sampling and the sensitivity of the existing system. If we can make worldwide estimates of fallout with acceptable accuracy, based simply upon a limited number of measurements of radionuclides in surface air, we can effect substantial savings in both time and funds. Further refinement of this relationship to smaller specific geographic areas should make possible estimates of local deposition and may lead to a greater understanding of hemispheric differences. In addition, the possibility of estimating the worldwide contamination from pollutants dispersed in a manner similar to that of global fallout is of great value.

MICHAEL T. KLEINMAN
HERBERT L. VOLCHOK
Health and Safety Laboratory, U.S. Atomic Energy Commission, New York, New York 10014

References

1. H. L. Volchok, *Nature* **206**, 4988 (1965); H. L. Volchok and M. T. Kleinman, *U.S. At. Energy Comm. Rep. HASL 204* (January 1969), p. I-17; *U.S. At. Energy Comm. Rep. HASL 207* (April 1969), p. C-1; H. L. Volchok, *U.S. At. Energy Comm. Rep. HASL 210* (July 1969), p. I-2.
2. H. L. Volchok and P. W. Krey, *J. Geophys. Res.* **72**, 20 (1967).
3. H. L. Volchok, *U.S. At. Energy Comm. Rep. HASL 200* (October 1968), p. I-2.
4. D. H. Pierson and R. S. Cambray, *Nature* **216**, 5117 (1967).
5. J. H. Harley, *U.S. At. Energy Comm. Rep. HASL 149* (October 1964), pp. 138–141.
6. H. L. Volchok, *U.S. At. Energy Comm. Rep. HASL 207* (April 1969).
7. K. Telegadas, *U.S. At. Energy Comm. Rep. HASL 204* (January 1969), p. I-2; personal communication.

16 June 1969

One Thousand Centuries of Climatic Record from Camp Century on the Greenland Ice Sheet

Abstract. *A correlation of time with depth has been evaluated for the Camp Century, Greenland, 1390 meter deep ice core. Oxygen isotopes in approximately 1600 samples throughout the core have been analyzed. Long-term variations in the isotopic composition of the ice reflect the climatic changes during the past nearly 100,000 years. Climatic oscillations with periods of 120, 940, and 13,000 years are observed.*

Use of the oxygen-18 concentration in glacier ice as an indicator of past climatic conditions was proposed in 1954 (*1*). The concentration of O^{18} in precipitation, particularly at high latitudes, is determined mainly by its temperature of formation. Decreasing temperature of formation leads to decreasing content of O^{18} in rain or snow. Analysis of stable isotopes has become an important method in studies of ice flow patterns in glaciers (*2–4*) and for determining past rates of accumulation in the upper layers (*3–5*).

In 1966, the U.S. Army Cold Regions Research and Engineering Laboratory (CRREL, now the U.S. Army Terrestrial Sciences Center) succeeded in drilling a core through the entire Greenland ice sheet at Camp Century (*6*). Certain physical and chemical studies have since been made over the upper 300 m of the 1390 m long, 12 cm in diameter, ice core (*7*).

We now show how a chronology can be obtained over nearly all of the vertical core profile by simple theoretical considerations; in particular, we show that climatic changes during the last 70,000 or 100,000 years are reflected as variations in the content of heavy isotopes in the core. The data represent nearly 1600 samples from 218 different core increments measured by mass spectrometry.

Few methods exist for dating a deep ice core. Some advances have been made with C^{14} dating (*8*) in which the carbon dioxide of occluded atmospheric air contained in huge blocks of glacier ice is measured for radiocarbon. Recently a technique for sampling C^{14} from ice *in situ* was tested in Greenland and Antarctica (*9*). These studies required a minimum of 1 ton of glacier ice.

Another way of dating an ice core is to determine the annual layers by measuring the seasonal oscillations in the O^{18} content in small vertical increments containing about 10 cm^3 of the core and then to count the summer maxima continuously from the surface downward. Unfortunately, various processes tend to diminish the isotopic gradients in snow and ice. For example, molecular diffusion in the firn and solid ice, hastened by the progressive plastic thinning of the layers with depth, gradually obliterates the oscillations that remain after firnification. This process alone establishes a limit of a few thousand years on the applicability of the stable isotope dating method on the Camp Century core (see *10*).

The approximate age of different sections of the core can be obtained by (i) considering the generally accepted glacier flow pattern outlined in Fig. 1, and (ii) making certain assumptions concerning the parameters that influence it. Correlation of the time as a function of depth (time scale) discussed here was calculated (*11*) on the basis of simple assumptions such as unchanged rate of accumulation a [35 cm of ice per year (*12*)], unchanged thickness of the ice sheet, H, and unchanged flow pattern back in time. Furthermore, any section of the core was assumed to originate from a location with essentially the same a and H as the Camp Century area. The following approximation to the horizontal velocity profile was used (see Fig. 1):

V_x proportional to $x \cdot y$ from
$y = 0$ to $y = h$ $(= 400$ m),

V_x proportional to x, but independent of
y, from $y = h$ to $y = H$ $(= 1367$ m),

x being the distance from the ice divide, y the distance from the bottom, and H the thickness of the ice sheet in meters of ice. Here H equals 1367 m of ice, instead of 1390 m of core, to correct for the low firn densities in the upper layers.

