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Late-Pleistocene Climates and Deep-Sea Sediments

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The shells of planktonic Foraminifera in deep-sea sediment cores provide the most trustworthy evidence on climatic variations of the Pleistocene because they vary in accordance with the temperature changes in the surface waters of the ocean and because the stages are represented in chronological sequence (*1–4*). The study of the vertical variations in frequencies of the most temperature-sensitive species of planktonic Foraminifera in hundreds of cores at Lamont Geological Observatory has shown that a rather sharp faunal change in the sediments of the major part of the Atlantic Ocean and adjacent seas marks a transition from a relatively cold to the present relatively mild oceanic climate. In order to correlate this change with continental events that occurred at the end of the last ice age, carbon-14 dates were made on a series of deep-sea cores representing different localities, sedimentation rates, and sediment types. The locations, geographical positions, and depths of water at the stations where the cores were obtained are shown in Fig. 1.

Several considerations determined the choice of these particular cores. Wide separation in longitude and latitude was desirable in order to make clear that the faunal variations from which the climatic changes are inferred are not due to local but to ocean-wide conditions and in order to detect any possible variations with latitude. The distance between the core from the most westerly station, A179-15, and the core from the most easterly station, A180-48, is more than 6000 kilometers. The south-north range, the distance between core A180-74 and core R10-10, is about 4500 kilometers. Cores from areas of varied bottom topography were chosen, for it is recognized that the configuration of the ocean floor strongly influences sedimentary processes, and particularly the rate of sediment accumulation (*3, 5*). The locations include a submarine canyon, core A180-48, a basin, R10-10, and the crest of a ridge, A172-6. A description of the topography at the core stations and the lithology of the cores is included as an appendix.

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Climate Determination

Since nearly all the species of Foraminifera which are found in abundance in the Pleistocene sediments of the Atlantic Ocean are still living in the Atlantic, it is possible to chart the present areal distributions of the various species either by collecting living assemblages with plankton nets or by noting the distribution of species in samples of the uppermost layer of sediment. Charts of the present areal distributions of the important planktonic species have been published by W. Schott (*1*) and by Phleger, Parker, and Pierson (*6*), and additional, as yet unpublished data have been collected by workers at Lamont Geological Observatory. The areal distributions indicate that some species are particularly sensitive to temperature.

Vertical variations in frequencies of such stenothermal species in sediment cores are the basis of the micropaleontological method of interpreting the climatic record. W. Schott (*1*) was the first to apply this method. He found in a series of short sediment cores from the equatorial Atlantic a well-defined faunal change which he interpreted as marking the end of the Last glaciation.

The species of most significance as climatic indicators are *Globorotalia menardii menardii* (d'Orbigny), *Globorotalia menardii flexuosa* (Koch), and *Globorotalia menardii tumida* (H. B. Brady), which are particularly characteristic of warm water, and *Globorotalia hirsuta* (d'Orbigny), *Globorotalia scitula* (H. B. Brady) and, especially, *Globigerina inflata* d'Orbigny, which are indicative of relatively cold water.

In drawing the climatic curves shown in Figs. 2 and 3, we plotted the top sample by convention on the mid-point of the column. Lower samples containing *Globigerina inflata* in abundance, but not *Globorotalia menardii* are plotted near the right margin of the column. Such assemblages are interpreted as indicating the extreme climate of a glacial age. Intermediate assemblages are plotted somewhere in between, depending on the degree of dominance of warm and cold water species. In many cores about halfway down in the Recent zone, there is a somewhat stronger dominance of *Globorotalia menardii* than in the top sample. This is taken to indicate that at that time the climate was a trifle warmer than it is now. A more complete discussion of these methods has been given by Ericson and Wollin (*4*).

A check of the validity of the method is provided by the recently published paleotemperature data obtained at the University of Chicago (*7*) on three cores from the Lamont collection. Paleotemperatures are obtained by relating variations in the oxygen-18/oxygen-16 ratio in the $CaCO_3$ of the Foraminifera shells to temperature of the water in which the deposition occurred. The theory and technique of this method have been discussed by Urey (*8*), Epstein *et al.* (*9*), and Epstein and Mayeda (*10*).

