because the group is reasonably well collected and studied, the diversity residuals of planktonic foraminifera can be almost wholly ascribed to distortions of the planetary temperature gradient by ocean currents (*4*).

Figure 3 shows contoured, positive (more diverse and, therefore, warm) and negative (less diverse and, therefore, cold) residuals from the computed regional surface for present-day planktonic foraminifera. The control was adequate to show the regional surface, as was originally desired (Fig. 2), but is, unfortunately, minimal for consideration of residuals (75 points for the entire world), and is poorly distributed (concentration in the North Atlantic). Nevertheless, a marked relationship to the surface circulation pattern is evident. Positive diversity residuals are divided into two arbitrary classes (greater than +5 is very warm; +5 to 0 is warm); negative diversity residuals are, likewise, divided (0 to −5 is cold; greater than −5 is very cold). A comparison of the residual surface (Fig. 3) with a simplified diagram of ocean circulation patterns (Fig. 4) shows clearly that even at the present level of resolution many of the major current systems can be recognized. All of the data were used in the original calculation of the regional and residual surfaces. In contouring the residual surface, however (Fig. 3), four points have been omitted because it is believed that they owe their value to factors other than ocean currents, such as poor collecting or deltaic influences. The four omitted points are marked with a special symbol; the residual value for each is given.

The data presented in the figures permit only a preliminary test of the technique suggested. A further and more critical test of the relationship of present-day diversity residuals and existing ocean currents is required. The preliminary test is sufficiently encouraging, however, to suggest some interesting possibilities for the method.

Planktonic foraminifera have been abundant since the late Mesozoic. Thus, it is theoretically possible to reconstruct patterns of oceanic circulation back into the Cretaceous, or for a period of about 100 million years. A practical limit is presently imposed by the extreme scarcity of oceanic cores which have penetrated early Tertiary or Cretaceous sediments. For the time being, the method could probably be applied only to Miocene and younger ages. Its greatest immediate possibilities appear to lie in detailed studies of the Pleistocene record.

Core coverage for the latter part of Pleistocene time is probably adequate for reconstruction of the oceanic circulation patterns that accompanied the extreme climates of the Pleistocene. It should be possible to recognize the circulation of both a fully glacial and a fully interglacial interval and, thus, establish boundary conditions within which other less extreme changes should be contained. It should also be possible to determine whether the Arctic Ocean has been ice-free during any part of the Pleistocene, thereby providing a test of some theories of glaciation. With a relatively few O^{18}/O^{16} paleotemperature measurements on suitable samples, it might even be possible to calibrate the residual surface for temperature.

If the surface oceanic currents bear the close relationship to atmospheric circulation suggested by Munk (*5*), a knowledge of the currents characterizing the extremes of the Pleistocene might reflect considerable information about the patterns of zonal wind circulation in the atmosphere. Such information might lead to a better understanding of the conditions attendant upon continental glaciation of the middle latitudes.

Exploration of the interesting possibilities for the use of taxonomic diversity residuals of planktonic foraminifera in the study of past patterns of oceanic circulation must await a critical and detailed test of the recent model.

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Is There Vegetation on Mars?

Abstract. *At least some of the changes in the color of Mars at different seasons are caused by color centers produced by electromagnetic and corpuscular solar radiation in solids on the surface. Calculated radiation flux, at appropriate energies and known temperature variation, could account for seasonal formation of color centers and bleaching if a simple trap model is assumed. In certain kinds of rhyolite (SiO_2, $NaAlSi_3O_8$), which has been suggested as one of the possible constituents of the martian surface, color centers can be produced. No color centers are expected in limonite, $Fe_2O_3 \cdot 3H_2O$, the other likely constituent.*

The hypothesis that vegetation exists on Mars is based on two kinds of observations: one is the presence of infrared absorption bands of aldehyde, the other is the seasonal variation of the darkness of bluish-green areas on the otherwise orange-rusty planet. Recently (*1*) the absorption bands have been shown to be caused by heavy water, HDO, of telluric origin. My purpose is to point out the possibility that the seasonal color variations may be explicable, at least in part, in terms of the well-known phenomenon of color-center formation and bleaching by incident ionizing radiation under varying temperature conditions. According to recent estimates (*2, 3*) the martian atmosphere (total pressure 13 to 20 mm-Hg) consists of 85 percent (by volume) of N_2, 14 percent (by volume) of CO_2, and the rest being essentially argon. These data, combined with the known absorption coefficients (*4*), indicate that no significant absorption of solar ultraviolet radiation occurs for wavelengths longer than about 2000 Å ($h\nu = 6$ ev). In the range of photon energies between 4 and 6 ev the total solar flux at the top of the martian atmosphere is about 10^4 erg cm^{-2} sec^{-1}. This gives 10^{15} photons cm^{-2} sec^{-1} as an estimate of the intensity of this radiation at martian surface in the subsolar region. On Earth ultraviolet radiation in this range of wavelengths is totally absorbed by atmospheric ozone. On Mars the upper limit of ozone concentration (*2*) is about 10^{-4} volume percent and thus it is ignored. No account is taken either of the unknown effect of the so-called "blue haze" which may decrease the ultraviolet flux. According to Gold (*5*) radiation-induced coloration is responsible for the dark areas on our moon. There, however, the absence of an atmosphere and of drastic color changes makes the situation considerably different.

