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A Perspective on Climatic Change

Climate responds rapidly and significantly to small changes of the independent variables.

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In recent years, with the heightened concern about the impact of human activities on the environment and the more immediate concern about the impact of the environment on world food supplies, there has been an upsurge of interest in climatic change at all scales, from local to global. Discussions of climatic change in the general press and professional journals have ranged from completely theoretical to completely descriptive, and from detailed analysis of certain aspects of the problem to general speculation. However, there has been a dearth of discussion of climatic change from an historical perspective and minimal attention to certain important questions, such as: (i) How large must a climatic change be to be important? (ii) How fast can the climate change? (iii) What are the causal parameters, and why do they change? (iv) How sensitive is the climate to small changes in the causal parameters? This article is an attempt to provide an element of perspective on these questions.

How Big Is an Important Change?

Climatic variations range in size from those which mark the onset or end of a glacial epoch down to those which are questionably detectable by the present meteorological observing network. Clearly, variations that bring on ice ages are important. An ice sheet 4000 kilometers wide and 3 kilometers high, reaching from the Arctic to Ohio and from the Rockies to the Atlantic, certainly represents a significant modification of the North American environment. A concomitant worldwide reduction of sea level of 100 to 150 meters is also significant if one looks at a bathymetric chart. But how much change of climate is necessary to initiate or terminate such conditions? There are a number of estimates based on field data to answer this question (*1*).

Oxygen isotope analyses of deep sea cores in the tropics suggest glacial to postglacial changes of temperature of the order of a few degrees Celsius (2). Similar analyses of Greenland ice cores suggest temperature differences at high latitude of about 10°C (*3*), and analyses of pollens from Minnesota give a July mean temperature difference of 3° to 4°C (*4*), as shown in Fig. 1. Most careful analyses suggest a mean global surface temperature difference from full glacial to the present of 4° to 6°C, but with smaller and larger changes in specific areas. Most of the changes deduced from the geologic record, for specific places, fall within the range of individual monthly anomalies for the historical record.

Figure 1 gives not only the magnitude of the temperature change between glacial and postglacial, but an indication of the rapidity with which the climate can change to extreme conditions. The time resolution of most records of the Pleistocene-Holocene break is two to five centuries, but the few records which have finer resolution suggest that most of that dramatic climatic change occurred in a century or two at the most. Indicators such as biotic assemblages and areal extent of continental glaciers have time constants that tend to mask the rapidity of the changes (*5*).

On the scale of Fig. 1, it is clear that the character of the past ten millennia (the Holocene) is quite different from that of the preceding millennia (the Pleistocene) in terms of temperature, and that the mean temperature fluctuations within the Holocene are an order of magnitude smaller than the differences between epochs. One should not assume, however, that the fluctuations of climate during the Holocene were unimportant. The records of summer temperatures reflect radiation conditions dominantly. Other climatic parameters show more differentiation of the Holocene into distinctive episodes.

Consider, for example, the length of the growing season, precipitation during the growing season, annual hours of sunshine, and annual snowfall that parallel the July mean temperature for a location near Minneapolis, Minnesota, represented in Fig. 1 (Fig. 2). These plots are for a particular location, but, considering the integrity of the hemispheric circulation pattern, the postglacial climatic changes indicated here must have been associated with changes over large parts of the world. Clearly, the Holocene has not been a time of unchanging climate, even though the changes were smaller than those which terminated the Pleistocene. The period around 8000 years ago in southeastern Minnesota was only slightly warmer in July, but it was cloudier, drier in summer, less snowy in winter, and had a somewhat longer growing season than at present. In fact, if these reconstructions are correct, the climate was rather like that of Huron, South Dakota, at present. Certainly, in

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terms of agriculture, this represents a significant difference.

The last 2000 years is not only a period for which we have a great deal of historical information, but it is also a distinct climatic period within the Holocene. However, the time resolution of the data for Figs. 1 and 2 is too coarse for us to explore the question of climatic changes over shorter historical periods and their significance. There is a suggestion of such changes in those figures but too much uncertainty as to timing. In order to get more resolution we may turn to the work of Lamb (*6*) and Bergthorsson (*7*).

Bergthorsson has reconstructed the decadal mean annual temperature for Iceland over the past 1000 years from the historic records of the duration and extent of sea ice on its shores. By calibrating the ice record against the observed climatic data for the past century and a half, he was able to derive a regression equation that makes possible a rather credible decadal temperature plot for the past millennium (Fig. 3).

Figure 3 puts the present and the climatic "normal" period 1931 to 1960 into perspective. The normal period is normal only by definition. There appears to be nothing like it in the past 1000 years.

Figure 3 also raises two pertinent questions: (i) How important are the mean temperature variations shown? (After all, the whole range is less than 2°C.) (ii) How representative is Icelandic temperature of climate other than that of Iceland? (It's a very small country.)

The significance of small mean temperature changes in Iceland is rather easy to demonstrate, thanks to the excellent documentation summarized by Thorarinsson (*8*) and others. Famines are noteworthy events in a subsistence economy, and if a chronicler describes the death of animals and people in winter and failure of the few crops and pastures in summer because of the cold, there is no a priori reason to doubt him.

