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Isotopic Paleotemperatures

Urey's method of paleotemperature analysis has greatly contributed to our knowledge of past climates.

Cesare Emiliani

As is well known, the temperature dependence of oxygen-isotope fractionation in the system carbon dioxide-water-calcium carbonate was proposed by Urey (*1*) as a basis for determining the temperature of precipitation of the carbonate by measuring its oxygen-isotope composition. With a precision mass spectrometer of the type developed by Nier (*2*) and modified by McKinney and his co-workers (*3*), and with the acid extraction technique developed by McCrea (*4*) and Epstein *et al.* (*5*), it is possible to measure the O^{18}/O^{16} ratio in any calcium carbonate sample with an analytical error of $\pm$ 0.1 per mil. According to the relationship between temperature and isotopic composition of the carbonate found empirically by Epstein *et al.* (*6*), the corresponding error in the temperature value would be $\pm$ 0.5°C. However, the isotopic composition of the carbonate does not depend on temperature alone but depends also on the oxygen-isotopic composition of the water.

The oxygen-isotopic composition of surface water in the open seas varies from about 0 per mil (with respect to the Chicago PDB-1 standard) at the equator to about +1 per mil in the tropical evaporation belts, and to about −1 per mil in the higher latitudes. The total range (2 per mil) is equivalent to a temperature change of 9°C; calcium carbonate deposited from water having an isotopic composition of −1 per mil will give an apparent temperature 9°C higher than a carbonate deposited at the same temperature from water having an isotopic composition of +1 per mil. The uncertainty resulting from this effect can be eliminated only by measuring directly the isotopic composition of the water (this can be done only for the modern ocean) or by measuring the oxygen-isotopic composition of a compound co-precipitated in equilibrium with the water and the carbonate (a technique which is still in the experimental stage). The uncertainty, however, can be minimized if the oxygen-isotopic composition of the sea water can be estimated within reasonably narrow limits (*7*). This is best done for the more recent geological time (Late Cenozoic) and becomes progressively more difficult with increasing age of the sample. Thus, the oxygen-isotopic composition of Pleistocene, Tertiary, and even Mesozoic fossils may be reasonably interpreted in terms of temperature, while that of Paleozoic fossils (*8*) may not be.

Paleotemperature Analysis of Fossil Shells

The oxygen-isotope method of paleotemperature analysis has been applied extensively to Mesozoic belemnites and Cenozoic mollusks and foraminifera. Belemnites are extinct cephalopods with a thick calcium carbonate rostrum as part of their endoskeleton. Isotopic analysis of these rostra (*9–12*) revealed two successive major temperature cycles having an amplitude of several degrees Celsius and extending across about 60 million years of Mesozoic time. While the temperature trend across the Mesozoic-Cenozoic boundary remains unknown, for lack of fossils suitable for analysis, isotopic analysis of Tertiary mollusks (*13, 14*) showed additional major temperature fluctuations. Figure 1 shows the temperature trend during the Mesozoic and Tertiary for the middle northern and southern latitudes. As may be seen, the amplitude of the temperature fluctuations is several degrees Celsius, and the wavelength is some 20 million years. In addition, there seems to be an overall average temperature decrease amounting, again, to several degrees Celsius (dashed line in Fig. 1). While this decrease may result from latitudinal changes of the land masses from which the fossils were collected, it is comparable to a temperature decrease of about 12°C for the high latitudes (and the abyssal waters of the oceans) during the past 75 million years, deduced from paleotemperature analysis of Cretaceous belemnites from Alaska and Siberia (*10*) and of Tertiary benthonic foraminifera from the equatorial Pacific (*15, 16*) (Fig. 2). Thus, the data so far available for the Mesozoic and Tertiary, although still very limited, suggest that the average temperature of the earth's surface decreased 5° to 10°C during the past 150 million years. Isotopic corrections to account for the preferential removal of O^{18} from the ocean by precipitation of carbonate and silica would augment the temperature decrease by about 2°C.

A possible explanation for the inferred decrease in overall temperature is a decrease in solar emission. This possibility cannot be excluded, because, in spite of the assumed great secular stability of the sun as a nuclear engine, 150 million years does not represent an entirely negligible fraction of its probable life.

Variation in solar emission could also explain the temperature fluctuations shown in Fig. 1. However, there ap-

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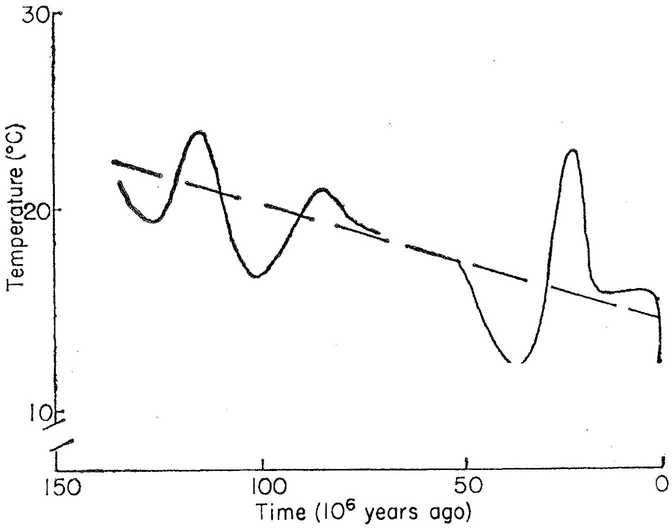

