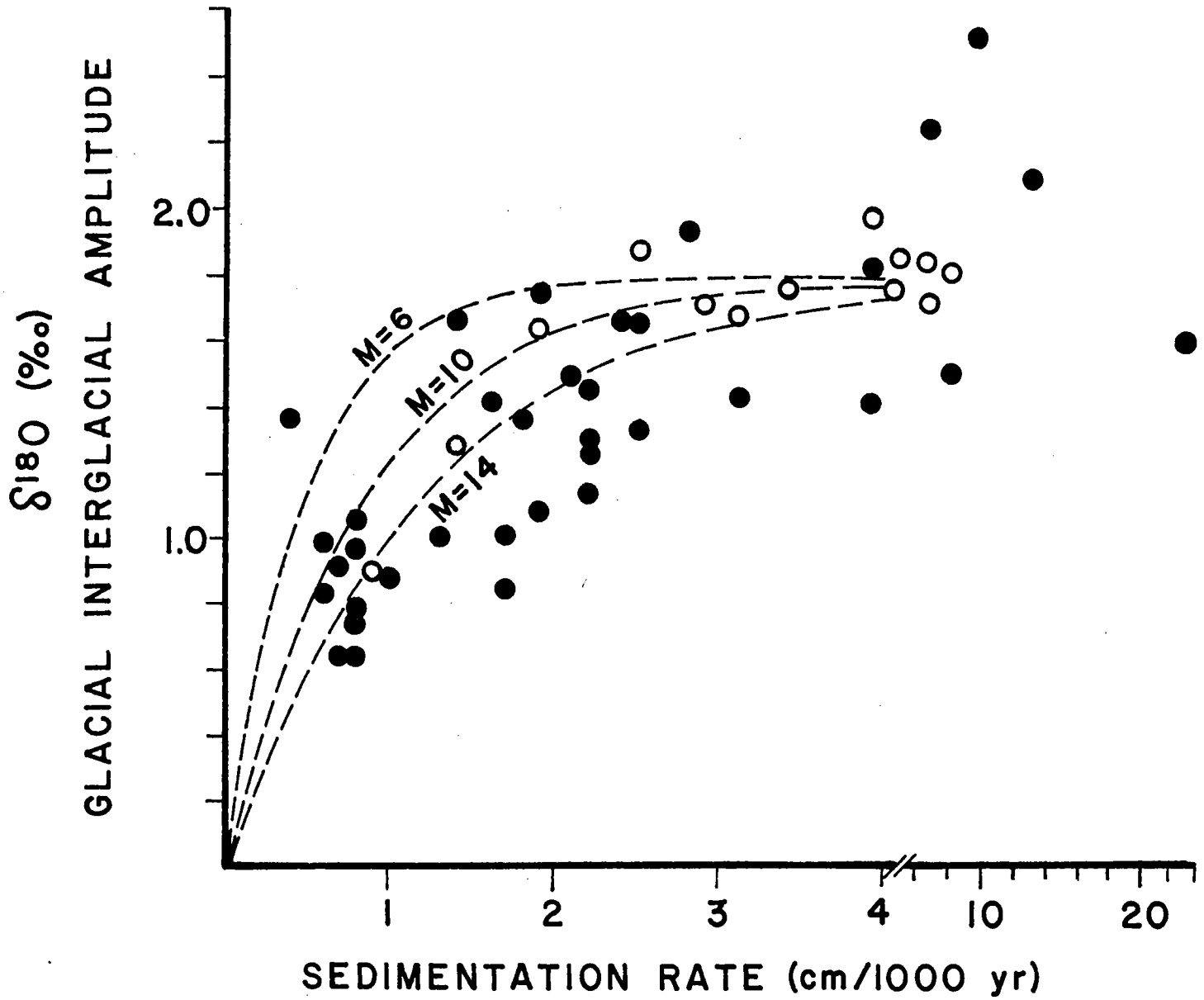

GLACIAL INTERGLACIAL AMPLITUDE
δ18O (‰)
2.0
1.0
M=6
M=10
M=14
1
2
3
4
10
20
SEDIMENTATION RATE (cm/1000 yr)

Figure 30.