According to Thoroddsen (*9*), there were 12 famine years between A.D. 975 and 1500, and 37 between A.D. 1500 and 1804. Of the first group of famine years, five were in the cold period between A.D. 1250 and 1390, and of the second group, 34 were between A.D. 1600 and 1804. The famines are clearly associated with the cold periods. This sensitivity of food production to small changes of climate may be illustrated by considering the nature of the agriculture in Iceland and the proportional effect of small mean temperature changes, as shown in Fig. 4.

The mainstay of Icelandic agriculture is grass and the flocks raised on it. In a climate of ample moisture the forage yield is approximately proportional to the number of growing degree-days above a base of 5°C. Figure 4 shows schematically that a decrease of 1°C in the mean annual temperature of Akureyri, Iceland, would reduce the length of the growing season by about 2 weeks, but would reduce the number of growing degree-days by 27 percent. A reduction of mean annual temperature by 2.4°C would reduce the number of growing degree-days by 54 percent and the growing season by 40 days (25 percent). This would be disastrous to a subsistence economy.

In the late 1950's, hay yields in Iceland averaged 4.33 metric tons per hectare with 2.83 kilograms of nitrogen fertilizer per hectare and a mean warm-half-year temperature of 7.65°C. In 1966 and 1967, the yields averaged only 3.22 metric tons per hectare with 4.83 kg of fertilizer, but with a mean temperature of 6.83°C. The climatic reduction of yield overshadowed the expected technological increase (*10*).

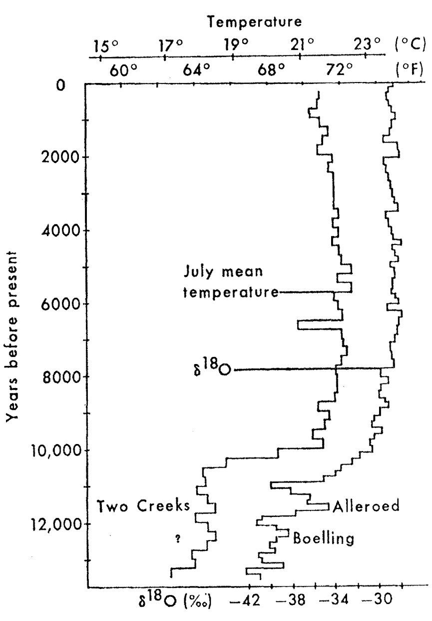


Fig. 1. July mean temperatures at Kirchner Marsh, Dakota County, Minnesota, compared with the oxygen isotope variation in Greenland over the past 13,000 years. The temperatures are from pollen data; the isotope variation is expressed as the per mil enrichment in ^{18}O, or $\delta^{18}O$, and should be proportional to the snow season mean temperature. Two Creeks, Alleroed, and Boelling denote late glacial intervals. Pollen data are from Wright *et al.* (*42*), analyzed by the method of Webb and Bryson (*4*), and isotope data are from Dansgaard *et al.* (*3*).

This discussion highlights the significance of small temperature changes in cold-limited, marginal northern agriculture, but does not show that climatic differences, such as those between the past century and the present, are important in a broader context. Along with changes of temperature, which are associated with changes of atmospheric circulation, there are also changes of precipitation. Wahl and Lawson (*11*) have documented the changes of seasonal temperature and precipitation from the period 1850 to 1870 to the modern "normal" period.

Most interesting is the change in the western half of the United States. In the earlier period the high plains and Rockies were up to 20 percent wetter for the whole year and 20 to 30 percent wetter in summer. Consider the high plains, which are currently semiarid range land, in terms of carrying capacity, and the vast bison herds of the last century. Assuming that the carrying capacity of the range for bison is proportional to the carrying capacity for cattle, figures given in the *Yearbook of Agriculture* (*12*) suggest that the bison herds would have diminished by 50 to 75 percent as the precipitation diminished in the late 19th century, even without overhunting.

At the present time of world food grain shortage, a return to the higher rainfall and range carrying capacity of the mid-1800's would be most significant, and yet such changes seem to be small compared to the intermillennial changes of the Holocene. A climatic change does not have to be large in absolute numbers to be important.

There is another way in which small climatic changes are important, and this is related to the physiology of trees and probably most perennials.

During one growing season a tree normally stores photosynthates in excess of that year's growth needs. This provides a reserve which is drawn on the following year and makes it possible for the tree to survive a single severe year. Pine trees also retain their needles for a number of years. These integrative mechanisms mean that the variance spectrum of tree response to climate should show suppressed response to high-frequency climatic variations, but accumulated response to small systematic changes in climate, and such spectra do (*13*). Thus, trees can somewhat ignore large seasonal anomalies but integrate the effect of small systematic changes into the biomass of the forest. It is not realistic to say that, because trees survive the large interannual variations, climatic mean changes are unimportant.

Glaciers provide a clear-cut example of the significance of small systematic changes in climate. If ablation over the year is equal to or potentially slightly greater than the snowfall, there is no glacier. But if accumulation is greater than ablation by even a tiny amount, the glacier will inexorably grow. In the Canadian Arctic, on the Keewatin ice divide north of Baker Lake, snowdrifts last until late summer and the snows begin in September. August is cool and often cloudy. A slightly cooler summer that would reduce ablation slightly, or somewhat heavier winter snows—just enough to make the snowdrifts last a month longer, would let the September snows begin to accumulate on the residual of the previous year. Continued, this would inevitably result in a glacier. The region is marginally interglacial at present.