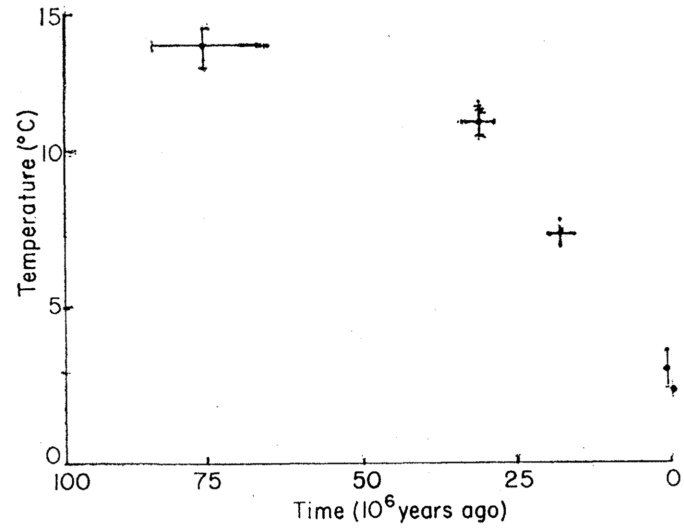

Fig. 1 (left). Temperature variations, indicated by isotopic analysis, for the middle northern and southern latitudes during the past 150 million years. [Data from *11, 12, 14*] Fig. 2 (right). Temperature decrease, indicated by isotopic analysis, for the high latitudes and the abyssal waters of the ocean during the past 75 million years. [From *16*]

pears to be a fair correlation between the times of occurrence of the temperature maxima and the times of occurrence of known, major marine transgressions. It would seem, therefore, that the observed temperature fluctuations could be accounted for by changes in the earth's albedo and, therefore, by terrestrial dynamics alone.

The most recent portion of the curve of Fig. 1 represents the Pleistocene. This epoch, characterized by the repeated occurrence of major glaciations, has been defined, by unanimous decision of the 7th INQUA Congress (Denver, Colorado, 1965), as the time that has elapsed since the first appearance of the benthonic foraminiferal species *Hyalinea (Anomalina) baltica* (Schroeter) in the late Cenozoic section at Le Castella, Calabria, southern Italy. The time of occurrence of this event is unknown at present, and, therefore, the duration of the Pleistocene is also unknown.

Shells of planktonic foraminiferal species have been used extensively in isotopic studies of Plio-Pleistocene paleotemperatures. In the low latitudes, two of the more common species, *Globigerinoides rubra* and *G. sacculifera*, deposit their shell material within 50 meters of the ocean surface (*17*). At these latitudes, therefore, isotopic analysis yields yearly average surface temperature. In the middle and high latitudes, where significant seasonal variations occur in the surface temperature of the ocean, the temperatures obtained by isotopic analysis of *G. rubra* and *G. sacculifera* are essentially summer averages.

Paleotemperature analysis of fossil shells of *Globigerinoides rubra* and *G. sacculifera* collected at close stratigraphic intervals from the section at Le Castella revealed marked temperature oscillations and an overall decrease in temperature from the Pliocene into the Pleistocene (*18*) (Fig. 3). The Pliocene minima are lower by several degrees Celsius than modern summer temperatures off Le Castella, but they are still considerably higher than the temperature minima of the more recent Pleistocene (see *19*). The occurrence of cold episodes during the late Pliocene is also demonstrated by the presence of pollen belonging to cold plants in deposits below the Pliocene Amphistegina limestone at Castell'Arquato, northern Italy (*20*) and by oxygen-isotope analysis of fossil shells of the molluscan species *Arctica (Cyprina) islandica* (Lamarck) from deposits below the Pliocene Amphistegina limestone at Castellanselmo, Pisa, Tuscany (*21*).

Analysis of Deep-Sea Cores

The temperature oscillations which occurred during the past million years have been studied by means of isotopic, chemical, and micropaleontological analysis of deep-sea cores of globigerina-ooze facies. These analyses, together with carbon-14 and protactinium-231/thorium-230 age measurements, revealed quasi-sinusoidal temperature oscillations having an apparent amplitude of 9° to 10°C (uncorrected for glacial-interglacial changes in the isotopic composition of sea water) in the Caribbean and the equatorial Atlantic, and an average wavelength of 50,000 years (*22–26*) (Fig. 4). Results for all Atlantic and Caribbean cores so far analyzed isotopically were cross-correlated by different criteria, in an effort to detect and account for stratigraphic discontinuities, and an "average core" was reconstructed by averaging, over all cores, the stratigraphic thicknesses of each successive temperature stage (*24*). A generalized curve for temperature relative to time was then reconstructed, based on the O^{18}/O^{16} data (corrected for the glacial-interglacial changes in the isotopic composition of sea water), together with the C^{14} and Pa^{231}/Th^{230} age measurements and reasonable extrapolations therefrom (Fig. 5). This curve, which extends from the present to about 425,000 years ago, shows eight temperature cycles which, with the single exception of the cycle extending from 65,000 to 15,000 years ago, have approximately the same amplitude.

