PLEISTOCENE TEMPERATURES[1]

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ABSTRACT

Oxygen isotopic analyses of pelagic Foraminifera from Atlantic, Caribbean, and Pacific deep-sea cores indicate that the temperature of superficial waters in the equatorial Atlantic and Caribbean underwent periodic oscillations during the Pleistocene with an amplitude of about 6° C. The temperature record of the Pacific cores was much affected by local oceanographic conditions.

Seven complete temperature cycles are shown by a Caribbean core. By extrapolating rates of sedimentation based on radiocarbon data, an age of about 280,000 years is obtained for the earliest temperature minimum. Correlation with continental events suggests that the earliest temperature minimum corresponds to the first major glaciation.

The chronology of Pacific cores proposed by Arrhenius (1952) must be modified if correspondence with the chronology of Atlantic and Caribbean cores is desired.

In one Pacific core which extends to the Pliocene, the 610-cm. level below top is believed to represent the Plio-Pleistocene boundary. About fifteen complete temperature cycles occur above this level, and the length of Pleistocene time is estimated at about 600,000 years. The so-called pre-Günzian stages appear to span a time interval about as long as the Günz and post-Günzian stages. A glacial lowering of sea-level of about 100 m. is indicated.

Closely spaced samples from short pilot cores furnish a detailed temperature record for postglacial times. A continuous temperature increase from about 16,500 to about 6,000 years ago is indicated, followed by a small temperature decrease. The temperature maximum at about 6,000 years ago is correlated with the "Climatic Optimum."

Isotopic analyses of calcareous benthonic Foraminifera show that the temperature of bottom water in the equatorial Pacific during glacial ages was similar to the present, but in the eastern equatorial Atlantic it was about 2.1° C. lower. This difference resulted from the large amount of marine ice present in the North Atlantic. Interglacial bottom temperature in the equatorial Pacific was not more than about 0.8° C. higher than glacial temperatures; interglacial data for the equatorial Atlantic are inconclusive with respect to temperature but indicate an influx of ice meltwater along the bottom larger than at present.

Correspondence in time between temperature variations in the low latitudes, as shown by the cores, and glacial events in the high northern latitudes indicates close correspondence between glacial or interglacial phases and wet or dry phases, respectively.

Good correlation exists between times of temperature minima as indicated by extrapolated rates of sedimentation and times of insolation minima in high northern latitudes. Control of world climate during the Pleistocene by insolation in the high northern latitudes is indicated. A retardation of about 5,000 years occurred between temperature and insolation cycles.

Complete revision of current correlations between the insolation curve and continental events is necessary.

The glacial epoch and its ages may be explained by a theory combining topographical and insolation effects.

Conditions may be suitable for the beginning of a new ice age in about 10,000 years.

INTRODUCTION

The idea of using as a thermometer the variations with temperature of the fractionation factors in isotopic exchange equilibria was first formulated by Urey (1947), in particular relation to the oxygen isotopes in the system CO_2-H_2O-$CaCO_3$.

Three important problems required solution before even preliminary testing of the method could be attempted:

1. The precision attainable at that time in the mass spectrometric measurements of oxygen isotope ratios had to be increased about ten times: an improved six-inch 60° deflection mass spectrometer, based on Nier's design (Nier, 1940, 1947), was constructed by McKinney, McCrea, Epstein, Allen, and Urey (1950); this instrument has now made some 12,000 analyses in Urey's laboratory without subsequent modification.

2. A method of extracting CO_2 from

CaCO₃ with constantly reproducible isotopic fractionation had to be devised: this problem was solved by McCrea (1950) in the course of his work which established experimentally the relationship between the temperature of precipitation of inorganic $CaCO_3$ and the isotopic composition of the precipitate and of the water.

3. An empirical temperature scale based on isotopic analyses of $CaCO_3^-$ depositing marine invertebrates grown in natural environments was required, in order to eliminate possible "vital effects" nonoperative in McCrea's experiments on inorganic precipitation: this empirical relationship and the modifications in analytical technique necessary for dealing with organically precipitated $CaCO_3$ were established by Epstein, Buchsbaum, Lowenstam, and Urey (1951, 1953); in particular, Epstein solved the very difficult problem of destroying organic material without altering the isotopic composition of the calcium carbonate and reduced the sample preparation to a series of routine operations which could be performed by laboratory technicians—thus making it possible to run the large number of analyses on which the present work is based.

PHYSICOCHEMICAL BASIS OF THE ISOTOPIC TEMPERATURE SCALE

The following discussion, based on a number of papers by various authors, is presented here for the convenience of readers who may not be familiar with the principles governing isotopic processes.

Molecular energy may be divided into (a) translational energy, (b) vibrational energy, (c) rotational energy, and (d) electronic energy. In chemical reactions, the various isotopes of the different elements present in a system tend to concentrate to different extents in different compounds, so as to produce the maximum decrease in the free energy of the system. At ordinary temperatures and above, and for all elements except hydrogen, only the vibrational energy of the molecules affects isotopic distributions.
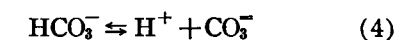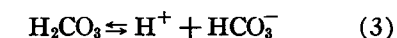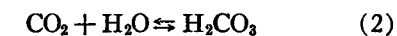
The hydration reaction

$$CO_2 + H_2O \leftrightarrows H_2CO_3 \qquad (1)$$
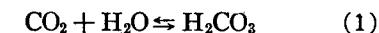
may be used as an example to illustrate the mechanism of exchange of isotopes. One oxygen atom in 500 is an O^{18}, one in 2,700 is an O^{17}. In the following, O^{17} will be disregarded because of its scarcity in nature. One couple of CO_2-H_2O molecules in 166 couples of the system (1) will contain an O^{18} atom. This may be either in the CO_2 or in the H_2O molecule. Both these molecules combine to form an H_2CO_3 molecule, and when the reverse reaction takes place, the O^{18} will have the choice of entering the CO_2 molecule or the H_2O molecule. The probability of entering either one of these two molecules is not exactly in the ratio 2:1 but is somewhat greater than 2 for the CO_2 molecule. This is because the total vibrational energy of the couple $CO^{18}O^{16}$-H_2O^{16} is somewhat smaller than the total vibrational energy of the couple CO_2^{16}-H_2O^{18}.

The O^{18}/O^{16} fractionation factor (a) between CO_2 and H_2O has been calculated and found equal to 1.045 at 0° C. This means that, at 0° C., the ratio O^{18}/O^{16} in the CO_2 molecules is 45 per mil greater than the same ratio in the water molecules.

Reaction (1) provides a mechanism for the attainment of isotopic equilibrium in the CO_2-H_2O system. Isotopic equilibrium may also be reached in the carbonate-water system, as shown by the following general equations:

$$CO_2 + H_2O \leftrightarrows H_2CO_3 \qquad (2)$$

$$H_2CO_3 \leftrightarrows H^+ + HCO_3^- \qquad (3)$$

$$HCO_3^- \leftrightarrows H^+ + CO_3^- \qquad (4)$$

$$CO_3^- + Ca^{++} \leftrightarrows CaC_3O \qquad (5)$$

The decrease of the free energy of a system, when exchange reactions take place and a given isotope is preferentially concentrated in a given compound, becomes less important with respect to the total energy of the system with increasing temperature, because of the increase of the total

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energy. Thus, the preferential concentration of a given isotope in a given compound will matter less and less with increasing temperature, and the fractionation factor α will decrease and tend toward unity.

Table 1 shows the O^{18}/O^{16} fractionation factors between various compounds and water at $0°$ C. and $25°$ C. The fractionation factor between calcium carbonate and water is not known exactly, but it may be seen from table 1 that at $0°$ C. it is 4 per mil greater than at $25°$ C. This corresponds to a variation of 0.2 per mil per degree centigrade and makes it possible to determine the temperature at which calcium carbonate was deposited by determining its α value.

TABLE 1

O^{18}/O^{16} FRACTIONATION FACTORS BE-
TWEEN VARIOUS COMPOUNDS
AND LIQUID WATER

Compound	$T = 0°$ C.	$T = 25°$ C.
CO_2	1.045	1.039
$CaCO_3$	1.025	1.021
CO_3^-	1.018	1.014
O_2	1.007	1.006
H_2O (vapor)	0.990	0.992

In practice, rather than the absolute value of the fractionation factor α of a given sample with respect to water, the per mil difference δ is measured between the O^{18}/O^{16} ratio in the sample and in a fixed standard. This difference is thus

$$\delta = 1,000 \frac{O^{18}/O^{16} \text{ (sample)} - O^{18}/O^{16} \text{ (standard)}}{O^{18}/O^{16} \text{ (standard)}}. \quad (6)$$

The difference δ is measured by alternately passing the sample and the standard through the mass spectrometer (McKinney et al., 1950). The calcium carbonate samples to be analyzed for oxygen isotopic ratios are treated with 100 per cent H_3PO_4 acid. In the reaction that takes place, the CO_2 gas produced contains only two out of the three oxygen atoms originally present in the $CaCO_3$ molecule. This introduces a fractionation which is constant and independent of the original isotopic constitution of the sample (McCrea, 1950). The presence of such fractionation is shown by the fact that CO_2 obtained from $CaCO_3$ precipitated from solution at $16.5°$ C. has the same oxygen isotopic composition as CO_2 equilibrated with the water of the solution at $25°$ C., whereas, if it were representative of the original oxygen isotopic composi-

tion of the $CaCO_3$, it should show an O^{18}/O^{16} ratio about 17 per mil lower.

The standard used in Urey's laboratory at the University of Chicago is CO_2 gas extracted with 100 per cent H_3PO_4 at $25.2°$ C. from a pulverized specimen of *Belemnitella americana* from the Upper Cretaceous Peedee formation of South Carolina.

The mass spectrometric analysis is performed by balancing the currents developed by the two ion beams in the dual ion collector when the standard is run, and then by measuring the current excess or deficiency from this null point when the sample is run. The difference is proportional to the per mil difference in the O^{18}/O^{16} ratios.

The standard deviation of 102 different analyses of the same calcium carbonate, a pulverized specimen of *Glycimeris violacea* from a modern beach a few kilometers south of Ceuta, Spanish Morocco, was calculated in order to establish an error for the isotopic analyses. The 102 analyses were performed over a period of two years, between July, 1953, and July, 1955. A standard deviation of 0.10 per mil was found, corresponding to $0.4°$ C. Thus, if a confidence limit of $\pm 2\sigma$ is taken, the error of the temperature data to be presented later will be $\pm 0.8°$ C.

The value δ (6) for calcium carbonate precipitated in equilibrium with water is inversely proportional to the temperature T at which the precipitation takes place. The exact relation between δ and T was established empirically by determining the δ values for calcium carbonate deposited by marine organisms at various temperatures ranging from $7.4°$ to $29°$ C. (Epstein et al., 1951, 1953).

EFFECT OF THE ISOTOPIC COMPOSITION OF WATER

The δ value depends not only on temperature T but also on the isotopic composition A of the original water. This is clearly shown by the empirical equation (7), which establishes the relationship between δ, A, and T:

$$T = 16.5 - 4.3 (\delta - A) + 0.14 (\delta - A)^2 \quad (7)$$

where

$$A = \frac{O^{18}/O^{16} (W') - O^{18}/O^{16} (W)}{O^{18}/O^{16} (W)}$$

is the correction to be applied if the oxygen isotopic composition of the water (W') from which a given sample was deposited is different from that of average marine water (W) (Epstein et al., 1951, 1953).

The oxygen isotopic composition of sea water varies considerably from place to place. This variation is caused by the fact that the vapor pressure of H_2O^{16} is higher than that of H_2O^{18}. Therefore, more H_2O^{16} evaporates from the two tropical and subtropical evaporation belts of the oceans, and surface-ocean waters in these areas are somewhat (about 1 per mil) richer in O^{18} than the average. The water vapor raised from the evaporation belts will generally travel poleward, discharging on the way a certain amount of water as rain or snow. Because isotopic equilibrium seems to obtain in atmospheric condensation processes (Craig and Boato, 1955), the residual water vapor will become progressively "lighter," that is, progressively depleted in O^{18}. The final fraction which precipitates over such areas as the Canadian Rockies, Greenland, the Scandinavian highlands, and Antarctica, is about 15–30 per mil lighter than average ocean water, and when it finally returns to the ocean (by melting of ice, etc.) it dilutes the ocean water in the neighboring areas. This ocean water, therefore, possesses an O^{18}/O^{16} ratio lower than average. Epstein and Mayeda (1953) have shown that marine waters in high latitudes, as well as marine water in the lower latitudes in areas affected by ocean currents from higher latitudes, have an O^{18}/O^{16} ratio about 0.5 to 1 per mil lower ($\delta = -0.5$ to -1.0) than average marine water.

Also, bottom water in the deep sea has an isotopic ratio about 0.2 to 0.4 per mil lower than average marine water (Epstein and Mayeda, 1953), because of circulation from the polar regions.

Sea water in marginal marine environments near melting ice or river mouths or in tropical lagoons, etc., may exhibit O^{18}/O^{16} ratios which deviate noticeably from average marine water.

Besides these geographical variations, the oxygen isotopic composition of the ocean as a whole has probably varied during geologic time. Data published by Baertschi (1950), Silverman (1951), and Epstein and Mayeda (1953) show that the O^{18}/O^{16} ratio in average ocean water is 6.5 per mil lower than in average basic igneous rocks. Assuming that equilibration at high temperature took place originally between ocean water and igneous rocks, the present difference may have resulted from organic and inorganic precipitation of carbonates and silica in the sea, a process which preferentially removes O^{18}. This process is too slow to affect the isotopic composition of the material considered in this study.

General variations of the oxygen isotopic composition of the ocean as a whole were certainly produced during the Pleistocene and previous glacial epochs by the formation of large quantities of ice on the surface of the earth. Since H_2O^{16} has a higher vapor pressure than H_2O^{18}, the former molecule evaporates preferentially from the oceans, and ice is considerably depleted in O^{18} with respect to average ocean water.

Epstein and Mayeda (1953) and Friedman (1953) have shown that isotopic equilibrium exists between water vapor and ocean water (see discussion in Craig and Boato, 1955). The oxygen isotopic composition of the vapor, for a single-stage distillation process, is -7 per mil.

Water condensing from such vapor will have, initially, an isotopic composition similar to that of the original ocean water; but if the process of condensation continues, it will become progressively depleted in O^{18} because the process itself will make the residual vapor progressively poorer in this isotope. Therefore, from a mass of water vapor originating from the ocean in the tropical areas and moving poleward, precipitation will fall with a progressively lower O^{18}/O^{16} ratio. This process is complicated by the fact that, if the vapor mass passes over a continental

TABLE 2

OXYGEN ISOTOPIC COMPOSITION (δ PER MIL) OF FRESH WATERS FROM VARIOUS PARTS OF THE WORLD*

Locality	Rain	Snow	Rivers, Lakes, and Glaciers
Valparaiso, Chile	0	....	....
Bombay	−2	....	....
Malaya	−4.5	....	....
Chicago	−7	−8 to −17	−6
Copenhagen	−9	....	....
Norway	.....	....	−12
NW. Canada	.....	....	−14 to −20
Greenland	.....	−27	−26 to −31
Maudheim, Antarctica	.....	.....	−19

* Data from Epstein and Mayeda (1953), Dansgaard (1954), and Craig (personal communication).

area, it will become more and more mixed with vapor produced by evaporation, from the ground, of previous and current precipitations, and, in the end, very little or nothing may remain of the original oceanic vapor. Addition of continental water vapor to the original mass will tend to decrease further its O^{18}/O^{16} ratio.

Table 2 shows the oxygen isotopic composition of rains, snow, and water from rivers, lakes, and glaciers in various parts of the world. The isotopic composition of rain and snow varies considerably in the same locality depending upon the history of the vapor mass from which the precipitation originated. The rain figures

of table 2 are averages for the given regions. The data indicate a general decrease of O^{18} with increasing latitude.