Basic mass balance considerations led to the following expressions for the vertical velocity components:

$$V_y = -\frac{a}{2H - h}(2y - h), \quad H \geqq y \geqq h$$

$$V_y = -\frac{a}{h(2H - h)}y^2, \quad h \geqq y \geqq 0$$

and, thereby, to the age of the ice at a distance y from the bottom:

$$t = \int_H^y \frac{dy}{V_y}$$

The results of the calculations are shown in Fig. 2, where the age of the core in years (before the present) is given together with the depth in meters from the 1966 surface.

The flow pattern discussed has recently been substantiated (*13*) insofar as it has explained the measured temperature profile (*7*) along the Camp Century borehole.

In the following, the isotopic composition of snow and ice will be denoted by the relative deviation $\delta(O^{18})$ in per mille, of the O^{18}/O^{16} ratio from that of Standard Mean Ocean Water (*14*). Unlike seasonal variations in $\delta(O^{18})$ of snow which are gradually obliterated in the ice by molecular diffusion, long-term variations are preserved for many tens of thousands of years. The mean isotopic composition of a section or long vertical increment of a core therefore reflects the average climatic conditions existing at the surface during the period of time when

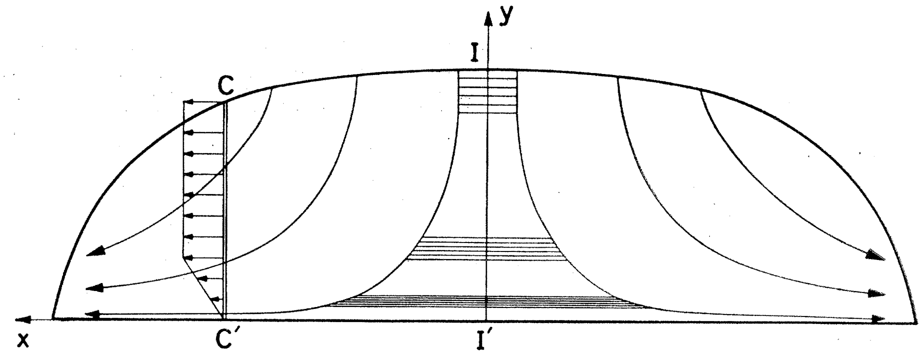


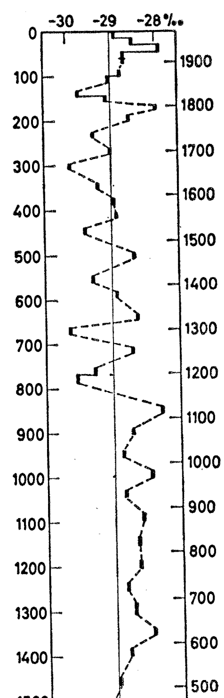
Fig. 1. Vertical cross section of an ice sheet resting upon a horizontal subsurface. Ice particles deposited upon the snow surface will follow lines that travel closer to the base the farther inland the site of deposition. An ice mass formed around the divide (*I–I′*) will be plastically deformed (thinned) with depth as suggested by the lined areas [compare (*35*)]. The horizontal arrows along the vertical ice core (*C–C′*) show the assumed profile of horizontal velocity component, V_x.

the original deposit formed. Increasing $\delta(O^{18})$'s along the core correspond to warming climate, and decreasing $\delta(O^{18})$'s correspond to cooling climate.

Figure 3 presents the $\delta(O^{18})$ measurements (precision better than ± 0.2 per mille) on sections from the upper 470 m of the Camp Century core plotted against the calculated age. If we go back in time from the top, we find high $\delta(O^{18})$'s (> -29 per mille) in the recorded climatic optimum of the 1920's and 1930's. Further back, the cold period in the 19th century and the so-called "little ice age" in the 17th and 18th centuries are recorded by low $\delta(O^{18})$'s (< -29 per mille). At approximately 1150 A.D. there appears a sudden shift in the mean $\delta(O^{18})$ value which remains > -29 per mille to about 550 A.D., corresponding to the favorable period of warmer climate when the Vikings settled in Iceland and Greenland. Comparison with Icelandic historical sources (*15*) indicates that the time scale is hardly more than 20 years off at this time.

During the last 1000 or 1400 years the $\delta(O^{18})$ value has apparently oscillated with a period of approximately 120 years (Fig. 3). Extrapolation of the curve suggests that the next $\delta(O^{18})$ minimum, corresponding to a climatic minimum, will occur about 20 or 30 years from now; the next $\delta(O^{18})$ maximum is projected for the middle of the 21st century.