Reviewing Figs. 1 and 2, and dozens of similar diagrams, one finds that the past analog of the present climate is to be found in the transition from Pleistocene to Holocene; that is, the present climate is marginally interglacial.

According to Mitchell's work (*14*), the present episode of warmer temperature in the Northern Hemisphere began about 1920. It appears to be terminating at present (*15, 16*). Before 1920, stations in northwestern India experienced years with less than half of normal rainfall with a probability of .116 or a "return period" of around 8.6 years. From 1920 through 1960 the probability of such a dry year was .071 for a return period of about 14 years. During the 1930's, the warmest decade of this century in terms of mean temperature in the Northern Hemisphere, the probability was .055 or about once in 18 years. If the climate returns to that typical of the period before 1920, will a half-again-as-probable recurrence of drought be significant in that hungry land?

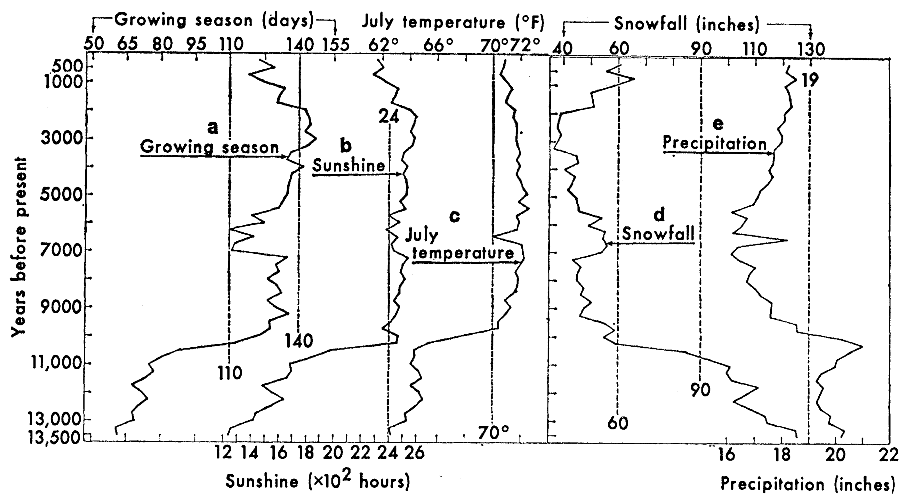


Fig. 2. Thirteen-thousand-year record of (a) length of the growing season, (b) hours of bright sunshine annually, (c) mean July temperature, (d) annual snowfall, and (e) precipitation during the growing season at Kirchner Marsh, Minnesota. The climatic data are derived from pollen data by canonical transfer functions.

How Big Are the Causes of Climatic Change?

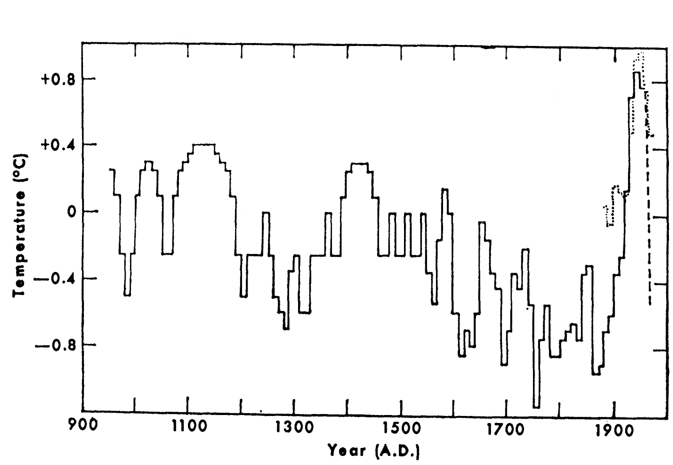
Considering the small absolute magnitude of significant climatic change, one might conclude prematurely that the climate is rather insensitive to changes in causal factors, or that there is some stabilizing influence, such as the oceans, which reduces the impact of changes in the causal factors. If this were the case, we could simply monitor those factors and anticipate when the changes were becoming big enough to be of concern. Let us consider whether this is the case.

As an extreme instance, assume that the entire Northern Hemisphere is covered by water and that the thermally active layer of this sea is 100 m deep. How much heat must be stored in this layer of water to produce a temperature rise of 0.5°C per 20 years? This figure is picked to match the increase of the mean surface air temperature in the Northern Hemisphere observed between 1920 and 1940. Simple arithmetic gives a figure of about 0.7 calorie per day per square centimeter of surface. Certainly, this is not currently measurable at any one point, and it is not likely that a rise in mean sea temperature of 0.1°C in 4 years would be readily monitored. Making the case more realistic by introducing continents of lower heat capacity reduces the heat storage required to simulate the observed change.

Reitan (*17*) estimates that the changes of solar radiation required to produce the observed temperature changes over the past century are of the order of 2 cal cm^{-2} day^{-1}. In order to monitor changes of 50 percent in this number, which might be produced by fluctuations in the solar constant, the solar constant would have to be measured with an accuracy of 1 part in 700. Judging from the variety of published values, one would conclude that the solar constant is not known to that accuracy, even though each published value may be assigned higher accuracy.