The Greenland and Antarctic ice caps are particularly noteworthy because they contain about 99 per cent of the total ice now present on the earth. The very light values for Greenland were obtained from samples of ice from the west coast at 69° N. and snow from the northern part about 400 km. east of Thule. Precipitation in these areas must have originated somewhere in the North Atlantic and traveled a long distance, a portion of the trip being over the ice cap. Hence, as great a decrease of O^{18} as is possible anywhere in the Northern Hemisphere should have taken place. Samples from the southeastern and southern parts of Greenland might be expected to contain more O^{18}. The value for Antarctica was obtained from a sample of firn probably collected not far from the coast, and the percentage of O^{18} is higher than for Greenland.

Although data available at present are few, an average isotopic composition for modern ice of −25 per mil may not be far from true.

The total volume of the hydrosphere, as water at STP, is estimated at about 1,377 × 10⁶ km³, divided as shown in table 3. The volume of ice is based on estimates of thickness of the Greenland and Antarctic ice caps.

Results of the recent French Polar Expeditions to the center of Greenland indicate an ice thickness between 1,900 and 2,200 m. (Cailleux, 1952). Thicknesses of 2,650–2,950 m. were found in the central part of northern Greenland by Bruce and Bull (1955). In consideration of the thinning of ice near the coasts, an average thickness of 2,000 m. may be assumed. The figures available for Antarctica (Robin, 1953) and the fact

that the morphology of the ice surface seems to be controlled to some extent by bedrock topography indicate that the average thickness may not be much greater than about 1,200 m. A value of 18.8 × 10⁶ km³ is obtained by applying these estimates to tables 1 and 21 in Flint (1947). Higher values have been proposed by Cailleux (1954), but his estimate of the thickness of Antarctic ice is possibly too high.

The weighted average oxygen isotopic composition of the whole hydrosphere, calculated from the data of table 3, is −0.3 per mil. Therefore, if all present ice melted, this would be the composition of the ocean, and a correction of −0.3 should be added to A in equation (7) if fossils from nonglacial epochs are analyzed. This correction reduces the isotopic temperature by 1.3° C.

Estimates of the amount of ice present at times of maximum glaciation vary considerably because, although the area covered by Pleistocene ice is rather well known, the thickness is a matter of conjecture. The average thickness probably was closer to the value mentioned for Greenland than to that for Antarctica. The present remarkable thickness of Greenland ice probably is related to the nearness of comparatively warm ocean water, a situation quite different from that prevailing near Antarctica. Similar relations probably existed with respect to the Scandinavian ice sheet and perhaps, to a somewhat less extent, to the Laurentide ice sheet. Therefore, an average thickness of 2,000 m. for the former and of 1,750 m. for the latter may be assumed. If Flint's (1947, table 25) estimate is corrected on the basis of these values and a value of 1,200 m. for the average thickness of Antarctic ice, a total of 59 × 10⁶ km³ is obtained for the maximum Pleistocene glaciation.

The average isotopic composition of Pleistocene ice probably was different from that of modern ice. The ice of Greenland and Antarctica is located at very high latitudes, and water vapor travels a long distance to reach and feed these areas. In contrast, Pleistocene ice extended to much lower latitudes, and water vapor traveled shorter distances. Therefore, precipitation feeding the Scandinavian and Laurentide ice sheets probably had an isotopic composition not far from that of the average snow now falling on these areas. If this is correct, the average isotopic composition of Pleistocene ice was about −15 per mil.

TABLE 3

OXYGEN ISOTOPIC COMPOSITION OF THE HYDROSPHERE (δ PER MIL)

	Volume, as H_2O at STP (×10⁶ km³)	Average Oxygen Isotopic Composition
Ocean	1,360.0	0
Ice (av. $\rho = 0.88$)	16.5	−25
Fresh water	0.5	−7
Water vapor	0.01	−7 to −16

The formation of 40.2 × 10⁶ km³ of ice in excess of the amount now present on the surface of the earth required the removal of 35.4 × 10⁶ km³ of water from the ocean, or 2.6 per cent of its present volume. This would have lowered sea-level 100 m. but, if isostatic adjustment was completely effective, the lowering might have been as little as 70 m.

Extraction from the ocean of 2.6 per cent of its water with an isotopic composition of −15 per mil would raise the O^{18}/O^{16} ratio in the remaining ocean water by 0.4 per mil. Therefore, a correction of 0.4 should be added to A in equation (7) if temperatures are obtained from fossils which lived during the glacial ages of the Pleistocene. Such correction increases the isotopic temperatures by 1.7° C.

The figure of 0.4 per mil depends upon an estimate of the amount of ice formed during the ice ages, and therefore it may not be correct. There are good reasons, however, to believe that it cannot be far from the true value. Apart from the geological evidence which is against a eustatic glacial lowering of sea-level by much more than 100 m. (Blanc, 1937; Cailleux, 1954, table 1), certain isotopic temperatures to be presented later indicate that the isotopic composition of sea water could not have undergone a change greater than that mentioned, or true glacial temperatures would be higher than interglacial ones.

In summary, isotopic temperatures depend not only upon the true temperature but also upon the isotopic composition of the original water. If this were different from the present composition of average marine water, the isotopic temperature would be different from the true temperature by a corresponding amount.

Certain corrections can be applied to the isotopic temperatures by introducing a suitable value for A in equation (7), so as to bring isotopic temperatures close to true temperatures. These corrections are: (1) a variable correction depending upon the geographic location of the sample and local conditions; (2) a correction of -0.3 per mil for nonglacial times; (3) a correction of $+0.4$ per mil for the glacial ages.

Correction 1 is the most difficult to evaluate. As mentioned before, marine waters from marginal environments may have an isotopic composition several per mil different from average marine water. On the other hand, marine waters in open oceanic areas do not generally show a variation of more than 1 per mil from the average, this variation being positive in the evaporation belts and negative in the high latitudes. Generally, shelf fossils, especially from the littoral en-

vironment, are less likely to give reliable temperatures than fossils from the open ocean. For the latter, the assumption is reasonable that, in the low and middle latitudes, the present geographic distribution of isotopic composition in the sea is much the same as in the past, and corrections given by present measurements may be applied to fossil determinations.

Absence of ice in high latitudes in nonglacial times should make marine waters in these latitudes heavier than at present, although still somewhat lighter than the average. On the other hand, bottom water was probably just average rather than 0.3 per mil lighter, as at present.

SOURCES OF ERROR IN OXYGEN ISOTOPIC TEMPERATURE DETERMINATIONS

Isotopic temperature analyses may be affected by errors other than those caused by variations in the isotopic composition of the original water. Postdepositional exchange between the fossil calcium carbonate and percolating waters will reduce the original O^{18} content of the carbonate because these waters generally have an isotopic composition similar to that of fresh water. Recrystallization is a sure index that exchange has taken place, and only well-preserved fossils should be used.

Some organisms in the euphotic zone, such as echinoderms and corals, do not seem to deposit calcium carbonate in equilibrium with the water, and, therefore, they are not suitable for isotopic temperature determinations. Many such organisms contain symbiotic *Zooxanthellae* in their tissues, and exchange may occur between O_2 liberated by the symbiotic organisms and the $CaCO_3$ which is being deposited. Problems of this type have been discussed by Urey *et al.* (1951) and by Lowenstam and Epstein (1954).

The carbonate of many marine organ-

isms contains an appreciable amount of magnesium. Values as high as 10–15 per cent have been reported by Chave (1954). While the fractionation factor between $MgCO_3$ and water is certainly somewhat different from that between $CaCO_3$ and water, this difference is thought to be small and probably beyond experimental detectability. The same may be said for strontium. Spectrographic and X-ray analyses of several species of pelagic Foraminifera from a number of localities (Emiliani, in press, *a*), both modern and Pleistocene, have shown that the tests are formed by virtually pure calcite, with no Mg and only about 0.13 per cent Sr.

RELATIONSHIP BETWEEN ISOTOPIC TEMPERATURES AND TEMPERATURE OF SURFACE OCEAN WATER

Oxygen isotopic analysis of marine carbonates gives at best the exact temperature at which the carbonate was deposited. In the ideal case of planktonic organisms living at the surface in areas where no appreciable seasonal temperature variations occur, the isotopic temperature will correspond to the yearly superficial oceanic average. If seasonal variations occur, the isotopic temperatures will be spread over part or all of the yearly range, because different animals may grow in different temperature ranges (Epstein and Lowenstam, 1953). Isotopic temperatures obtained from pelagic Foraminifera are probably close to the yearly mean at low latitudes, where seasonal temperature variation is small. Beyond the tropics, where this variation becomes important, the isotopic temperature is probably close to the summer mean. In fact, *Globigerinoides rubra* from a part of the Gulf of Mexico where temperature ranges between 22° and 29° C. gave a temperature of

27.5° C. (Emiliani, 1954a).

In the case of benthonic organisms from the littoral environment, microenvironmental temperatures may be some degrees different from prevailing temperature, and a picture having regional significance can be obtained only by analyzing a number of different organisms from different places.

Pelagic organisms which do not live at the surface and benthonic organisms from the neritic and bathyal environments introduce complications because of the temperature variation with depth. If isotopic analysis is based upon extant species, the rather arbitrary assumption must be made that the depth habitats have not changed with time. On the other hand, if extinct species are used, interpretation of isotopic data may be difficult and sometimes impossible.

If pelagic Foraminifera are analyzed, several hundred tests generally are required to make up the 5 mg. of calcium carbonate necessary for each temperature determination. Therefore, the temperature values obtained represent the weighted average temperature of shell deposition of the foraminiferal population. A previous study (Emiliani, 1954a) showed that different species of pelagic Foraminifera live at different depths and, therefore, register different temperatures. It was also shown that *Globigerinoides rubra* and *Globigerinoides sacculifera* live closest to the surface (probably within 30–40 m.), and, therefore, they are in the best position to register whatever temperature variations may have occurred. These two species are rare or absent in Pacific cores and in the middle Atlantic core 280A. Consequently it has been necessary to use different species, namely *Globorotalia tumida*, *Pulleniatina obliquiloculata*, and *Globigerina inflata*, which occupy a deep-

er habitat (down to about 140 m.) and register temperature variations smaller than those occurring at the surface.

Because pelagic Foraminifera seem to be adjusted to water densities (Emiliani, 1954a) one might expect that density variations related to temperature would cause the foraminiferal populations to move vertically, smoothing or erasing the temperature record. That this is not so is indicated by the temperature graphs to be discussed later. The reason is probably twofold. First, temperature variations would affect the density of protoplasm in the Foraminifera similarly to that of sea water; second, since pelagic Foraminifera are restricted to the euphotic zone, vertical migration would vacate certain eutrophic niches which would be immediately reoccupied by proper mutants of the migrating populations. Temperature records from the cores indicate that the specific populations maintain the same depth habitats through time, in spite of rather wide temperature variations (Emiliani, 1954a). Food and ultimately the pattern of solar light penetration seem to be the controlling factors.

DEEP-SEA CORES

INTRODUCTION

The deep sea is generally thought of as an environment in which sedimentation is largely continuous and undisturbed. When this is true, the deep-sea sediments furnish a continuous record of conditions in the different environments contributing to the sediments themselves. These environments are essentially (a) the ocean bottom, (b) the entire water column above it and especially the euphotic zone, and (c) neighboring land areas.

The most important constituents of deep-sea sediments of "*Globigerina*-ooze" facies are biogenous carbonates formed in

the euphotic zone, which constitute an abundant record of the conditions at or near the surface, and their variations through time. This, of course, was recognized long ago, and much work has been done based on this fact.

Methods have been developed only recently to sample appreciable lengths of the sedimentary column of the sea bottom. Schott (1935) worked with cores less than a meter long. A few years later Piggot succeeded in raising cores 3 m. in length. The most important step in sampling deep-sea sediments was the invention of the piston corer by Kullenberg (1947), which made possible the recovery of cores more than 20 m. long and eliminated compaction due to coring. The Kullenberg corer was extensively used by the Swedish Deep-Sea Expedition of 1947–1948, which brought back about 300 long cores. With the Kullenberg corer, however, up to 40–50 cm. of sediments may be lost at the top. This incompleteness was compensated by sampling the top 50 cm. of submarine sediments with short pilot cores, usually raised at the same locations. Recently, Ewing has modified the original Kullenberg design so as to prevent the loss of surface material (Ewing *et al.*, 1954; Ericson and Wollin, in press).

PREVIOUS WORK

Studies of submarine cores for past climatic changes are based mainly on (1) variations in carbonate content, and (2) variations in the relative abundances of cold- and warm-water pelagic Foraminifera. As is well known, submarine solution may affect both variables importantly.

Solution is proportional to the influx along the ocean bottom of undersaturated waters from high latitudes. The bottom water which forms in the high latitudes is probably more aggressive during deglacial phases when mixed with large amounts of ice meltwater. Solution at times of temperature maxima, as shown by Caribbean core A172-6 (p. 554), the Caribbean core described by Phleger (1948), and Pacific cores described by Arrhenius (1952), indicate a delay of some thousand years between the time in which formation of aggressive bottom waters in high latitudes was most intensive and the time in which solution in low latitudes was greatest. Reduction by

solution of the carbonate percentages of the warm stages in the low-latitude cores may lead to false conclusions as to climatic changes deduced from such percentages. This matter has been discussed by Wiseman (1953).

Solution may also influence conclusions based on micropaleontology, because different species are affected differently, and fossil faunas different from the original living faunas may be produced.

The stratigraphically longest cores of the *Globigerina*-ooze type which have been described (Phleger, 1948; Arrhenius, 1952; Schott, 1952, 1954; Ericson, 1953; Hough, 1953; Phleger, Parker, and Peirson, 1953) clearly show a number of alternating warmer and colder stages. It is convenient, following the system introduced by Arrhenius (1952), to designate the present, warm stage as no. 1 and the preceding stages with increasing positive integers, which are odd for warm stages and even for cold ones. Ideally, in a sinusoidal change of temperature with time (fig. 1) the boundaries between succeeding stages may be established at the midpoints between temperature maxima and minima. The terms "anathermal" and "catathermal" (abbreviated respectively to An and Ct) may be conveniently used to define the intervals of rising and declining temperatures, which include adjacent halves of neighboring stages.

The beginning of the last anathermal (An 2-1) is a well-marked point in most cores. Its depth below the sedimentary surface depends on the local rate of sedimentation. It generally occurs 20–75 cm. below the surface in eupelagic areas (Schott, 1935; Phleger and Hamilton, 1946; Arrhenius, 1952; Schott, 1952, 1954; Ericson, 1953; Phleger, Parker, and Peirson, 1953), but in areas where terrigenous sedimentation was appreciable it rarely occurs above 50 cm. (Phleger, 1939, 1942, 1947; Bramlette and Bradley, 1940; Cushman and Henbest, 1940; Phleger and Parker, 1951).

Below stage 1, a colder stage has been recognized in most cores. Ionium dating on three North Atlantic cores and a Caribbean core (Piggot and Urry, 1942) indicates that the base of An 2-1 corresponds to an age of about 15,000 years, while the base of the colder stage corresponds to an age of about 60,000–65,000 years. These figures are in agreement with evidence presented later.

In his monumental work on Pacific cores, Arrhenius (1952) described two (nos. 58 and 62) which appear to extend into sediments of Plio-

cene age. The Plio-Pleistocene boundary was tentatively identified at 395 cm. below the top in core 58 and at approximately 1,050 cm. below top in core 62. It is indicated in both by a marked change in the accumulation rate of calcium carbonate. Nine warm and nine cold stages were recognized in the Pleistocene section, and a tentative correlation was suggested between the last four cold stages (which are more pronounced than the preceding ones) and the four recognized glaciations in higher northern latitudes.