The prime cause of the oscillations in $\delta(O^{18})$ is probably related to fluctuations in solar radiation. Solar variations are also considered to cause changes in the C^{14} concentrations in the atmosphere. During periods of high sun spot activity an increased amount of short wave radiation reaches the earth and, at the same time, the magnetic field of plasma emitted from the sun shields off the cosmic radiation to a relatively high degree, causing low production of C^{14} in warm periods associated with the high $\delta(O^{18})$'s. This effect gives rise to temporal oscillations in the uptake of C^{14} by plants (*16–18*). Five C^{14} production minima have been suggested (17) within the last millennium (at 1790, 1600, 1460, 1110, and 980 A.D.). In all these cases the $\delta(O^{18})$ curve shows a maximum. Two additional C^{14} minima, at 820 A.D. and 680 A.D. (*17*) apparently fall in a climatically quiet period and are not revealed in the $\delta(O^{18})$ curve.



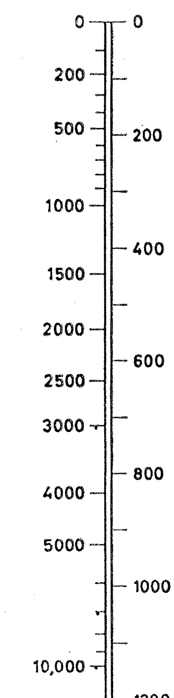


Fig. 2 (left). A depth-age nomograph for the 1390 m long Camp Century ice core. Fig. 3 (right). Variations in $\delta(O^{18})$ in the upper 470 m of the Camp Century ice core plotted against the calculated age of the ice. The lengths of the small vertical lines correspond to the number of years of accumulation represented and calculated as $-L/V_y$, L being the length of the core section. A 120-year climatic cycle is observed.

Figure 4 shows the $\delta(O^{18})$ variations during the last 15,000 years (down to the 1218-m core depth). The minimum between 2100 and 2500 B.P. reflects a cold period that caused considerable glacier advance (*19*). The maximum between 4400 and 7000 B.P., where only one out of 20 samples measured has a $\delta(O^{18})$ of < -29 per mille, reflects the post-Wisconsin climatic optimum. At 10,000 years B.P. the $\delta(O^{18})$ values fall off rapidly, corresponding to the final stages of the last glaciation, in agreement with practically all other time estimates. The lower part of Fig. 4 shows the $\delta(O^{18})$ variations in a continuous sequence of measured samples that extends over the last 5000 years of the Wisconsin (10,000 B.P. to 15,000 B.P. in our time scale). The two peaks at 11,900 to 11,100 B.P. and 12,500 to 12,100 B.P. coincide with short-lived interstadials, Alleroed and Boelling, that are well known from European climatic records (Alleroed: 11,800 to 11,000 B.P.; Boelling: 12,400 to 12,000 B.P. in the C^{14} scale). This is the first suggestion of the occurrence of both Alleroed and Boelling in the Western Hemisphere. If correct, the identification of these peaks provides us with precise index horizons at great depths and, in fact, dates the core better than a direct radiocarbon measurement could with the techniques available at present (*8, 9*). From the preceding evidence, our time scale appears correct at 1000 B.P., 2500 B.P., and 12,000 B.P.

The Alleroed and Boelling interstadials seem to have had three very short-lived forerunners at approximately 13,100 B.P., 14,100 B.P., and 14,900 B.P., and a successor at 10,850 B.P. (partly blurred by the general increasing trend of the curve). Although the $\delta(O^{18})$ curve is not continuous in the postglacial time, the curve suggests ten $\delta(O^{18})$ oscillations, shown by the arrows in the upper part of Fig. 4. If we choose the two $\delta(O^{18})$ maxima at 12,300 B.P. and 850 B.P. as fixed points for our time scale, we count 11 maxima in between, corresponding to periods averaging 940 years. Oscillations in the C^{14} production, possibly caused by solar activity variations, have been observed by investigators (*17, 18*) who suggest periods of 1200 and 1000 years. The long-term $\delta(O^{18})$ oscillations might, therefore, also be ascribed to variations in the radiation from the sun. If the period of this variation has been constant back in time, the corresponding $\delta(O^{18})$ variations may be used as a tool for stepwise dating of the core at great depths, where other methods may fail, and also for correcting our time scale in the post-Wisconsin.