The cores on which this comparison was made were carefully selected on the basis of micropaleontological, lithological, calcium carbonate, and size-fraction analyses by Ericson and Wollin (*4*). The oxygen istotope measurements made on tests of *Globigerinoides sacculifera* (H. B. Brady) and *Globigerinoides rubra* (d'Orbigny) indicated that the climatic change involved a 6°C change in the temperature of surface ocean water (*7*). Fig. 2 illustrates that a comparison be-

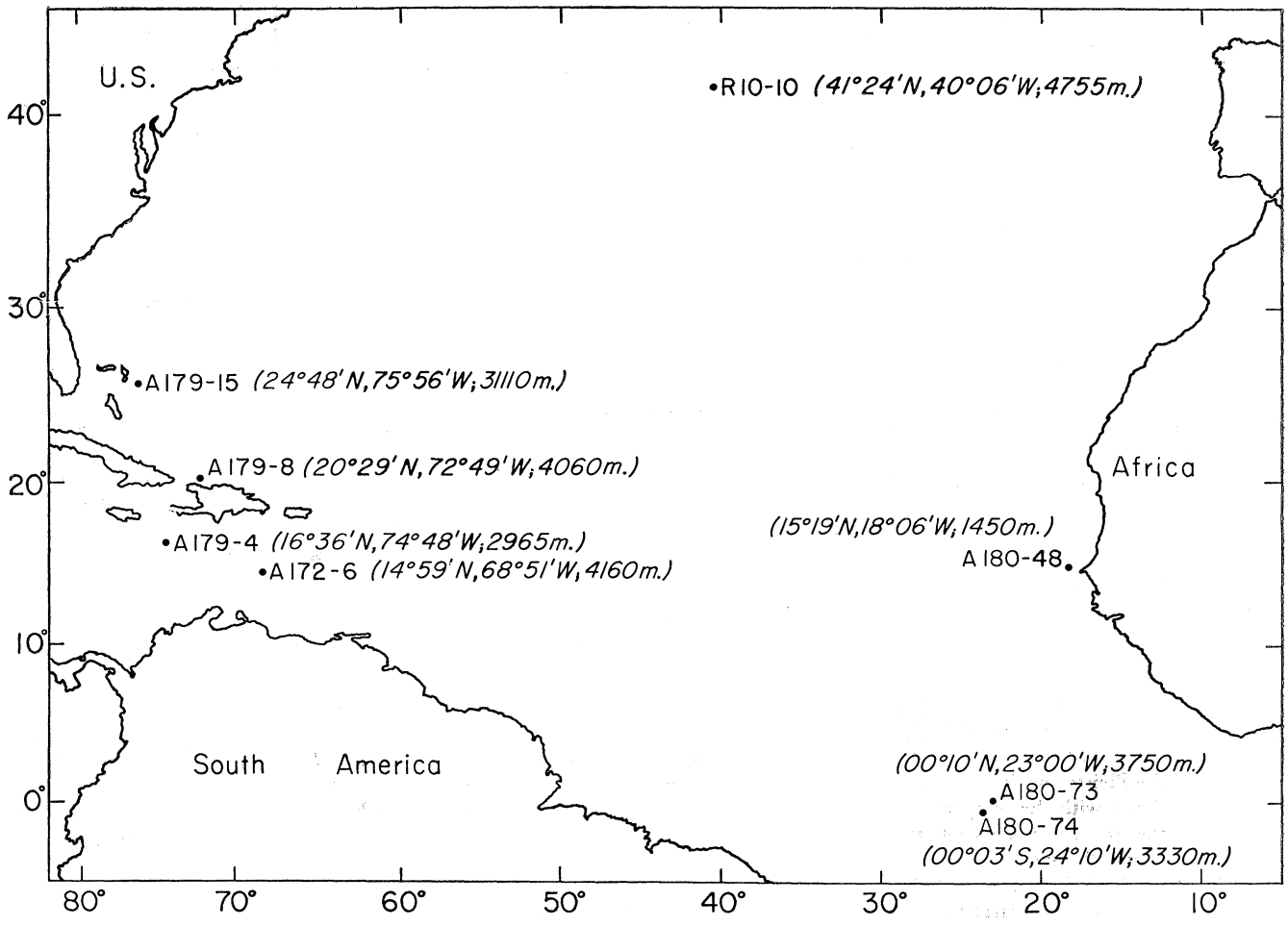

Fig. 1. Locations of and depths (in meters) of water at the core stations.

