The Astronomical Theory of Climatic Change on Mars

OWEN B. TOON AND JAMES B. POLLACK

Theoretical and Planetary Studies Branch, Ames Research Center, NASA, Moffett Field, California 94035

WILLIAM WARD

Jet Propulsion Laboratory, Pasadena, California 91103

JOSEPH A. BURNS

Theoretical and Applied Mechanics Department, Cornell University, Ithaca, New York 14853

AND

KENNETH BILSKI

Informatics, Incorporated, Palo Alto, California 94303

Received May 8, 1980; revised October 21, 1980

We examine the response of Martian climate to changes in solar energy deposition caused by variations of the Martian orbit and obliquity. We systematically investigate the seasonal cycles of carbon dioxide, water, and dust to provide a complete picture of the climate for various orbital configurations. We find that at low obliquity (15°) the atmospheric pressure will fall below 1 mbar; dust storms will cease; thick permanent CO₂ caps will form; the regolith will release CO₂; and H₂O polar ice sheets will develop as the permafrost boundaries move poleward. At high obliquity (35°) the annual average polar temperature will increase by about 10°K, slightly desorbing the polar regolith and causing the atmospheric pressure to increase by not more than 10 to 20 mbar. Summer polar ground temperatures as high as 273°K will occur. Water ice caps will be unstable and may disappear as the equilibrium permafrost boundary moves equatorward. However, at high eccentricity, polar ice sheets will be favored at one pole over the other. At high obliquity dust storms may occur during summers in both hemispheres, independent of the eccentricity cycle. Eccentricity and when the longitude of perihelion is close to the location of solstice hemispherical asymmetry in dust storm generation and in polar ice extent and albedo will occur.

The systematic examination of the relation of climate and planetary orbit provides a new theory for the formation of the polar laminae. The terraced structure of the polar laminae originates when eccentricity and/or obliquity variations begin to drive water ice off the dusty permanent H_2O polar caps. Then a thin (meters) layer of consolidated dust forms on top of a dirty, slightly thicker (tens of meters) ice sheet and the composite is preserved as a layer of laminae composed predominately of water ice. Because of insolation variation on slopes, a series of poleward- and equatorward-facing scarps are formed where the edges of the laminae are exposed. Independently of orbital variations, these scarps propagate poleward both by erosion of the equatorward slopes and by deposition on the poleward slopes. Scarp propagation resurfaces and recycles the laminae forming the distinctive spiral bands of terraces observed and provides a supply of water to form new permanent ice caps. The polar laminae boundary marks the furthest equatorward extension of the permanent H_2O caps as the orbit varies. The polar debris boundary marks the furthest equatorward extension of the annual CO_2 caps as the orbit varies.

The Martian regolith is now a significant geochemical sink for carbon dioxide. CO_2 has been irreversibly removed from the atmosphere by carbonate formation. CO_2 has also been removed by regolith adsorption. Polar temperature increases caused by orbital variations are not great enough

to release large quantities of adsorbed CO_2 . Early in Martian history, before the formation of carbonates and an extensive clay regolith, large quantities of CO_2 may have resided in the atmosphere-polar cap system. Orbital variations could have initiated a high-pressure climate by driving CO_2 out of early permanent CO_2 polar caps. Prior to the formation of high-albedo iron oxides on the Martian surface, 500 mbar of atmospheric CO_2 was enough to maintain the global average temperature at the freezing point of water. Such high temperatures may have played a role in creating some of the fluvial features observed on Mars.

For the present orbital configuration and climate of Mars we show that: regolith damping of the seasonal CO_2 cycle is unlikely; the mean atmospheric pressure is probably in equilibrium with the regolith; the low albedo of the north H_2O polar cap can be explained by an admixture of 85% ice and 15% dust; and the albedo of the polar caps as well as the polar heat budget are very sensitive to small variations in dust deposition. We suggest that a small variation in dust content of the H_2O and CO_2 caps is mainly responsible for the difference in H_2O and CO_2 deposition in the north and south polar regions at present.

INTRODUCTION

Mariner 9 spacecraft observations of Mars during 1971 and 1972, as well as more extensive Viking observations beginning in 1976, provide evidence that current Martian conditions are not typical of all of Martian geologic history. The Martian polar regions are covered with extensive layered terrain (Fig. 1), probably composed mainly of wind-deposited dust and water ice, suggesting cyclic polar processes (Cutts et al., 1976). The existence of hundreds of desiccated stream beds indicates that liquid water once flowed over and under the Martian surface even though Mars is extremely cold and dry today (Masursky et al., 1977; Sharp and Malin, 1975). The abundance ratios of nitrogen isotopes and noble gases found in the current Martian atmosphere require that earlier Mars had an atmosphere with much greater N_2 abundance than at present and that large quantities of CO₂ and H₂O have been outgassed but now lie in hidden reservoirs (McElroy et al., 1977; Owen et al., 1977; Anders and Owen, 1977; Pollack and Black, 1979).