Assuming that the solar input of heat to the earth and back radiation from the earth were equal in 1920 and again in 1940, we may calculate how much

Fig. 3. Mean annual temperature in Iceland over the past millennium [data from Bergthorsson (*7*)]. The dashed line indicates the rate of temperature decline in the period 1961 to 1971, and the dotted line shows the variation of mean temperature in the Northern Hemisphere plotted to the same scale.



the albedo of the earth-atmosphere system would have to have diminished to account for the mean temperature rise of 0.5°C. A decrease from 35 to 34 percent would be more than adequate, yet we cannot measure the overall albedo with sufficient accuracy to know whether such changes have occurred or are occurring. To know whether the observed temperature change at the earth's surface was the result of changes in the greenhouse effect would require knowing the difference between the upward radiation from the surface of the earth and the upward radiation from the top of the atmosphere to an accuracy of about 1.6 percent, or about 0.004 cal cm^{-2} min^{-1}. This is probably beyond our present capability of measurement, but perhaps within the range of calculation as a relative change, if we assume that the parameters needed for calculation are known with sufficient accuracy.

The point of all this discussion is to emphasize the small magnitude of the variations in climatic causal factors needed to make significant, though small, climatic changes.

Of course, it may be argued that the global mean temperature is not really very important compared to changes in circulation pattern that might affect rainfall patterns.

Some light may be thrown on this question by considering a criterion derived by Smagorinsky (*18*). The defining equation is:

$$\tan \phi_z = -(H/R)\left(\frac{\partial\theta}{\partial z}\bigg/\frac{\partial\theta}{\partial y}\right) \qquad (1)$$

where ϕ_z is the latitude at which wave number one of the zonal flow becomes dynamically unstable (which in turn approximates, in practice, the latitude of the transition from Rossby regime to Hadley regime); H is a scale height, roughly the height of the 500-millibar level of the atmosphere; R is the radius of the earth; $\partial\theta/\partial z$ is the vertical lapse rate of potential temperature; and $\partial\theta/\partial y$ is the north-south gradient of potential temperature.

Empirically, ϕ_z is a good approximation to the subtropical anticyclone latitude at the surface as well as the latitude of the Rossby-Hadley transition aloft. Since the latitude of the subtropical anticyclones is critical to the great subtropical desert regions, the monsoon regions, and the equatorward portions of the Mediterranean climates, ϕ_z is an important circulation parameter.

The greenhouse effect of water vapor

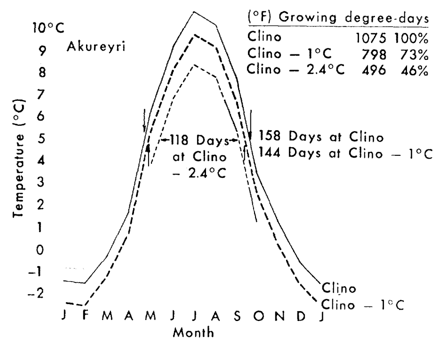


Fig. 4. Effect of reduction of the annual mean temperature on the length of the growing season and the number of Fahrenheit growing degree-days. The figures apply to Akureyri, Iceland. *Clino* means climatic normal, here the normal annual mean temperature. The figures at the upper right show that at 1°C below *clino*, for example, there are 798 growing degree-days, or 73 percent of the normal number.

and carbon dioxide affects the surface temperature and upward radiation from the earth's surface but at equilibrium with the solar input cannot affect the effective radiation out of the top of the atmosphere. Defining the magnitude of the greenhouse effect as the difference between upward infrared radiation from the earth's surface and upward infrared radiation from the top of the atmosphere gives a number of the order of 0.024 cal cm^{-2} min^{-1}. Thus, a 1 percent change in the greenhouse effect would change the surface temperature about 0.3°C and the vertical lapse rate about 0.8 percent.

Logarithmic differentiation of Eq. 1 shows that a 1 percent change in either the vertical lapse rate or the horizontal temperature gradient changes the tangent of ϕ_z by 1 percent. For summer subtropical anticyclones, this is about 0.2° of latitude.

Now if we plot the latitude of the northern boundary of monsoon rains in West Africa [which is at the intertropical discontinuity (ITD) between moist monsoon air and dry desert air] against the latitude of the subtropical anticyclones, we find that there is about a three-to-one amplification, so that moving the subtropical anticyclone 0.2° farther south moves the ITD 0.6° southward (*19*). Since Ilesanmi (*20*) has shown that the annual rainfall gradient south of the ITD is about 180 mm per degree of latitude, it would appear that a 1 percent change in the greenhouse effect would move the ITD southward by 0.6° and reduce the West African monsoon rainfall by over 100 mm.

By similar reasoning, we can examine the effect of a change of 0.4 percent in the solar radiation intensity, which was suggested by Reitan (*17*) as adequate to explain the rise of mean hemispheric temperature (0.5°C) from 1920 to 1940. This change was mostly in the higher latitudes, with essentially no change in the tropics. Assuming that there was a 1.0°C rise at 60°N and none at 20°N gives a decrease in the north-south gradient of about 0.24°C per 1000 km, for a change of about 2.4 percent in the mean. In turn, a 2.4 percent change in gradient would mean a 2.4 percent change in tan ϕ_z, or a change in the latitude of the anticyclones of about half a degree and a change of the West African rainfall of 270 mm/year.