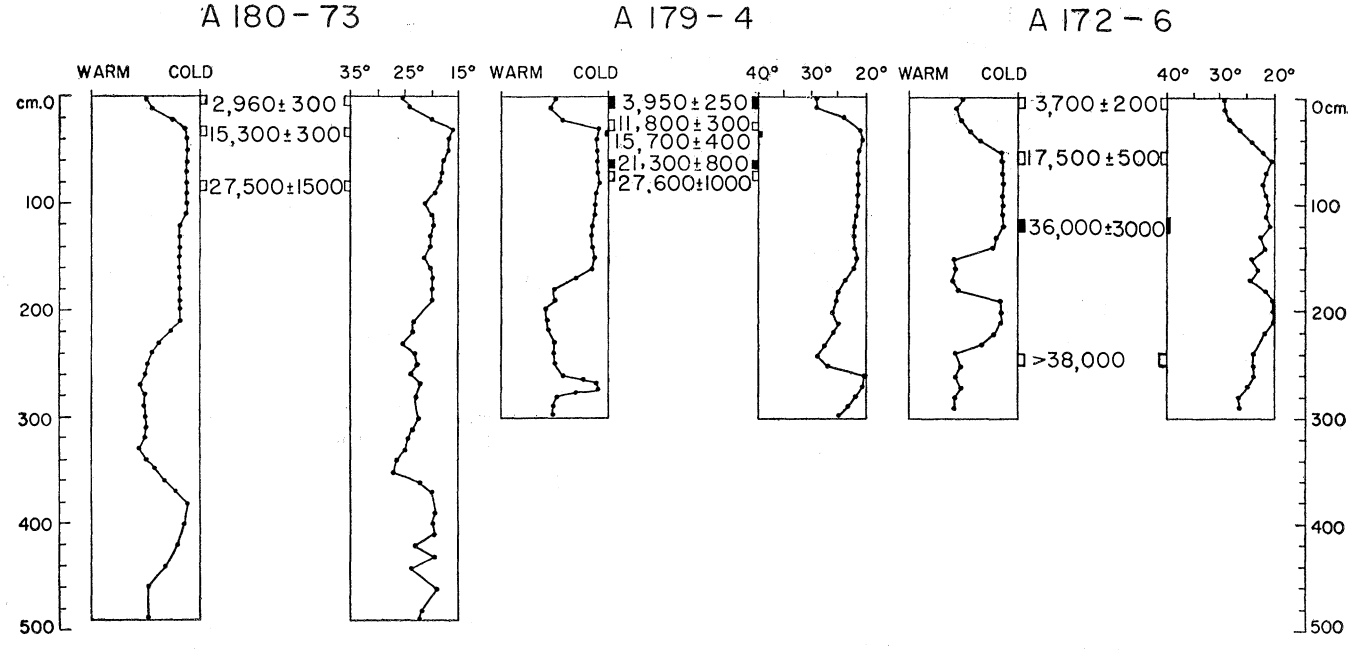

Fig. 2. Comparison of climatic curves derived from variations in Foraminifera by Ericson and Wollin (*4*) with oxygen isotope paleotemperature curves for *Globigerinoides sacculifera* by Emiliani (*7*) for three deep-sea cores. The change, near the top of the curves, from cold to warm indicates the end of the last ice age. The scales are in centimeters of core lengths. Between each pair of curves, radiocarbon ages (in years) by Rubin and Suess (*11, 12*) are given with boxes indicating the sections of the cores used for dating. Where total carbonate was used, areas are blackened; open areas represent measurement of only the coarse fraction (greater than 74 microns).

tween the isotopic temperatures and the curves derived from variations in planktonic Foraminifera shows good correlation for the part of the curves which represent the change from the cold climate of the last ice age to the relatively mild climate of the present time.

