Small-area balance records (*18, 19*) show year-to-year variations that are consistent with random local variations due to chance positions of buried snow sastrugi and drifts. At "Old Byrd Station," 11 km beyond the southwestern end of the traverse, Gow (*18*) found no significant net trend in balance since A.D. 1650, although at the South Pole Giovinetto and Schwerdtfeger (*19*) found that balance increased significantly after A.D. 1830. Certainly the climate must have changed at the end of the Wisconsin/Würm glaciation (*8*). A real decrease in balance near Byrd Station probably implies a change in the nature and movement of air masses onto the continent, which in turn may be due to an increase in the extent of sea ice (*20*) some time after the end of the Wisconsin/Würm glaciation. If the decrease in mass balance was associated with a major change in the temperature of the air over the surface of the ice sheet, then from the oxygen isotope record of Johnsen *et al.* (*8*) the nonequilibrium must originate in the end of the Wisconsin/Würm glaciation.

Changes in the form of the West Antarctic ice sheet will have important effects on sea level and on the East Antarctic ice sheet. A drop in ice level may allow more ice to exit from East Antarctica through the bedrock trough west of the Pensacola Mountains (*6*). The present nonequilibrium, if applicable to all of West Antarctica, would raise sea level by 6 mm per century, a figure very much smaller than the observed rate of sea level rise of 120 mm per century (*21*).

Several factors could have caused the nonequilibrium, and it is not possible to discriminate between these factors on the basis of the present data. This result cannot be automatically applied to other ice sheet drainage systems, especially in view of the reported (*22*) thickening in central Greenland and the reported (*23*) thinning in northern Greenland.

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Glacial-Postglacial Temperature Difference Deduced from Aspartic Acid Racemization in Fossil Bones

Abstract. *The magnitude of the temperature increase that occurred in continental regions following the termination of the last glaciation has been determined from the degree of racemization of aspartic acid in fossil bones of known age. The results indicate an increase of 4°C for the Mediterranean coast and 5° to 6°C for East Africa. These estimates are believed to be reliable within 1°C.*

The two methods generally used to estimate the climatic changes that occurred on the earth during glacial cycles are determination of $^{18}O/^{16}O$ ratios in calcareous fossils and analysis of pollen and faunal distributions. However, because these methods reflect the influence of a number of factors, the absolute magnitude of the temperature difference in various regions between Pleistocene glacial and interglacial cycles cannot be unequivocally ascertained in most cases. Both temperature fluctuations and isotopic changes in seawater resulting from storage of greater amounts of isotopically light water in continental ice sheets during times of extensive glaciation affect the $^{18}O/^{16}O$ ratios found in calcareous fossils (*1*). Since both effects are in the same direction, it is not at present clear what combination of these factors produced the observed $^{18}O/^{16}O$ ratios. Pollen (*2–4*) and faunal (*5*) studies provide a general indication of the types of vegetation and animals that were present at a particular time, and therefore suggest what kind of general climate might have prevailed in the region. However, paleoecological evidence cannot give accurate estimates of prevailing temperatures, but only a general indication of whether the climate was warmer or cooler.

Recent studies have shown that the racemization reaction of amino acids can be used to estimate paleotemperatures (*6, 7*). Only L-amino acids are usually found in the proteins of living organisms, but over long periods of geological time these L-amino acids undergo slow racemization, producing

Table 1. Carbon-14 ages (based on collagen fraction), D/L aspartic acid ratios, and average racemization rate constants (k_{asp}) in fossil bones. Analyses were carried out on primary (dense) pieces of bone. Porous bone pieces (such as vertebrae) have a greater surface area and thus are more likely to be contaminated with amino acids introduced by groundwaters, and so forth. Indeed, D/L ratios were found to be generally lower in porous pieces.