The stratigraphically longest deep-sea cores of globigerina-ooze facies so far available are part of a suite of five cores (Nos. 58, 59, 60, 61, and 62) collected in the eastern equatorial Pacific by the Swedish Deep-Sea Expedition of 1947–1948. These cores were studied in considerable detail by Arrhenius (*27*) and, in part, analyzed by the oxygen-isotope method (*22*). Some of the cores (Nos. 59, 60, and 61) exhibit very clear and well-defined oscillations in the percentages of carbonate, while others (Nos. 58 and 62) do not (in part because of postdepositional solution). While the shallower pelagic foraminiferal species are insufficiently abundant in these cores for isotopic analysis, analysis of the deeper subspecies *Globorotalia menardii tumida* confirmed the inverse relationship between carbonate content and temperature predicted by Arrhenius. Thus, each carbonate cycle in the Pacific cores is likely to be the equivalent of a temperature cycle in the Atlantic and Caribbean cores. As a result, cores 59, 60, and 61, containing, respectively, nine, eight and five carbonate cycles (*28*) reach sediments which should be, respectively, 440,000, 400,000, and 260,000 years old.

Cores 59, 60, and 61 can be easily correlated with each other by means of the size distribution of the diatom species *Coscinodiscus nodulifer* (Fig. 6); and the bottom layer of core 59 can be correlated with the layer at 400 centimeters below the top in the stratigraphically longer core 58, from the percentages of the foraminiferal species *Pulleniatina obliquiloculata* and *Sphaeroidinella dehiscens* + *Sphaeroidina bulloides* (Fig. 7). The rate of accumulation of TiO_2 in core 58, when averaged across several tens of thousands of years, remains almost constant (*27*, plate 2.58), indicating an almost constant rate of bulk sedimentation. As a result, the age of about 440,000 years for the layer at 400 centimeters below the top may be used for estimating an age of about 1.1 million years for the bottom of the core (991 centimeters below the top).

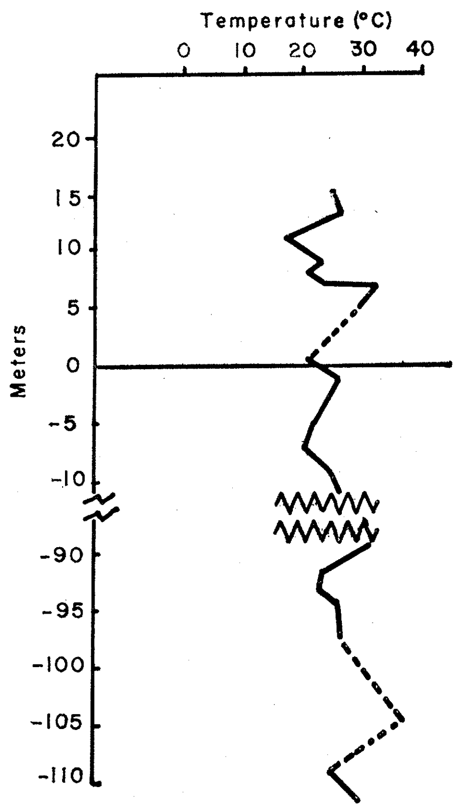

Fig. 3 (left). Temperatures, indicated by isotopic analysis, across the Plio-Pleistocene boundary (0 meters) at Le Castella, Calabria, southern Italy. [Data from *18*]

The carbonate cycles continue through core 58, and, in fact, one of the highest carbonate peaks, presumably representing a strong glaciation, occurs near the bottom of the core, at an estimated date of about one million years ago. Oxygen-isotope analysis of the core, however, shows average temperatures higher and amplitudes of the oscillations smaller in the lower portion of the core than in the upper portion (Fig. 8). It is not known whether or not core 58 crosses the Plio-Pleistocene boundary, because the age of this boundary is unknown and paleontological correlations cannot yet be made between core 58 and the section at Le Castella.

Glacial, Interglacial, and Nonglacial Sea Water

In order to translate the temperature values obtained through isotopic analysis into the approximately true temperatures of Fig. 5, a correction has been applied for the amount and the isotopic composition of (i) the sea water removed during glacial ages and (ii) the sea water added, in excess of the present amount, during interglacial ages (*24*). The average isotopic composition of this water was estimated at −15 per mil (*22, 24*), a