Age Determination

Several assumptions must be made in the calculation of carbon-14 ages on core materials. One of these is the choice of the carbon-14 concentration of the material at the time of deposition. For shell, the value should be equal to the concentration in the carbonate and bicarbonate of the ocean water in which the shell grew. Recent data obtained at Lamont indicate that surface ocean water has a value about 3 percent greater than contemporary wood and that deep ocean water has values that range down to at least 10 percent lower than the same wood. In view of this range, a value equal to modern wood was arbitrarily chosen. In any event, the error involved should be less than 300 years.

The possibility of exchange or recrystalization of the carbonate after deposition must also be considered. Since ages as great as 35,000 years have been obtained on core material, it appears that this effect should be negligible in the 11,000-year range under consideration in this article.

The most serious possibility of error is the incorporation of carbonate from preexisting sediment. Since this material would have a lower carbon-14 content than the normal material, the ages obtained would be anomalously low. That this is certainly a real problem, at least in Caribbean cores, has been demonstrated by Rubin and Suess (*11*), who found that the fine fraction from a sample of core A179-4 gave an age of 1300 years, or 10 percent older age than the coarse material. Another set of measurements made at Lamont on another Caribbean core, A172-2, is in agreement with this difference.

One might suspect that, for cores obtained from ridges well separated from the continents, the lack of a source of such material would insure more reliability. One set of measurements made on a sample from an equatorial core, A180-76, which fits the afore-mentioned requirements, did indeed show agreement between the fractions. The results of the coarse- and fine-fraction measurements are given in Table 1.

In four of the cores discussed in this article, the amount of coarse fraction (particles greater than 74 microns in diameter), which is composed almost entirely of the shells of planktonic organisms, was insufficient for checks to be made. The fifth core, as mentioned in the preceding paragraph, was taken from an ideal area, and no difference between the fractions was noted. An added confirmation of this point is seen in comparing the dates obtained at the U.S. Geological Survey (*12*) on the coarse fraction of core A180-73 with dates obtained at Lamont on the bulk material from a lithologically identical core A180-74. This comparison is shown in Fig. 3 with the other carbon-14 dates obtained at Lamont Geological Observatory.

The excellent agreement concerning the date of the mid-point of the climatic change in the five cores gives confidence that the problem of incorporating dead material has caused less-than-1000-year errors in the dates obtained. In any case, the ages obtained on bulk material must be considered to be maximum.

Summary and Discussion

The radiocarbon dates, the climatic curve deduced from the Foraminifera, and the paleotemperature measurements shown in Figs. 2 and 3 indicate that the mid-point of the the major change from glacial to postglacial conditions occurred about 11,000 years before the present. This major change appears to be broadly simultaneous throughout the North Atlantic and adjacent seas, and it probably occurred in a time interval on the order of 1000 years. The Caribbean cores (A172-6 and A179-4) and the equatorial cores (A180-73 and A180-74) show evidence of a gradual beginning of climatic change 13,000 to 15,000 years ago, with the major break occurring close to 11,000 years ago. In other cores where sedimentation was more rapid, the change is somewhat sharper.

On the basis of the carbon-14 dates shown in Fig. 2, Emiliani concluded that the major climatic change began 16,500 years ago (*7*). As can be seen by examining the curves, the beginning of the climatic change is not well defined. In cores showing rapid deposition, the change occurs in less than 2000 years, whereas in the cores showing a slow deposition rate, it is spread out from 2000 to as much as 10,000 years (in core A172-6). The spreading out of the break in core A172-6 may represent a more rapid deposition at the close of the glacial period. The increase in deposition rate may well represent the addition of reworked material to the normal sediment, in which case the age obtained is too great.

A more easily defined point, and perhaps a more significant one, is the mid-point of the steep portion of the curve. Examination of the curves shows that, in all cases except possibly core A172-6, the age at this point lies within 1000 years of 11,000 years ago. It should be noted that the isotope temperature curves as well as the foraminiferal climatic curves show this apparent variation in the shape of the climatic change. This eliminates the possibility that the explanation lies in an error in the foraminiferal method itself.