Figure 5a shows the $\delta(O^{18})$ variations from the present back 110,000 years in our time scale. The peak between 17,000 and 20,000 B.P. agrees with the Lascaux optimum (*20*) and also appears in most of the deep-sea core measurements reported (*21*). Extremely low $\delta(O^{18})$'s (< -40 per mille) are found in the interval 13,000 to 17,000 B.P., reflecting the coldest portion of the last advance of the Wisconsin. The deep and long-lasting minimum between 21,000 and 25,000 B.P. agrees with the minimum sea level dated at approximately 20,000 B.P. (*22*). Two long-lasting periods with relatively high $\delta(O^{18})$'s are found about 29,000 to 35,000 B.P. (Plum Point) and 41,900 to 49,000 B.P. (Port Talbot). Other $\delta(O^{18})$ maxima occur at 59,000 B.P. (Broerup) and 63,000 to 66,000 B.P. (Amersfoort); all coincide with well-known interstadials.

From 73,000 B.P. on back in time the $\delta(O^{18})$ values are essentially the same as post-Wisconsin. We interpret these high $\delta(O^{18})$'s as representing the Sangamon (Eemian) interglacial. A minimum is shown at approximately 100,000 B.P., but it only extends over some 40 cm of ice, which can hardly be all that is left from the last ice advance during the preceding Illinoian (Riss) glaciation.

In the lowest 20 m of ice core other minima, not shown in Fig. 5a, cover layers, several meters thick, which may be remains of previous glaciations. However, it is difficult to even roughly estimate the age of this ice, because of the strong influence that the subsurface topography has on the ice-flow pattern close to the bed.

Three additional curves are shown in Fig. 5: a C^{14}-dated pollen study (Fig. 5b) (*23*); a smoothed curve showing advances and retreats of the Laurentian ice sheet in the Ontario-Ohio region (Fig. 5c) (*24*); and part of a generalized temperature curve for the surface water of the central Caribbean (Fig. 5d) (*25*). The agreement of the main features of the four curves (down to 70,000 or 100,000 B.P.) is notable, particularly in view of the uncertainties of the different time scales applied. Curves b and c are based upon C^{14} dating, the uncertainty of which increases rapidly with age; curve d is based jointly on C^{14} and Pa^{231}/Th^{230} datings. Since an uncertainty of 5,000 to 10,000 years for ages about 50,000 years may be expected for any of the radioactive dating methods, it is remarkable that all four chronologies shown in Fig. 5 place the beginning of the Wisconsin within 68,000 to 75,000 years ago. For the same reason, the slight displacements of the Port Talbot maximum shown in Fig. 5, a and c, do not involve a serious disagreement.

Figure 5a also suggests a 13,000-year oscillation with $\delta(O^{18})$ maxima at 6, 19, 32, 45, 59 (?), and 74 (?) millennia B.P. A similar oscillation might be interpreted in Fig. 5d. The 13,000-year period suggests that it may be related to the earth's precession. The mean $\delta(O^{18})$ for annual Arctic precipitation might well be influenced in the same direction, both when the minimum distance between sun and earth occurs in June (relatively short, warm summers and long, cold winters) and when it occurs in December (relatively long, cool summers and short, mild winters). Extremes of the June sun-earth distance occurred approximately 11, 21, 34, 46, 61 and 71 millennia ago, the period being influenced by changes in the eccentricity of the earth's orbit (*26*). This cycle is in phase with the $\delta(O^{18})$ cycle mentioned above and suggested in Fig. 5a, except for the position of the youngest $\delta(O^{18})$ maximum, which is obscured by the rapidly increasing trend of the curve.

We have no means of checking the validity of the individual assumptions (such as thickness and velocity) in connection with the calculation of our time scale. However, we conclude from the above that if they were not valid, the parameters involved must have