Before about 1945 these two effects should have been opposed, but since then they would be additive in reducing sub-Saharan West African rainfall.

If we estimate the warming of the earth's surface due to increased carbon dioxide as 0.1°C between 1957 and 1970, and the increase of the north-south temperature gradient as 0.05°C per 1000 km during the same time, the figures above would suggest a decrease of precipitation of 86 mm annually in the Sahel of West Africa. According to Winstanley (*21*), the decrease averaged for five Sahelian stations was 96 mm between 1957 and 1970. This is a disastrous decline in an arid region, but with miniscule causes if the reasoning above is correct. It suggests that our climatic pattern is fragile rather than robust.

What Changes the Climate?

Practically every elementary textbook of meteorology or climatology contains a description of the atmosphere as a heat engine driven by the sun. A primary or general circulation results from the distribution of heating and cooling, and this, interacting with the terrain (in the broadest sense) and generating large-scale waves and eddies, produces the climatic pattern of the world. The dynamic internal mechanism and details are the essence of meteorology, but we shall focus here on the heat engine controls extrinsic to these internal mechanisms of the atmosphere itself. The extrinsic variables are (i) the intensity of sunlight reaching the earth; (ii) the transmittance of the atmosphere as modified by processes not internal to the atmosphere; (iii) the albedo of the earth-atmosphere system or the earth's surface, again as modified by noninternal processes; and (iv) the greenhouse control of infrared fluxes from the earth, as modified by gases and particulates not depending directly on atmospheric processes.

Intensity of sunlight—the solar constant. The evidence is weak that the intensity of the sunlight leaving the sun is variable on the scale of years to thousands of years. There may be changes of solar output on this scale, and indeed there are measured changes in some wavelengths, but the total energy that ultimately drives the atmosphere does not seem to be variable within the limits of observational accuracy. There is now an opportunity to establish the fact of the matter from space platforms, but past measurements —even from mountain observatories— are not unequivocal. Each earth-based observation must be corrected for large atmospheric effects, and thus the source of the variance is difficult to determine.

Whether or not there are short-term variations in solar output, there are calculable variations of solar input to the top of the atmosphere arising from the varying distance from the sun to the earth, and changes in seasonal and latitudinal distribution of the input depending on the variation in inclination of the earth's axis to the plane of the ecliptic and the axis of the earth's orbit. These variations were studied extensively by Milankovich (*22*) and more recently by Broecker (*23*) and Kutzbach *et al.* (*24*), among others. Such variations are of great interest in the study of recurrent ice ages, but the time scale of the cycles involved is of the order of 10^4 to 10^5 years, and they contribute little to our understanding of decadal to millennial changes.

There is a modifiable heat input, however, that simulates, to a certain extent, variations in solar input; that is, "waste heat." The term itself is rather anthropocentric, however, for essentially all energy used by man is ultimately degraded to heat, which is added to the atmosphere. The local effects of this waste heat on city climates, for example, is widely known and has been extensively discussed [for example, see (*25*)]. Less is known about global climatic effects of waste heat (*26*), but it is estimated as 3.8×10^{-5} times the average solar input to the atmosphere-earth system. However, it is mostly released as sensible heat and is about 5.8×10^{-4} times the natural sensible heat flux at the earth's surface, and is released almost entirely over the continental areas. For midlatitude regions of the continents, this figure should probably be multiplied by a factor of 6 to 10. This is still a very small fraction of the heat from natural sources, but secondary effects of this energy use have a larger impact, as will be indicated below.

Transmittance. Since the climate is primarily determined by processes near the earth's surface, the solar energy arriving at the top of the atmosphere is really not as important as that reaching the lower atmosphere. If the physical properties of the atmosphere and the earth were invariant, then about the only cause of climatic variation would be variations in the input from the sun at the top of the atmosphere. However, the transmittance of the atmosphere varies with time, and this variation modulates the receipt of solar energy in the lower atmosphere and at the surface.

If changes of water vapor and cloud are regarded as internal to the climate and part of the climatic response rather than a direct cause, the major variation of transmittance due to external causes is the result of addition of particulate material to the atmosphere.

There has been a great deal of discussion of the relative importance of various sources of particulate material in the atmosphere, both primary particles and those derived from gaseous or liquid precursors. It is hard to see how the particulate turbidity of the atmosphere would vary on a time scale of decades to millennia without climatic change being the cause of the variation, if the sole source of the particulates were natural atmospheric processes operating at the surface of the land or sea, such as deflation of soil material from deserts, evaporation of sea spray, or emission of terpenes from forests.

I assume in the following discussion that there is a level of natural turbidity normal to each climatic state, and that variations from this level are due to external sources such as volcanoes, direct man-made sources such as fossil fuel fires and machines, and human modification of natural sources such as increased soil disturbance, increased forest fires, and slash-and-burn agriculture. For example, Rex and Goldberg (*27*) have shown that there has been much eolian transport of dust throughout the Pleistocene and Holocene, but there have been no adequate studies of the increase in the atmospheric loading of soil material resulting from human disturbance of the surface. There are indications that it is of considerable magnitude (*16*).