The oxygen paleotemperature curves and the climatic curves based on variations in frequencies of species of Foraminifera (Fig. 2) agree in showing that the transition from glacial to postglacial climate was marked by a pronounced temperature rise of the upper layer of water in which the planktonic Foraminifera live. This world-wide change of surface-water temperature at the end of the glacial period argues against any theory of continental glaciation which calls on increased precipitation only without lowering of the world mean annual temperature.

Although the late Wisconsin glacial maximum may have occurred 18,000 years ago (*13*), it is quite likely that rapid deglaciation did not set in until after the Mankato readvance 11,400 years ago. The difference in the total ice accumulated on the continents between the Tazewell (18,000 years before the present) and Mankato (11,400 years before the present) advances probably did not exceed 10 percent. The cores indicate that no large climatic change occurred

Table 1. Radiocarbon ages of the coarse and fine fractions from one deep-sea core, A179-4, determined at the U.S. Geological Survey (*12*) and from two cores determined at Lamont Geological Observatory (*15*). The coarse fraction (greater than 74 microns) consists almost entirely of the shells of planktonic organisms. The positions and depths are as follows. A179-4: 16°36′N, 74°48′W, 2965 meters; A172-2: 16°12′N, 72°19′W, 3070 meters; A180-76: 00°46′S, 26°02′W, 3510 meters.

Core No.	Sample position (centimeters from top)	Age of coarse fraction (years)	Age of fine fraction (years)
A179-4	23–30	11,800 ± 300	13,500 ± 400
A172-2	14–29	11,500 ± 250	14,400 ± 400
A180-76	10–22	9,800 ± 200	10,300 ± 350

over the North Atlantic Ocean between 18,000 and at least 13,000 years before the present. The data indicate a rather sudden change from more or less stable glacial conditions to postglacial conditions. This is opposed to the usual view of a gradual change.

The 11,000-year date for the major rise in the temperature of ocean water is consistent in that it slightly postdates the last ice advance on land at Two Creeks, Wis., 11,400 years ago (*14*) and in the Puget Sound and southern British Columbia areas 11,000 to 12,000 years ago (*15*), which was followed by the rapid retreat. Further evidence is found in a change in sedimentation of sand to silt in the Mississippi Delta region, and a rapid drying up of pluvial Lake Lahontan (*16*), both very close to 11,000 years ago.

The core data point definitely to the period immediately before and after 11,000 years as a very critical period in glacial history. Further correlation of events both in the ocean and on land during this interval may lead to an understanding of some of the factors causing glaciation (*17*).

Appendix

In describing the cores, the term *lutite* has been used to indicate sediments composed of particles smaller than the silt fraction, whether they consist of clay minerals, calcite, or any other mineral. The terms *calcareous sand* and *calcareous silt* are used when the coarse sediment is dominated by the shells and debris of pelagic organisms. Simplified graphic lithological descriptions of six of the cores are shown in Fig. 3.

Core A172-6 was taken on the crest of an eastern extension of the Beata Ridge in the Caribbean. The sediment is a brown foraminiferal lutite, uniform throughout except for several zones of black manganese oxide speckling. It contains no distinct bedding or layers of good particle size sorting.

Core A179-4 is from a gentle slope southeast of Albatross Bank in the Caribbean. The upper half is a dark brown unsorted foraminiferal lutite with zones of manganese oxide speckling.

Core A179-8 was raised from the bottom of a basin north of Hispaniola. It contains a series of layers of well-sorted calcareous silt and sand alternating with brown, unsorted foraminiferal lutite. The layers of calcareous sand are graded—that is, the particles are coarse in the lower part and gradually become finer in the upper part. They contain remains of calcareous organisms of shallow-water environment, particularly fragments of calcareous algae. It is believed that the graded layers containing transported material were deposited by turbidity currents. The interbedded layers of brown lutite are of typical deep-water facies.

Core A179-15 was taken on the Eleuthera Island rise. The upper 100-centimeter section is a light gray calcareous lutite without bedding. At a depth of 100 centimeters there is gradational change to light tan calcareous lutite of the same texture and also without bedding. The lower part of the core is similar except for a graded layer that extends from 355 to 520 centimeters.