In this paper we explore the details of one process that may be important for Martian geologic history—climate changes caused by variations in the orbit and obliquity of Mars. Solar energy is the basic factor controlling climate on the terrestrial planets, and its distribution with latitude and with season is determined by the orbit and spin orientation of the planet. Ward (1973, 1974, 1979) and Murray *et al.* (1973) have previously shown that the orbital elements of Mars now undergo very large changes due to the interaction of the torque of the Sun on the aspherical body of Mars and the torque of the planets on the inclined and eccentric orbit of Mars. Ward *et al.* (1979) have shown that geologic processes acting to change the figure of Mars may have modified the solar torque so that the variations in the orbital elements of Mars may once have been somewhat different from the variations as they are now.

The orbital elements of Mars are subject to much greater variation than are those of Earth. The idea that terretrial ice ages might be caused by changes in the Earth's orbital elements is very old, although Milankovitch (1930) is most widely known for developing it mathematically. Several empirical studies (Hays et al., 1976) have shown that the periodicity of observed terrestrial ice ages is well correlated with the periodicity of the Earth's orbital elements. There have been few theoretical attempts to model terrestrial climate changes caused by orbital element variation. Several investigators (Mason, 1976; Pollard, 1978) have presented simple theoretical arguments suggesting that the energy imbalance caused by the changing orbit of the Earth was sufficient to cause ice sheet growth, but no model has yet shown an ice age development with the time history observed. Prior to the last few million years Earth was



FIG. 1. High-resolution picture of laminated terrain in the south polar region of Mars. Terraces are exposed on steep slopes. Each individual terrace is about 30 m thick and can be traced horizontally for hundreds of kilometers. The width of the set of terraces exposed in the figure is about 5 km at the widest. The laminae are thought to contain both dust and H_2O ice. Their regular structure and the fact that they occur in both polar regions indicate that their formation may be controlled by climatic changes (Mariner 9 picture 4213-21).

warmer than at present. Large-scale ice sheets have been generally absent except during a few glacial ages (Pollack, 1979). The astronomical theory of ice ages is unable to account for such long-time-scale changes in climate.

There are two difficulties in modeling terrestrial ice ages caused by orbital changes that do not occur for Mars. First, the solar insolation changes caused by the orbital variations for Earth are very small. Second, the Earth's climate is extremely complex because of latent heat release and the transport of heat by the oceans and by the atmosphere. For Mars the orbital changes are large, oceans are absent, latent heat release is unimportant except at the poles, and the atmosphere is not dense enough to carry much heat. Thus the theoretical problem is more tractable for Mars than for Earth. Unfortunately, our limited observational knowledge of Mars does present problems in assigning boundary conditions and in knowing the importance of various processes even on present-day Mars. In addition, as for the Earth, variations in seasonal cycles are important for Mars, yet the seasonal cycles are much more difficult to model than the annual means.

The Martian climate is primarily defined by the seasonal cycles of temperature, CO_2 , H₂O, and atmospheric dust. We systematically explore the variations in these cycles as the orbital elements change. In Section I we review the solar insolation changes caused by orbital element changes. The temperature and carbon dioxide cycles are considered in Section II for the simple case of CO₂ polar caps containing limited amounts of CO₂. In Section III we discuss the CO_2 cycle in the more complex case when CO₂ interacts with the regolith. In Section IV we discuss the dust cycle which depends upon the CO₂ cycle. In Section V we consider the water vapor cycle and the interactions of water vapor with permafrost.

One conclusion from our study of the cycles of CO_2 , H_2O , dust, and temperature

is that the Martian climate has been rather similar to the present climate since the formation of large CO_2 reservoirs in the regolith. Early in Martian history, before the formation of these reservoirs, large amounts of CO_2 may have resided in the atmosphere. In Section VI we discuss the nature of the climate when the atmospheric pressure was greater.

Our discussions of the theory of the variations of the volatile, temperature, and dust cycles are summarized at the end of each section. In Section VII we present an overview of the Martian climate for various orbital configurations and discuss the observational evidence for the predicted climate changes. In particular a new theory of the formation and evolution of the polar laminae is suggested.