Historically, by far the greatest extrinsic source of particulates has been volcanic activity. Volcanic activity is variable in the right frequency range to match climatic change, the magnitude of the injections of volcanic dust is right, and significant changes of atmospheric transmittance have been measured following eruptions. The evidence of Hamilton and Seliga (*28*) is quite convincing. They have shown that the temperatures over the Greenland and Antarctic ice sheets have been inversely proportional to the amount of volcanic dust falling on those ice sheets over the past hundred millennia or so. Bryson (*16*) and Reitan (*17*) have shown that in the past century a major control of the mean temperature in the Northern Hemisphere has been volcanic dust augmented by a human contribution.

A particular feature of volcanic activity is that it is sporadic, occurring in pulses, although the frequency and intensity of these pulses vary greatly (*29*). If the colder periods of recent earth history have been due to pulsed volcanic injections that reduce atmospheric transmittance, then these colder periods should have more temperature variance than the warm periods of volcanic quiescence. This is evident in Fig. 3.

Direct measurement of the solar radiation normally incident at the ground under a cloudless sky (*30*) shows a variation in transmittance during the past century, with the highest values in the warm period of the 1920's and 1930's, when volcanic activity was at a minimum. However, both the mean temperatures and the measured radiation intensity in the Northern Hemisphere began to decrease before the resumption of volcanic activity in the 1950's and 1960's. This suggests another source of particulates, which should have become significant by 1940.

If the industrialization and mechanization of the world are a source of particulate material, then there should have been a nearly exponential increase in the atmospheric loading of the by-products of this trend in the middle third of this century. The lead fall on the Greenland ice cap shows this effect (*31*) (see Fig. 5). The most rapid increase began about 1940. A

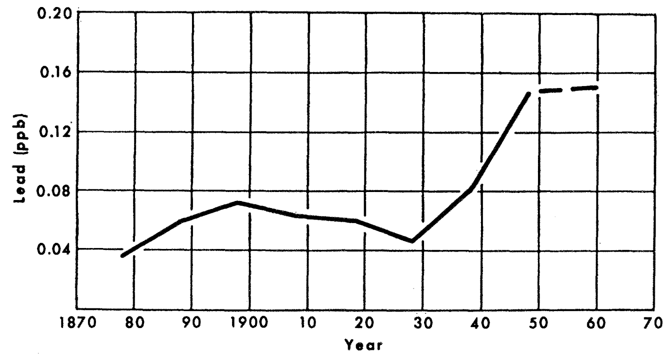


Fig. 5. Variation of anthropogenic lead dust fall at Camp Century, Greenland, from about 1880 to 1960, showing the rapid recent increase in man-made particulates in the atmosphere. Lead is measured as parts per billion (ppb). [After Murozumi *et al.* (*31*)]

similar rapid increase of dust fall on the Caucasian snowfields began about 1930 to 1940, indicating at least a rapid increase in soil deflation in eastern Europe or the Near East, probably due to increased mechanization of agriculture (*32*). The population of the world has doubled, with especially rapid growth in the subtropics, since 1940. In the drier monsoon lands this has led to overuse of the land and extreme dust loading of the atmosphere (*16*), and, in the wetter regions, an increased slash-and-burn rotation rate with increased production of smoke.

Again assuming that the climatic complex of patterns, and its internal oscillations and variations, changes in response to extrinsic variables, then the major control of secular changes of atmospheric transmittance should be injection of particles of volcanic origin and of direct and indirect human origin. Robinson and Robbins (*33*) estimate the extrinsic particulate inputs into the atmosphere annually as 4×10^6 metric tons of fine volcanic dust and 296×10^6 metric tons from human pollutant emissions. To these figures may be added my own estimates, based on field observation and measurement, of (40 to 60) $\times 10^6$ metric tons from smoke from slash-and-burn agriculture and (100 to 250) $\times 10^6$ metric tons from deflation of soil disturbed by agriculture and construction.

Of course, the actual atmospheric loading of particles and its variation with time is the critical parameter in climatic change. Peterson and Junge (*34*) quote an estimated 4×10^6 metric tons of volcanic "dust" particles in the troposphere and an equal amount, averaged over the past 120 years, in the stratosphere. Since volcanic eruptions are sporadic, the amount of volcanic dust in suspension must vary from nearly zero during the volcanically quiet period between 1920 and 1955 to very large quantities during the years immediately following the eruptions of volcanoes like Tambora (1815), Krakatoa (1883), or Agung (1963). Such volcanoes can eject as much as 100×10^6 metric tons of fine ash in a very short time. This variable input produces a variable transmittance of the atmosphere, which produces climatic variation.

Human inputs of particulates into the atmosphere are not sporadic—nearly everyone on the earth works at it nearly every day—and the number of individual sources is so myriad that the day-to-day fluctuations are averaged out. It is a variable input, however, for the number of people keeps growing, as do the extent of mechanization and industrialization and the extent and intensity of agriculture. Pollutant emission tends to be proportional to energy use and economic level (*35*), and both of these parameters are growing rapidly.

