ditions of the experiment may not duplicate exactly those at which the bediasites formed. There is no evidence in the work concerning the linearity of the fractionation trend. It is possible that the O^{18}/O^{16} ratio reaches a maximum at intermediate values for SiO_2.

However, these data show that the vapor fractionation of "dry" melts can result in an increase of the O^{18}/O^{16} ratio. This provides an explanation of the correlation of this ratio with the content of silica in bediasites and suggests that the compositional trend in these tektites is due to a process of vapor fractionation, which is in accord with conclusions based on the compositional trends of the major oxides (*5*).

Note added in proof: Recently, one of us (R.N.C.) performed oxygen-isotope analyses on soda-lime glasses that had been heated for 16 hours at 1600°C. Since this material did not exhibit any measurable change in the O^{18}/O^{16} ratio, the possibility, mentioned in this report, that the O^{18}/O^{16} increase in the bediasites is due to equilibration with the atmosphere, is unlikely. Thus, the other alternative of enhanced volatility of O^{16}-bearing species seems more likely.

LOUIS S. WALTER
Laboratory for Theoretical Studies, Goddard Space Flight Center, Greenbelt, Maryland

ROBERT N. CLAYTON
Enrico Fermi Institute for Nuclear Studies, University of Chicago, Chicago, Illinois

References and Notes

1. H. P. Taylor, Jr., and S. Epstein, *J. Geophys. Res.* **67**, 4485 (1962).
2. ———, in *Isotopic and Cosmic Chemistry*, H. Craig, S. L. Miller, G. J. Wasserburg, Eds. (North-Holland, Amsterdam, 1964).
3. L. S. Walter and J. E. Giutronich, *Solar Energy J.*, in press.
4. E. C. T. Chao, in *Tektites*, J. A. O'Keefe, Ed. (Univ. of Chicago Press, Chicago, 1963).
5. L. S. Walter, in preparation.
6. Helpful discussion on the problem of oxygen isotope fractionation with S. Epstein of the California Institute of Technology is gratefully acknowledged. Work was supported, in part, by NSF grant GA-514.

7 February 1967

Climate Modification by Atmospheric Aerosols

Abstract. *Theoretical considerations and empirical evidence indicate that atmospheric turbidity, a function of aerosol loading, is an important factor in the heat balance of the earth-atmosphere system. Turbidity increase over the past few decades may be primarily responsible for the decrease in worldwide air temperatures since the 1940's.*

The possibility of deliberate or inadvertent modification of weather and climate from the emission of man-made pollutants into the atmosphere is receiving increased attention by the scientific community (*1, 2*). Of those constituents whose concentration and distribution have a bearing on the heat balance of the earth-atmosphere system, particular attention was drawn to CO_2 in a report (*3*) which concluded that the atmospheric CO_2 has been increasing over the past century and will continue to do so because of the continuing increase in combustion of fossil fuels. However, the worldwide cooling of mean annual air temperature, reported (*4, 5*) to have started between 1940 and 1950, discouraged the conclusion that there has been a direct effect on climate, that is, atmospheric warming. This enigma led to the further conclusion that " . . . climatic 'noise' from other processes has at least partially masked any effects on climate due to past increases in atmospheric CO_2 content" (*3*). Nevertheless, it was suggested that the possibility of inducing "countervailing climatic changes" should be explored by such means as raising the earth's albedo through the spreading of small buoyant reflective particles over large oceanic areas.

We may already be achieving this increased planetary reflectivity by the emission of fine particulate and aerosol pollutants into the atmosphere. The climatic significance of variations in the solar "energy albedo" of the earth by changes in the atmospheric turbidity was studied in some detail by Angstrom (*6*). Turbidity was expressed in terms of the Angstrom "turbidity coefficient" β, derived some years earlier (*7*) and defined by the empirical formula,

$$P_\lambda = \exp - \beta/\lambda^\alpha \qquad (1)$$

where P_λ is the solar transmission factor whose values range from 0 to 1 in accordance with the degree of scattering and absorption of solar radiation by atmospheric aerosol, α is called the wavelength factor (apparently related to a particle size), the reference wavelength λ is taken to be 1.0 μ, and β with limiting values lying between ∞ and 0 is related to the "dust" loading of the atmosphere. The quantity β is little affected by selective absorption of solar radiation by permanent gaseous components of the atmosphere, for example, water vapor, CO_2, and O_3, because of the techniques of determination (*7–9*).