Location	Sample	D/L	Radiocarbon age (years)	k_{asp} (10^{-5} year^{-1})
Muleta Cave, Majorca, Spain	UCLA 1704C	0.222 0.206	8,570 ± 350	1.72*
	UCLA 1704D	0.273	16,850 ± 200	1.25
	UCLA 1704E	0.293	18,980 ± 200	1.22
Lukenya Hill, Kenya	UCLA 1709C	0.154	2,120 ± 60	4.02
	UCLA 1709B	0.500	17,700 ± 760	2.71
Olduvai Gorge, Tanzania	UCLA 1695	0.316	17,550 ± 1000	1.48

* This rate constant represents the average value obtained from analyses of two different pieces of bone.

the nonprotein D-amino acids (8). This racemization reaction depends on both time and temperature; thus, if one of these variables is known, the reaction can be used to calculate the other. If paleotemperatures are evaluated by using other information, the extent of racemization of amino acids can be used to estimate the age of fossil materials found in certain environments (9–11). Conversely, with radiocarbon-dated fossils, the racemization reaction can be used to estimate the average temperature a sample has experienced since it was deposited. By determining the amount of racemization in a radiocarbon-dated hippopotamus bone, it was possible to show that the average temperature of a warm spring (Florisbad) in South Africa has been nearly constant for the past 40,000 years (7). A similar study with radiocarbon-dated shells from the southeastern United States has suggested a very large drop (approximately 15°C) in the average temperature in this area during the last glacial (6).

The most useful amino acid for paleotemperature studies is aspartic acid since it has one of the fastest racemization rates of any of the stable amino acids (12). At 20°C, the half-life in bone for the aspartic acid racemization reaction is 15,000 to 20,000 years (11, 13). Thus, in the time interval datable by radiocarbon, a substantial amount of racemization of aspartic acid takes place. In comparison, at 20°C only a small amount of epimerization of isoleucine occurs during the time span covered by radiocarbon (10). We report here paleotemperatures deduced from the aspartic acid racemization reaction for a cave on the Mediterranean island of Majorca and for two areas in East Africa.

The racemization of aspartic acid can be written as

$$\text{L-aspartic acid} \underset{k_{asp}}{\overset{k_{asp}}{\rightleftharpoons}} \text{D-aspartic acid} \quad (1)$$

where k_{asp} is the first-order rate constant for the reversible interconversion of the D and L enantiomers. The kinetics and mechanism of amino acid racemization in bone has been studied and reported elsewhere (10). The equation expressing the D/L aspartic acid ratio as a function of time (t) has been shown to be (11)

$$\ln\left(\frac{1 + \text{D/L}}{1 - \text{D/L}}\right) - 0.14 = 2k_{asp}t \quad (2)$$

Sample preparation and D/L aspartic acid analyses were carried out as described by Bada and Protsch (11). The radiocarbon ages were determined at the radiocarbon laboratory of the University of California, Los Angeles, on the collagen fraction, which was isolated and purified from the bones by the method described by Protsch (14).

The procedures used to calculate paleotemperatures can be divided into three steps. First, an "average" rate constant is determined for each sample by substituting the measured D/L ratio and ^{14}C age in Eq. 2. Next, the "average" temperature the sample has experienced throughout its history is calculated from the known temperature dependence of the racemization rate constant. And finally, the difference in "average" temperature for two samples spanning the termination of the last glaciation is fitted to a simple model of the earth's temperature history for the past 20,000 years in order to determine the magnitude of the temperature increase which has occurred. Table 1 gives the locations, ^{14}C ages, D/L ratios, and rate constants for the samples studied.