An important question is how much of the man-generated particulate material remains in the atmosphere at any given time. Estimates range from 10^{-3} (*36*) to 4×10^{-2} (*26*, p. 201). The former appears far too small, for it requires that the atmosphere be completely scavenged of particulates every 8 or 9 hours on the average. A figure of 2×10^{-2} to 4×10^{-2} is more reasonable. This would yield a particulate loading of the atmosphere due to human activities of roughly (8 to 24) $\times 10^6$ metric tons; that is, an amount comparable to that produced by moderate volcanic activity. According to Barrett (*37*), an increase of 2×10^6 metric tons in atmospheric loading is capable of reducing world mean temperatures by 0.4°C. Thus, both men and volcanoes appear adequate to account for the hemispheric cooling since 1940.

According to Machta (*38*), most of the recent increase of turbidity has been at middle to high latitudes. This, plus the greater path length of sunlight through the atmosphere at higher latitudes, should have produced selective cooling of these latitudes, and that is what has been observed (*26*, p. 44).

The major effect of particles in the atmosphere (turbidity) on transmittance is through the backscatter of solar radiation. This affects the albedo of the earth (*39*). There are other factors also which may modify the albedo.

Albedo. Much has been made of the effect that variable cloudiness would have on the climate of the earth (*40*). However, cloudiness is part of the climate and responds to climatic change in a feedback loop. It is not an extrinsic factor, except perhaps for the condensation trail.

Other than variations in cloudiness and turbidity, two factors may modify the albedo: snow cover variations and changes in the surface characteristics of the continents. Snow cover, like cloudiness, is a response to climate and is part of the climate. Man, as an extrinsic factor, can change the albedo of the earth's surface by changing the nature of the surface from natural cover to crops and cities, and so forth. However, a rough calculation suggests that this effect is of minor significance. About .05 of the overall albedo of the earth is contributed by the surface, of which about 30 percent is land. Of the land, about 12 percent is agricultural or urban. A reasonable change of albedo due to changed land use is 15 percent, and a generous estimate of the rate of change of albedo with changing land use is 5 percent per decade. This gives $.05 \times .30 \times .12 \times .15 \times .05$, or 1.35×10^{-5} per decade, or a little over a thousandth of a percent per decade as the albedo change. According to Manabe and Wetherald (*41*), this would produce a mean surface temperature change of about 0.001°C per decade. It does not appear to be comparable to other extrinsic climate-changing factors.

Greenhouse effect. The solar energy which is not reflected from the earth-atmosphere system (65 percent) is absorbed at the earth's surface (47 percent) or in the atmosphere (18 percent). This absorbed portion must be reradiated if the global climate is to remain constant. Since the infrared reradiation to space is dependent on the effective temperature of the earth-atmosphere system, if the solar energy arriving at the top of the atmosphere and the overall albedo remain constant, then the effective radiating temperature of the system must remain constant also. This does not mean a constant surface climate of the earth, however.

A change in the composition of the atmosphere may change the distribution of temperature within the atmosphere (*41*).

The constituent gases of the atmosphere that are important variables are water vapor and carbon dioxide. They both have an important role in modifying the vertical distribution of temperature in the atmosphere by controlling the flux of infrared radiation. It is this control which is called the greenhouse effect. However, the water vapor content of the atmosphere responds to the climate (with feedback, of course), and Manabe and Wetherald (*41*) imply this in the assumption of constant relative humidity. Carbon dioxide, on the other hand, is an extrinsic variable, generally regarded as responding not to climate but to consumption of fossil fuels (*38*). If this is indeed the major source of carbon dioxide variation, then it must be a cause of climatic change of rather recent origin in the perspective of climatic history.

The concentration of carbon dioxide in the atmosphere is currently about 320 parts per million (ppm) and is increasing about 1 ppm annually (*38*). The evidence suggests that this rate will also increase, and is greater now than a decade ago, and the estimated effect of this is that the global mean surface temperature will rise roughly 0.01°C per ppm carbon dioxide increase (*41*). In the last century the carbon dioxide content of the atmosphere changed from about 292 to about 320 ppm. If this were the only cause of climatic change, the mean global surface temperature should have risen steadily and smoothly, at an increasing rate, by about 0.25°C. It has not, and in fact has decreased by about that amount since 1940. There clearly are other factors of greater magnitude that dominate climatic change in terms of mean global surface temperature.

Probably more important is the role of carbon dioxide variation in changing the static stability of the atmosphere. Since the effect of changing carbon dioxide becomes nil at an altitude of about 12 km, a change of 0.01°C/ppm at the surface will change the lapse rate about 0.001°C km^{-1} ppm^{-1}. As pointed out in a preceding section, the latitude of the subtropical anticyclones depends in part on the vertical stability. So do the vertical motions of the Hadley circulation. These minute changes appear to be capable of producing very important shifts in the distribution of rainfall.

Summary

Throughout the preceding discussion certain assumptions have been made, and certain others were implicit. It has been assumed that variations of the gross energy flow controls extrinsic to the atmosphere are adequate to explain climatic variations without seeking some "trigger" mechanism of small size, such as a burst of solar particles which modifies the high atmosphere and, in turn, the low-level climate. This does not imply that there is no short-term atmospheric effect, but that there is simply no climatic effect of significance compared to those which relate to the all-important solar energy which drives the atmosphere.

Also implicit is the assumption that climate is thermodynamically forced and that the overall hydrodynamic pattern of waves, eddies, momentum fluxes, and the like is an internal response to this forcing.