No Atlantic core so far described appears to reach the Pliocene boundary. A rather consistent record of much of the Pleistocene, however, is furnished by the cores described by Bramlette and Bradley (1940), Cushman and Henbest (1940), Phleger and Hamilton (1946),
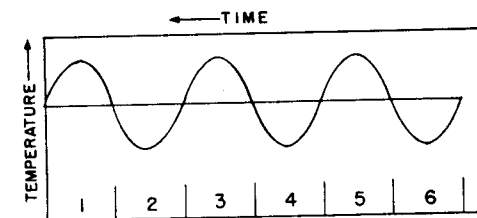

FIG. 1.—Ideal variation of temperature with time, and stage numbers.

Schott (1952, 1954), Ericson (1953), and Phleger, Parker, and Peirson (1953). Several warm and cold stages have been recognized in these cores on micropaleontological evidence, and correlation between widely separated cores seems possible and has been attempted (Bramlette and Bradley, 1940; Ericson, 1953; Phleger, Parker, and Peirson, 1953; Ericson and Wollin, in press). Evidence presented later will help to clarify the conclusions of these authors.

MATERIAL FOR THE PRESENT STUDY

Isotopic analyses have been made of 12 cores. Eight were raised by the Swedish Deep-Sea Expedition of 1947–1948 and samples were furnished by Hans Pettersson of the Oceanografiska Institutet of Göteborg. The other four, identified by prefix "A," were obtained from David B. Ericson, of the Lamont Geological Observatory. Pilot cores are distinguished by the suffixes "A" or "TW." All cores consist largely or entire-

ly of sediment known as *Globigerina* ooze. The location, depth, and length of cores are given in table 4.

SAMPLING AND TREATMENT OF SAMPLES

Half-core samples 2 cm. thick were taken at 10-cm. intervals from each of the Atlantic and Caribbean cores. The Pacific cores were sampled by G. and O. Arrhenius, mostly on the basis of the carbonate percentage curves published by G. Arrhenius (1952). These samples were 3 cm. thick for core 58 and 6–8 cm. thick for cores 60 and 61. The pilot cores were sampled every other centimeter, with half-core samples 1 cm. thick, except core A179-TW4, which was sampled continuously with half-core samples 2 cm. thick.

TABLE 4

LOCATION, DEPTH, AND LENGTH OF THE DEEP-SEA CORES

Core No.	Lat.	Long.	Depth	Length (m.)
A172-6	14°59′N	68°51′W	4,160	9.35
A179-TW4	16°36′N	74°48′W	2,965	0.14
A179-4	16°36′N	74°48′W	2,965	6.90
A180-73	0°10′N	23°00′W	3,749	4.90
234A	5°45′N	21°43′W	3,577	0.44
234	5°45′N	21°43′W	3,577	14.70
235A	3°12′N	28°28′W	4,560	0.69
246	0°48′N	31°28′W	3,210	8.70
280A	34°57′N	44°16′W	4,256	0.43
58	6°44′N	129°28′W	4,440	9.91
60	1°35′N	134°57′W	4,540	10.13
61	0°06′S	135°58′W	4,437	10.29

The samples were dried at room temperature, weighed, and washed in distilled water through a 230-mesh sieve (side length of the square openings, $62\,\mu$). The residue retained by the sieve was dried in an oven at 90° C., weighed, and split into two equal fractions. This preparation of cores A172-6, A179-4, A179-TW4, and A180-73 was done by Ericson at the Lamont Geological Observatory, using a 200-mesh sieve (side length of the square openings, $74\,\mu$). One of the two fractions was again split into three different size fractions ($>500\,\mu$, 500–$250\,\mu$, and $<250\,\mu$).

A sufficient number of shells of the foraminiferal species desired for isotopic analysis were separated under the microscope from the largest or the two largest fractions (and occasionally also from the smallest fraction). Depending upon their size, about 100–400 tests are required to make up the 5 mg. of calcium carbonate necessary for each isotopic analysis.

The foraminiferal samples were crushed, washed in distilled water in an agitator for 15 minutes to loosen whatever foreign material might be inside the tests, dried at 90° C., pulverized in an agate mortar, heated at 475° C. in a stream of helium for an hour and fifteen minutes to remove the organic material, treated with 100 per cent H_3PO_4 at 25° C., and the CO_2 gas thus obtained analyzed by mass spectrometer (cf. Epstein *et al.*, 1953).

The weight percentage of the fraction larger than 62 or $74\,\mu$ was determined for all samples of the Atlantic and Caribbean cores. This was done for the fraction larger than $74\,\mu$ by Ericson at the Lamont Geological Observatory for the Caribbean cores and Atlantic core A180-73. No such determinations were made on the Pacific core samples, for which accurate carbonate percentages are available (Arrhenius, 1952). The fraction larger than 62 or $74\,\mu$ consists almost exclusively (i.e., more than 99 per cent by weight) of foraminiferal shells or fragments in all samples examined.

ISOTOPIC TEMPERATURE RESULTS

INTRODUCTION

Figures 2–6 and 11–13 show the isotopic temperature results for the longer

TABLE 5

CORRECTIONS FOR ISOTOPIC COMPOSITION OF SEA WATER AT LOCATION OF CORES

Core No.	Water Correction Per Mil
A172-6	+0.92
A179-TW4, A179-4	+0.92
A180-73	+0.48
234A, 234 (cm. 1-700)	+0.48
235A	+0.48
246	+0.48
280A	+0.95
58	+0.22
60	+0.22
61	+0.22

cores, and figures 7–10 for the pilot cores. The temperature scale, in degrees centigrade, is on the ordinate, and the depth, in centimeters below top of core, is on the abscissa. The foraminiferal species analyzed are also indicated.

Correction 1 (p. 544) was applied to all temperature results obtained from pelagic Foraminifera, and a correction of
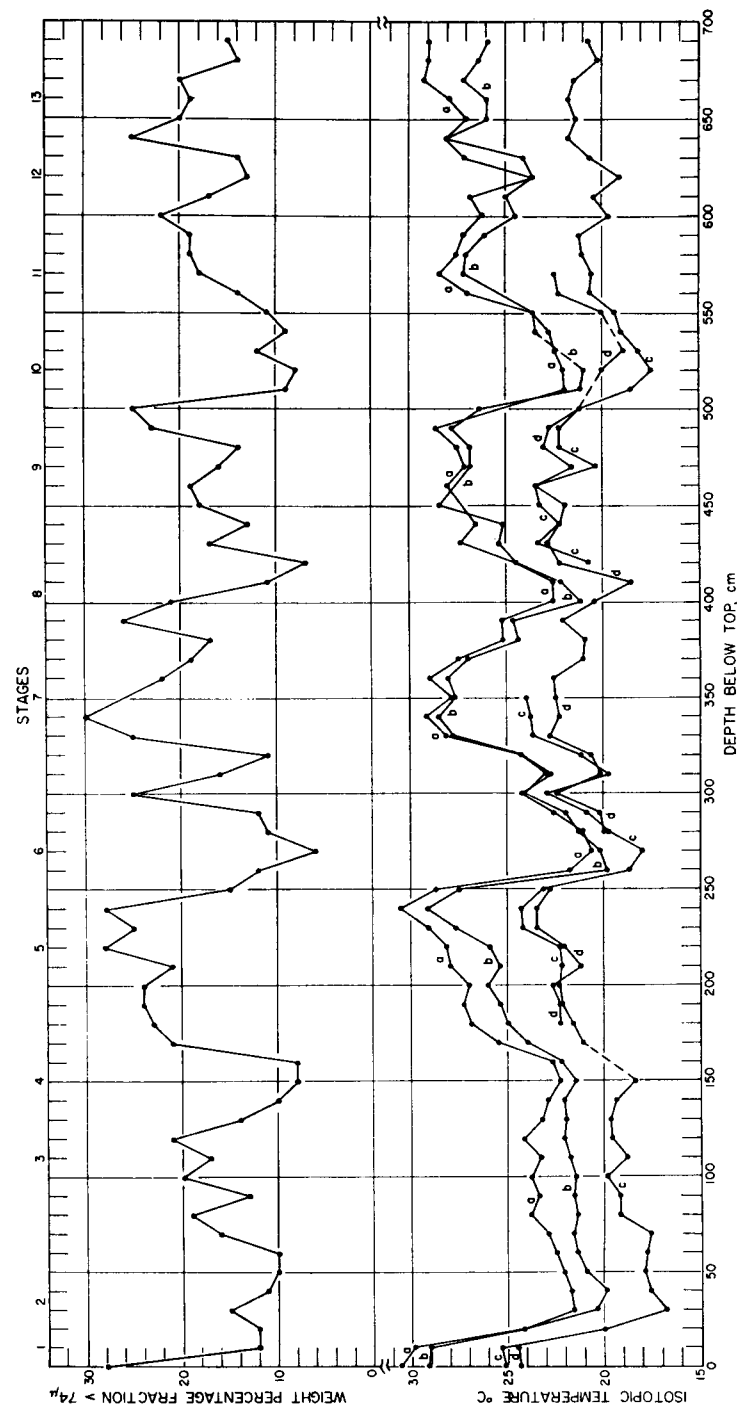

FIG. 2.—Core A179-4: percentages of the fraction larger than $74\,\mu$ and isotopic temperatures obtained from *Globigerinoides rubra* (a), *Globigerinoides sacculifera* (b), *Globigerina dubia* (c), and *Globorotalia menardii* (d).
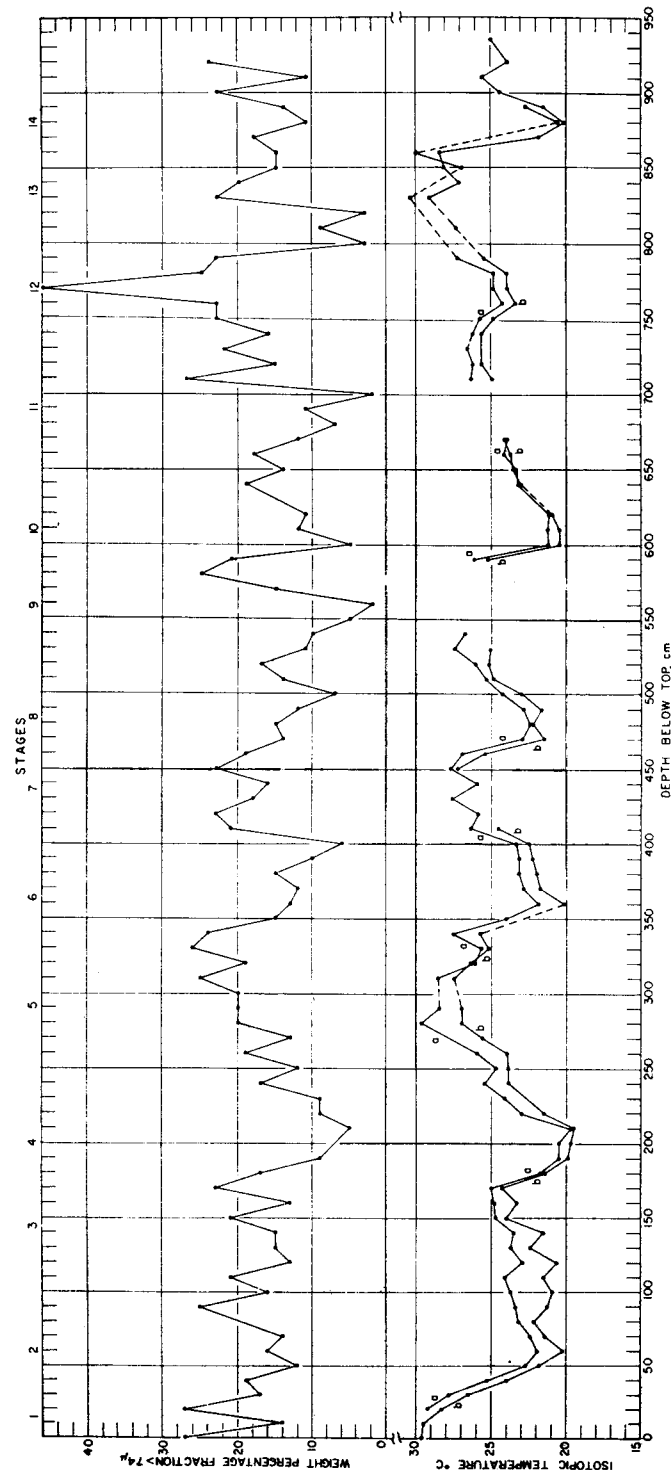
FIG. 3.—Core A172-6: percentages of the fraction larger than 74 μ and isotopic temperatures obtained from *Globigerinoides rubra* (a) and *Globigerinoides sacculifera* (b).

−0.30 per mil (average for present deep water; see Epstein and Mayeda, 1953) was applied to all results obtained from benthonic Foraminifera. The values for correction 1 (table 5) were obtained from data published by Epstein and Mayeda (1953) on samples of ocean water at or near the location of each core. Application of modern values to fossil samples seems reasonable on the basis of the apparent constancy of oceanic circulation through time (Schott, 1935; Arrhenius, 1952; Ericson, 1953). Corrections 2 and 3 were not applied to temperature maxima and minima, as they should have been, because of the practical difficulty of estimating proper reduction of these corrections for temperatures near the maxima and minima. Application of corrections 2 and 3 would lower the maxima by about 1.3° C. and raise the minima by about 1.7° C.

Temperature graphs of Atlantic and Caribbean cores are accompanied by

graphs showing the variation of the weight percentage of the fraction larger than 62 or 74 μ, and the temperature graphs of Pacific cores 60 and 61 are accompanied by graphs showing the percentage of carbonates, based on Arrhenius' (1952) data. When considering the temperature records shown in the graphs, one must keep in mind the relation of isotopic temperatures and the tempera-
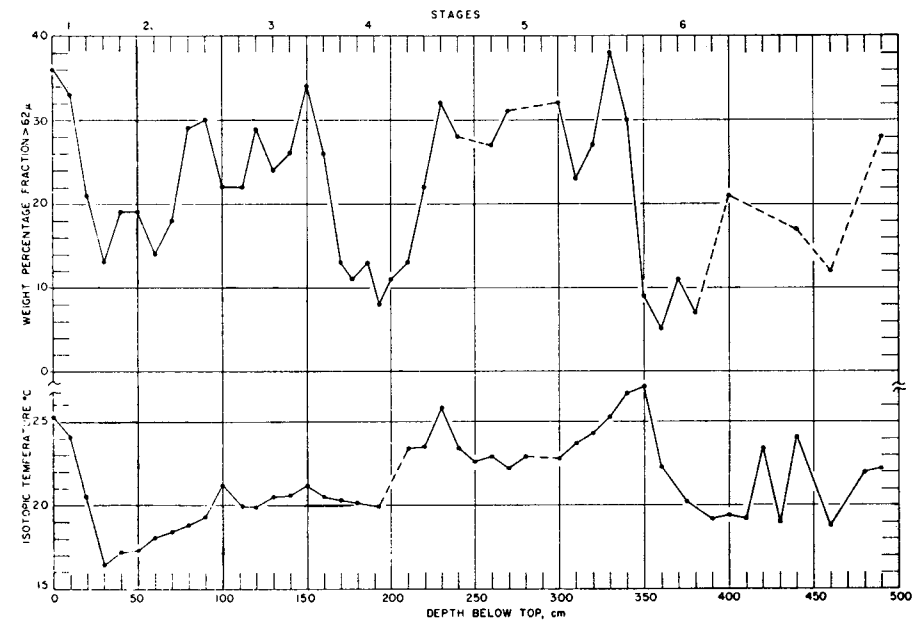

FIG. 4.—Core A180-73: percentages of the fraction larger than 74 μ and isotopic temperatures obtained from *Globigerinoides sacculifera*.

tures of surface ocean water discussed earlier (p. 545). Also, one must remember that mud-eating animals commonly rework the sediments, and that even apparently undisturbed cores may have been affected by such reworking, bottom currents, or other agents. Minor internal inconsistencies in the temperature records of different cores should be interpreted accordingly.