While recognizing the dominating influence of the amount and albedo of clouds on the variations of the planetary albedo, Angstrom presented arguments to the effect that if the cloud amount remains constant, " . . . a change of 10 percent in the turbidity produces a change of about 1.5 percent in the albedo value, or about 0.8 percent in the energy available to warm the earth." Independent empirical evidence from data on solar transmission for the United States supports this conclusion (*10*). From 1962 through 1965, on days with 100 percent possible sunshine and no clouds reported (daylight hours) over Washington, D.C., and Cincinnati, Ohio, a 100 percent increase in turbidity produced a 5 percent reduction, on the average, in the transmission of solar radiation to the ground. This loss of transmission was not just that of the direct solar beam, but also that of the total hemispherical solar radiation incident on a horizontal surface, as measured by a 180-degree pyrheliometer.

The direct effect of the above process in lowering the temperature of the earth-atmosphere system cannot be determined within known limits of precision. Angstrom made a "rough estimate" (*6*) that a " . . . change of 1 percent in the albedo (from 0.40 to 0.41) corresponds to a change . . . (in the mean temperature of the earth) . . . of close to 1°C." Humphreys made similar calculations (*11*) with about the same results and also showed that the interception of outgoing radiation by fine atmospheric dusts " . . . is wholly negligible in comparison with the interception of solar radiation." Temporal and spatial changes in the atmospheric turbidity of 100 percent, that is, albedo change of 10 to 15 percent, from one day to the next or from one locality to another are very commonplace. Even though these figures may well overestimate the actual changes brought about in atmospheric temperatures, the course of atmospheric turbidity over the earth is an important climatic factor.

There are available data upon which the trend in turbidity during this century can be estimated. Angstrom (*7*) gave 0.098 as the value of the mean annual turbidity β at Washington, D.C. (1903–1907), and 0.024 as the value at the Davos Observatory, Switzerland (1914–1926). The values for Washington were determined from data on solar transmission (by wavelength) published by the Smithsonian Institution; those for Davos were from data attributed to Lindholm on dust absorption. In 1962, determinations of the atmospheric turbidity were begun at the Continuous Air Monitoring Program station (*12*) of the Public Health Service, just a few blocks away from the Smithsonian Institution; the mean annual turbidity for 1962–1966 was 0.154. From 1957 to 1959, determinations of atmospheric turbidity were again made for Davos by Valko (*13*) and were given as 0.043. The $\Delta\beta$ was 0.056 for Washington and 0.021 for Davos; the percentage of increase was 57 for Washington and 88 for Davos. The later Washington and Davos data are published in terms of the "Schüepp coefficient" B, but were transformed into β by the relation

$$B = 1.07\beta \qquad (2)$$

for average conditions (*8*).

When the scattering theory with a Junge distribution of particle size (*14*) is used, the values of $\Delta\beta$ imply an increase in the average annual number of aerosol particles, in the range of 0.1 to 1.0 μ radius, of 2.8×10^7 cm^{-2} and 1.05×10^7 cm^{-2} over Washington and Davos, respectively, during the periods shown. As Davos is at an elevation of somewhat over 1600 m, nearly two-thirds of the Washington increase might be attributed to the increased population and urbanization of the district since the turn of the century. A significant remainder, however, as judged by the Davos increase, may be indicative of a worldwide buildup of atmospheric aerosol.

The increase in atmospheric turbidity due to volcanic eruptions may have temporary effects on atmospheric temperatures. Mitchell (*4*) concluded that temperatures may be depressed " . . . as much as 0.5°F or more in the first or second year following an unusually violent eruption. . . ." However, we suggest that the effects of man's pollution of his environment are monotonically increasing along with the world population. The emission of long-lived aerosol, keeping pace with the accelerated worldwide production of CO_2 (*3*) may well be leading to the decrease in worldwide air temperature in spite of the apparent buildup of CO_2 In any case, it is clear that in this "large-scale geophysical experiment" in which human beings are engaged (*1*), the course of atmospheric turbidity must be documented with concern.