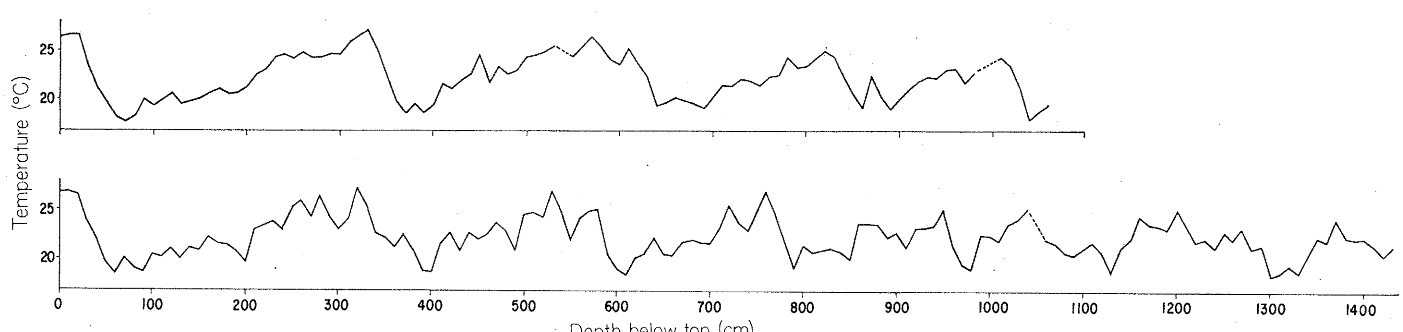

Fig. 4. Temperatures indicated by isotopic analysis of two Caribbean deep-sea cores: (top) P6304–8; (bottom) P6304–9. [Data from *24*]

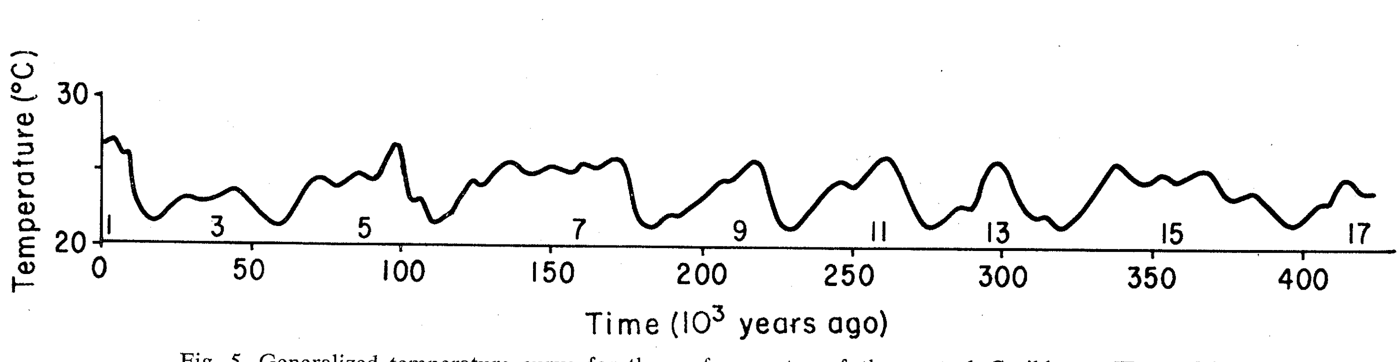

Fig. 5. Generalized temperature curve for the surface water of the central Caribbean. [From *24*]