ATLANTIC AND CARIBBEAN CORES

The best evidence of past temperature variations is furnished by the two Caribbean cores

FIG. 5.—Core 234: percentages of the fraction larger than 62 μ and isotopic temperatures obtained from *Globigerinoides sacculifera*. (A sedimentary disturbance may be indicated by the high value of the upper curve at 473.5 cm.; therefore, identification of stages has not been carried beyond stage 6.)
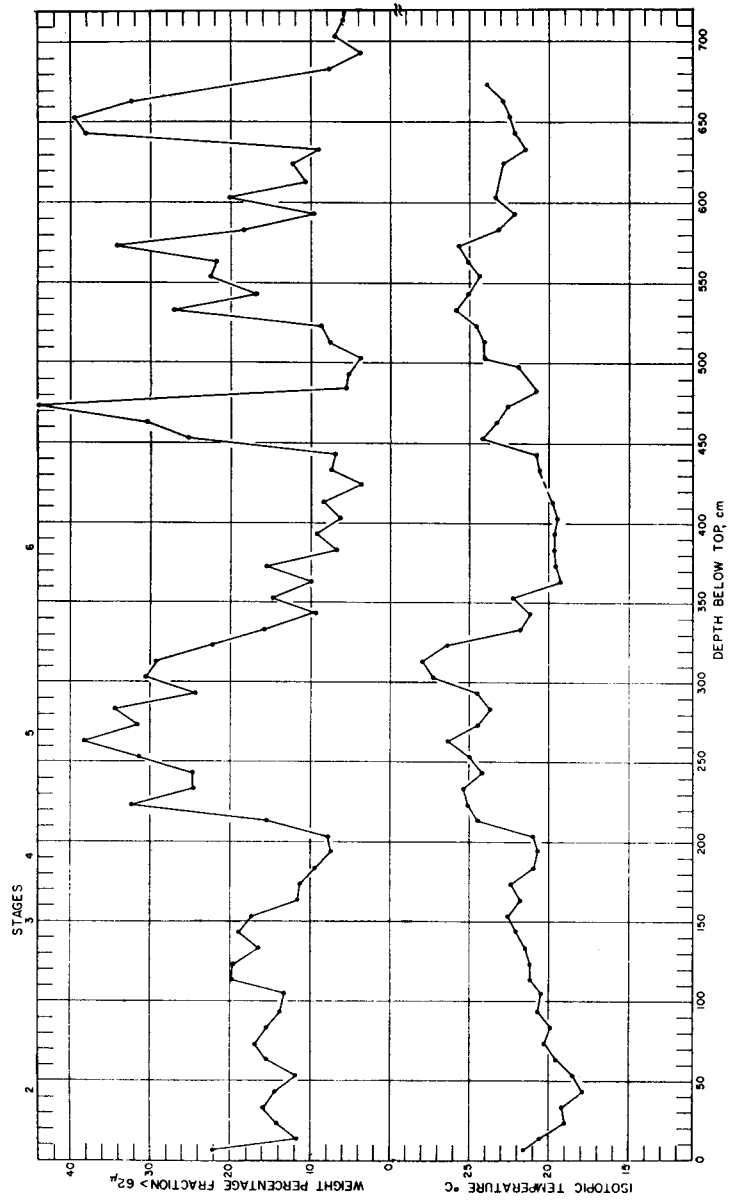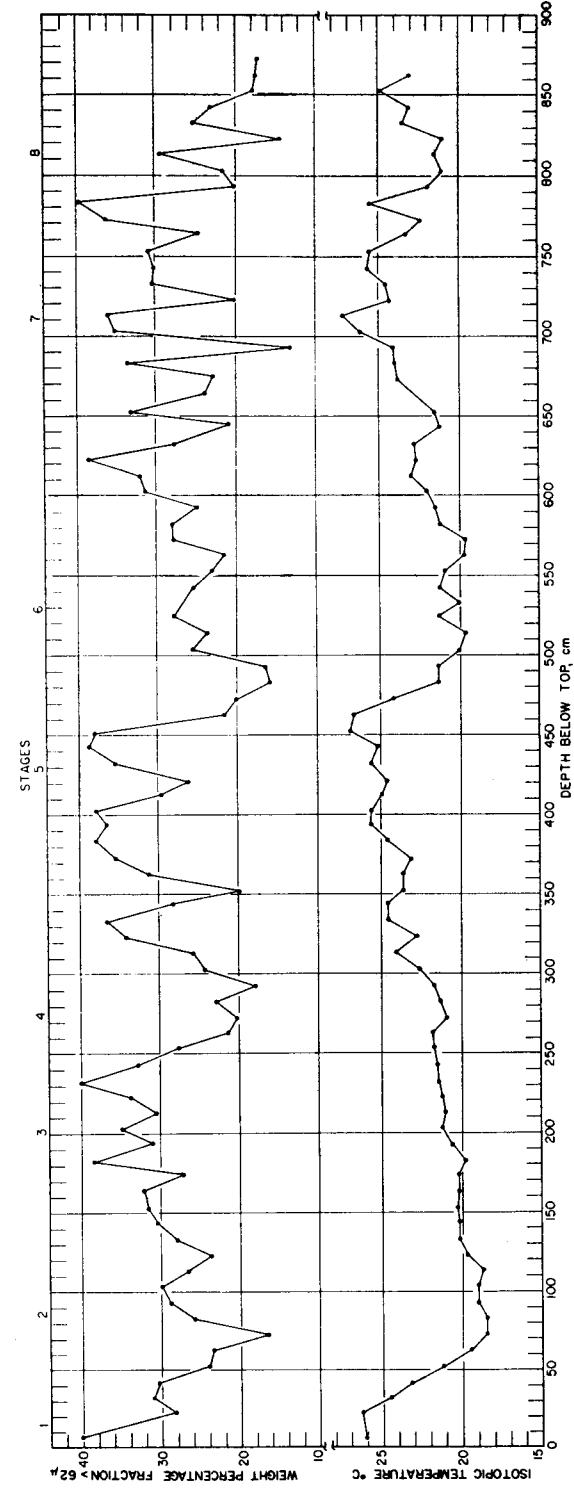
552

FIG. 6.—Core 246: percentages of the fraction larger than 62 μ and isotopic temperatures obtained from *Globigerinoides sacculifera*

553

A179-4 and A172-6 (figs. 2–3). The Atlantic cores (figs. 4–6) are stratigraphically shorter. Only the upper 700 cm. of core 234 are considered here because the lower part is of Miocene age (Phleger, Parker, and Peirson, 1953) and is dealt with elsewhere (Emiliani, in press, b). The two Caribbean cores are the stratigraphically longest cores of almost undisturbed sedimentation, from which pelagic *Globigerinoides sacculifera* and *Globigerinoides rubr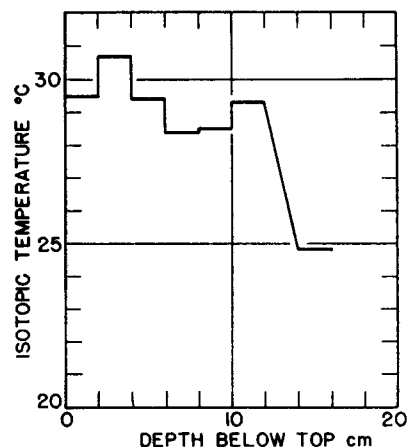a* were available. A small sedimentary hiatus is probable between 250 and 260 cm. in core A179-4, as indicated by comparison with the temperature graph of core A172-6, in which it is represented by the 300–360-cm. section. A certain amount of solution occurred in core A172-6 at 300 cm., between 540 and 580 cm., and between 670 and 710 cm. Samples from these levels correspond to temperature maxima in core A179-4 and do not contain enough Foraminifera for isotopic analysis. Reasons for solution at temperature maxima have been presented in the section on previous work. The lithologic character and micropaleontology of the two Caribbean cores and of the equatorial Atlantic core A180-73 are described by Ericson and Wollin (in press).

Both Caribbean cores show a remarkable periodicity of about equally spaced temperature maxima and minima. The temperature difference is about 8°–9° C., but if corrections 2 and 3 are applied, this difference is reduced to about 6° C. Similar values obtain for the equatorial Atlantic cores. The temperature maxima have

all about the same value and are similar to present temperatures. Also the temperature minima are about equal to each other.

Taking 25° C. as average temperature for the two Caribbean cores, the average thickness of sediment deposited at higher temperatures is 75 cm. in core A179-4 and 85 cm. in core A172-6, and the average thickness of sediment deposited at lower temperatures is 60 cm. in both cores. The smaller thickness of colder stages does not necessarily mean less time, because calcium carbonate precipitation from superficial waters may have been much slower, so as to compensate for possible less solution and more terrigenous sedimentation.
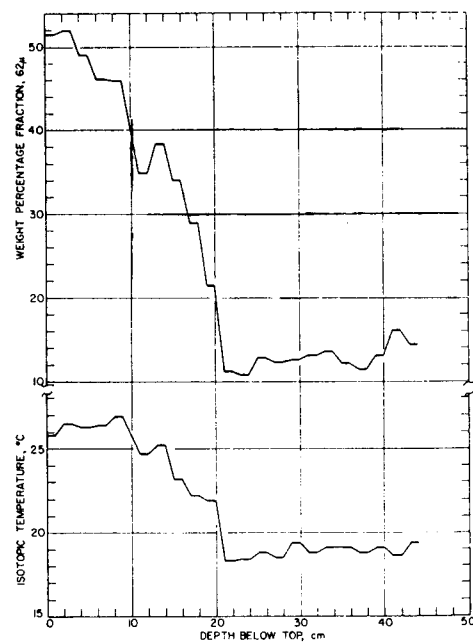
If the assumption is made that each complete temperature cycle was of equal duration, differential compaction does not seem to be important because thicknesses do not decrease downward (table 6). No compaction was recognized by Arrhenius (1952) in his Pacific cores. Probably sediments that can be reached by present coring techniques are partly hydrostatically supported. This lessens the risk of extrapolating to lower parts of cores the rates of sedi-

FIG. 7.—Core A179-TW4: isotopic temperatures obtained from *Globigerinoides sacculifera*.

FIG. 8.—Core 234A: percentages of the fraction larger than 62 μ and isotopic temperatures obtained from *Globigerinoides sacculifera*.

mentation which have been calculated for the upper parts.

Temperatures determined for the zero level of all pilot cores and those raised with the Ewing corer should correspond to modern temperatures. These are in fact close to the superficial temperatures now obtaining in areas where the cores were raised, although for the Caribbean cores they are slightly higher than present August means. Agreement probably would be better if superficial water samples were available for isotopic analysis from the exact location of each core.

Below the modern levels, and going back in time, all cores show a tempera-

ture decrease to the first minimum (stage 2). As previously mentioned, this, as well as all other minima, should be about 1.7° C. higher because of the change in isotopic composition of the sea water. A minor temperature rise (stage 3) and another minimum (stage 4)

FIG. 9.—Core 235A: percentages of the fraction larger than 62 μ and isotopic temperatures obtained from *Globigerinoides sacculifera*.

TABLE 6

STRATIGRAPHIC THICKNESSES OF COMPLETE TEMPERATURE CYCLES (CM.)

Core No.	Increasing Depth→					
A179-4	110	110	140	110	110	...
A172-6	140	160	120	130	150	120

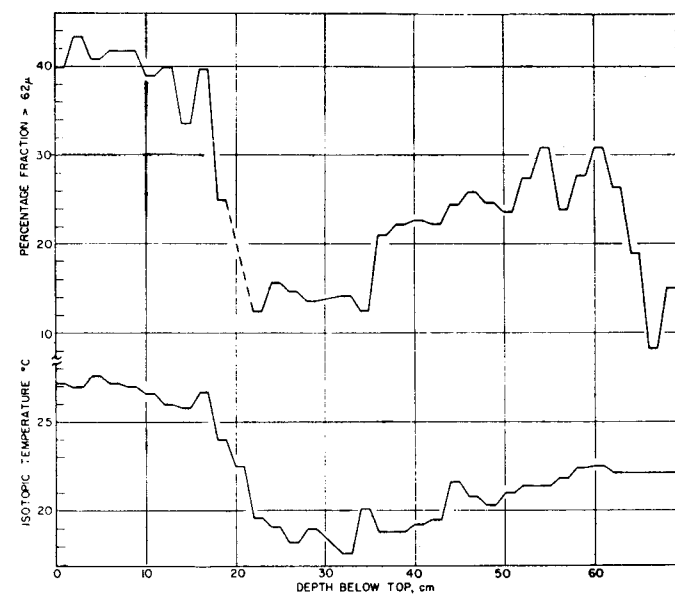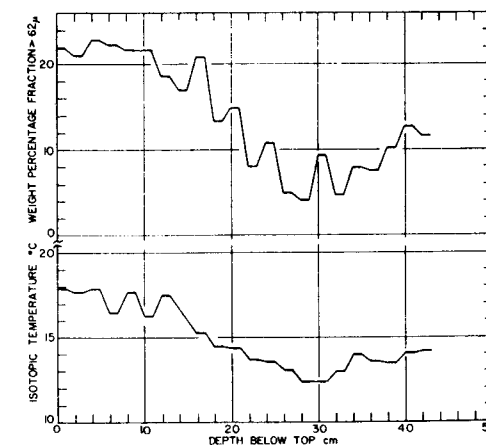FIG. 10.—Core 280A: percentages of the fraction larger than 62 μ and isotopic temperatures obtained from *Globigerina inflata*.

follow, after which the temperature rises again to values similar to modern ones (stage 5).

Stage 3 has temperatures intermediate between the major maxima and minima but has a thickness of sediment similar to that of the major warm stages. It is more evident in some cores (e.g., A172-6) than in others (e.g., 246) but generally has escaped micropaleontological detection because of the low value of its temperature maximum. Thus, the last cold stage as previously recognized (Cushman and Henbest, 1940; Phleger and Hamilton, 1946; Schott, 1952, 1954; Ericson, 1953; Phleger, Parker, and Peirson, 1953) generally corresponds to combined stages 2, 3, and 4 as defined here.

In Pacific cores 59 and 60, described by Arrhenius (1952), best showing the alternance of stages, the average thickness and the average total accumulation of TiO_2 for stage 2 as recognized by him are respectively 163 cm. and 48.5 mg/cm², while the averages are 62.6 cm. and 27.1 mg/cm² for the other stages. A longer duration in time is indicated, therefore, for his stage 2, and correlation with our stages 2, 3, and 4 as identified here is suggested. Only in core 59, however, the curve for carbonate accumulation does indicate the presence of our stage 3 within Arrhenius' stage 2.

Radiocarbon dates were determined by Rubin and Suess (1955a, b) for various levels in cores A179-4, A172-6, and A180-73 (table 7). The first determinations were made on bulk core material. Later ones based on foraminiferal shells only showed some discrepancies with the earlier determinations, but they are be-

lieved to be more reliable. The levels were selected to obtain as exact an age as possible for the beginning of An 2-1, and to obtain information as to rates of sedimentation both above and below this point.

The beginning of An 2-1 occurs between 20 and 30 cm. in cores A179-4 and A180-73, and between 50 and 60 cm. in core A172-6. This point is better defined in the latter core because of the greater sedimentation rate. The 55-cm. level of core A172-6, which is certainly very close to or coincides with the point in question, was found to be 17,500 years old, but the supposedly corresponding 26.5-cm. level of core A179-4 was found to be 11,800 years old. In the latter core this is only 4 cm. above the surely cold 30-cm. level, which therefore should be about 13,200 years old. This discrepancy may indicate a sedimentary hiatus between 26 and 30 cm. in core A179-4. Another probable hiatus occurs in the same core between 250 and 260

cm., the missing part being represented in core A172-6 by the section between 310 and 360 cm.

In core A180-73, the surely cold 34-cm. level was found to be 15,300 ± 300 years old. A discrepancy of about 2,000 years thus exists between this core and core A172-6.