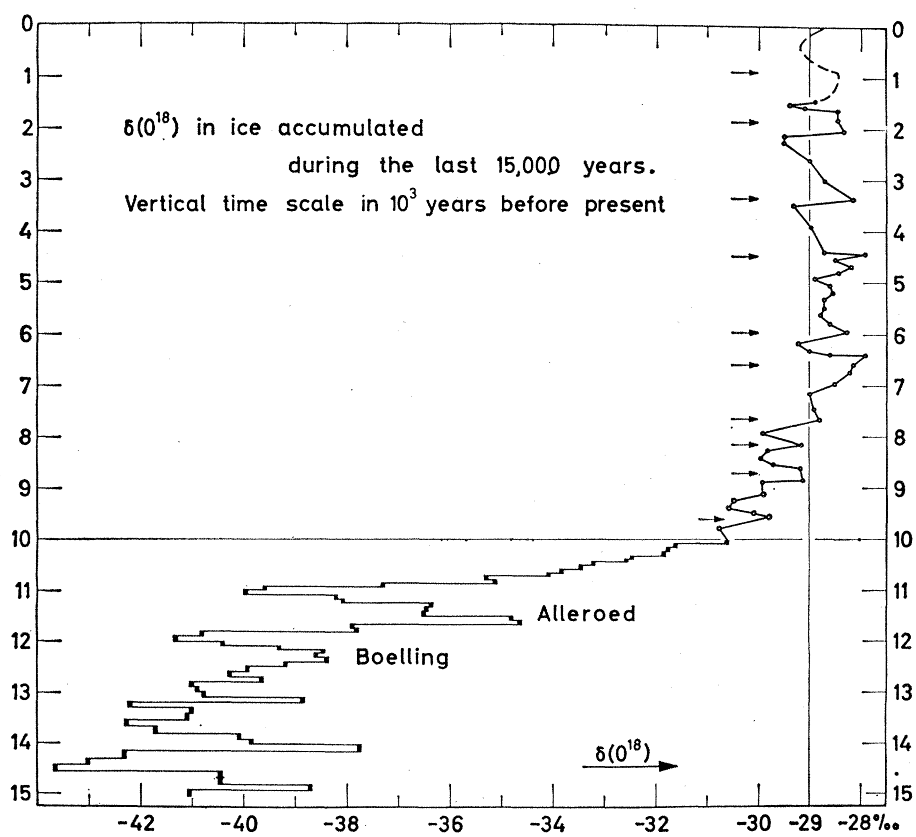


Fig. 4. Climatic variations reflected by $\delta(O^{18})$ variations in the ice from the present to the late Wisconsin glacial period. The data points in the upper portion of the curve represent time periods of from 25 years to 50 years. The step curve in the lower portion of the curve represents a continuous sequence of measured samples, each extending over approximately 100 years. A 940-year climatic period is suggested, which is indicated by the arrows in the upper curve and the maximum peaks in the lower curve.

changed so as to counterbalance each other; for example, in a period with decreasing temperature and snow accumulation, the upper part of the ice would become more viscous, causing slower thinning of the layers.

In spite of the mean $\delta(O^{18})$ of precipitation and the mean air temperature being closely correlated in Greenland today (*27*), no attempt has been made in this study to convert the temporal $\delta(O^{18})$ variations into a temperature curve. There are several reasons for this: (i) the deeper strata originated further inland, where perhaps slightly different climatic conditions existed; (ii) the isotopic composition of seawater, which provides the moisture for the precipitation, changed; (iii) the ratio of summer to winter precipitation possibly changed; (iv) the main meteorological wind patterns possibly changed; (v) the flow pattern of the ice in the accumulation area possibly changed; and (vi) the thickness of the ice sheet changed (*28*).