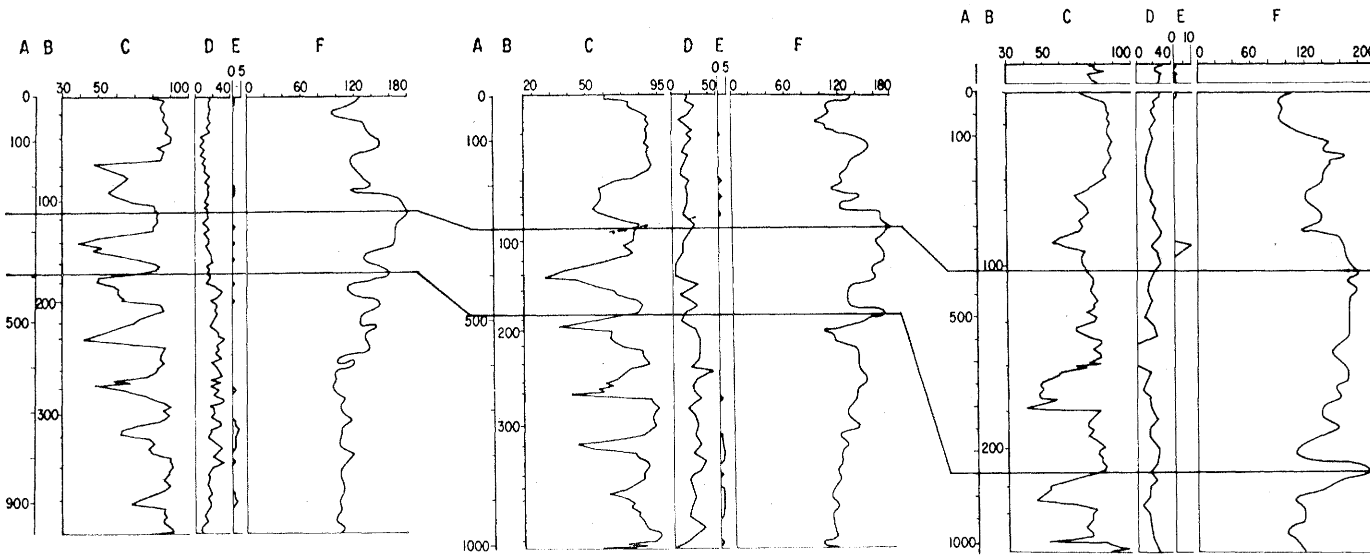

Fig. 6. Correlation among three cores from the eastern equatorial Pacific: (left) core 59; (center) core 60; (right) core 61. *A*, Depth below top (in centimeters); *B*, cumulative amount of TiO_2 (in milligrams per square centimeter); *C*, percentage of $CaCO_3$; *D*, relative abundance of *Pulleniatina obliquiloculata; E*, relative abundance of *Sphaeroidinella dehiscens* + *Sphaeroidina bulloides; F*, size distribution of *Coscinodiscus nodulifer*. [Data from *27*]

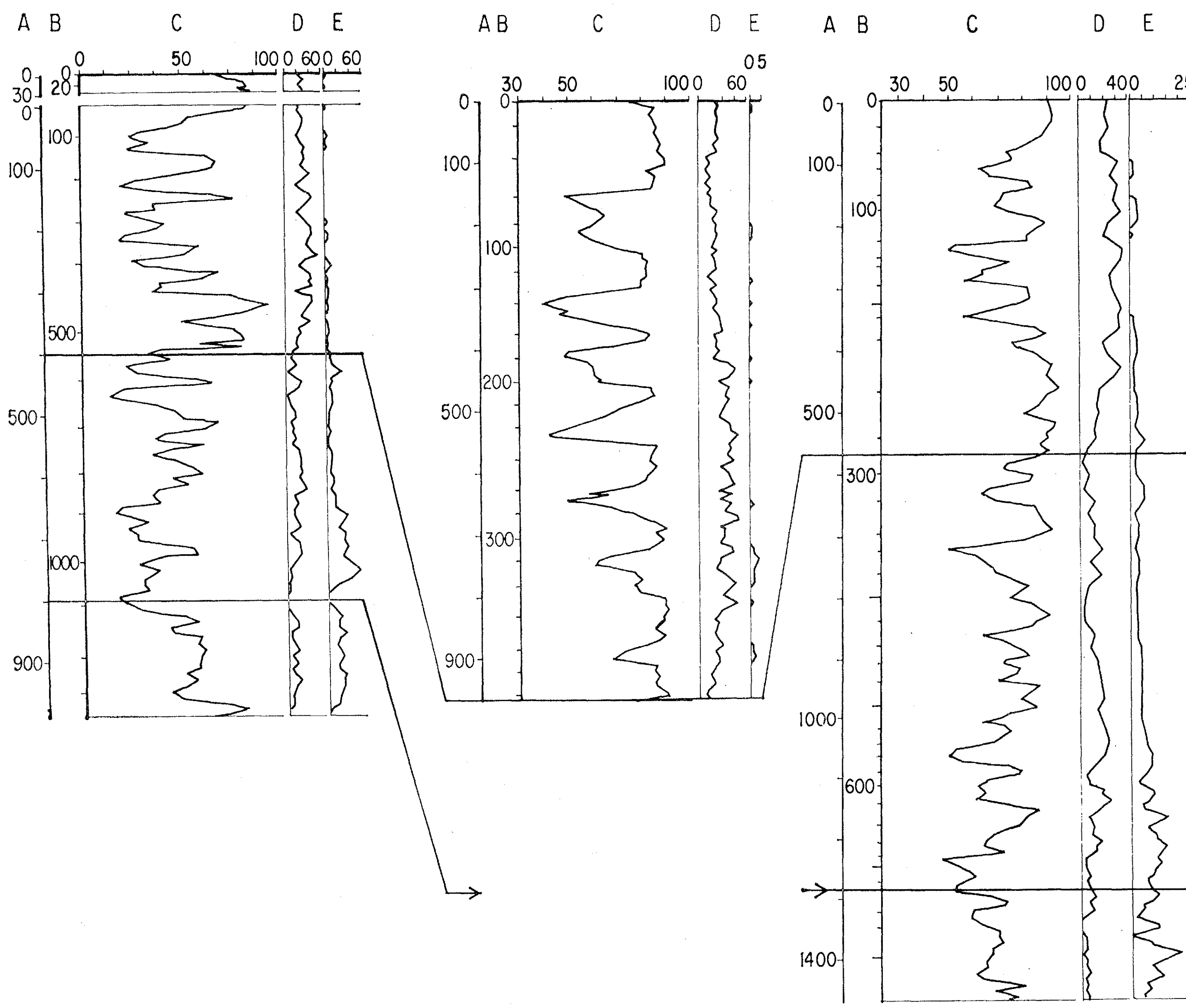

Fig. 7. Correlation among three cores from the eastern equatorial Pacific: (left) core 58; (center) core 59; (right) core 62. *A, B, C, D, E*, and *F* as in Fig. 6 [Data from *27*]

value which is in close agreement with Craig's estimate of −17 per mil (*7*). The volume of sea water removed during glacial ages was calculated on the basis of a probable decrease in sea level of 130 meters, and the volume of sea water added during interglacial ages was calculated on the basis of a probable increase in sea level of 30 meters (*24*).