Ionium dates for the beginning of An 2-1 obtained by Piggot and Urry (1942) for North Atlantic cores P-126 and P-130 are, respectively, about 16,500 and 18,100 years. The former age corresponds to a level which appears to be within An 2-1, while the latter appears to be slightly earlier. The beginning of An 2-1 should lie between these two levels.

An ionium age of about 14,000 years (Urry, 1949) was considered by Hough (1953) to correspond to the end of the Wisconsin in a core from the southeast Pacific. This should correspond to the end of stage 2 and postdate the beginning of An 2-1. Therefore, no disagreement with the previously mentioned ages may be apparent here.

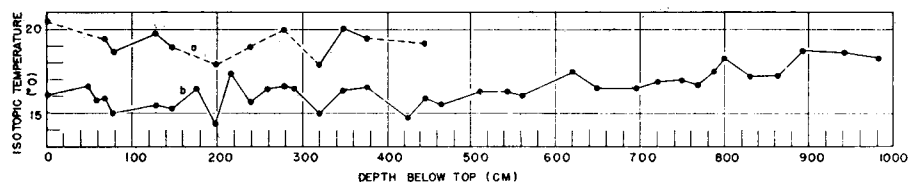On the basis of the foregoing figures, an age of about 16,500 years may be assumed for the

FIG. 11.—Core 58: isotopic temperatures obtained from *Pulleniatina obliquiloculata* (a) and *Globorotalia tumida* (b).
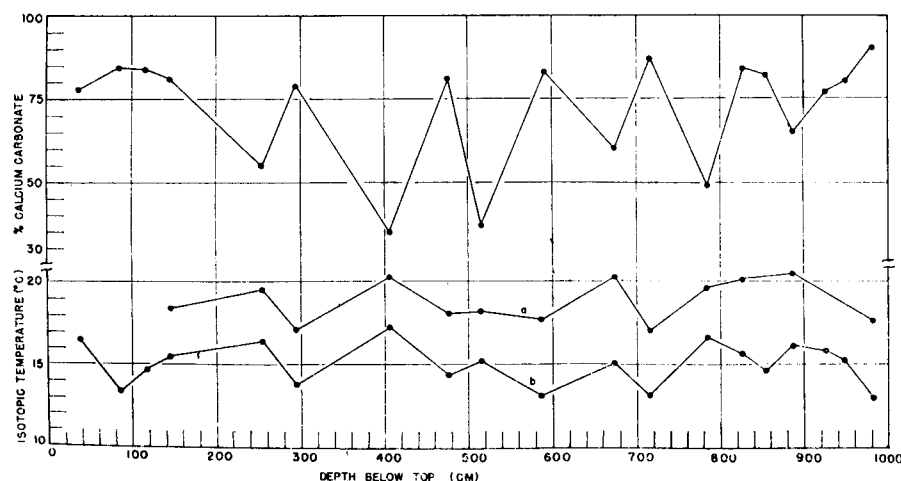

FIG. 12.—Core 60: percentages of calcium carbonate and isotopic temperatures obtained from *Pulleniatina obliquiloculata* (a) and *Globorotalia tumida* (b).
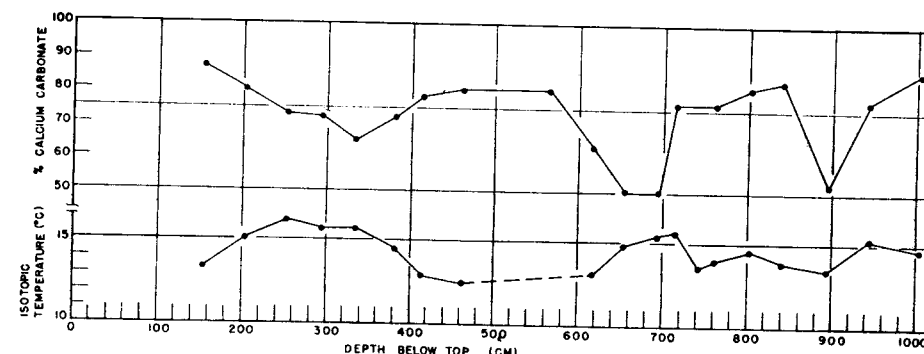
FIG. 13.—Core 61: percentages of calcium carbonate and isotopic temperatures obtained from *Globorotalia tumida*.

TABLE 7

RADIOCARBON DATES FOR CRITICAL LEVELS OF THREE CORES, BASED ON FORAMINIFERA ONLY

Core No.	Stratigraphic Range of C-14 Samples (Cm.)	Level Dated (Cm.)	C-14 Age	Rate of Sedimentation (Cm/1,000 Years)
A179-4	23–30	26.5	11,800 ± 300 ⎱	
A179-4	70–77	73.5	27,600 ± 1000 ⎰	3.0
A172-6	0–10	5	3,700 ± 200 ⎱	
A172-6	51–61	56.0	17,500 ± 500 ⎰⎱	≤ 3.6
A172-6	114–125	119.5	≥ 34,000 ⎰	≤ 3.4
A180-73	0–8	4	2,960 ± 200 ⎱	
A180-73	30–38	34	15,300 ± 300 ⎰⎱	2.4
A180-73	80–88	84	27,500 ± 1500 ⎰	4.3

beginning of An 2-1. This age, although a little lower, as expected, is close to the 18,000 years accepted as near the end of the maximum extent of Wisconsin ice in North America (Flint and Rubin, 1955; Flint, 1955).

Thicknesses of sediment deposited since the end of Ct 5-4 and since the beginning of An 2-1, as shown by various Atlantic and Caribbean cores, are presented in table 8. The average since the beginning of An 2-1 is 26.6 cm., and if a corresponding time duration of 16,500 years is assumed, an average rate of sedimentation of 1.6 cm/1,000 years is obtained for that interval (table 9). However, the sedimentary column is shortened by friction in all cores except those raised with the piston corer. A shortening of about 50 per cent has been reported by Ericson and Wollin (in press). If only cores of table 8 obtained with Ewing's piston corer are considered, the average thickness of sediments deposited since the beginning of An 2-1 is 32.9 cm. and the average rate of sedimentation is 2.0 cm/1,000 years. Thus, a shortening considerably lower than that mentioned by Ericson and Wollin (in press) is indicated for the other cores.

The end of Ct 5-4 has been dated by the ionium method at approximately 60,700 and 61,000 years for North Atlantic core P-126 and Caribbean core P-137 (Piggot and Urry, 1942). Also, the beginning of the "Wisconsin," tentatively identified from the sedimentary sequence in a core from the southeast Pacific (Hough, 1953) was placed at about 65,000 years on the basis of a series of ionium dates determined by Urry (1949). The rounded average of the two former ages, which refer to levels more reliably identified from the stratigraphic point of view, is 61,000 years.

The average thickness of sediment deposited between the end of Ct 5-4 and the beginning of An 2-1, covering a complete temperature cycle, is 185.4 cm. for the cores of table 8 or 190 cm. if only Ewing's cores are used (table 9). If this corresponds to 44,500 years (61,000–16,500), the

TABLE 8

THICKNESS OF SEDIMENTS DEPOSITED IN VARIOUS ATLANTIC AND CARIBBEAN CORES SINCE THE BEGINNING OF AN 2-1 AND SINCE THE END OF CT 5-4

Core No.	Author	Location	Beginning An 2-1	End of Ct 5-4
7-41	Phleger and Hamilton, 1946	North Atlantic	31	?
9-41	Phleger and Hamilton, 1946	North Atlantic	55	?
241	Ovey, 1950	Equat. Atlantic	50	
22	Schott, 1952	Subtrop. Atlantic	20	80
29	Schott, 1952, 1954	Tropical, Atlantic	30	135
227	Schott, 1952, 1954	Tropical, Atlantic	50	270
A180-72	Ericson, 1953	Equat. Atlantic	30	220
A180-74	Ericson 1953	Equat. Atlantic	20	290
A180-76	Ericson, 1953	Equat. Atlantic	40	250
48 cores	Schott, 1935	Equat. Atlantic	24	...
A179-4		Caribbean	26.5	160
A172-6		Caribbean	56	210
A180-73		Equat. Atlantic	25	210
234, 234A		Equat. Atlantic	20	205
235A		Equat. Atlantic	25	...
246		Equat. Atlantic	70	300
280A		Subtrop. Atlantic	27	...
Average			*26.6*	*212*
Average (only cores prefixed "A," raised with Ewing's corer)			*32.9*	*223*

TABLE 9

AVERAGE THICKNESSES AND AVERAGE RATES OF SEDIMENTATION FOR VARIOUS STRATIGRAPHIC INTERVALS OF CORES LISTED IN TABLE 8

Stratigraphic Interval	Average Thickness (Cm.)		Average Rates of Sedimentation (Cm/1,000 Years)	
	All Cores	Ewing's Cores	All Cores	Ewing's Cores
Since beginning of An 2-1	26.6	32.9	1.6	2.0
Since end of Ct 5-4	212	223	3.5	3.7
End of Ct 5-4 to beginning of An 2-1	185.4	190	4.2	4.3

rate of sedimentation was 4.2 (or 4.3) cm/1,000 years.

Rates of sedimentation for a portion of the sedimentary column deposited during Ct 3-2 in cores A179-4, A172-6, and A180-73 can be determined on the basis of the radiocarbon dates for the levels of 73.5, 119.5, and 84 cm., respectively. Rates of 3.0, ≤3.4, and 4.3 cm/1,000 years are obtained, and by extrapolating to the interval between the end of Ct 5-4 and the beginning of An 2-1 time intervals of 44,500, ≥45,300, and 43,000 are obtained, which agree with the average of 44,500 years mentioned previously.

The cores described here and in the literature indicate that the rate of sedimentation was fairly constant at each locality, although it varied from place to place. This probably accounts for the agreement among the figures presented which substantiates the results obtained from radiocarbon and ionium dating and from the extrapolation of sedimentation rates calculated from radiocarbon dates.

The temperature cycle between the end of Ct 5-4 and the beginning of An 2-1 is probably a little less than the 44,000 years estimated above, because the sediments extend a little beyond the temperature minima of stages 2 and 4 in opposite directions. A duration of about 40–41,000 years may be assumed.

Core A172-6 is stratigraphically the longest one studied and shows 6 complete temperature cycles. The oldest minimum occurs at 880 cm. below the top. Because all cycles in an individual core seem to be represented by about the same thickness of sediment, the assumption may be made that they also represent approximately equal time intervals. If this is true, the age of the 880-cm. level of core A172-6 should be about 260,000 years, including the time interval for the top 60 cm. of the core.

The pilot cores, which have been sampled every other centimeter, provide a more detailed record of temperature variations from the upper part of stage 2 to the present. The beginning of An 2-1 occurs at 20 cm. in core 234A, at 22 cm. in core 235A, and at 27 cm. in core 280A (table 8; figs. 8–10). Temperature appears to have risen rather uniformly during An 2-1, until a maximum was reached a few centimeters below the top of the cores (table 10). During this time, ice retreated in North America, with the Cary, Mankato, and Cochrane readvances. These events are not indicated by obvious temperature fluctuations in the cores, showing that they were com-

TABLE 10

DEPTHS OF THE "CLIMATIC OPTIMUM" AND OF THE MINOR TEMPERATURE MINIMUM BELOW IT

Core No.	"Climatic Optimum" (Cm. below Top)	Minor Temperature Minimum below "Climatic Optimum" (Cm. below Top)
A179-TW4	2-4	6-10
234A	8-9	11-12
235A	4-5	14-15
280A	4-5	10-11

paratively minor episodes in a period of generally increasing temperatures. Possibly this resulted from increased evapo-precipitation produced by increasing temperature. Agreement between cores makes improbable total smoothing of record by reworking. A very minor temperature minimum occurs in the upper part of An 2-1 (table 10) which might tentatively be correlated with the Mankato substage. A similar general temperature pattern, inferred from accurate chemical analyses, was discussed by Wiseman (1954) for the Swedish pilot core 241A.[2]

An 2-1 ends with a temperature maximum a few centimeters below the top of the cores (table 10) and correlation with

[2] All O^{18} temperatures mentioned by Wiseman (1954) are uncorrected for the oxygen isotopic composition of the sea water. They should be about 2° C. higher.

the postglacial "Climatic Optimum" is probable. A radiocarbon age of 6,000 years was determined for the English Atlantic phase and for the end of the Pine zone in Maine (Arnold and Libby, 1951), and the age of the minor Cochrane advance may be estimated at about 7,000 years (cf. Rubin and Suess, 1955).

If one may be confident that nothing has been lost during and after coring from the top of the pilot cores (Ericson and Wollin, in press), a sedimentation rate of 1 cm 1,000 years or less is deduced for the top few centimeters of the cores from the age of 6,000 years for the "Climatic Optimum." This rate agrees with 1.3 cm/1,000 years derived from radiocarbon dates of 3,700 and 3,000 years for the 5- and 4-cm. levels of cores A172-6 and A180-73 (table 7), if compaction produced by coring is taken into consideration. These rates are lower than the average from the beginning of An 2-1 and much lower than the average between the end of Ct 5-4 and the beginning of An 2-1 (table 9).

Apparent lower rates of sedimentation during the last few thousand years together with the fact that plankton hauls usually yield much fewer planktonic Foraminifera than needed for a normal rate of sedimentation may possibly indicate that planktonic Foraminifera are becoming rarer.

Curves showing the percentage of sedimentary fractions larger than 62 or 74 μ closely agree with the temperature record of all Atlantic and Caribbean cores. In fact, such curves probably can be used for quick, preliminary investigations of the stratigraphic and climatic record of deep-sea Globigerina-ooze cores. More than 99 per cent by weight of the sedimentary fraction larger than 62 or 74 μ consists of foraminiferal shells in all samples examined. Benthonic Foraminifera usually constitute less than 1 per cent of all Foraminifera, both by weight and by number. Assuming rough constancy in the rates

of sedimentation of terrigenous material and in other pertinent factors (such as postdepositional shell fragmentation etc.), the percentage of the coarser fraction may be taken as an index of planktonic foraminiferal productivity. The correlation coefficients between temperature and the percentage of the coarser fraction determined by Ericson are 0.65 ($P < 0.01$) and 0.44 ($P < 0.01$) for cores A179-4 and A172-6, respectively. The lower correlation for core A172-6 is believed to reflect the effect of postdepositional solution previously discussed. In some cores, a sudden, large increase of the percentage of the coarse fraction has been observed, which probably identifies layers in which the finer sediment has been swept away, and may indicate sedimentary disturbances. Such levels occur in core A172-6 at 770 cm. and in core 234 at 421 and 640 cm.

PACIFIC CORES

Sampling of Pacific cores was based principally on the curves of carbonate percentages (Arrhenius, 1952), and most samples were taken to correspond with carbonate maxima and minima. The samples consisted of vertical halves of core segments varying in length between 2 and 13 cm. The average sample length was 3 cm. for core 58, 6 cm. for core 60, and 8 cm. for core 61. As previously mentioned the superficial species *Globigerinoides rubra* and *Globigerinoides sacculifera* are not sufficiently abundant in the Pacific cores for isotopic analysis. Therefore, the deeper species *Globorotalia tumida* and *Pulleniatina obliquiloculata* have been used, for which depth habitats of 130 m. and 110 m., respectively, were determined (Emiliani, 1954a).

Temperatures obtained from *Globorotalia tumida* and, for cores 58 and 60, also from *Pulleniatina obliquiloculata* are shown in figures 11-13. Graphs of cores 60 and 61 include values of the carbonate percentages as determined by Arrhenius (1952).