Figure 31. Schematic presentation of modifications in the $\delta^{18}O$ glacial-interglacial amplitude (GIA) due to $CaCO_3$ dissolution cycles in the Atlantic and the Pacific Oceans. The original GIA is that which would have been displayed in a given core if dissolution/preservation cycles would have not existed.

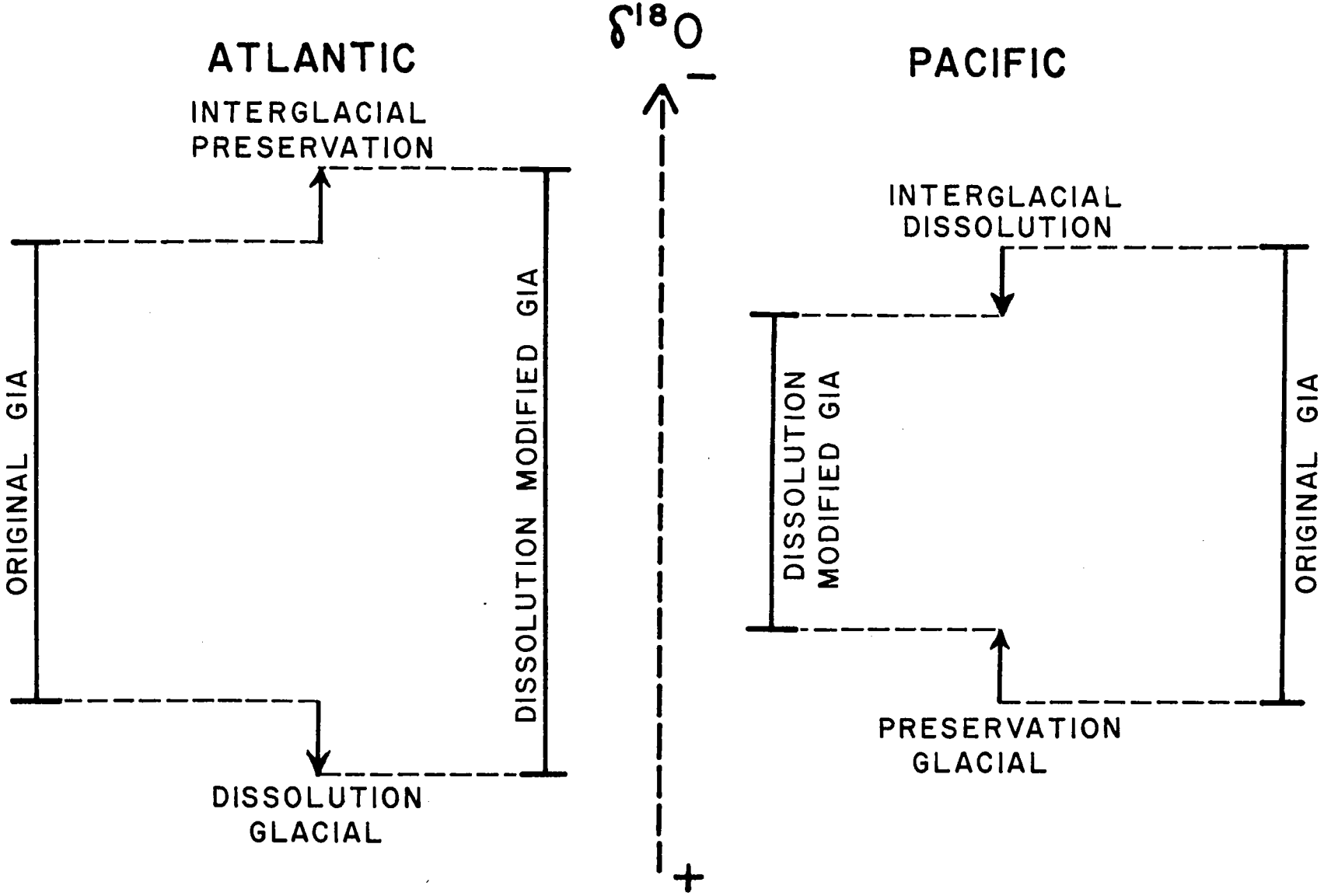

ATLANTIC
INTERGLACIAL
PRESERVATION
ORIGINAL GIA
DISSOLUTION MODIFIED GIA
DISSOLUTION
GLACIAL
δ18O
−
+
PACIFIC
INTERGLACIAL
DISSOLUTION
DISSOLUTION
MODIFIED GIA
ORIGINAL GIA
PRESERVATION
GLACIAL

Figure 31.