ROBERT A. MCCORMICK
JOHN H. LUDWIG
National Center for Air Pollution Control, Cincinnati, Ohio 45226

References and Notes

1. "Weather and climate modification: problems and prospects," *Nat. Acad. Sci. Nat. Res. Council Publ. 1350* (1966), vol. 2, pp. 82–108.
2. R. A. Bryson, *Saturday Review*, 1 April 1967, pp. 52–55.
3. "Restoring the quality of our environment," *Report of the Environmental Pollution Panel* (President's Science Advisory Committee, 1965), pp. 111–131.
4. J. Murray Mitchell, Jr., *Ann. N.Y. Acad. Sci.* **95**, 235 (1961).
5. ———, in "Changes of climate," *Arid Zone Research XX* (UNESCO, Paris, 1963).
6. A. Angstrom, *Tellus* **14**, 435 (1962).
7. ———, *Geogr. Ann.* **11**, 156 (1929); *ibid.* **12**, 130 (1930).
8. ———, *Tellus* **13**, 214 (1961).
9. ———, *ibid.* **16**, 64 (1964).
10. E. C. Flowers, R. A. McCormick, P. A. Starke, in preparation.
11. W. J. Humphreys, *Bull. Mt. Weather Observ.* **6**, 1 (1913).
12. *Atmospheric Turbidity Report* (quarterly mimeo.), National Center for Air Pollution Control, Cincinnati.
13. P. Valko, *Arch. Meterol. Geophys. Bioklimatol. Ser. B*, **11**, 143 (1961).
14. R. A. McCormick and D. M. Baulch, *J. Air Pollut. Contr. Ass.* **12**, 492 (1962).

2 May 1967

Atmospheric Ions and Germination of Uredospores of Puccinia striiformis

Abstract. *Atmospheric ions, identified by mobility characteristics, were associated with germination of lyophilized uredospores of* Puccinia striiformis *West. at Bozeman, Montana. Ions of intermediate size were highest in concentration, and percentage germination of spores was lowest during periods conducive to air pollution. In duplicate experiments at an isolated site near Barrow, Alaska, essentially all atmospheric ions were small ions and the fungus spores were consistently germinated near maximum.*

Uredospores of *Puccinia striiformis*, causal agent of stripe rust of wheat, are extremely sensitive to environment (*1*). They often germinate poorly in a so-called controlled environment during about a 4-month period centered on winter solstice, with wide fluctuations in day-to-day values (*2*). Furthermore, the daily germination values are related to external meteorologic conditions (*2*), and certain meteorologic factors can influence natural levels of ions (*3*). There is considerable information on the effects of generated ions on biologic systems (*4*), but there are few quantitative data on the biologic effects of naturally occurring air ions. In earlier studies, exposure of the spores to generated small ions of either charge did increase percentage germination (*2*). My study indicates that the germination process of uredospores of *P. striiformis* is influenced by naturally occurring atmospheric ions.

Techniques for measuring atmospheric ions have been described (*5*). In my study, Royco-412 ion collectors and accessories were used to allow monitoring within a range of mobilities from 0.0064 to more than 0.274 cm^2 volt^{-1} sec^{-1}. This range includes the intermediate and small ions of both polarity groups. Most comparisons were made between negative ions having a mobility of 0.03 to 0.274 cm^2 volt^{-1} sec^{-1} and negative ions having a mobility exceeding 0.274 cm^2 volt^{-1} sec^{-1}.

Spores from two collections of the fungus were dusted onto polyethylene membranes at the same time each day and were exposed to simulated dew at 5°C over a 30-day period from 5 January to 3 February 1967. Concentrations were adjusted to give 20 to 40 spores per microscope field at × 100. After 6 or 24 hours, percentage germination was determined by microscope examination of a total of more than 800 spores contained on two membranes per treatment. Any spore having a germ tube longer than the spore diameter of about 20 μ was considered germinated. All spores within a collection were harvested at the same time, were lyophilized shortly thereafter, and had the same germination potential throughout the assay period. Spores contained within separate hermetically sealed lyophil tubes were used for each day's assay. External air introduced to the germination chambers and to the ion collectors was not filtered. Duplicate trials were conducted on the university campus at Bozeman, Montana, and in a wanigan located on the tundra about 1.6 km from the Arctic Research Laboratory, Barrow, Alaska.

In the Bozeman trials, the ratio of small to intermediate ions varied greatly