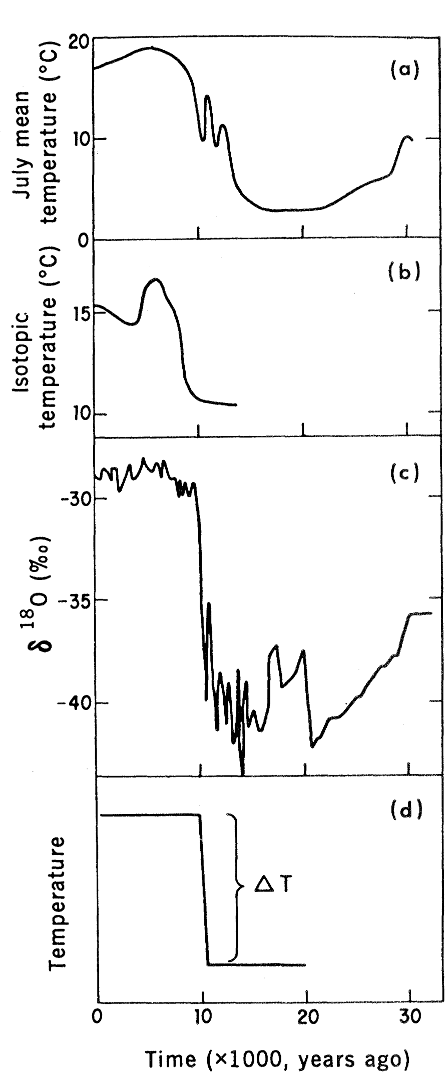

Fig. 1. Climatic variations during the last 30,000 years estimated by using various techniques. (a) July mean temperature in the Netherlands deduced from pollen data (2). (b) Mean winter temperature for the Mediterranean derived from $^{18}O/^{16}O$ ratios in marine mollusks (20). [A similar curve is given in (20) for the mean summer temperature.] (c) Oxygen isotopic changes in an ice core from Camp Century, Greenland (24); $\delta^{18}O$ is the per mil enrichment in ^{18}O. (d) Temperature history model used in our calculations. The shape of this model is based on the results given in Fig. 1, a to c; ΔT is the glacial-postglacial temperature difference.

Since we are interested in determining differences rather than absolute temperatures, the following equation can be used to calculate the difference in average temperature ($\Delta\overline{T}$) of any two samples (15)

$$\Delta\overline{T} = \overline{T}_2 - \overline{T}_1 = \frac{2.303R\overline{T}_1\overline{T}_2}{E_a}\ln(k_2/k_1) \quad (3)$$

where E_a is the Arrhenius activation energy for the racemization reaction, R is the gas constant, and k_1 and k_2 are the rate constants for the two samples. The temperature dependence of the rate of racemization of amino acids in bone has been determined. By studying the epimerization of isoleucine in bone at elevated temperatures, Bada (10) found $E_a = 33.4$ kcal/mole. Furthermore, recent results (13) indicate that the Arrhenius activation energies for the amino acid racemization reaction in bone are the same for aspartic acid, alanine, glutamic acid, and isoleucine. Hence, the value determined for E_a from the isoleucine data is applicable to the aspartic acid racemization reaction.

To solve Eq. 3 it is necessary to substitute in precisely known values for the temperatures ($\overline{T}_1$ and $\overline{T}_2$). The temperature difference ($\Delta\overline{T}$) is, however, quite insensitive to the actual values used (approximately 0.01°C for a 1°C change in absolute temperature). Since we are only interested in temperature differences, an approximate estimate of temperature suffices to solve the equation. For the Muleta Cave site, the mean temperature at present (and, by implication, for the past 10,000 years) is 19°C (16). Substitution of this value and the rate constants from Table 1 for samples UCLA 1704C and UCLA 1704D into Eq. 3 gives the difference in average temperature that these two fossil bones have experienced throughout their history as 1.60°C.

The final step in our calculation is

to fit these average temperatures to a simple model and solve for the magnitude of the temperature changes in that model. The average temperature $\overline{T}_i$ of sample i is defined by

$$\overline{T}_i = \frac{\int T_i(t)\,dt}{\int dt} \qquad (4)$$

in which the integral is taken throughout the history of the sample. The difference in average temperature for two samples is then

$$\Delta\overline{T} = \overline{T}_2 - \overline{T}_1 = \frac{\int T_2(t)\,dt}{\int dt} - \frac{\int T_1(t)\,dt}{\int dt} \qquad (5)$$

The value of $\Delta\overline{T}$ is already known from Eq. 3 and the model we select gives us a general idea of temperature as a function of time. The solution of Eq. 5 will give the magnitude of the temperature changes in the model.