Another implicit assumption has been that there is only one climatic pattern that is appropriate to each configuration of the extrinsic control variables, at equilibrium. In the perspective of climatic history as we know it at present, there appears to be no need to adduce alternate quasi-stable patterns as responses to a given set of inputs.

At the beginning of this article I posed certain questions about the size of important climatic changes and the magnitude of changes in the causal parameters needed to produce these changes. I showed that numerically small changes in climatic variables may produce significant environment changes and that rather small changes in the extrinsic control variables are adequate to explain these responses. A significant feature of recent paleoclimatic research is that significant climatic pattern changes are surprisingly small—explanation of past environments does not require drastic modification of the general circulation.

I also suggested that several of the extrinsic control variables may be significantly modified by human activity. These include the turbidity of the atmosphere and its carbon dioxide content.

The data presented in several of the figures show that climate can change very rapidly. While some causal parameters, such as earth-sun geometry, change slowly, others, such as volcano-induced turbidity, may change rapidly and sporadically. Apparently the only controls of the speed of climatic change are the time constant of the active layer at the surface of the earth and the time constant of glaciers. Ice age climates may end (and probably start) in a century or two, although glacial and oceanic response and a new equilibrium may take millennia. Holocene climatic changes, smaller in magnitude, may be accomplished in decades. The overriding present question, of course, is how the present climatic change will develop. In perspective, such changes do not appear to be random fluctuations from some long-term "normal."

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Biosynthesis of Natural Products

Problems of substrate nonpermeability and location of isotopic label are described.

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Natural products, for many years the touchstones of structural and synthetic organic chemists, continue to be discovered from various sources; and 1000 structures are added annually to a storehouse of some 12,000 described molecules whose molecular weights are between 20 and 2000. These compounds are usually referred to as "secondary" metabolites because their biochemical function in the host organism is in most cases obscure. Yet in the few cell-free systems that have been developed to study the biosynthesis of natural products, the enzymology appears to follow the operation of mechanisms identical or closely allied to "primary" metabolism. In view of the considerable practical difficulties in the preparation of the enzymes of secondary metabolism, knowledge of this field has mainly been derived from the results of administration of ^{14}C-, ^{2}H-, and ^{3}H-labeled substrates to the intact plant or organism, with subsequent appropriate degradative techniques to locate the label, and, more recently, from the use of isotope ratio techniques to ensure nonrandomization. Although great progress in experimental techniques has been made, it seems to be an intrinsic property of many higher plant systems that specific incorporations of between 0.01 and 0.1 percent represent an average level of success. The purpose of this article is to suggest that several techniques can be used to improve a given situation which at first sight might be indicative of negative incorporation of a suspected precursor. In addition, in systems where incorporation of 5 percent or more can be achieved (usually in fungi or bacteria) the use of ^{13}C as biosynthetic label offers unique advantages and can be combined very successfully with the results of other kinds of radioactive labeling experiments.

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Identification of Intermediates by Short-Term Incubation in Vinca rosea

Of the plants in which indole alkaloids occur, perhaps none holds more interest for the chemist or the biochemist than *Vinca rosea* in that the periwinkle produces a profuse array of indole alkaloids including some 100 representatives of all the important structural classes (*1–3*). It is also an ideal plant with which to begin cell-free studies, for the viable seeds are readily available, easy to cultivate, and the seedlings possess a remarkable capacity for synthesizing alkaloids with a vigor that is perhaps matched only by the chemotaxonomists who have bestowed such complex names on these compounds.

Earlier biosynthetic studies (*2*) with *V. rosea* utilized either intact shoots or seeds. After some experimentation it was decided to use young seedlings (9 to 17 days from germination) grown from a mixed strain (Burpee) at 33°C in an environmental chamber with full light. Intact seedlings were removed and placed in groups of four (the average weight was 16 milligrams) in ¼-dram vials (1 dram = 1.8 grams). The seedlings remained healthy and continued to develop root growth for about 9 to 10 days in water at 33°C. The range of alkaloids present in the 9-day seedlings approximated to that of mature *V. rosea*, and the more predominant compounds were easily detectable by thin-layer chromatography (TLC) (*4*). These were vindoline (**13**), coronaridine (**12**), catharanthine (**11**), akuammicine (**5**), and vinervine (**6**) (see chart 1). One of the main difficulties in carrying out cell-free studies with *V. rosea* is the absence (or virtual absence) in the mature plant of many alkaloids which correspond to the intermediates of the various pathways. In practical terms this will probably require preparative scale cell-free incubations to isolate and characterize these dynamic compounds. However, with the use of short-term incubation techniques with the above seedlings, some progress can be made toward the solution of this problem.

The main events of indole alkaloid biosynthesis based on earlier incorporation data (*2*) are summarized in chart 1. Here the structures marked with an asterisk (*) correspond to alkaloids identified below. The remaining alkaloids have already been described from mature plants and germinating seeds of *V. rosea* (*2*).

Incorporation of tryptophan into the alkaloids takes place in two distinct phases. For the first 2 hours of incubation it is linear (Fig. 1), but after this interval a rapid increase in the rate of incorporation is observed. Between 12 and 48 hours, a maximum of 3 percent is reached and this level is maintained during the full time of the experiment