unknown) normal variations in tissue density need to be determined. Although the pictures already taken show that useful data can be obtained with water immersion, more work needs to be done on the effectiveness of water or other suitable liquids used to compensate for variations in body thickness. Nevertheless, it is clear that heavy particle beams probably represent the most sensitive tool presently available for the measurement of internal density distributions of objects. Indeed, the technique is general and is applicable in areas other than medicine.

E. V. Benton
R. P. Henke

Physics Department, University of San Francisco, San Francisco, California 94117

C. A. Tobias

Donner Laboratory, University of California, Berkeley 94720

References and Notes

1. C. A. Tobias, T. L. Hayes, H. D. Maccabee, R. M. Glaeser, *Report UCRL-17357* (University of California Radiation Laboratory, Berkeley, 1967), p. 108; C. A. Tobias and H. D. Maccabee, in *Elementary Particles: Science, Technology, and Society*, L. C. Yuan and C. S. Wu, Eds. (Academic Press, New York, 1971), chap. 5.
2. H. L. Anderson, U.S. Patent No. 2,953,510 (1960); A. M. Koehler, *Science* **160**, 303 (1968); D. West and A. C. Sherwood, *Nature* **239**, 157 (1972); V. W. Steward and A. M. Koehler, *Science* **179**, 913 (1973).
3. C. A. Tobias *et al.*, *Science* **174**, 1131 (1971); E. V. Benton, R. P. Henke, C. A. Tobias, *Report LBL-2016* (Lawrence Berkeley Laboratory, Berkeley, Calif., 1973), p. 1.
4. R. L. Fleischer, H. W. Alter, S. C. Furman, P. B. Price, R. M. Walker, *Science* **178**, 255 (1972).
5. M. Goitein, *Nucl. Instrum. Methods* **101**, 509 (1972).
6. J. W. Boag, A. J. Stacey, R. Davis, *Brit. J. Radiol.* **45**, 633 (1972); A. H. Dowdy, W. F. Barker, L. D. Lagasse, L. Sperling, L. J. Zeldis, W. P. Longmire, *Cancer* **28**, 1558 (1971); R. L. Egan, *ibid.*, p. 1555.
7. We thank J. Lyman, J. Howard, G. Welch, D. Palmer, W. Shimmerling, R. Walton, J. McGinnis, and the bevatron crew at Lawrence Berkeley Laboratory for their generous help. This work is supported in part by the Atomic Energy Commission.

9 July 1973

State of Equilibrium of the West Antarctic Inland Ice Sheet

Abstract. *Data from a traverse connecting Byrd Station, Antarctica, with the local ice divide allow calculation of the mean volume flux at points along the traverse. This is compared with current rates of surface accretion upstream from each point. Near the ice divide the ice sheet seems to be in equilibrium, but near Byrd Station the volume flux is in excess of that needed for ice sheet equilibrium by at least 15 percent. The discrepancy may exist because the traverse does not follow the ice flow exactly or because ice flow at depth is very complex. Although neither of the foregoing possibilities can be disproved, it seems most likely that the discrepancy is due to ice sheet thinning, as has previously been suggested by work on oxygen isotope ratios and temperature in the boreholes at Byrd Station. This thinning probably started at the end of the Wisconsin/Würm glaciation. Causes for the thinning are discussed.*

The large Antarctic inland ice sheets have very low rates of movement and surface accretion which makes them very sluggish in response to climatic change. Measurements in central West Antarctica (Fig. 1) suggest that the ice sheet regime is not in equilibrium with the present climate, and that the ice sheet is thinning and that it is, at least in part, a relic of a former climate, probably that of the Wisconsin/Würm glaciation.

Surface mass balance, strain rates, and ice thicknesses have been obtained along a traverse upstream from Byrd Station, Antarctica (*1*). The traverse crosses the ice divide, and, by a process of successively adding balances and correcting for strains, it has been possible to calculate the equilibrium condition of this portion of the ice sheet.