The model for temperature history of the past 20,000 years is shown in Fig. 1d. Its form is based on the curves deduced from pollen and isotopic analyses (reproduced in Fig. 1, a to c) by other investigators. The quantity we want to solve for is ΔT, the increase in average temperature which occurred at the termination of the last glaciation. In our model, the equation for temperature as a function of time is simply

$$T(t) = \begin{cases} T_p & t < 10{,}000 \text{ years} \\ T_p - \Delta T & t > 10{,}000 \text{ years} \end{cases} \qquad (6)$$

where T_p is the present temperature. For the pair of samples, UCLA 1704C and UCLA 1704D, substitution into Eq. 5 gives

$$-1.60 = \frac{\int_0^{10{,}000} T_p dt + \int_{10{,}000}^{16{,}850} (T_p - \Delta T)\,dt}{\int_0^{16{,}850} dt} - \frac{\int_0^{8{,}570} T_p dt}{\int_0^{8{,}570} dt} = -\left(\frac{6{,}850}{16{,}850}\right)\Delta T \qquad (7)$$

the solution of which yields $\Delta T = 3.9°C$. A similar calculation for the other pair of Muleta samples (UCLA 1704, C and E) gives $\Delta T = 3.7°C$, and the Lukenya Hill pair gives $\Delta T = 4.7°C$.

Only a single sample is given for Olduvai Gorge in Tanzania, so this calculation differs slightly from the others. The rate constant for this sample is calculated to be 1.48×10^{-5} year^{-1}. It is known that the present mean temperature for the cave on Majorca is 19°C (*16*) and that the aspartic acid rate constant at this temperature is 1.72×10^{-5} year^{-1} (from UCLA 1704C). Substitution of these values into Eq. 3 gives an average temperature at Olduvai Gorge over the past 17,550 years of 18.25°C. The present mean temperature at Olduvai is 21°C (*17*) so the drop in average temperature was 2.75°C. Substitution into Eq. 5 gives a temperature change of 6.4°C. This ΔT is less reliable than the Muleta or Lukenya Hill results since it depends on a knowledge of the current mean temperature at the site.

In principle, any pair of samples can be used to solve the foregoing series of equations. There is, however, a practical limit, so that the best results should be obtained when one sample in the pair has a substantial postglacial history while the other goes well back into the last glaciation. On the other hand, if the length of time is too great, short-term fluctuations in the actual temperature might cause us to underestimate the magnitude of the temperature change because the calculations are based on average temperatures. The samples chosen are about the best that could be selected, although a larger number of samples would, of course, produce better temperature estimates.

A quantitative evaluation of errors in these results is possible, although it will necessarily be slightly different for each sample, depending on its age and the extent of racemization. We have stated that the error introduced by the mathematical approximations probably amounts to less than 0.1°C. Another principal source of error is limitations on the precision and accuracy of age and D/L aspartic acid measurements (*18*). The statistical counting errors in the ^{14}C age determination result in an error of $\pm$ 0.2°C. However, these are difficult to interpret, since ^{14}C determinations require large samples and several bones while amino acid analyses use much smaller quantities. A precision of 5 percent in the measured D/L ratio corresponds to an error of $\pm$ 0.5°C. Finally, there are errors which likely arise from the model we used in the calculations. On the basis of deep-sea sedimentation rates, Broecker *et al.* (*19*) have estimated the termination of the last glaciation at about 11,000 years ago. Using 11,000 years as the termination of the last glaciation in our model would increase the estimated ΔT by 0.5°C. A change in the model to allow a gradual increase in temperature from 12,000 to 8,000 years ago will have a negligible effect on the estimate. Overall, we believe our results give an estimated temperature increase which is reliable to 1°C.

The results obtained here indicate an increase in average temperature of about 4°C for the western Mediterranean coastal regions, following the termination of the last glacial. Using the oxygen isotopic ratios of marine shells (food refuse) found in caves along the Mediterranean coast, Emiliani *et al.* (*20*) estimated an increase of 5° to 6°C in the average yearly temperature. Duplessy *et al.* (*21*) estimated a 4°C difference between the glacial and interglacial temperatures from the oxygen isotopic composition of a stalagmite found in a cave in southern France. Our results from the Muleta Cave bones agree fairly well with both of these estimates.

On the basis of ancient snow lines of equatorial East African mountains, Flint (*22*) roughly estimated a difference of 5° to 7°C in mean annual temperatures between glacial and postglacial times. The vegetation history of various East African highland regions suggests that the mean annual temperatures during the last glacial may have been 5° to 8°C lower than present-day temperatures (*4, 23*). Our estimate is 5°C for the more reliable sample (Lukenya Hill) and as much as 6°C for the less reliable one (Olduvai Gorge).