Because *Pulleniatina obliquiloculata*, *Globigerina dubia*, and *Globorotalia menardii* occupy similar depth habitats (Emiliani, 1954a), the temperature graph of *Pulleniatina obliquilocu-*

lata in cores 58 and 60 (figs. 11-12) may be compared with the graphs of *Globigerina dubia* and *Globorotalia menardii* in Caribbean core A179-4. The far less conspicuous temperature variations in the Pacific cores are probably due to the increased vertical circulation in superficial layers during colder stages, as proposed by Arrhenius (1952), which resulted in higher carbonate production. Present superficial water temperatures (about 25° C.) are similar in the areas from which the Pacific and the equatorial Atlantic cores were taken. The average temperature during colder stages was 20° C. in the equatorial Atlantic (correction 3 included), and a similar value probably obtained in the Pacific area. Increased vertical circulation probably caused the thermocline to move downward, with the result that foraminiferal populations occupied higher thermal levels than at present during colder stages. A relative increase of temperature at depth was superimposed upon the general trend of decreasing temperatures, with the consequence that the temperature differences between "high-carbonate" and "low-carbonate" stages were very small or insignificant (table 11).

Correlation coefficients between carbonate percentages and temperatures for cores 58, 60, and 61 are 0.07, −0.70, and −0.11, respectively. This indicates that increased vertical circulation during colder stages at the locations of cores 58 and 61 carried enough warmer superficial water downward to prevent a decrease of temperature at depths occupied by the foraminiferal species here considered. This effect apparently was smaller at the location of core 60, and there the general temperature change was not completely masked. Local oceanographic conditions may explain the difference between the temperature record of core 60 and that of the other two cores.

The vanishing temperature difference between high- and low-carbonate stages indicates that not more than approximately 40×10^6 km³ of ice were formed during the glacial ages, corresponding to about a 100-m. lowering of sea-level. A larger amount would increase the value of correction 3, with the consequence that significantly higher temperatures would correspond to glacial ages and lower ones to interglacial ages. While

this is possible as a result of increased vertical circulation, cooling of superficial water to approximately 20° C. makes it rather improbable.

Although the temperature record furnished by the Pacific cores has been greatly altered by vertical circulation of superficial water, the correlation suggested by Arrhenius (1952) between high-carbonate stages and low general temperature appears correct.

Arrhenius (1952) established a chronology for cores 58, 59, 60, 61, and 62, based on the rate of accumulation of TiO_2, which was found to be constant in the area of these cores. The

TABLE 11
MEAN TEMPERATURES OF HIGH-CARBONATE AND LOW-CARBONATE STAGES

	MEAN TEMPERATURE (° C.)	
CORE No.	High Carbonate Stages	Low Carbonate Stages
58	14.7	15.5
60	13.2	15.2
61	13.4	14.3

rate was calibrated by a radiocarbon determination on pilot core 61B, which gave an age of 14,200 ± 900 years for the bottom of the core at a depth of 45 cm. (Arrhenius, Kjellberg, and Libby, 1951). The total amount of TiO_2 was found to be 7.59 mg., and the rate of accumulation 0.535 mg/cm² per 1,000 years.

Arrhenius' chronology was based on the assumption that the rate of accumulation of TiO_2 remained constant during the time represented by the cores. However, there is a difference between the amount of TiO_2 which accumulated during the high- and low-carbonate stages. This is best exemplified by cores 59 and 60, where the alternation of high- and low-carbonate stages is the clearest. The results are shown in table 12. Arrhenius' stage 2 of these cores has been excluded from the calculation because of its possible multiplicity.

If TiO_2 deposition was constant in time, larger values for high-carbonate stages would mean that these stages represent about 50 per cent more time than low-carbonate stages. The Atlantic cores, however, do not indicate that colder stages, which correspond to the high-carbonate stages of the Pacific cores, represent more time than warmer stages. If the assump-

tion is made that warmer and colder stages were about equally long, a higher rate of TiO_2 accumulation is indicated for the high-carbonate stages of the Pacific cores. This may have been caused by the presence of a larger amount of terrigenous lutite in the ocean water, following glacial weathering and glacial eustatic lowering of the sea-level which uncovered portions of the continental shelves.

The Atlantic pilot cores (figs. 7–10) show a remarkably regular temperature increase since the beginning of An 2-1. A corresponding decrease in carbonate percentages should occur in the Pacific cores. However, core 61B, on which TiO_2 chronology was based, does not show this, nor does it seem to connect well with core 61. It is possible that the sediments of core 61B have been disturbed, and, if so, the radiocarbon determination mentioned may be invalid. Also,

TABLE 12

AVERAGE AMOUNT OF TiO_2 IN HIGH- AND LOW-CARBONATE STAGES

	TiO_2 (MG/CM²)	
CORE NO.	High-Carbonate Stages	Low-Carbonate Stages
59	28.2	20.3
60	34.8	21.8
Weighted average	31.5	21.0

this determination was based on bulk core material, which may have introduced an error (cf. Rubin and Suess, 1955).

Core 58B, however, shows a much more regular decrease of carbonate in the top 16 cm. It is probable that the 16-cm. level, where the carbonate rate of accumulation and percentage begin to decrease, marks the beginning of An 2-1, for which an age of 16,500 years was obtained. TiO_2 to the amount of 16 mg/cm² is present in the top 16 cm., indicating a rate of accumulation of 0.97 mg/cm² per 1,000 years, which is considerably higher than the rate obtained for core 61B on the basis of the radiocarbon age determination.

As shown by Arrhenius, stage 1 (including the postglacial) is characterized in the Pacific cores by low carbonate, and from this point of view, is essentially similar to older, low-carbonate stages. Therefore, sedimentary conditions probably also were similar, and extrapolation of the rates of accumulation of TiO_2, as previously determined, to the older, low-carbonate stages, seems reasonable. From the total TiO_2 amount

of 21.0 mg/cm² deposited during low-carbonate stages (table 12) and a rate of accumulation of 0.97 mg/cm² per 1,000 years, an average duration of 21,600 years is indicated. Assuming equal duration for the high- and low-carbonate stages, a complete temperature cycle appears to correspond to a time interval of 43,200 years. This figure agrees with that obtained from the Atlantic and Caribbean cores by different methods.

The temperature graph of core 58 (fig. 11) shows a general temperature decrease from bottom to top. The 395-cm. level was selected by Arrhenius (1952) as the Plio-Pleistocene boundary, but the low-temperature value at 580–584 cm. seems to indicate that this boundary should be established at some lower level. The 610-cm. level seems a good choice because it immediately follows a high-temperature value and immediately precedes a major carbonate maximum.

Comparison with cores 59, 60, 61, and 62 indicates the occurrence of about fifteen complete carbonate-temperature cycles above this level. A time interval of about 600,000 years is represented, which should constitute the entire length of the Pleistocene epoch. Correlation of the 610-cm. level of core 58 with "the horizon of the first indication of climatic deterioration in the Italian Neogene succession," officially proposed as the beginning of the Pleistocene (Internat. Geol. Cong., 1950) is hypothetical.

Atlantic cores of Tertiary age (Emiliani, in press, b) show rather uniform temperatures and absence of cyclic changes.

BENTHONIC FORAMINIFERA

A number of samples of calcareous benthonic Foraminifera from the Atlantic cores 234A, 234, and 280 and the Pacific cores 58 and 60 have been analyzed. The samples were selected from levels representing temperature maxima and minima. The 117- and 522.5-cm.

levels of core 280 represent the temperature minimum of stage 2 and the temperature maximum of stage 5. Core 280, which comes from the same location as core 280A, is not included in the present study except for the two samples mentioned.

TABLE 13

PRESENT BOTTOM TEMPERATURES AT LOCATIONS OF CORES FROM WHICH BENTHONIC FORAMINIFERA HAVE BEEN ANALYZED

Core No.	Present Bottom Temperature (° C.)	Location
234A, 234	2.5	Eastern equatorial Atlantic
246	0.9	Western equatorial Atlantic
280	2.5	North Atlantic
58	1.7	Eastern equatorial Pacific
60	1.7	Eastern equatorial Pacific

Present bottom temperatures at the locations of the cores are shown in table 13. Bottom temperature at the location of core 246 is lower than at the location of cores 234A and 234.

Core 246 was taken from an area west of the Mid-Atlantic Ridge separated from the Antarctic Basin by the Rio Grande Ridge; the latter is not high enough to prevent the influx of cold bottom water from the south. Cores 234A and 234 were taken at places east of the Mid-Atlantic Ridge, in an area somewhat shielded from such influx by the higher Walfisch Ridge. The Romanche Trench, which allows communication between the Brazil and Sierra Leone basins, is evidently too narrow to alter significantly the regional thermal pattern. Core 280 was taken too far north to be reached by any significant amount of deep Antarctic water.

Table 14 shows the isotopic temperatures obtained from the benthonic Foraminifera. These temperatures have been corrected for the isotopic composition of the deep water ($A = -0.30$; see Epstein and Mayeda, 1953).

Comparison between tables 13 and 14 shows that interglacial bottom temperatures in the Pacific are similar to present temperatures, but in the Atlantic they are about 2°–3° C. higher. Particularly, the 2-3-cm. level of core 234A, which should not be older than 2,000–3,000

TABLE 14

O^{18}/O^{16} RATIOS (δ PER MIL) AND TEMPERATURE (° C.) OF BENTHONIC FORAMINIFERA FROM ATLANTIC AND PACIFIC CORES

	GLACIAL TEMPERATURE MINIMA				INTERGLACIAL TEMPERATURE MAXIMA		
Core	Level (Cm.)	δ Per Mil	T (° C.)	Core	Level (Cm.)	δ Per Mil	T (° C.)
				ATLANTIC OCEAN			
234	43.5+ 53.5	+4.33	−0.4	234A	2–3	+2.72	4.8
234	153.5	+3.96	0.7	234	303.5	+3.00	3.8
234	194.5	+3.96	0.7	246	5–15	+2.48	5.7
234	203.5	+4.06	0.5	280	522.5	+3.16	3.3
234	363.5+373.5	+4.04	0.4				
280	117	+4.08	0.4				
Average		+4.07	0.4	Average		+2.84	4.4
				PACIFIC OCEAN			
58	145–149	+3.52	2.1	58	346–350	+3.70	1.5
58	197–200	+3.55	2.1	58	460–470	+3.70	1.5
58	237–240	+3.88	0.9	60	251–257	+3.27	3.0
58	276–280	+3.55	2.1	60	672–674	+3.53	2.0
58	318–322	+3.78	1.3	60	882–890	+3.05	3.7
58	509–513	+3.46	2.3				
60	582–588	+3.85	1.1				
60	712–717	+3.48	2.3				
60	977–985	+4.05	0.4				
Average		+3.68	1.6	Average		+3.45	2.4

years, indicates a temperature of 4.8° C., and the 5–15-cm. level of core 246, which should not be older than about 8,000 years, gave a temperature of 5.7° C.

If these were true temperatures, a 2°–3° C. decrease of the temperature of equatorial Atlantic bottom water during the last few thousand years would be indicated. That this is unlikely is shown by the fact that the superficial temperature decreased only 1°–1.5° C. since the climatic optimum (figs. 7–10) and that interglacial bottom temperatures in the equatorial Pacific were at most 0.8° C. higher than at present. Probably, therefore, high values obtained for interglacial bottom temperatures in the Atlantic indicate that the isotopic composition of bottom water there and at those times was considerably lighter than at present and that a correction larger than −0.30 per mil should be applied.

About 95 per cent of the ice which accumulated in middle and high northern latitudes during the various glacial ages entered the Atlantic when it melted. Part of the meltwater reached the Atlantic by way of the Mississippi River or from the Mediterranean, acquiring a higher temperature during its journey and becoming part of the superficial or intermediate Atlantic water. A large part, however, reached the Atlantic more directly along the northeastern coast of North America and the western coast of northern Europe. Formation of bottom water in the northern part of the North Atlantic was probably more active than at present, and a greater influx of isotopically lighter water flowing southward along the bottom probably occurred. An isotopic composition of −15 per mil for meltwater (see p. 543) and −0.85 per mil for bottom water of the equatorial Atlantic during interglacial ages (required to lower isotopic temperatures to modern values) would indicate the presence of about 5 per cent of meltwater. Present isotopic composition of about −0.30 per mil (Epstein and Mayeda, 1953) indicates the presence of only about 2 per cent of meltwater.

The presence of an appreciable amount of meltwater at the bottom of the Atlantic Ocean as recently as 2,000–3,000 years ago, i.e., several thousand years after the ice had largely disappeared, suggests that movement and mixing of bottom oceanic waters may be slow processes.

Conditions were quite different in the Pacific. This ocean received little interglacial meltwater from the Northern Hemisphere, and the amount received from Antarctica in excess of the amount now contributed probably was small.

This and the great mass of the Pacific Ocean account for the small change in the isotopic composition of its bottom water. Assuming constant bottom temperature during glacial and interglacial ages, the change would be only 0.23 per mil (difference between averages; see table 14).

Corrected temperatures for glacial ages (table 14) are close to modern values for Pacific samples but about 2.1° C. lower for the Atlantic samples. This is believed to have been a true temperature difference, resulting from glacial cooling of Atlantic bottom water by ice in and around the North Atlantic. In fact, all glacial Atlantic samples of table 14 are from areas not easily reached by bottom Antarctic water, but open toward the north. Similar cooling did not affect the Pacific because of its greater mass and the comparatively smaller amounts of ice which formed along North Pacific shores during glacial ages.

In summary, isotopic temperatures obtained from benthonic Foraminifera indicate that (1) bottom temperatures of glacial ages were the same as today in the equatorial Pacific but about 2.1° C. lower in the eastern equatorial and northern Atlantic; (2) interglacial bottom temperatures were not more than 0.8° C. higher than at present in the equatorial Pacific, while data for the Atlantic are inconclusive with respect to temperature but indicate a larger influx of meltwater along the ocean bottom.

CORRELATION WITH CONTINENTAL EVENTS

HIGHER LATITUDES

Table 15 shows the probable correlation of core stages with continental stratigraphy of Europe and North America.

Correlation between Mankato and Fennoscandian and between Two Creeks and Allerød substages in North America and Europe, respectively, seems safely established on the basis of numerous radiocarbon age determinations. Correlation between Cary and Pomeranian substages is also probable, but correla-

tion between older substages is more uncertain. A radiocarbon date of more than 36,000 years (Rubin and Suess, 1955) for Belgian peat underlying Younger Loess II, of Weichsel age, and overlying deposits with a cold fauna and Mousterian artifacts, of probable Warthe age, suggests correlation between the Weichsel and the whole lower Wisconsin (Farmdale-Iowan-Tazewell).

Correspondence between the temperature maximum of stage 1 of the cores and the "Climatic Optimum" seems obvious and has already been mentioned. Radiocarbon and ionium dates indicate that the Cary-Mankato and Pomeranian-Fennoscandian correspond to An 2-1. Because the temperature graphs do not show any secondary temperature decrease during An 2-1, there is possible evidence that these substages resulted from higher evaporation following rising temperature, a mechanism advocated by Simpson (1934, 1940) for odd glacial ages. The Cary and Mankato substages and their European counterparts appear to be mere readvances of the continental ice superimposed on a phase of general retreat (cf. Flint, 1955).

Correspondence of stage 2 with the Farmdale-Cary = Weichsel-Pomeranian appears probable. The cool Rixdorf horizon seems to correspond to the cool stage 3 and the important Warthe phase with

TABLE 15

PROBABLE CORRELATION BETWEEN CORE STAGES AND CONTINENTAL STRATIGRAPHY OF EUROPE AND NORTH AMERICA

CORE STAGES	STAGES AND SUBSTAGES		YEARS* (×1,000 B.P.)
	Europe	North America	
	"Climatic Optimum"		
1	?	Cochrane	— 7 —
	Fennoscandian	Mankato	— 8 —
	Allerød	Two Creeks	— 11 —
	Pomeranian	Cary	— 12 —
2	Masurian	Brady	— 14 —
		Tazewell	— 17 —
	Weichsel	Iowan	— 20 —
		Farmdale	— 22 —
3	Rixdorf	?	— 30 —
4	Warthe	?	— 55 —
5	Riss/Würm	Sangamon	— 75 —
6	Riss	Illinoian	— 103 —
7, 8, 9	Mindel/Riss	Yarmouth	— 125 —
10	Mindel	Kansan	— 175 —
11, 12, 13	Günz/Mindel	Aftonian	— 200 —
14	Günz	Nebraskan	— 265 —
			— 290 —

* Based on radiocarbon ages and the insolation data discussed in the next section.

stage 4. The radiocarbon age of 25,000 years for the Farmdale, which is the earliest recognized substage of the Wisconsin, clearly indicated that the whole Wisconsin as presently defined should correspond to stage 2 and to much of An 2-1. Deposits corresponding to stages 3 and 4 have been recognized in North America (Flint and Rubin, 1955), but they are, so far as known, very scarce. Some deposits now considered to be of Sangamon or Illinoian age may, in the future, be shown to belong to these stages.