These values were chosen because they are widely reported in the literature (*29*) and because they are, if anything, excessive. Using these figures, one obtains a total glacial-interglacial volume change of 57.7×10^6 cubic kilometers. Craig, using estimates by various authors of the volume of the Pleistocene ice sheets, suggests a glacial-interglacial volume change of 100×10^6 cubic kilometers for the ocean water. The estimates used, however, can easily be in error. Some convincing experimental evidence bearing on the amount of lowering of sea level during glacial ages has been recently obtained by Hoyt *et al.* (*30*), along a vast tract of the western coast of South Africa. Using an electrosonic sub-bottom profiler, these workers found strikingly consistent evidence, over a 650-kilometer-long tract, that Pleistocene sea level decreased about 100 meters. This is very close to an early estimate made by me (*22*) and to the result recently obtained by Markov and Suetova (*31*).

Table 1. Ages of continental stages (Late, Middle, and Early Wisconsin by C^{14} dating; Pelukian, Kotzebuan, and Middletonian, by Th^{230}/U^{238} dating) and ages of the corresponding temperature stages of the deep-sea cores (by C^{14} and Pa^{231}/Th^{230} dating). Ages of oceanic stages 5, 7, and 9 refer to the temperature maxima. [Data from *22, 24–26, 35, 40*].

Continental stages		Oceanic stages	
Name	Age (number of years ago)	No.	Age (number of years ago)
Late Wisconsin	10,000–25,000	2	10,000–25,000
Middle Wisconsin (Port Talbot–Plum Point, Worozonfian)	25,000–55,000	3	25,000–55,000
Early Wisconsin	55,000–70,000	4	55,000–70,000
Pelukian	100,000	5	96,500
Kotzebuan	172,500	7	171,500
Middletonian	217,500	9	218,000

As to high interglacial stands of sea level, the detailed field analysis of Bonifay and Mars (*32*) indicates that interglacial sea level did not rise, eustatically, higher than 10 meters above the present level. Thus, a total glacial-interglacial volume change of about 40×10^6 cubic kilometers is indicated. The concomitant change in the average isotopic composition of sea water is 0.43 per mil (for the construction of the curve shown in Fig. 5, a value of 0.5 per mil was chosen), indicating that more than 70 percent of the glacial-interglacial range in the isotopic ratio of the foraminiferal shells is due to temperature alone. The occurrence of significant temperature variations in the tropics is also demonstrated by the marked variations in the foraminiferal faunas (*33*, Figs. 6 and 7; *34*).

Correlation of the Oceanic and Continental Records

The temperature oscillations which occurred during the past 425,000 years, illustrated by the generalized temperature curve of Fig. 5, are established rather firmly with respect to both amplitude and age. Accurate comparisons and correlations of this curve with the glacial and interglacial events of the continents are assured whenever abso-

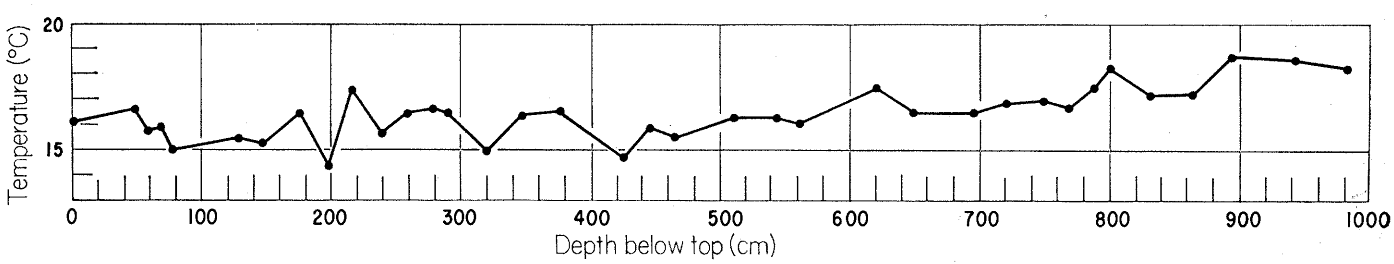

Fig. 8. Core 58; temperatures obtained from *Globorotalia menardii tumida*, through isotopic analysis. [Data from *22*]