Consider a small prism of ice (Fig. 2) bounded on two faces by flow lines and on two other faces by the top and bottom surfaces of the ice sheet. If the strain rate (measured only at the surface) does not vary with depth, equilibrium requires that, at a distance x from the ice divide

$$Q_{x+\Delta x}\,\bar{h}_{x+\Delta x}\,w_{x+\Delta x} = Q_x\bar{h}_x w_x + \tfrac{1}{2}(b_t + b_b)(w_x + w_{x+\Delta x})\Delta x \qquad (1)$$

where Q_x is the equilibrium velocity at x or the mean velocity with depth needed for ice sheet equilibrium; Δx is the length of the prism; h_x is the mean ice thickness at x, taken to be constant in time; w_x is the spacing between flow lines at x; and b_t and b_b are the top surface and bottom balances, respectively.

For an ice sheet in perfect equilibrium, the amount of ice moving through a section is equal to the amount of snow accumulation upstream from the section plus or minus the amount of basal freezing or melting. All these quantities are expressed in equivalent volumes of ice. The volume flux density then has the units of a velocity (cubic meters per square meter per year or meters per year). This velocity has been called the "balance velocity" (*2*) but, as a reviewer (Dr. C. R. Bentley) has pointed out, this term is confusing and so I will use the term "equilibrium velocity." The nonequilibrium of the ice sheet is expressed by any difference in value between this theoretical equilibrium velocity, Q, and the actual velocity, $\bar{V}$, with the mean taken through the thickness of the ice sheet. In principle, the equilibrium velocity can be calculated by means of Eq. 1 and other simple relationships at any distance from the ice divide, where $Q_{x=0} = 0$.

The surface velocity, V, is found from the relative movements of poles along the traverse and the assumption that the crest of the ice sheet is also the ice flow divide. At the crest the surface slope is zero and there can be no horizontal gravity force gradients, and, hence, in the absence of significant longitudinal stresses or structural asymmetry, there can be no horizontal motion. Owing to internal shear, the measured V is greater than $\bar{V}$, and a correction must be applied.

The traverse, however, does not follow the flow line exactly, and it is necessary to work with components resolved onto the traverse axis. Let θ be the angle between the flow lines and the traverse axis. Then, with primed quantities being measured with respect to the traverse axis, the following equalities hold:

$$Q'_{x'} = Q_x \cos\theta$$
$$x' = x\cos\theta$$
$$\Delta x' = \Delta x\cos\theta$$
$$w' = w/\cos\theta$$

and let

$$\Delta w' = w'_{x'+\Delta x'} - w'_{x'}$$

From Eq. 1

$$Q'_{x'+\Delta x'} = \frac{Q'_{x'}\bar{h}_{x'} + (b_t + b_b)\left(1 + \tfrac{1}{2}\dfrac{\Delta w'}{w'_{x'}}\right)\Delta x'}{\bar{h}_{x'+\Delta x'}\left(1 + \dfrac{\Delta w'}{w'_{x'}}\right)}$$

where $\Delta w'/w'_{x'}$, the transverse strain, is equal to the transverse strain rate, $\dot{e}_{y'y'}$, times the time, t, taken to travel the distance $\Delta x'$,

$$\frac{\Delta w'}{w'_{x'}} = \dot{e}_{y'y'}\,t$$
$$t = \frac{\Delta x'}{\bar{V}'}$$

where $\bar{V}'$ is the component of the mean volume flux in the x' direction. For simplicity in computation, the surface velocity component, V', is used instead of $\bar{V}'$; the error thus introduced is small and, for example, makes Q' less than 0.4 m year^{-1} (or 4.5 percent) too large at kilometer 150 of the traverse.

In Fig. 3 I have plotted the equilibrium velocity component, Q', for the case of zero bottom balance ($b_b = 0$), using the data from the traverse. Owing to shear within the ice sheet, $\bar{V}'$ is less than V'. Surface slopes are about 0.002, and the ice temperature is about $-30°C$ for the upper 1000 m of the ice sheet and increases to the melting point at the bottom (*3*). Integration of the "flow laws" for ice compiled by Budd (*4*) for Byrd Station indicates that about 4 m year^{-1}, or 30 percent, of the surface velocity is due to shear in the upper 2000 m of ice (the remainder being basal sliding and intense shearing near the base), and $\bar{V}'$ is about 0.7 m year^{-1} less than V'. At kilometer 150, $\bar{V}'$ exceeds Q' by 1.3 m year^{-1}, or 15 percent. More ice is leaving the region upstream from Byrd Station than is being replaced by snow accumulation.