Temperatures deduced from aspartic acid racemization appear to provide an independent verification of the result derived from other evidence that continental temperatures were several degrees cooler during the last glacial. Although we have processed a limited number of samples, the results are in reasonable agreement with what we believe to be the most reliable estimates of other methods. The aspartic acid racemization reaction should provide an important new tool in quantitative evaluations of the temperature variations that occurred in continental areas during the upper Pleistocene.

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Ocean Current Monitoring Employing a New Satellite Sensing Technique

Abstract. *The very-high-resolution radiometer on the NOAA-2 (National Oceanic and Atmospheric Administration) satellite has recently obtained imagery in the visible channel containing sunglint over a major portion of the coastal waters off the eastern seaboard of the United States. An abrupt change in surface roughness has been observed at the shoreward edge of the Gulf Stream Current from Florida to Cape Hatteras that results from the opposition of waves propagating against the flow of the Gulf Stream.*

The Gulf Stream has been observed by satellite radiometers under relatively cloud-free conditions since 1966 (*1*). In virtually every case the differentiating signature of this current has been a thermal boundary measured in one of the thermal infrared (IR) water-vapor windows (that is, at wavelengths near 4 or 11 μm). The NOAA-2 (National Oceanic and Atmospheric Administration) environmental satellite obtains global imagery operationally from a scanning radiometer (SR) sensitive in both the visible and IR wavelengths. A very-high-resolution radiometer (VHRR) being developed by NOAA for future operational use is also carried on NOAA-2 (*2*). From the satellite's near-polar, sun-synchronous orbit, the VHRR provides limited coverage of the earth in the IR region with a 1-km resolution; observation times are during daylight at about 0900 local time and at night at about 2100 local time. During daylight, visible imagery with a resolution of 1 km is also provided. The visible and IR detectors are sensitive at 0.6 to 0.7 μm and 10.5 to 12.5 μm, respectively. The VHRR-IR imagery is particularly useful in coastal areas where fine detail is needed to define coastal surface temperatures.

On 29 April 1973, the NOAA-2 satellite obtained imagery of the eastern seaboard of the United States. The simultaneous visible and IR VHRR images are shown in Fig. 1. The IR display (Fig. 1B) shows the relatively cold clouds and land surfaces as light areas (less radiant energy reaches the radiometer) and the warmer regions, such as the Gulf Stream (temperature about 25°C), as darker areas. Landward of the thermal front of this current, the cooler coastal water (about 18°C) contrasts with the warmer land of the southeastern and mid-Atlantic states.

What is truly remarkable, however, is that it is also possible to locate this current boundary in the adjoining visible image (Fig. 1A). The entire front of the Gulf Stream from Florida to Cape Hatteras is delineated by a roughness contrast, highlighted by the solar reflection. Contrasting reflectances have been observed in satellite photography in the past but over very limited areas. For example, it has been shown that anomalous dark patches are indicative of calm surface conditions where short-wavelength capillary waves are absent (*3*). These conditions prevail beneath high pressure ridges and occasionally in conjunction with upwelling. Attempts have been made to relate the size and intensity of the sunglint pattern to surface wind speed (*4, 5*). Under more extreme conditions, brightness contrasts have been observed where local winds are quite strong. This phenomenon results when winds that originate overland, where topography channels the flow offshore, cause rough and choppy seas. Such conditions are well known along the Pacific Ocean shores of Mexico and Central America (for example, the Gulf of Tehuantepec) (*6*).

Scanning radiometer images of sunglint are much different from those obtained by satellite cameras (*7*). The vidicon cameras on the earlier weather satellites obtained approximately 12 essentially instantaneous pictures along the sunlit portion of each pass. In about nine of these photographs, wherever cloud cover permitted and a water surface was being viewed, one nearly circular sunglint would appear at the specular point. On NOAA satellites presently in use a SR obtains data as it scans from left to right through the nadir. An image is built up scan line by scan line in a fashion analogous to that used to generate a television picture. The geometry of the scanning mode of the SR and VHRR and the near-polar