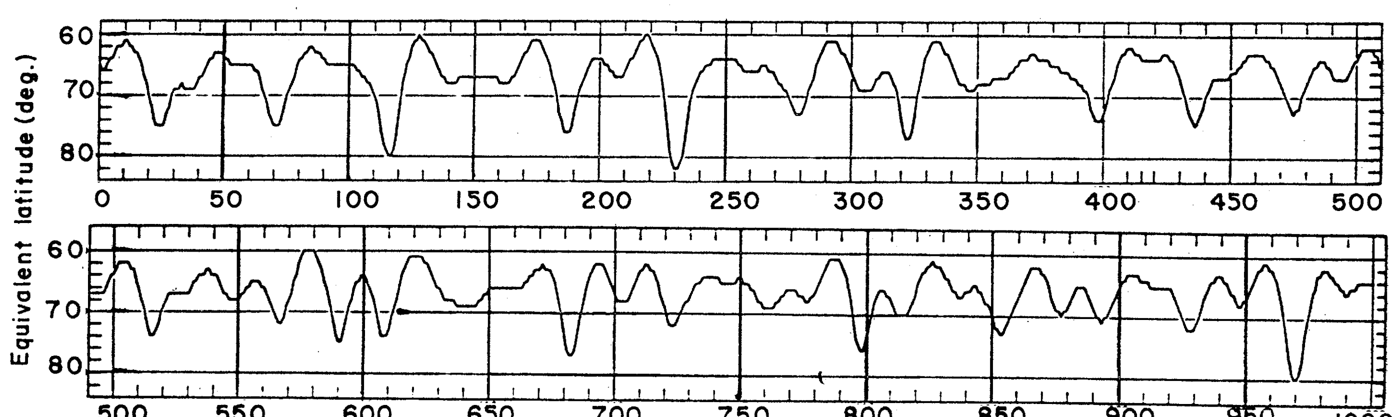

Fig. 9. Variation of summer insolation at 65°N during the past million years (in degrees of equivalent latitude). [From *22*, based on data calculated by D. Brouwer and A. J. J. van Woerkom]

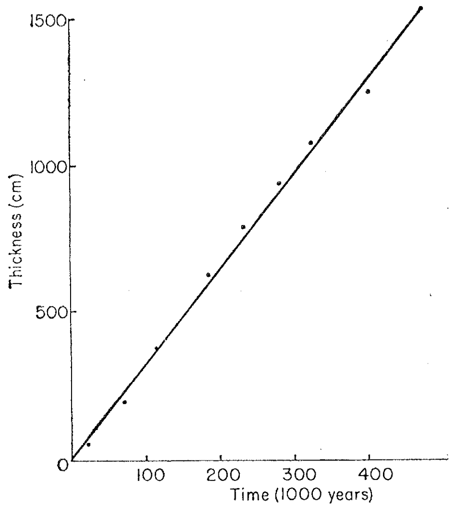

Fig. 10. Average stratigraphic position of temperature minima in globigerina-ooze cores from the eastern Caribbean and equatorial Atlantic plotted against ages of summer insolation minima at 65°N. [From *24*]

lute ages of sufficient precision are available for the continental events. Thus, the continental stages known as Late, Middle, and Early Würm [or Late, Middle, and Early Wisconsin (*35*)] have been shown, by C^{14} dating, to correlate with the oceanic stages 2, 3, and 4; and the Alaskan stages Worozonfian, Pelukian, Kotzebuan, and Middletonian have been shown, by Th^{230}/U^{238} dating, to correlate with the oceanic stages 3, 5, 7, and 9 (Table 1). In part because of sampling uncertainties, the ages so far obtained by K^{40}/Ar^{40} dating of materials from glacial and periglacial areas (see *36*, Table 8) are generally not sufficiently accurate to allow unequivocal correlation with the oceanic stages. However, recent analyses by Frechen and Lippolt (*37*), which assign an age of 350,000 to 400,000 years to the Günz (compare also *24* and *25*), indicate that the temperature curve shown in Fig. 5 represents the entire glacial Pleistocene as classically understood.

In view of the fact that the continental deposits which have been dated accurately appear to correlate very closely with the oceanic temperature stages, there is little doubt that close correlations will also be obtained for the continental deposits which have not yet been dated, or which have not been dated with sufficient accuracy. In this context, it appears almost certain that the simple scheme of four or five major glaciations, still widely accepted, will have to be abandoned. In its place is apparently emerging a picture of many alternating high- and low-temperature stages. At least some of these are probably the "interstadials" within the classic glacial and interglacial stages of the Pleistocene.

Causes of Glaciation

The study of the oceanic record has provided significant evidence toward solution of the problem of the cause of the ice ages. When the first cores were analyzed by the oxygen-isotope method, an apparent periodicity of 40,000 to 50,000 years was noted, a periodicity similar to that of the variations of summer insolation in the high latitudes (Fig. 9) (*22*). Now that additional cores have been analyzed, statistical treatment has become feasible. With this approach, it was found that the correlation coefficient between the astronomically calculated ages of the summer insolation minima at 65°N and the stratigraphic position of the temperature minima in the "average core" previously mentioned is 0.997 (Fig. 10). Although the correlation coefficient refers to two incremental sequences, a causal relationship is strongly suggested (*24*).