The comparatively good agreement in velocity components between the ice crest and kilometer 60 supports the assumption of no horizontal movement at the ice crest and indicates that the ice sheet is more nearly in equilibrium in this region.

Apparent nonequilibrium can result from changes in ice flow characteristics at depth and by advection of ice of different origin across the traverse. For example, the increase in the equilibrium velocity component at kilometer 105, where the ice sheet thins, is not reflected in the surface velocity component, and the changes in the velocity components at kilometer 160 correspond to sudden and large changes in surface slope and in strain rates. The importance of these effects in equilibrium considerations cannot be properly evaluated with the available data, but such phenomena could produce a significant portion of the apparent nonequilibrium.

However, it is most likely that these effects are not so important, especially for the upper portion of the traverse. From the ice crest to kilometer 100, the ice flow is very nearly parallel to the traverse axis, and between kilometer 40, where nonequilibrium begins, and kilometer 100 there are no large bedrock irregularities that may cause cross-traverse ice advection near the bottom.

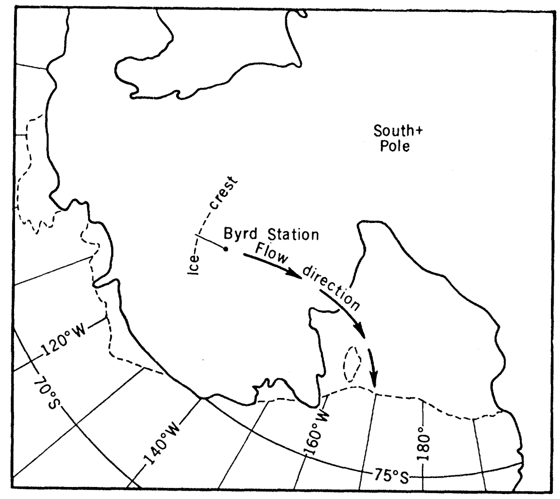


Fig. 1. West Antarctica, showing the location of the traverse (solid straight line).

The reader can thus place most confidence in the nonequilibrium result at kilometer 100 (a place where the ice sheet is even more out of equilibrium than at kilometer 150). Future field studies are designed to investigate these possible effects, but I assume here that they are not important.

The data used in the calculation are accurate to within a few percent. The 6-year mean surface balance is very little different from 2-year means within that interval, which suggests that the balances represent present long-term conditions. The seismic thicknesses have been adjusted by G. Dewart to agree with the directly measured thickness in the Byrd core hole but, because of higher surface balance and reduced frictional heat, ice temperatures up the flow line are probably colder and hence the mean seismic velocities are faster than at Byrd Station. The ice depths upstream are thus probably too small

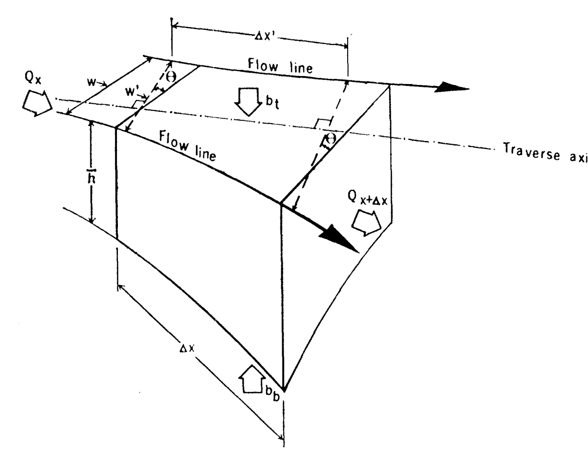


Fig. 2. Oblique view of a prism of an ice sheet bounded by the top and bottom surfaces of the ice sheet, two flow directions, and two arbitrary sections perpendicular to the flow lines. In this study each side of the prism is about 3 km long. Wide arrows show schematically the components of the continuity equation (Eq. 1). Notation is defined in the text.