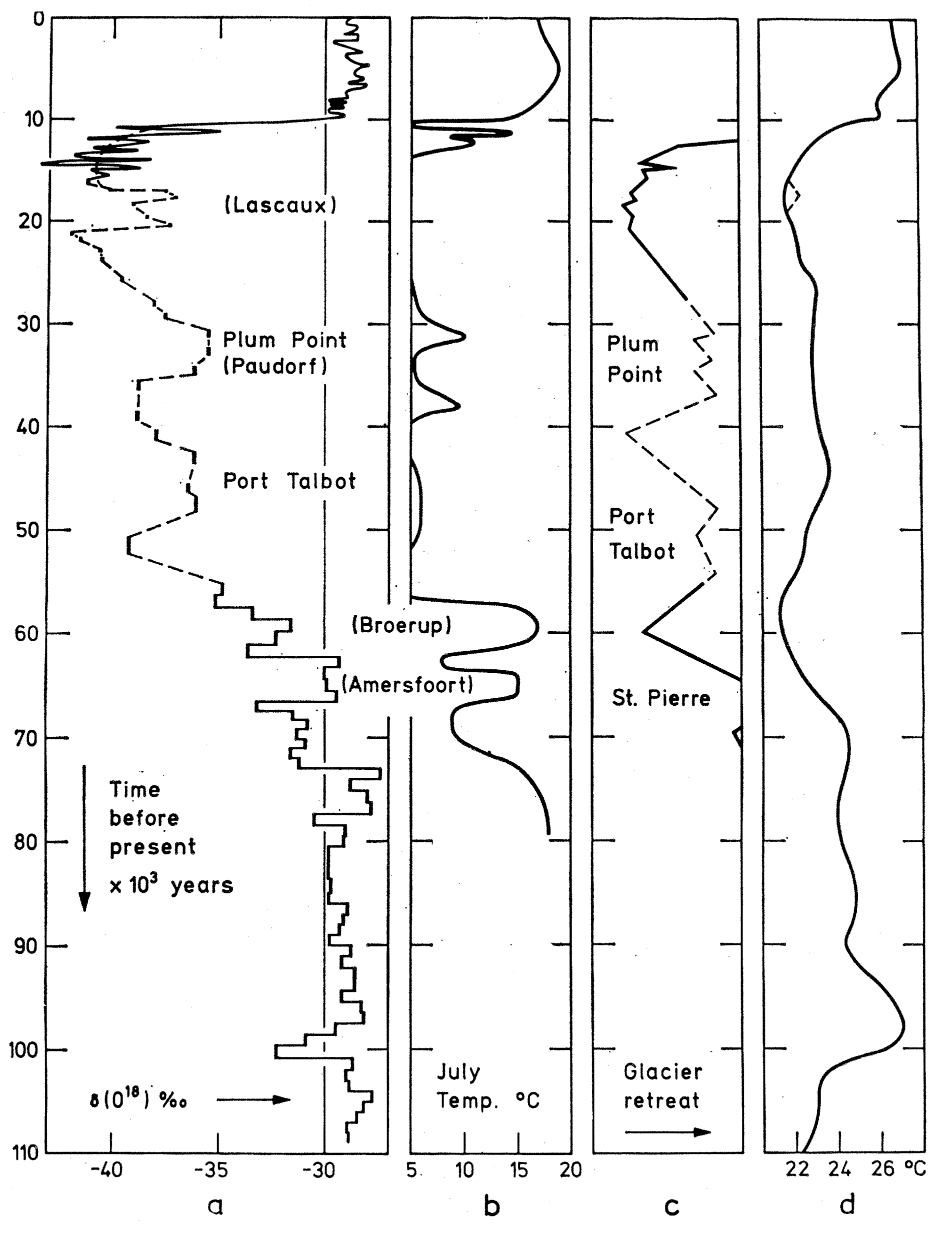


Fig. 5. Climatic variations during the last 70,000 or 100,000 years estimated by various methods. (a) The $\delta(O^{18})$ variations in the Camp Century ice core. (b) A C^{14}-dated pollen study from Holland (*23*). (c) A C^{14}-dated stratigraphic study of Pleistocene deposits in the Ontario and Erie basins showing the measurements of the edge of the Laurentian ice sheet (*24*). (d) An oxygen-isotope study on deep-sea cores showing part of a generalized temperature curve for the surface water of the central Caribbean (*25*). The dashed part of the curve is added by the present authors in view of the data given in (*21*). Dates established by C^{14} and Pa^{231}/Th^{230}. A 13,000-year climatic cycle is suggested. European names of interstadials are given in parentheses.

The changes in the isotopic composition of seawater are less important in changing ice $\delta(O^{18})$'s than changes in the temperature of precipitation formation (*29–34*). The extremely low $\delta(O^{18})$'s shown during the Wisconsin do not necessarily correspond to temperatures that were many degrees centigrade lower than the present—if the thickness of the ice sheet were considerably greater than it is today, this would, in itself, cause lower surface temperatures and lower $\delta(O^{18})$'s of precipitation at a given geographical location.

For the same reason, the $\delta(O^{18})$'s during the climatic optimum between 4500 B.P. and 7000 B.P. would probably have been even higher had the ice sheet not still been thicker than at present. However, a major increase (or decrease) of the surface altitude could hardly occur without a general cooling (or warming) of the climate. Both of these effects influence the $\delta(O^{18})$'s in the same direction, but it is not yet possible to distinguish between the individual contributions to an observed change in $\delta(O^{18})$.

Deuterium isotope analyses made on a selection of samples from the core show a close linear correlation with the oxygen isotope data

$$\delta(D) = 8.0\ \delta(O^{18}) + 12$$

independent of the time of formation. Measuring both isotopes, therefore, adds no further climatological information.

In conclusion, although the complete $\delta(O^{18})$ curve is primarily valid for the North Greenland area, the general trend of the curve agrees with known and reported climatic changes in other parts of the world, at least in the course of the last 75,000 years. It appears that ice-core data provide far greater, and more direct, climatological detail than any hitherto known method; furthermore, unlike other terrestrial deposits, ice cores from some dry-snow zones can provide continuous sedimentary records spanning perhaps several hundred millennia.

W. Dansgaard
S. J. Johnsen
J. Møller
Physical Laboratory II, H. C. Oersted Institute, University of Copenhagen, Copehagen, Denmark
C. C. Langway, Jr.
Cold Region Research and Engineering Laboratory, U.S. Army Terrestrial Sciences Center, Hanover, New Hampshire 03755