Core stage 5 should correspond to the Sangamon and core stage 6 to the Illinoian = Riss. Assignment of core stage 5 to the last glaciation as a warm substage and of core stage 6 to the preceding interglacial as a cold substage is contradicted by their temperatures, which reach values similar to the postglacial maximum and to the Wisconsin minimum, respectively.

The Yarmouth = Mindel Riss interglacial stage has generally been estimated, on geological grounds, to have been much longer than any other interglacial stage. Core stage 7, however, is not represented by thicker sediment than the other warm stages in any of the cores, and it is preceded by a temperature minimum, stage 8, which appears to be about as low as other minima. But the temperature graphs of the two Caribbean cores show that the thickness of stage 8 sediments is considerably less than that of either stages 6 or 10. Therefore, the Yarmouth perhaps is actually represented by stages 7, 8, and 9, stage 8 being only a cold phase within the interglacial. Some geologic evidence for such an interpretation has been reported in the literature (cf. Zeuner, 1945). If stages 7, 8, and 9 represent the Yarmouth, a duration of about 82,000 years may be estimated for this interglacial, which is not in too great disagreement with the 120,000 years estimated from the depth of leaching of Kansan drift (Kay, 1931; estimate revised for the shorter duration of post-Mankato time). The temperature maxima of stages 9 and 11 in core A172-6 can be inferred only by comparison with core A179-4.

Stage 10 may correspond to the Kansan = Mindel. Stage 12, represented by a minor temperature minimum, may correspond to only a minor cold phase, and stages 11, 12, and 13 may represent an interglacial about 82,000 years long. This interglacial may be the Aftonian, for which a duration of about 80,000 years has been estimated on the basis of depth of leaching in Nebraskan drift (revised estimate). Finally, cold stage 14 may correspond to the Nebraskan = Günz glaciation.

Correlation of core stages and continental stratigraphy seems fairly well established up to and perhaps including stage 5. Interpretation for older stages is highly speculative at present and needs much substantiating evidence.

LOWER LATITUDES

Evidence of Pleistocene climatic fluctuations in nonglaciated areas of middle and low latitudes is furnished by ancient lake strand lines, stream features, and soil formation. Thus, several wet and dry phases have been recognized in various parts of the world. The number of the major wet phases, however, is generally less than the number of major glaciations, but, as Flint (1947) points out, the whole Pleistocene record need not be preserved at each locality. Correlation of wet phases and glacial stages generally has been considered probable, although some authors have expressed dissenting views (Sandford, 1935; Zeuner, 1945; etc.).

Because all deep-sea cores here considered, except core 280A, are from low latitudes, the previously demonstrated correlation of at least the last five core stages with glacial and interglacial events in high latitudes strongly indicates that the climatic pattern in high latitudes determined the climatic pattern in lower latitudes, and that wet and dry phases in equatorial and tropical regions correlate with glacial and interglacial ages. Also, contemporaneity of colder and warmer ages the world over is indicated.

CORRELATION WITH INSOLATION

If solar radiation is constant, variations in the insolation of different parts of the earth result from the three periodic elements shown in table 16. The first and third elements produce insolation changes which are opposite in the Northern and Southern Hemispheres. The second, on the other hand, produces changes which are similar in both hemispheres. Increase of obliquity results in increased summer insolation for almost all latitudes and decreased winter insolation for all latitudes. Also, total yearly insolation increases for latitudes higher than 43° and decreases for lower latitudes.

Interaction of the three elements produces insolation changes varying in period for different latitudes. Students have concluded that changes around 65° N. are the most important because they probably controlled the formation and disappearance of ice sheets, and that summer insolation, rather than winter, seems to have been the more important factor (Zeuner, 1945).

Detailed curves showing the variation of insolation at 65° N. and other latitudes during the last 600,000 years have been calculated by Milankovitch (1920,

1930, 1938). Recently, the curve for 65° N. has been recalculated by Brouwer and van Woerkom on the basis of new data for the planetary masses (van Woerkom, 1953). The new curve, which is not much different from the older one, has been redrawn inverted from data kindly furnished by van Woerkom, and is shown in figure 14.

An insolation maximum at −10,000 years and a minimum at −23,000 years may be observed. If these are correlated respectively with the "Climatic Optimum" (−5,000 years) and the time of maximum extent of Wisconsin ice in North America (−18,000 years), a re-

TABLE 16

ASTRONOMICAL ELEMENTS HAVING CLIMATIC IMPLICATIONS

Elements	Period (Years)
Precession of the equinoxes	21,000
Obliquity of the ecliptic	40,000
Eccentricity of the orbit	92,000

tardation of 5,000 years is evident. Such an effect is to be expected and was foreseen a long time ago (see Zeuner, 1945). It seems reasonable to assume that a similar retardation obtained also for earlier times.

Table 17 shows the ages of the even core stages and the ages of the last seven insolation minima of figure 14. Ages of core stages 2 and 4 were discussed previously, and ages of earlier even stages were estimated by assuming a constant sedimentation rate and extrapolating backward in time, from stage 4, the rates of 3.0 and 3.6 cm/1,000 years in cores A179-4 and A172-6, respectively (table 7). A correction of 5,000 years has been applied to the ages of insolation minima because of apparent retardation.

General agreement between ages of insolation minima, and even core stages, supports the conclusion that summer in-
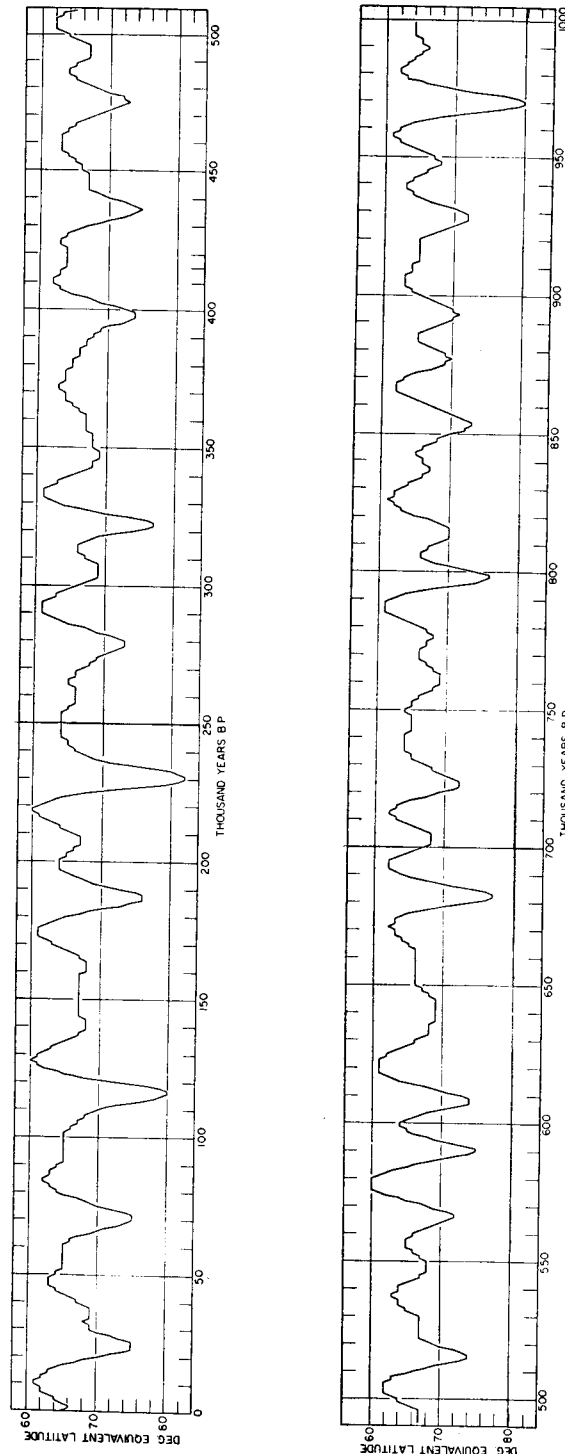
FIG. 14.—Variation of the insolation at 65° N. during the last million years, in degrees of equivalent latitude (based on data calculated by D. Brouwer and A. J. J. van Woerkom).

solation at high northern latitudes and Pleistocene temperatures may be related.

Disagreement between the calculated age of stage 14 and the supposedly corresponding insolation minimum may possibly be explained by sedimentary hiatus between stages 12 and 14. Sedimentary disturbance is suggested by the high coarse fraction percentage at the 740-cm. level, indicating that the finer fraction may have been washed away by some

TABLE 17

AGES OF EVEN CORE STAGES AND THE LAST SEVEN INSOLATION MINIMA

CORE STAGE	LEVEL (CM.) A179-4	LEVEL (CM.) A172-6	AGE OF CORE STAGES (×1,000 YEARS B.P.) A179-4	AGE OF CORE STAGES (×1,000 YEARS B.P.) A172-6	AGES OF INSOLATION MINIMA (×1,000–5,000 YEARS B.P.)
2	40	60	18	18	18
4	150	200	61	61	67
6	260	360	98	105	111
8	405	480	146	139	148 (?)
10	515	610	183	175	183
12	620	770	218	219	226
14	—	880	—	250	275

sort of submarine current, and the sedimentary thickness reduced.

Closer comparison of the insolation curve and the temperature curves reveals some additional supporting evidence and also some inconsistencies. For example, the lower value of the insolation peak at −45,000 years corresponds to the low-temperature maximum of stage 3, and the high insolation minimum around 143,000 years corresponds to the thinner sediments of stage 8. On the other hand, the most pronounced insolation minimum (−231,000 years) is opposed by the least pronounced temperature minimum at stage 12, and the apparently unimportant insolation minimum of −280,000 years is opposed by the important temperature minimum of stage 14.

Thus, a time correspondence seems to exist between temperature and insolation, but correspondence in amplitude is not exact. A causal connection is suggested but is not proved. Additional evidence undoubtedly will be obtained from future isotopic analyses of some stratigraphically longer cores than are now available from the equatorial Atlantic or the Caribbean.

Figure 15 shows estimated temperature variations with time in the low latitudes, based on the temperature records
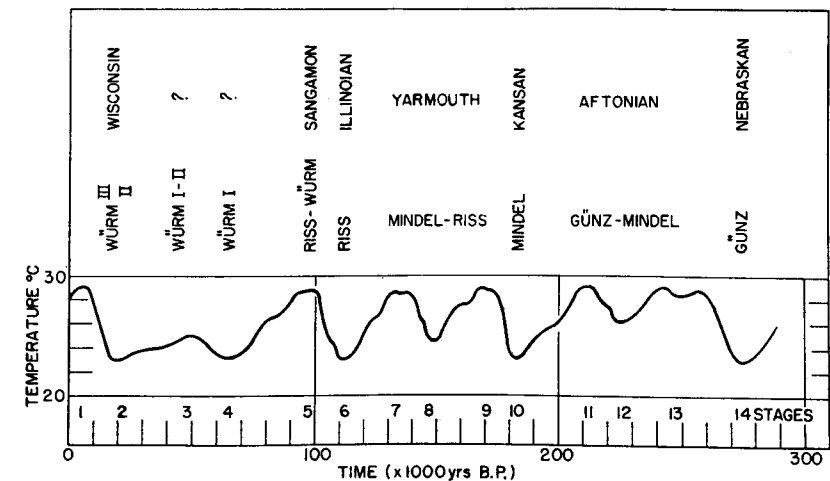

FIG. 15.—Generalized temperature variation, based on the temperature graphs of the cores and on the astronomical time scale.

of the cores and including all corrections (p. 544). If correlation between core stages and insolation cycles as shown in table 17 is correct, current correlations between the insolation curve and continental events (Zeuner, 1945, 1952) must be modified as shown in table 18 and figure 15.

All Pleistocene time since the beginning of the Günz glaciation appears to

TABLE 18

CORRELATION BETWEEN CORE STAGES, INSOLATION MINIMA AND MAXIMA, AND CONTINENTAL STAGES AND SUBSTAGES

CORE STAGE	AGE OF INSOLATION MINIMA AND MAXIMA (×1,000-5,000 YEARS B.P.)	CONTINENTAL STAGES AND SUBSTAGES	
		Europe	North America
1	5	"Climatic Optimum"	
2	18	Würm II–III	Wisconsin
3	45	Rixdorf	?
4	67	Würm I	?
5	79	Riss/Würm	Sangamon
6	111	Riss	Illinoian
7	124	} Mindel/Riss	Yarmouth
8	138		
9	170		
10	183	} Mindel	Kansan
11	214		
12	226	} Günz/Mindel	Aftonian
13	245		
14	275	Günz	Nebraskan

be about 280,000 years, which is not in disagreement with the estimates of Penck and Brückner (1909) and Kay (1931), if these are corrected for the younger age of the Würm 3 and the shorter duration of the post-Mankato time, respectively.

The insolation curve preceding the minimum at −280,000 years, tentatively correlated with the Günz glaciation, shows several more or less pronounced minima extending back a million years or more. Some may correspond with pre-Günzian stages or core stages earlier than stage 14 (in the Pacific cores described by Arrhenius, 1952), but correlations are too uncertain at present to be attempted,

especially as a close quantitative relation between insolation and world temperatures does not seem to have occurred.

CAUSES OF GLACIAL EPOCH AND AGES

The evidence presented in the preceding section seems to indicate that variable insolation in high northern latitudes probably caused the alternation of colder and warmer ages. As is well known, this factor cannot have been the cause of the glacial epoch as such.

At present, the least objectionable explanation of both glacial epoch and its ages seems to be a theory calling upon a combination of topographical and insolation effects. This is close to Flint's (1947) "solar-topographic" hypothesis, with the difference that insolation variations are substituted for solar emission variations. Increasing continentality and mountain-building during the Tertiary made the transfer of heat from lower to higher latitudes more difficult and caused a general lowering of the earth's temperature (cf. Emiliani, 1954b). When this lowering reached a certain value, accumulation of permanent snow and consequent glaciation occurred in Greenland and Antarctica, possibly in pre-Pleistocene times. Glaciation of Greenland and Antarctica further lowered temperature. At this point, apparently, insolation variations at high northern latitudes became critical, and temperatures in the Laurentide and Scandinavian regions, which stood 200–250 m. higher than at present, were such that the first insolation minimum caused permanent snow to accumulate there.

A 5,000-year lag between temperature and insolation was suggested in the preceding section, but probably temperature was promptly reflected by the amount of ice. The time lag of isostatic movements of northern glaciated lands with respect

to temperature and ice may be estimated at about 10,000–15,000 years, on the basis that the Laurentide and Scandinavian regions have recovered about two-thirds of their former height since ice disappeared from these regions about 8,000 years ago. The recovery should be largely completed in about 4,000 years.

A glimpse of future climate might be obtained by extending the insolation curve toward the future; if its trend during the last 10,000 years (fig. 14) continues, an insolation minimum may be expected to occur in about 10,000 years, when the northern lands will have completed their isostatic recovery. Conditions will then be favorable for the inception of a new ice age.