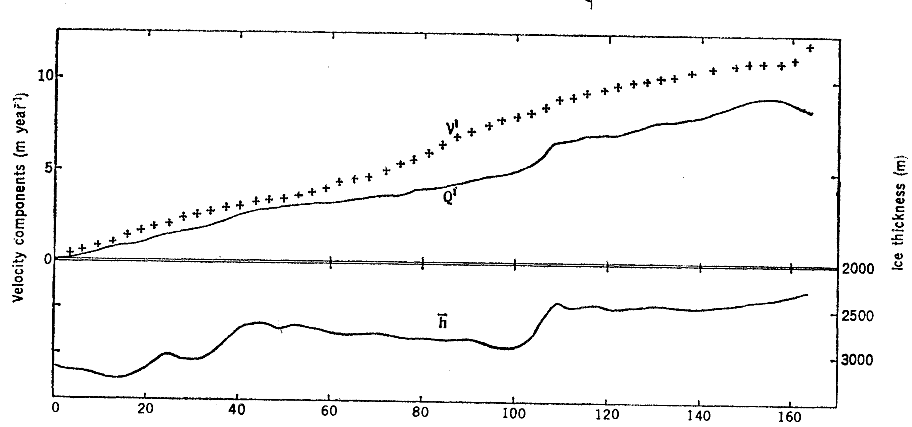
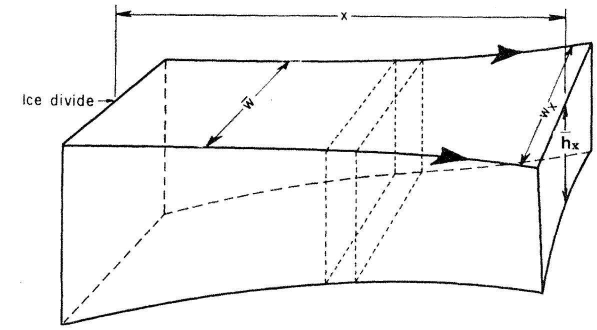


Fig. 3. Surface velocity component V' and equilibrium velocity component Q', compared with the distance from the ice crest x'. The ice thicknesses $\bar{h}$ were obtained by G. Dewart. "New" Byrd Station is at kilometer 163.

Fig. 4. Oblique view of a portion of an ice shee bounded by the top and bottom surfaces, two flow lines, a section under the ice divide, and an arbitrary section perpendicular to the flow lines. Figure 1 could represent the dashed-line portion of this segment.



and the calculated equilibrium velocities too fast. Thus, if ice thickness errors were corrected, the nonequilibrium would be more than 15 percent.

The bottom balance has been set at zero in the calculations, and a basal freezing rate of about 20 mm year^{-1} would increase the equilibrium velocity so that it would be equal to the actual mean velocity. Such a freezing rate on the bottom would, however, require a basal temperature gradient four times steeper than that found in the Byrd core hole (*3*). This rate of bottom freezing would result in a layer of dirty basal ice much thicker than the 4 to 5 m reported by Gow (*5*) at Byrd Station. The bottom balance must thus be, at most, close to zero.

The nonequilibrium implies that the ice sheet is thinning, contrary to the results of Giovinetto (*6*). Three lines of evidence support the idea that the thinning is occurring in an orderly fashion: (i) the ice velocity is only 15 percent more than that needed for equilibrium; (ii) shear stresses at the base of the ice sheet are generally greater than 0.25 bar; and (iii) there is no geologic evidence for a recent and sudden ice sheet advance or for a sudden rise in sea level. Furthermore, Robin (*7*) from the temperature profile and Johnsen *et al.* (*8*) from the oxygen isotope profile at Byrd Station have suggested that the ice sheet has been thinning for several thousand years.