References and Notes

1. W. Dansgaard, *Geochim. Cosmochim. Acta* **6**, 241 (1954).
2. R. P. Sharp and S. Epstein, *Proc. Int. Ass. Sci. Hydrol. Chamonix Publ.* **47**, 359 (1958); W. Dansgaard, *Medd. Grønland* **165**, No. 2, 1 (1961); W. Ambach, W. Dansgaard, H. Eisner, J. Møller, *Tellus* **20**, 595 (1968).
3. C. Lorius, *Comité National Français des Recherches Antarctiques (CNFRA) Rept., No. 8* (1963).
4. C. Lorius, R. Hagemann, G. Nief, E. Roth, *Earth Planet. Sci. Lett.* **4**, 237 (1968).
5. S. Epstein and R. P. Sharp, *Trans. Am. Geophys. Union* **40**, 81 (1959); R. Gonfiantini, V. Tagliatti, E. Tongiorgi, W. de Breuck, E. Picciotto, *J. Geophys. Res.* **68**, 3791 (1963); S. Epstein, R. P. Sharp, I. Goddard, *J. Geol.* **71**, 698 (1963); S. E. Wilgain, E. Picciotto, W. de Breuck, *J. Geophys. Res.* **70**, 6023 (1965); S. Epstein, R. P. Sharp, A. J. Gow, *ibid.*, p. 1809; C. C. Langway, Jr., *U.S. Army Cold Region Research and Engineering Laboratory (CRREL) Rep.* **77**, 1 (1967); E. Picciotto, R. Cameron, G. Crozaz, S. Deutsch, S. Wilgain, *J. Glaciol.* **7**, 273 (1968).
6. H. Ueda and D. Garfield, *U.S. Army Cold Region Research and Engineering Laboratory (CRREL) Spec. Rep.* **126**, 1 (1968).
7. B. L. Hansen and C. C. Langway, Jr., *Antarctic J. U.S.* **1**, 207 (1966); temperature profile published by J. Weertman, *J. Geophys. Res.* **73**, 2691 (1968).
8. P. F. Scholander, W. Dansgaard, D. C. Nutt, H. de Vries, L. K. Coachman, E. Hemmingsen, *Medd. Grønland* **165**, 1 (1962); C. C. Langway, Jr., H. Oeschger, A. Renaud, B. Alder, *Nature* **206**, 500 (1965); H. Oeschger, B. Alder, H. Loosli, C. C. Langway, Jr., *Earth Planet. Sci. Lett.* **1**, 49 (1966).
9. H. Oeschger, B. Alder, C. C. Langway, Jr., *J. Glaciol.* **6**, 939 (1967); H. Oeschger, B. Stauffer, C. C. Langway, Jr., B. L. Hansen, International Quaternary Antarctic Congress, Paris, France, abstr. (1969).
10. S. J. Johnsen and W. Dansgaard, in preparation.
11. W. Dansgaard and S. J. Johnsen, *J. Glaciol.* **8**, 215 (1969).
12. G. Crozaz and C. C. Langway, Jr., *Earth Planet. Sci. Lett.* **1**, 194 (1966).
13. W. Dansgaard and S. J. Johnsen, *J. Geophys. Res.* **74**, 1109 (1969).
14. H. Craig, *Science* **133**, 1833 (1961).
15. P. Bergthorsson, *Vedrid*, Reykjavik **12**, 55 (1967).
16. W. S. Broecker, E. A. Olson, J. Bird, *Nature* **183**, 1582 (1959); H. de Vries, *Kon. Ned. Akad. Wetensch. Proc.* **13**, 61 (1958).
17. E. H. Willis, H. Tauber, K. O. Münnich, *Am. J. Sci. Radiocarbon Suppl.* **2**, 1 (1960).
18. H. E. Suess, *J. Geophys. Res.* **70**, 5937 (1965).

→ J. Mercer, *Geogr. Rev.* **55**, 390 (1965).

20. F. Bourdier, *Le Quaternaire du Bassin du Rhône* (C.N.R.S., Paris, 1963), vol. 2.
21. J. N. Rosholt, C. Emiliani, J. Geiss, F. F. Koczy, P. J. Wagersky, *J. Geol.* **69**, 162 (1961).
22. R. Fairbridge, Ed. *Physics and Chemistry of the Earth* (Pergamon, New York, 1961), vol. 4, chap. 3.
23. T. V. D. Hammen, G. C. Maarleveld, J. C. Vogel, W. H. Zagwijn, *Geol. Mijnbouw* **46**, 79 (1967).
24. R. P. Goldthwait, A. Dreimanis, J. L. Forsyth, P. F. Karrow, G. W. White, in *The Quaternary of the United States*, H. E. Wright, Jr., and D. G. Frey, Eds. (Princeton Univ. Press, Princeton, N.J., 1965), p. 737.
25. C. Emiliani, *Science* **154**, 851 (1966).
26. A. van Woerkom, in *Climate Change*, H. Shapley, Ed. (Harvard Univ. Press, Cambridge, Mass., 1953), p. 147; W. S. Broecker, *Science* **151**, 299 (1966).
27. W. Dansgaard, *Tellus* **16**, 436 (1964).
28. L. A. Lliboutry, "Ice flow law from ice sheet dynamics," *Int. Symp. Antarctic Glaciol. Expl. 1st* (Hanover, N.H., Sept. 1968), in press.
29. This does not mean that isotopic change of seawater (due to binding of large amounts of isotopically light ice) is unimportant for the paleotemperature scale developed by Urey (*30*), Epstein *et al.* (*31*), and Emiliani (*32*); on the contrary, Emiliani's estimate of this effect seems to be much too low (*33*; see *34*).
30. H. C. Urey, *J. Chem. Soc.* **47**, 562 (1947).

→ S. Epstein, R. Buchsbaum, H. Lowenstam, H. C. Urey, *Bull. Geol. Soc. Amer.* **64**, 1315 (1953).