Two materials have been suggested (*6, 7*) as possible constituents of the martian surface: orange-rusty opaque limonite ($Fe_2O_3 \cdot 3H_2O$) and rhyolite (an igneous felsitic mixture of SiO_2 and various silicates). It is known that SiO_2 can be darkened by ionizing radiation and that these color centers are sensitive to impurity content. Experiments made in our laboratory have shown that no visibile color changes can be produced in the same way in limonite but that $NaAlSi_3O_8$, which is a known constituent of certain rhyolites, does indeed change from colorless to greenish under irradiation.

The seasonal color variation can be accounted for in various ways. One can assume, for instance, that besides color centers there are a large number of shallow electron traps which are easily thermally ionized during summer but are mostly occupied during winter. Another, more sophisticated, model is based on the assumption that there exists a set of recombination centers which produce supralinearity (*8*) of photocurrent in a narrow temperature range.

In this case it is necessary that the ultraviolet radiation is hard enough to produce holes in the valence band of the solid. Both models lead to a depletion of electrons from the color centers during winter and to an increased occupation of these centers by electrons during summer. This is just what is necessary to account qualitatively for the seasonal color changes. Whatever the model, it is essential that, in order to suppress the influence of daily variations of local temperature on coloration, the time constant for establishing equilibrium distribution of electrons in a given radiation flux be not less than several hours. Such long time constants occur in certain phenomena associated with color centers. The seasonal variations of the average daily temperature range from zero at the equator up to 120°C at increasing latitudes. The average blackbody temperature of Mars is 208° ± 10°K, depending upon the distance from the sun. It should be mentioned too that the seasonal increase of optical absorption of the surface material is in the right direction to account for the parallel increase of negative polarization of the reflected light (*6*). The theory of the latter effect is too uncertain at the present time to permit a more detailed comparison.

Goldstein and Gillmore (*9*) have observed that the dark areas of Mars have a much higher radar reflectivity that the rest of the planet. Usually such differences are interpreted in terms of surface roughness or other permanent properties. However, in this case it is tempting to associate this high reflectivity with the presumably high photoconductivity induced by the solar radiation. An unambiguous estimate of the magnitude of this effect would require a knowledge of the detailed mechanism and of the trap and carrier distribution. It appears, however, that under favorable conditions the skin depth can be much smaller than the wavelength of the incident radar, and the photoconductivity may play a significant role. It is hoped that the present opposition of Mars will provide a check of this conclusion.

The shapes of certain dark areas of Mars vary considerably from one year to another. On the proposed model these variations may be the result of solar flares (*10*) which produce exceedingly high fluxes of photons and corpuscular matter. A statistical correlation between these two phenomena would be very instructive.

It is not my intention to imply that there is no vegetation on Mars but rather to point out that some of the "organic" observations may have "inorganic" explanations.

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High-Pressure Single-Crystal Studies of Ice VI

Abstract. *By means of a precession camera incorporating a diamond-anvil high-pressure cell, x-ray diffraction data can be obtained from single crystals of ice VI produced and maintained under high pressures. The cell constants for ice VI at room temperature and approximately 9 kilobars are:* a = *8.38 Å,* b = *6.17 Å,* c = *8.90 Å. The unit cell is orthogonal and the space-group aspect is compatible with* P**a. *These data for single crystals agree with previously reported unindexed data obtained for polycrystalline ice VI, within the limits of experimental error. The single crystals of ice VI were grown in a diamond-anvil pressure cell, distilled water and a metal gasket being used.*

To obtain definitive data on the structures of high-pressure polymorphs, it is essential to obtain x-ray diffraction data on single crystals of the high-pressure phase. In a few instances, high-pressure phases may be quenched at low temperatures and single crystals recovered for study at 1 bar in a metastable condition. The number of such quenchable polymorphs appears to be limited, and it is desirable to have a method for producing and studying single crystals by x-ray diffraction while they are under high pressures. Such a method has been developed in our laboratory and has been found suitable for the study of high-pressure phases obtained by liquid-to-solid transformations as in ice VI, and by favorable solid-to-solid transitions —that is, transitions which exhibit relatively small volume changes, thereby retaining their single-crystal character. In this report we present preliminary data on ice VI obtained by this method. So far as we know, such data have not been obtained previously.

Single crystals of ice VI were ob-