An idea of the time involved can be obtained from a consideration of a portion of the ice sheet bounded by two flow lines, the top and bottom surfaces of the ice sheet, and a section under the ice divide (Fig. 4). The equilibrium velocity, Q_x (calculated above), is the mean velocity at a distance x from the ice divide if the ice sheet is in perfect equilibrium. In that case the volume flowing through a section at x would equal the volume being added vertically in the form of surface mass balance (neglecting basal balance):

$$\bar{h}_x w_x Q_x = \bar{b} \bar{w} x \qquad (2)$$

where $\bar{h}_x$ is the mean thickness at x, w_x is the flow line spacing at x, $\bar{b}$ is the mean surface mass balance corrected to the density of ice, and $\bar{w}$ is the mean flow-line spacing.

If $\bar{V}_x$ is the actual mean velocity at x, then the rate of change in elevation of the surface of the ice sheet, $\dot{h}$ (positive for thickening and assumed constant everywhere), is given by:

$$\bar{h}_x w_x \bar{V}_x = (\bar{b} - \dot{h}) \bar{w} x \qquad (3)$$

Subtracting Eq. 2 from Eq. 3 and expressing the nonequilibrium in terms of:

$$\Delta V = \bar{V}_x - Q_x$$

we have

$$\bar{h}_x w_x \Delta V = \dot{h} \bar{w} x$$

whose solution is of the form:

$$\bar{h}_x = \bar{h}_{0,x} \exp\left(-\frac{w_x \Delta V}{\bar{w} x}\right) t$$

where $\bar{h}_{0,x}$ is a constant. The time constant for ice sheet thinning ($\Delta V > 0$) is:

$$\frac{\bar{w} x}{w_x \Delta V}$$

which for the studied region in upper Marie Byrd Land, and for $\Delta V = 1.3$ m year^{-1}, is 10^5 years. Thus, at the present rate, the ice sheet thins by 1/2.7, or 40 percent, in 10^5 years or by 10 percent, or 250 m, in 10^4 years.

This result is in good agreement with Robin's (*7*) temperature profile analysis and also with the suggestion by Johnsen *et al.* (*8*) of thinning by a few hundred meters since the last glaciation. This result does not, however, support the contention of Wilson (*9*) or of Hollin (*10*) that the ice sheet is now building up to a surge, or that this portion of the ice sheet is presently in a surge condition. Hughes (*11*) has suggested that the ice sheet downstream from the study area is disintegrating: there is no evidence here for this, except that the ice accelerates rapidly at the lower end of the traverse close to Byrd Station (Fig. 3) and this might be the upper end of the disintegrating portion of the ice sheet.

The thinning may be due to mechanical or climatic causes. Since the thinning seems to be generally orderly, catastrophic causes (*11, 12*) may be discounted. Most of the ice motion is due to shear at and near the ice base. The ice sheet is probably resting on volcanic rocks (*13*), and exposed volcanic rocks of this geologic province are hyaloclastic rocks (*14*) which have high water permeabilities. Perhaps changes in water pressure at the base have reduced basal friction and allowed faster ice motion (*15*). Robin *et al.* (*16*) have reported ponded water at the ice base farther down the flow line and in the same geologic province.

A possible climatic cause for the thinning is the surface warming of $\sim 1.2 \times 10^4$ years ago (*8*) and subsequent warming of the ice mass. Warming of the ice mass would reduce the effective viscosity, enhance the internal shear, and increase V. However, even if all of the present-day internal shear of 0.7 m year^{-1} is a result of surface warming, this mechanism cannot account for the 1.3 m year^{-1} nonequilibrium.

By computer modeling the ice sheet Budd *et al.* (*2*) have found that the basal conditions in this region are marginal between melting and freezing, and Robin's (*7*) calculations indicate that a surface temperature change 10^4 years ago could have penetrated to the base of the ice sheet. This may have increased the basal sliding velocity and caused the present nonequilibrium.