SUMMARY

1) Shallow dwelling planktonic foraminifera (*G. ruber, G. conglobatus, G. siphonifera, O. universa, G. sacculifer,* and *H. pelagica*) that were collected by plankton tows exhibit light oxygen isotopic compositions, sometimes in slight deviation from isotopic equilibrium. Carbon isotope composition is out of equilibrium by roughly 1.5 to 6 °/₀₀. It is suggested that this deviation from equilibrium is probably caused by incorporation of light metabolic CO_2 into the skeleton which is enhanced by the existence of symbiotic algae.

2) Radioactive tracer experiments on hermatypic corals and associated benthonic foraminifera shows that photosynthesis of symbiotic algae enhances the incorporation of metabolic CO_2 in the skeleton and thus the deviation from equilibrium for carbon and oxygen isotopes increases. I suggest that symbiotic algae in shallow dwelling planktonic foraminifera are responsible at least in part for the deviation from isotopic equilibrium.

3) Shallow dwelling planktonic foraminifera continue to calcify when they sink down through the water column. This additional skeleton deposition is roughly 50% of the total skeleton weight and occurs below the photic zone from 100 to 400 m. At these depths the symbiotic algae cannot photosynthesize and the temperatures are lower. Therefore, the additional skeleton is isotopically heavier than the original

skeleton deposited in the photic zone. The combined isotopic composition seems therefore to be in equilibrium for calcite deposited in the upper 25 to 100 m of the water column as shown by foraminifera collected in sediment traps at 400 m, 1000 m and 4000 m.

Deep dwelling planktonic foraminifera (Globorotaliids) seem to continue to calcify at deeper depths, certainly down to 900 m. Their $\delta^{18}O$ and $\delta^{13}C$ shows positive correlation which again suggest a dilution by metabolic CO_2 that reduces with depth.

4) Dissolution of planktonic foraminifera on the ocean floor removes first the light weight, thin walled, non-calcified individuals from a species population. Because these individuals are isotopically lighter, the remainder population becomes isotopically heavier as well as in average weight per individual. In-situ dissolution experiment shows that the shift in the isotopic composition does not proceed beyond the stage when the fragile individuals are removed. Further dissolution decreases the average weight/individuals but $\delta^{18}O$ and $\delta^{13}C$ remain the same.

5) The scheme developed above can be applied to understand the effect of $CaCO_3$ dissolution cycles on the isotopic record of oxygen isotopes in the upper Pleistocene. It is suggested that the Atlantic and the Pacific are out of phase with respect to $CaCO_3$ dissolution cycles. In the Pacific dissolution is more intense during interglacials and therefore

the isotopic composition of planktonic foraminifera is somewhat heavier than expected. In the glacial preservation of foraminifera shifts the relatively heavy $\delta^{18}O$ somewhat towards lighter composition. The net effect is a reduction in the glacial-interglacial amplitude of $\delta^{18}O$. In the Atlantic dissolution is more intensive during the glacials. Therefore, relatively heavy $\delta^{18}O$ at that time becomes even heavier. In the interglacial the light $\delta^{18}O$ becomes lighter due to preservation. The net effect of dissolution cycles in the Atlantic is to increase the glacial-interglacial amplitude. Comparison between the amplitudes in the Atlantic and the Pacific for existing cores indeed shows that the Atlantic glacial-interglacial amplitude is higher than the Pacific one by roughly 0.6-0.7 °/₀₀. This difference matches the expected effect of dissolution on the isotopic record. Therefore, previous suggestions that the difference in isotopic amplitude between the two oceans is caused by temperature fluctuations in the Atlantic become unlikely.

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