Core A180-48 was taken near the bottom of a submarine canyon northwest of Cape Verde, French Equatorial Africa. The sediment is a dark green lutite, very uniform from top to bottom except for a few small silt lenses and silt films in the lower part. There is no well-defined change in color or texture anywhere in the core.

Cores A180-73 and A180-74 are very similar. They were taken on the gently sloping flanks of the Mid-Atlantic Ridge. The distance between stations was 60 nautical miles. Both cores have grayish tan foraminiferal lutite at the top. This very gradually changes, downward, to darker tan. At a depth of about 35 centimeters there is a fairly abrupt, but not sharply defined, color change to light gray foraminiferal lutite that continues, with minor changes in shade, to a depth of 225 centimeters in core A180-73 and to a depth of 300 centimeters in core A180-74, below which there are layers of brown and gray foraminiferal lutite. The texture is quite uniform throughout.

The various color layers, distinguished by letters on the graphic logs of these cores (Fig. 3) correlate from core to core. This is good evidence that the sedimentary sections included in these cores have not been disturbed by local catastrophic deposition of sediment by turbidity currents or by removal of sediment by slumping.

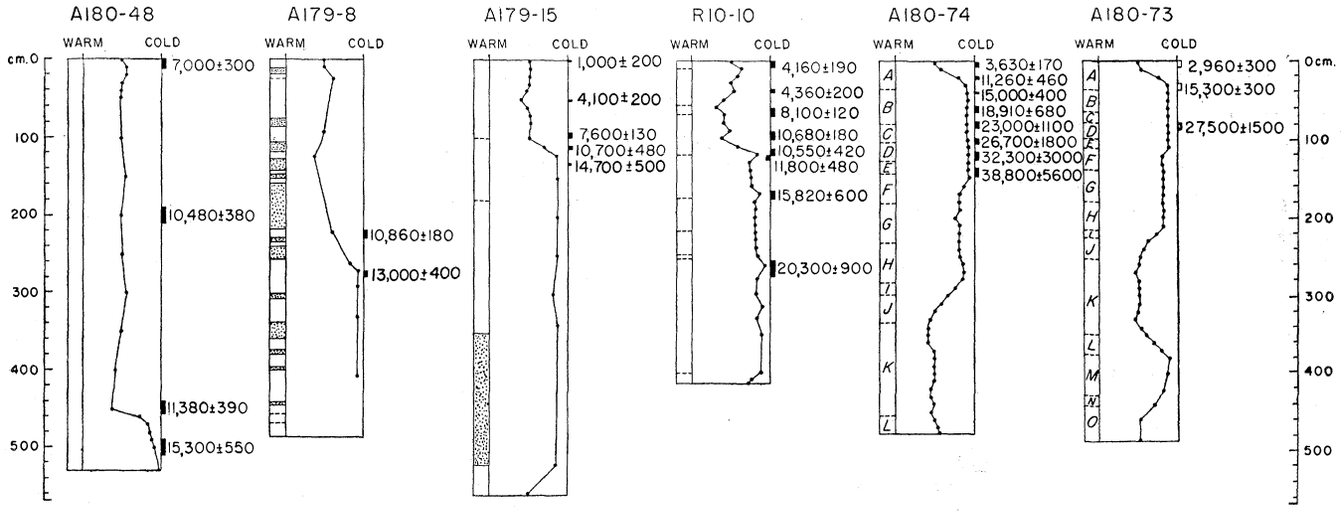

Fig. 3. Climatic curves derived from variations in Foraminifera and lithologic and chronologic data for six deep-sea cores ranging from the equator to latitude 41°N. Present climate is plotted on the mid-point between warm and cold, and inferred past climates are plotted with respect to it. The change from cold to warm in the curves indicates the end of the last ice age. The scales are in centimeters of core lengths. The numbers to the right of each curve are radiocarbon ages (in years) of carbonate of total sample obtained by Broecker, Kulp and Tucek (*15*) from five of the cores and of carbonate of the coarse fraction (greater than 74 microns) by Rubin and Suess (*12*) from core A180-73. Sections of the cores used for dating are indicated by boxes. The radiocarbon dates give rates of sedimentation ranging from 2.2 centimeters per 1000 years (part of core A180-74) to 270 centimeters per 1000 years (part of core A180-48). In the graphic descriptions of the lithology to the left of each curve, open zones indicate lutite, dashed lines the change in color of the sediment, and dotted zones calcareous sand and silt. The correlating zones marked by color changes in cores A180-73 and A180-74 are distinguished by letters.