I. SOLAR INSOLATION VARIATION

There are three orbital elements which affect solar insolation and which are each apt to change significantly: obliquity, θ ; eccentricity, e; and longitude of perihelion, ω . The obliquity is the angle between a planet's spin axis and its orbit normal. As θ increases, the planet's pole tilts further away from the orbit normal so that the polar energy supply in the summertime is increased. Changing θ causes neither variation in the planetwide energy received nor hemispherical asymmetry. The Earth's obliquity is now 23.5° and varies between 22.1 and 24.5° with a period of about 4×10^4 years (Berger, 1976; Hays et al., 1976). The obliquity of Mars is now 25.2° and oscillates by about $\pm 13^{\circ}$ with a dominant periodicity of 1.2×10^5 years and a modulation period of about 1.3×10^6 years (Ward, 1979). The Martian obliquity is forced by torques near a resonance condition so that the mean obliquity also slightly oscillates. Major geologic events, such as the formation of the Tharsis ridge, may have altered the proximity to the resonance condition, and earlier Mars may have had a mean obliquity near 32° with maxima and minima of 45 and 25°. Even more extreme values are possible under dynamically allowed motions that have not been fully explored theoretically (Ward *et al.*, 1979).

The eccentricity of a planet's orbit describes the departure from circularity. Since the solar flux at a planet depends upon the square of the planetary distance from the Sun, an eccentricity change alters the energy received annually by a planet. However, the magnitude of the alteration is negligible for the climate changes to be discussed. On an annual average changes in e do not redistribute energy with latitude and do not create asymmetry between the hemispheres. The eccentricity of the Earth's orbit varies from 0 to 0.06, with a period near 9.5×10^4 years (Berger, 1976); it is presently 0.0167. For Mars the present eccentricity is 0.09 and its minima and maxima are 0 and 0.14 (Ward, 1979). The oscillation has two superposed periods: one of 1.8 to 2.3×10^6 years with an amplitude of about 0.1 and the other of 9.5×10^4 years with an amplitude of 0.04.

The longitude of perihelion is the angle between the data of equinox and the time of perihelion passage. For Earth and Mars, the equinoxes precess through 360° with periods of about 2.1×10^4 and 1.75×10^5 years (Hays *et al.*, 1976; Ward, 1974). The precession of the equinoxes causes no change in the annual average solar insolation. However, for an eccentric orbit, both the diurnal and seasonal average insolation are strongly affected by the time of perihelion passage so that latitudinally and hemispherically asymmetric energy changes occur.

. Ward (1974) has previously developed formulas to calculate the daily and annually averaged insolation from the orbital elements. The daily averaged insolation at latitude δ , with solar constants S_0 at the mean Mars-Sun distance is

$$\bar{I} = S_0 \left[\frac{(1 + e \cos \nu)^2}{(1 + e^2)^2} \right] \left(\frac{\eta \sin \epsilon \sin \delta + \sin \eta \cos \epsilon \cos \delta}{\pi} \right).$$
(1)

Physically, the term in the first bracket represents the change in solar intensity caused by the r^{-2} variation in the Mars-Sun distance. The true anomaly of the Sun, ν , measures orbital position from perihelion. The second bracket contains two geometric effects: the length of the Martian day at latitude δ , as well as the angle between the local surface normal and the noontime solar beam. These depend directly on solar latitude ϵ , and solar longitude η , and indirectly on δ , θ , ν , and ω (Ward, 1979).

Ward (1974) presented plots of \overline{I} for various latitudes, eccentricities, and obliquities. For obliquities and eccentricities within the range of Martian values the equatorial insolation varies moderately through the year while the polar insolation varies from zero to a value *larger* than that at the equator (for $\theta > 17^{\circ}$).

The poles can receive more summertime

insolation than the equator because the distance around the latitude circle in the polar regions is much less than that in the equatorial regions. The equatorial regions receive sunlight during the day but not at night, so large daily temperature fluctuations occur. During the polar summer the Sun never sets, giving no daily temperature fluctuations.

Figure 2 presents calculations by Ward (1974) of the annual average solar insolation for various obliquities. It is important to note that the energy supplied to the equator is a weak function of obliquity compared with the energy supplied to the pole. For obliquities larger than 54° the poles receive more annual average energy than the equator. Even so, the poles only receive the energy during summer, so the equatorial energy will undergo a smaller seasonal oscillation than does the polar energy. The



FIG. 2. The latitude dependence of the annual average solar insolation expressed as a fraction of the solar constant at 1.52 AU for various obliquities. Ward (1974) pointed out that for obliquities less than 50° the polar insolation is a strong function of obliquity, but the equatorial energy deposition is a weak function of obliquity. As obliquity increases, polar insolation increases but equatorial insolation declines.

annual average insolation is a very weak function of e.

It should be noted that time series for the insolation reflect the basic periodicities in θ , e, and ω in a complex manner. It is partly for this reason that for the Earth comparisons of simple climate theories with ice age data have proved elusive (Hays *et al.*, 1976).