Related to this is the possibility of subglacial volcanism. LeMasurier (*14*) has reported subglacially erupted volcanic rocks from the Marie Byrd Land coast that may date from geologically recent times, and Gow and Williamson (*17*) have found ash layers from probably local eruptions 3×10^4 to 1.6×10^4 years ago. Basal heating, especially of an ice base previously colder than the melting point, could explain the computed nonequilibrium. Accounting for the bottom melting due to such volcanism would increase the nonequilibrium figure.

Alternatively, a change in surface mass balance could cause the nonequilibrium. Such a change must have occurred before 1963, when this study began. The possibility that the balance since 1963 is abnormally low for present long-term conditions cannot be discounted, and long-term balance data for central Antarctica that are applicable to a large area are not available.

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.

IAN M. WHILLANS
Institute of Polar Studies,
Ohio State University,
Columbus 43210

References and Notes

1. R. P. Southard, Jr., *Antarct. J.* **3**, 111 (1968); G. Dewart and I. Whillans, *ibid.* **5**, 111 (1970); I. M. Whillans, *ibid.* **7**, 111 (1972).
2. W. F. Budd, D. Jenssen, U. Radok, *Univ. Melbourne Meteorol. Dep. Publ. No. 18* (1971).
3. H. T. Ueda and D. E. Garfield, *Int. Ass. Sci. Hydrol. Publ. No. 86* (1970), p. 53.
4. W. F. Budd, *Aust. Nat. Antarct. Res. Exped. Publ. No. 108* (1969), p. 21.
5. A. J. Gow, *Int. Ass. Sci. Hydrol. Publ. No. 86* (1970), p. 78.
6. M. B. Giovinetto, *ibid.*, p. 347.
7. G. deQ. Robin, *ibid.*, p. 141.
8. S. J. Johnsen, W. Dansgaard, H. B. Clausen, C. C. Langway, Jr., *Nature* **235**, 434 (1972).
9. A. T. Wilson, *ibid.* **201**, 147 (1964).
10. J. Hollin, *Int. Ass. Sci. Hydrol. Publ. No. 86* (1970), p. 363.
11. T. Hughes, *J. Geophys. Res.*, in press.
12. L. Lliboutry, *Nature* **202**, 77 (1964); J. Weertman, *Can. J. Earth Sci.* **6**, 929 (1969); T. Hughes, *Science* **170**, 630 (1970).
13. C. R. Bentley and N. A. Ostenso, *J. Glaciol.* **3**, 882 (1961).
14. W. E. LeMasurier, in *Antarctic Geology and Geophysics*, R. J. Adie, Ed. (Universitetsforlaget, Oslo, 1972), pp. 251–259.
15. L. Lliboutry, *J. Glaciol.* **7**, 21 (1968).
16. G. deQ. Robin, C. W. M. Swithinbank, B. M. E. Smith, *Int. Ass. Sci. Hydrol. Publ. No. 86* (1970), p. 104.
17. A. J. Gow and T. Williamson, *Earth Planet. Sci. Lett.* **13**, 210 (1971).
18. A. J. Gow, *U.S. Army Cold Regions Res. Eng. Lab. Res. Rep. 197* (1968).
19. M. B. Giovinetto and W. Schwerdtfeger, *Arch. Meteorol. Geophys. Bioklimatol. Ser. A* **15**, 227 (1966).
20. J. O. Fletcher, *Bull. At. Sci.* **26**. 40 (1970).
21. R. W. Fairbridge, in *Encyclopedia of Oceanography* (*Encyclopedia of Earth Sciences Series*, vol. 1), R. W. Fairbridge, Ed. (Reinhold, New York, 1966), pp. 479–485.
22. H. Seckel and M. Stober, *Polarforschung* **6**, 215 (1968).
23. D. Raynaud and C. Lorius, *Nature* **243**, 283 (1973).
24. I thank G. Dewart for permission to use his ice thickness results and C. Bentley, C. Bull, T. Hughes, and M. McSaveney for their helpful comments and criticisms. This work was supported under NSF grant GV-26137X awarded to the Ohio State University Research Foundation and the Institute of Polar Studies. Contribution No. 250 of the Institute of Polar Studies.

20 April 1973; revised 15 August 1973

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.