Variations in summer insolation in the high northern latitudes have been used by many workers, during the past 100 years, as a basis for theories of glaciation. The modern treatment by Emiliani and Geiss (*38*) provides a plausible explanation for the inception of glaciation in the late Cenozoic and the occurrence and timing of the glacial and interglacial events of the Pleistocene. In this treatment, in fact, it was suggested that the inception of glaciation was made possible by Tertiary cooling caused by an albedo increase associated with increased continentality and the Alpine orogenesis. It was further suggested that the successive Pleistocene glaciations were started by quasi-periodic decreases of summer insolation in the high northern latitudes but proceeded largely as a self-sustaining process; and that complete deglaciation outside Antarctica and Greenland resulted from surface freezing of the northern North Atlantic, with plastic flow of glaciers, heat absorption by ice melting, and crustal warping providing the necessary impedance. More recently it was noticed that, while variations in summer insolation at high northern latitudes apparently modulate the frequency of the Pleistocene temperature oscillations, insolation variations in the northern evaporation belt (about 25°N) may determine the amplitude (cf. *39*). Thus, it is to be expected that high insolation at 25°N coupled with low insolation at 65°N will result in a strong glaciation; and that low insolation at 25°N coupled with high insolation at 65°N will result in a warm interglacial. Intermediate conditions will produce milder glacials and interglacials.

If this treatment is correct, it is to be expected that glacial and interglacial ages will continue to alternate in the future for some millions of years, or until erosion of the modern high relief and marine ingression have sufficiently reduced the earth's albedo to bring glaciation to an end. Meanwhile, it is to be expected that a new glaciation will begin within a few thousand years and reach its peak about 15,000 years from now.

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Three-Dimensional Map Construction

Use of the crossed-slit anamorphoser simplifies the technique and shortens construction time.

George F. Jenks and Dwight A. Brown

Three-dimensional maps are perspective representations of obliquely viewed statistical or topographic surfaces. Many readers are familiar with isometric topographic or geologic block diagrams, but these comprise only a fraction of the three-dimensional maps which have been, or could be, made. Maps showing some of the potential of this type of illustrative device are shown in Fig. 1. Why not, however, use maps of this kind to illustrate submarine or fish environments; airport facilities and landing obstacles; the development of hurricanes; intersecting statistical surfaces; the movement of fog, smog, and other air pollutants; and any other areal phenomenon which can be thought of as a three-dimensional surface?

At the turn of the century, block diagrams were common in geologic and geographic literature, but they are not often used today. Why have such useful illustrations been, in large part, discarded by modern authors? The answer seems to lie in both the economic fact and the mistaken belief that preparation of such maps requires special training. True, conventional three-dimensional map construction is so time-consuming that the cost of individual illustrations has become prohibitive, and it requires considerable skill on the part of the draftsman. However, the construction of three-dimensional maps need be neither costly nor overly demanding of artistic skill, as we hope to show. More specifically, this article contains a brief statement on the basics of map perspective, a section in which a unique construction technique is introduced, and discussion of a new method of optical transformation of map coordinates. Furthermore, the possibility of using the transformation technique in fields other than cartography is discussed.

The techniques introduced here are not based upon computer-driven plotout devices, but mention should be made of experimental work in this area. Computerized perspective illustrations have been developed to a high degree by the computer and aerospace industries, which have produced three-dimensional views of aircraft and aircraft parts, rendezvous, cosmic radiation belts, and so on, to bridge the information gap between the scientist and engineer and the nonscientist (*1*). It is quite apparent that computer technology will be refined to the point where many kinds of maps will be produced by the computer in final drafted form. Unfortunately, however, the time lag between what is to be and what is now technologically or economically feasible may be considerable. During this interim period the need for three-dimensional maps will have to be met by other means, such as those presented here.

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Basics of Three-Dimensional Map Perspective

Three-dimensional maps present earth surfaces as viewed from a real or assumed position in space. The viewing point can be any compass position, at any angle above or below the horizon (*2*). The map image that is presented for any viewing point is theoretically formed on a transparent perspective plane. This plane is perpendicular to the central line of vision and lies at a point between the eye of the viewer and the earth surface being mapped. In cartographic presentation the perspective map image, or transformed map, is constructed geometrically, but it can be visualized as being formed by converging projectors (in angular-perspective presentation) or by parallel projectors (in parallel-perspective presentation) (see Fig. 2).

Three-dimensional maps are conceptualized views of what the distributive surface *might look like,* and they cannot be regarded as oblique photographs of earth models. As a result, it is best to think of the process of construction as that of proceeding from a planimetric map to a transformed or foreshortened view of this flat map. The steps in construction are illustrated in Fig. 3. The planimetric map (Fig. 3A) is transformed by angular perspective to Fig. 3B, and then the various features are elevated on the vertical axis (Fig. 3D). In Fig. 3C the same map (Fig. 3A) is transformed by parallel perspective (*3*), and in Fig. 3E it is elevated, with a vertical scale equal to that used in the foreground of Fig. 3D.

The three-dimensional maps of Fig. 3 illustrate some of the differences between angular- and parallel-perspective presentations of earth surfaces. In terms of visual characteristics the angular-perspective map (Fig. 3D) is superior because the convergence cues contribute to the illusion of depth without detracting from the overall configuration (*4*). The parallel-perspective map (Fig. 3E) appears to be a contrived version of the