Core R10-10 was taken in the Newfoundland Basin. The upper half is a slightly silty lutite in various shades of gray and rose-gray. At a depth of 120 centimeters there is a gradational change from dark gray lutite to light gray calcareous lutite. The lower half contains several sharply defined color changes and two zones of glacial marine sediment—that is, lutite containing ice-rafted pebbles.

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Kinetic Aspects of Assembly and Degradation of Proteins

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Recent years have seen the development of highly refined techniques applicable to the study of protein synthesis. As a result, productive research on this formidable problem has grown considerably. Yet even today virtually nothing is known, except by inference, of the pathway between free amino acids, which are generally accepted as the ultimate precursors, and proteins. The catabolic aspects of protein metabolism have been almost completely neglected, and it is even questioned by some whether controlled intracellular degradation plays any role at all in normal protein metabolism (*1*).

Speculations with respect to the mechanism of protein biosynthesis have tended to focus on the problem of how a specific and definitive amino acid sequence is achieved. The result has been a deemphasis of the problem of the "pathway of biosynthesis" in the sense that this term has been applied in studies of other complex molecules. Many have concluded either that there are no intermediates in protein synthesis or that the intermediates must be so poorly defined or so transient that they defy characterization. It may be useful to consider the analogous situation that prevailed for a number of years in the field of fatty-acid synthesis. Even after isotopic evidence pointed overwhelmingly to a synthetic mechanism involving stepwise two-by-two condensation, intermediates were not detected by direct tissue analysis.

Studies from this laboratory have suggested the existence of intermediate compounds between free amino acids and completed protein molecules. Specifically, these studies have shown the nonidentity in terms of metabolic history of the different residues of a given amino acid species at different loci along the peptide chains of the proteins investigated.

Most of this work has been reported in detail elsewhere (*2–5*), and so we shall only outline here the general plan of the experiments and summarize the results for comparison with the findings of others. The two schemes shown in Fig. 1 serve as a basic outline for these considerations. In scheme A of Fig. 1, protein synthesis is pictured as an "all-at-once" assembly of individual amino acid residues on a sequence-determining template. Scheme B in Fig. 1 differs in that it includes intermediate compounds on the path between the free amino acid pool and the completed protein molecule. Many variants of these basic schemes can be devised by introducing reversible reactions, combining some of the features of each, multiplying the number of steps, and so on.

The design of our experiments can be made clear, however, without these refinements. We label amino acid A with carbon-14, incubate it with tissue *in vitro* or *in vivo*, and isolate from the tissue a sample of a pure protein. Synthesis according to scheme A should result in a protein having, for amino acid species A, the same specific radioactivity at all points along the peptide chain (*uniform labeling*). With scheme B, however, different residues of this amino acid species may have different specific radioactivities (*nonuniform labeling*). It should be emphasized that scheme B could also lead to uniform labeling if the pool sizes of the intermediates and the rates of the various reactions involved were favorable for this result. The finding of uniform labeling would therefore be ambiguous.

Experiments of this sort have been carried out in the case of three crystalline proteins—ovalbumin, insulin, and ribonuclease. In each case, the protein synthesized *in vitro* in the presence of a labeled amino acid has been rigorously purified. Samples of labeled amino acid derived from different loci in the peptide chain have been obtained by partially degrading the protein and separating the fragments. Ovalbumin has been studied most extensively of the three; the data obtained are summarized in part A of Table 1. Of particular interest is the comparison of the labeling data obained on the plakalbumin and hexapeptide fractions of ovalbumin which result from the treatment of this protein with the bacterial protease, subtilisin (*6*). Three of the residues in the hexapeptide—namely,

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