32. C. Emiliani, *J. Geol.* **63**, 538 (1955).

→ E. Olausson, *Progr. Oceanogr.* **3**, 221 (1965);
→ N. Shackleton, *Nature* **215**, 15 (1967); W. Dansgaard and H. Tauber, *Science*, in press; W. S. Broecker, in *Quaternary of the United States*, H. E. Wright, Jr., and D. G. Frey, Eds. (Princeton Univ. Press, Princeton, N.J., 1965), p. 737.

34. C. Emiliani, *J. Geol.* **74**, 109 (1966).
35. R. Haefeli, *J. Glaciol.* **4**, 1133 (1961).

36. Financial support was provided by the Carlsberg Foundation, Copenhagen, and the U.S. National Science Foundation, Office of Antarctic Programs. We thank Dr. H. Tauber for help and suggestions.

3 March 1969; revised 23 May 1969

Atmospheric Aerosols

Abstract. *Measurements of particle counts and size distributions of atmospheric aerosols have been made at various locations by use of an instrumented aircraft. The number of atmospheric particulates is related to the visibility.*

Investigations have been made to determine the fate of atmospheric pollutants subsequent to their release from man-made sources such as power stations, smelters, and pulp mills as well as from natural sources such as forest fires. Gaseous and particulate concentrations have been measured in effluent plumes and in relatively clear air by flying an instrumented Cessna 182 aircraft in a series of plume-tracking patterns. We report here some of the findings with respect to particle numbers and sizes for various atmospheric regimes.

Sample air is taken in through a rubber tube protruding from the leading edge of the aircraft wing, and airflow rates are reduced by use of a number of inverted Y's. A dust counter (Bausch & Lomb) detects, classifies according to size, and counts the particulates. The dust counter was calibrated in the laboratory by the use of polystyrene latex particles with a refractive index of 1.56. Thus, the reported particle diameters are "equivalent diameters" based upon the laboratory calibrations (*1*).

Figure 1 shows an array of curves depicting particle size distributions found in various flights. Instrumental capabilities allowed for particle counting in the size ranges noted on the abscissa (Fig. 1). The upper counting limit is 36 particles per cubic centimeter, and the lower limit, although not so clearly established, is about 0.03 particle per cubic centimeter.

Curves 1, 2, and 3 illustrate the particle sizes and counts observed under conditions of excellent visibility in Santa Fe, New Mexico; Cedar Key on the Gulf of Florida; and Douglas, Arizona. Curve 4 shows the particle size distribution found at Bronson, Florida, which is about 35 miles inland from Cedar Key.

Data for Cedar Key and Bronson were taken within a 20-minute interval and illustrate the buildup in particulate concentrations that occurred over the Florida land mass. In this instance, there was a 12-fold increase in the number of particles 1 μm in diameter or larger from the coast to this inland location. Winds were from the south parallel to the coast at 15 knots, and there was substantial turbulence at the sampling altitude of 760 m.

Curve 5 shows the influence of urban air pollution in the metropolitan area of Houston, Texas. Measurements were made about 37 km downwind and at an altitude of 1800 m. These early-afternoon measurements were below the top

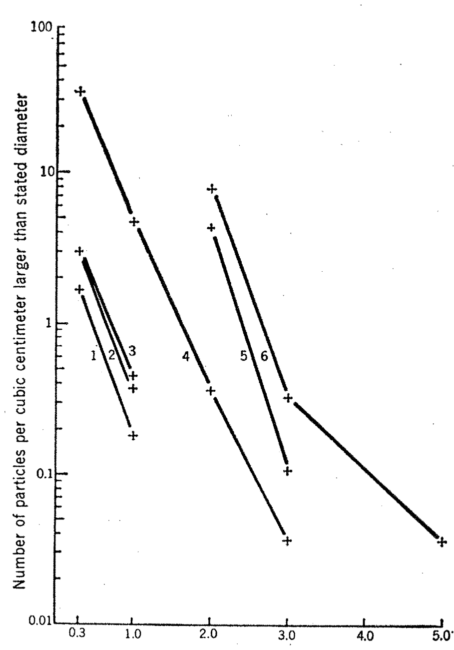


Fig. 1. Particle size distribution of natural and man-made aerosols. Data represent airborne samples from the following sources: curve 1, Santa Fe, New Mexico, 24 September 1968, 0920 hours, 2250 m; curve 2, Cedar Key, Florida, 19 December 1968, 1035 hours, 760 m; curve 3, Douglas, Arizona, 19 September 1968, 0715 hours, 1700 m; curve 4, Bronson, Florida, 19 December 1968, 1045 hours, 760 m; curve 5, 32 km north of Houston, Texas, 25 September 1968, 1300 hours, 1800 m; curve 6, Four Corners, New Mexico (in plume 6.5 km from power station), 23 September 1968, 0900 hours, 1800 m.