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APPENDIX[3]

OXYGEN ISOTOPIC ANALYSES (δ PER MIL)

TABLE 19

CORE 58

Depth below Top (Cm.)	Globorotalia tumida	Pulleniatina obliquiloculata	Depth below Top (Cm.)	Globorotalia tumida	Pulleniatina obliquiloculata
1–3	+0.30	−0.73	443–448	+0.34	−0.42
47–50	+0.18	—	460–470	+0.44	—
57–60	+0.38	—	509–513	+0.25	—
67–70	+0.35	−0.48	541.5–544.5	+0.25	—
87–90	+0.57	−0.31	580–584	+0.29	—
127–130	+0.44	−0.56	619–622	0.00	—
145–149	+0.51	−0.39	646–652	+0.21	—
175–178	+0.21	—	694.5–697.5	+0.22	—
197–200	+0.71	−0.12	717–725	+0.12	−0.47
215–218	0.00	—	747–750	+0.09	—
237–240	+0.41	−0.38	766.5–769.5	+0.17	—
257–260	+0.21	—	786.5–789.5	−0.02	—
276–280	+0.19	−0.60	797–803	−0.23	—
288–293	+0.23	—	829–833	+0.06	—
318–322	+0.57	−0.12	861.5–864.5	+0.06	—
346–350	+0.25	−0.62	890–893	−0.29	—
375–378	+0.19	−0.48	941.5–944.5	−0.28	—
425–428	+0.64	—	980–984	−0.23	—

[3] In tables 19–30, an asterisk (*) refers to samples containing a certain amount of tests of *Globigerinoides rubra*, usually not more than 20 per cent by weight.

TABLE 20

CORE 60

Depth below Top (Cm.)	Globorotalia tumida	Pulleniatina obliqui-loculata	Depth below Top (Cm.)	Globorotalia tumida	Pulleniatina obliqui-loculata
34–41.5	+0.22	—	672–674	+0.54	−0.63
81.5–87.6	+0.96	—	712–717	+1.00	+0.11
114–118	+0.65	—	782–786	+0.19	−0.52
140–150	+0.44	−0.22	822–830	+0.42	−0.60
251–257	+0.29	−0.50	853.5–857	+0.68	—
290–298	+0.84	+0.07	882–890	+0.30	−0.67
402–408	+0.02	−0.69	921.5–927.5	+0.39	—
473–479	+0.70	−0.16	944–950	+0.53	—
514–517	+0.52	−0.19	977–985	+1.03	−0.03
582–588	+1.01	−0.08			

TABLE 21

CORE 61

Depth below Top (Cm.)	Globorotalia tumida	Depth below Top (Cm.)	Globorotalia tumida	Depth below Top (Cm.)	Globorotalia tumida
150–160	+0.72	459–466.5	+1.17	758–766	+0.82
201.5–207.5	+0.56	614.5–622	+1.03	801–807	+0.73
251.5–254	+0.31	652–660	+0.62	837–846.5	+0.85
292–298	+0.41	690–699	+0.48	892–900	+0.94
330–340	+0.41	714–717	+0.44	940–950	+0.53
379.5–387	+0.69	740–746	+0.93	1002.5–1009	+0.67
412–420	+1.08				

TABLE 22

CORE A172-6

Depth below Top (Cm.)	Globigerinoides rubra	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides rubra	Globigerinoides sacculifera
Top	—	−1.87	470	−0.52	−0.21
10	—	−1.86	480	−0.41	−0.38
20	−1.81	−1.61	490	−0.50	−0.26
30	−1.51	−1.24	500	−0.80	−0.52
40	−1.01	−0.75	510	−1.03	−0.93
50	−0.47	−0.27	520	−1.17	−0.99
60	−0.29	+0.09	530	−1.44	−0.97
70	−0.41	−0.19	540	−1.31	−1.84
80	−0.56	−0.36	550	—	—
90	−0.61	−0.16	560	—	—
100	−0.68	−0.07	570	—	—
110	−0.77	−0.22	580	—	—
120	−0.51	−0.01	590	−1.19	−1.00
130	−0.68	−0.40	600	−0.13	+0.03
140	−0.63	−0.22	610	−0.16	+0.02
150	−0.88	−0.75	620	−0.14	−0.09
160	−0.91	−0.59	630	−0.35	—
170	−0.95	−0.80	640	−0.60	−0.54
180	−0.25	−0.20	650	−0.61	−0.66
190	+0.02	+0.16	660	−0.78	−0.69
200	+0.02	+0.21	670	−0.74	−0.78
210	+0.20	+0.24	680	—	—
220	−0.53	−0.21	690	—	—
230	−0.76	−0.46	700	—	—
240	−1.04	−0.71	710	−1.22	−0.95
250	−0.89	−0.71	720	−1.21	−1.09
260	−1.15	−0.73	730	−1.26	−1.08
270	−1.53	−1.05	740	−1.20	−1.08
280	−1.90	−1.34	750	−1.10	−0.92
290	−1.66	−1.34	760	−0.80	−0.62
300	—	—	770	−0.93	−0.75
310	−1.67	−1.45	780	−0.92	−0.74
320	−1.17	−1.22	790	−1.40	−1.04
330	−1.09	−0.99	800	—	—
340	−1.47	−1.10	810	—	−1.43*
350	−0.76	—	820	—	—
360	−0.29	+0.08	830	−2.03	−1.79
370	−0.50	−0.27	840	—	−1.38
380	−0.57	−0.31	850	−1.32	−1.60*
390	−0.56	−0.38	860	−1.94	−1.65*
400	−0.62	−0.41	870	—	−0.30
410	−1.23	−0.86	880	+0.02	+0.12
420	−1.12	−1.12	890	−0.46	−0.23
430	−1.48	−1.48	900	—	−0.85*
440	−1.14	—	910	—	−1.09
450	−1.50	−1.21	920	—	−0.73
460	−1.34	−1.05	935	—	−0.96

TABLE 23

CORE A179-TW4

Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera
Top	−1.85	6–8	−1.64	10–12	−1.81
2–4	−2.10	8–10	−1.66	12–14	−0.91
4–6	−1.83				

TABLE 24

Core A179-4

Depth below Top (Cm.)	Globigeri-noides rubra	Globigeri-noides sacculifera	Globigerina dubia	Globorotalia menardii	Depth below Top (Cm.)	Globigeri-noides rubra	Globigeri-noides sacculifera	Globigerina dubia	Globorotalia menardii
Top	-2.04	-1.80	-1.01	-0.82	350	-1.53	-1.49	-0.76	-0.42
10	-1.92	-1.78	-1.04	-0.85	360	-1.75	-1.57	—	-0.46
20	-0.77	-0.81	+0.19	—	370	-1.45	-1.36	—	-0.12
30	-0.23	+0.02	+0.86	—	380	-0.98	-0.84	—	-0.09
40	-0.26	+0.16	+0.67	—	390	-1.00	-0.90	-0.57	-0.34
50	-0.33	-0.09	+0.61	—	400	-0.44	-0.15	—	+0.02
60	-0.42	-0.20	+0.62	—	410	-0.55	-0.35	—	+0.42
70	-0.51	-0.23	+0.57	—	420	-0.83	-0.83	-0.07	-0.38
80	-0.69	-0.18	+0.31	—	430	-1.42	-1.03	-0.61	-0.51
90	-0.61	-0.25	+0.31	—	440	-1.27	-1.00	-0.39	-0.35
100	-0.69	-0.23	+0.15	—	450	-1.41	-1.66	-0.60	-0.33
110	-0.59	-0.27	+0.39	—	460	-1.57	-1.49	-0.63	-0.63
120	-0.78	-0.34	+0.22	—	470	-1.39	-1.34	+0.03	-0.26
130	-0.56	-0.32	+0.20	—	480	-1.47	-1.35	-0.40	-0.57
140	-0.51	-0.34	+0.27	—	490	-1.69	-1.54	-0.39	-0.48
150	-0.37	-0.22	+0.48	—	500	-1.08	-1.24	-0.16	-0.16
160	-0.47	-0.38	—	—	510	-0.30	-0.15	+0.42	-0.21
170	-1.05	-0.75	-0.14	—	520	-0.33	-0.12	+0.69	+0.09
180	-1.32	-0.95	-0.23	-0.39	530	-0.42	—	+0.52	+0.31
190	-1.41	-1.03	-0.36	-0.39	540	-0.49	-0.63	+0.34	—
200	-1.35	-1.17	-0.39	-0.43	550	-0.65	-0.67	+0.27	+0.09
210	-1.55	-1.03	-0.36	-0.15	560	-1.34	-0.99	-0.03	-0.39
220	-1.59	-1.13	-0.38	-0.35	570	-1.65	-1.41	-0.01	-0.42
230	-1.77	-1.50	-0.79	-0.64	580	-1.47	-1.38	-0.12	—
240	-2.07	-1.80	-0.81	-0.63	590	-1.39	-1.20	-0.19	—
250	-1.70	-1.45	-0.55	-0.50	600	-1.19	-0.83	+0.20	—
260	-0.27	+0.14	+0.41	-0.53	610	-1.30	-0.96	+0.01	—
270	-0.02	+0.07	+0.56	—	620	-0.66	-0.66	+0.34	—
280	-0.12	-0.16	+0.17	+0.11	630	-1.36	-0.77	-0.04	—
290	-0.44	-0.31	-0.08	+0.07	640	-1.60	-1.58	-0.27	—
300	-0.75	-0.81	-0.53	-0.41	650	-1.35	-1.16	-0.20	—
310	-0.53	-0.47	+0.18	+0.07	660	-1.52	-1.16	-0.28	—
320	-0.79	-0.82	-0.16	-0.04	670	-1.67	-1.40	-0.22	—
330	-1.60	-1.52	-0.68	-0.50	680	-1.75	-1.24	+0.05	—
340	-1.80	-1.68	-0.72	-0.38	690	-1.73	-1.13	-0.05	—

TABLE 25

Core A180-73

Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera
Top	-1.44	170	-0.38	340	-1.73*
10	-1.21	180	—	350	-1.80
20	-0.36	190	-0.30*	360	-0.82*
30	+0.51	200	—	370	—
40	+0.32	210	-1.06*	380	-0.36*
50	+0.31	220	-1.07*	390	-0.19*
60	+0.12*	230	-1.55*	400	-0.17*
70	+0.05	240	-1.26*	410	-0.14
80	-0.04	250	-0.88	420	-1.06*
90	-0.16	260	-0.95*	430	-0.10*
100	-0.59	270	-0.79	440	-1.20*
110	-0.29	280	-0.94	450	-0.63*
120	-0.30	290	—	460	-0.04*
130	-0.41	300	-0.90	470	-0.40*
140	-0.44	310	-1.11	480	-0.76*
150	-0.57	320	-1.24*	490	-0.80*
160	-0.42	330	-1.45*		

TABLE 26

Core 234A

Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera
0-1	-1.55	17-18	-0.80	31-32	-0.04
2-3	-1.68	19-20	-0.74	33-34	-0.10
4-5	-1.65	21-22	+0.08	35-36	-0.12
6-7	-1.67	23-24	+0.04	37-38	-0.03
8-9	-1.76	25-26	-0.05	39-40	-0.11
11-12	-1.33	27-28	+0.03	41-42	+0.01
13-14	-1.42	29-30	-0.18	43-43	-0.17
15-16	-1.00				

TABLE 27

Core 234

Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera
6.5	-0.68	233.5	-1.46	473.5	-0.88
13.5	-0.44	243.5	-1.22	485.0	-0.48
23.5	-0.08	253.5	-1.39	493.5 }	
33.5	-0.13	263.5	-1.67	502.5 }	-0.75
43.5	+0.15	273.5	-1.29	513.5	-1.19*
53.5	+0.04	283.5	-1.12	523.5	-1.29
63.5	-0.22	293.5	-1.27	533.5	-1.56
73.5	-0.37	303.5	-1.85	543.5	-1.41
83.5	-0.28	313.5	-1.98	554.5	-1.27
93.5	-0.46	323.5	-1.67	563.5	-1.43
105.0	-0.42	333.5	-0.71	573.5	-1.50
113.5	-0.58	343.5	-0.57	583.5	-1.00
123.5	-0.57	353.5	-0.82	593.5	-0.80
133.5	-0.66	363.5	-0.15	603.5	-1.05
143.5	-0.78	373.5	-0.21	613.5	-1.00
153.5	-0.87	383.5	-0.24*	624.5	-0.95*
163.5	-0.72	393.5	-0.24	633.5	-0.65
174.0	-0.85	403.5	-0.21	643.5	-0.79
183.5	-0.53*	413.5	-0.27	653.5	-0.86
194.5	-0.47*	433.5	-0.45	663.5	-0.94
203.5	-0.52*	443.5	-0.48	673.5	-1.14
213.5	-1.27	453.5	-1.21	683.5	—
223.5	-1.41	463.5	-1.02	693.5	—

TABLE 28

Core 235A

Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera
0-1	-1.83	24-25	-0.11	48-49	-0.37
2-3	-1.79	26-27	+0.10	50-51	-0.52
4-5	-1.90	28-29	-0.08	52-53	-0.62
6-7	-1.82	30-31	+0.03	54-55	-0.63
8-9	-1.79	32-33	+0.24	56-57	-0.72
10-11	-1.71	34-35	-0.33	58-59	-0.83
12-13	-1.60	36-37	-0.03	60-61	-0.86
14-15	-1.54	38-39	-0.05	62-63	-0.78
16-17	-1.72	40-41	-0.13	64-65	—
18-19	-1.17	42-43	-0.20	66-67	—
20-21	-0.86	44-45	-0.69	68-69	-0.77
22-23	-0.22	46-47	-0.49		

TABLE 29

Core 246

Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera	Depth below Top (Cm.)	Globigerinoides sacculifera
7.0	−1.61	303.0	−0.88	593.0	−0.66
23.0	−1.64	313.5	−1.19	603.0	−0.75
32.5	−1.28	323.5	−0.92	612.5	−0.96
42.0	−1.00	333.5	−1.30	623.0	−0.91
52.5	−0.57	344.5	−1.31	632.5	−0.95
63.0	−0.19	352.5	−1.10	643.0	−0.59
73.0	+0.02	363.0	−1.10	652.5	−0.65
83.0	+0.04	372.5	−0.99	664.5	−0.93
93.0	−0.09	384.0	−1.31	673.0	−1.13
104.0	−0.09	394.0	−1.50	683.0	−1.18
113.5	−0.02	402.5	−1.51	693.0	−1.19
123.0	−0.25	413.0	−1.37	703.0	−1.63
133.0	−0.35	421.0	−1.30	713.0	−1.84
144.0	−0.36	432.5	−1.50	722.5	−1.24
153.0	−0.38	443.0	−1.45	732.5	−1.31
163.0	−0.36	452.5	−1.79	742.5	−1.52
174.0	−0.36	463.0	−1.72	753.0	−1.50
182.5	−0.26	473.0	−1.21	764.0	−1.02
192.5	−0.44	483.0	−0.61	772.5	−0.83
203.0	−0.59	493.0	−0.60	783.0	−1.50
213.0	−0.52	503.0	−0.31	793.5	−0.74
223.0	−0.58	514.0	−0.23	803.0	−0.54
232.5	−0.64	525.0	−0.60	813.5	−0.65
243.0	−0.66	533.0	−0.31	823.0	−0.53
254.0	−0.69	542.5	−0.57	833.0	−1.08
263.0	−0.71	553.0	−0.51	842.5	−0.99
272.5	−0.50	563.5	−0.25	853.0	−1.37
283.0	−0.61	573.0	−0.23	862.5	−0.99
292.5	−0.69	582.5	−0.58		

TABLE 30

Core 280A

Depth below Top (Cm.)	Globigerina inflata	Depth below Top (Cm.)	Globigerina inflata	Depth below Top (Cm.)	Globigerina inflata
0–1	+0.64	16–17	+1.25	32–33	+1.78
2–3	+0.67	18–19	+1.43	34–35	+1.54
4–5	+0.63	20–21	+1.45	36–37	+1.63
6–7	+0.97	22–23	+1.61	38–39	+1.65
8–9	+0.68	24–25	+1.63	40–41	+1.52
10–11	+1.00	26–27	+1.76	42–43	+1.49
12–13	+0.72	28–29	+1.92		
14–15	—	30–31	+1.91		