II. THE SEASONAL CYCLES OF CO₂ AND TEMPERATURE WHEN CO₂ POLAR DEPOSITS ARE LIMITED

In previous models of climate change on Mars it was assumed that Mars had large permanent CO₂ polar caps (Ward et al., 1974; Gierasch and Toon, 1973; Sagan et al., 1973). Latent heat exchange with solid CO_2 polar caps would maintain the polar regions at a constant temperature throughout the year. Therefore, previous climate models were based primarily upon annual average calculations. Viking observations, however, show that the north permanent polar cap does not contain CO₂ and the south permanent polar cap probably contains no more than a small amount of CO_2 , if any (Kieffer et al., 1976; James et al., 1979). Therefore, we have made new calculations of the CO₂ and temperature cycles assuming that not enough CO₂ exists to

form permanent CO_2 polar caps under present Martian conditions.

(a) Model Description

The current Martian atmospheric pressure is so low that atmospheric heat transport is negligible (Gierasch and Toon, 1973). Hence, the ground temperature at any location is determined primarily by local radiative and latent heat sources. At present water does not supply much latent heat, nor do globally significant water clouds form. Hence, a model considering only CO_2 is a good description for the current climate and for small changes about this condition.

In this paper we employ a model that is essentially identical to those of Leighton and Murray (1966) and Kieffer et al. (1977). Our model balances absorbed solar radiation, infrared radiation from the surface, and heat conduction into the ground. CO₂ condenses whenever the ground temperature falls below the temperature at which the atmospheric pressure equals the vapor pressure above solid CO₂. The heat diffusion equation is solved with a semi-implicit scheme developed by Cuzzi and Muhleman (1972). The ground is divided into ten 1.5cm layers and ten 15-cm layers; the total depth is approximately two seasonal thermal wavelengths. The model has 5° latitude grid points and it calculates the energy balance every half hour, updating the changing solar insolation every 5 days. The model was run for 10 years to achieve stability, with the lower-layer temperature updated to the annual average after each year to avoid a long lag time. The 9th- and 10th-year temperatures and pressures were consistent to within 0.01°K and 0.01 mbar.

The model surface albedo is assumed to be 0.25, within the error bars of the Viking value of 0.214 \pm 0.063 (Kieffer *et al.*, 1977). The Viking mean value itself may be somewhat low because of incomplete coverage of Mars by the spacecraft observations. We have not treated the albedo of the permanent H₂O polar caps since they are of small geographic extent. We assume that: the CO_2 frost albedo is 0.65; the soil density is 1.6 g cm⁻³; the thermal inertia is 0.006 cal $cm^{-2} sec^{-1} deg^{-1}$; and, the specific heat is 0.17 cal g^{-1} de g^{-1} . All these values are close to those adopted by Kieffer et al. (1977). The thermal balance of the atmosphere is not considered in detail, so the energy received by the ground from atmospheric infrared radiation cannot be precisely calculated. We treat the atmospheric radiation as 1% of the noon solar radiation. We use a ground emissivity of 0.9. Kieffer et al. (1977) assume that the atmosphere radiates 2% of the noon solar radiation and that the ground emissivity is 1. These differences between the two models are mutually compensating and neither affects surface temperature by more than a few degrees.

An important assumption of our model is that the total amount of CO_2 in the atmosphere and seasonal polar caps is about 9 mbar. We chose this value because it makes the CO_2 caps disappear near solstice under present conditions, so permanent CO_2 polar caps are not found at either pole. The atmosphere is known to contain about 6 mbar of CO_2 and models of the seasonal caps suggest that they contain no more than a few millibars of CO_2 (Briggs, 1974; Davies *et al.*, 1977).

There are several features of the present Martian climate that are not well treated by models such as ours. For example, the models are not meant to follow accurately the details of polar cap recession, so they do not have albedos which are functions of frost thickness. Neither do the models consider the polar hood which affects the polar radiation balance. Of course, all parameters, such as thermal inertia and albedo, vary strongly with location about the mean values we use; in addition some soil properties may be functions of depth. Kieffer et al. (1977) report very low polar night temperatures, well below those that should occur if CO₂ latent heating maintains the temperature at the frost point. They suggest that the winter polar atmosphere becomes depleted in CO_2 and enriched in minor species which would influence the polar energy balance. However, the observed low temperatures may instead reflect low surface emissivity or the presence of high clouds. Until the observations are more completely understood they cannot be easily considered in models.

We have modeled two orbital configurations. In the first, $\theta = 15$, 25, 35, and 45° for e = 0. These models have complete symmetry between the hemispheres and the seasonal cycle is independent of ω . The values of θ represent the extreme reasonable values for the current oscillation cycle (15 to 35°) and the hypothetical oscillation (Ward *et al.*, 1979) prior to the formation of the Tharsis ridge (25 to 45°). Values of θ slightly smaller than 15° may actually be reached (e.g., 11°), but the corresponding climates will probably not be substantially different from those at 15°.

The other situation we have modeled is the case when e = 0.14, its maximum value. Again we study $\theta = 15, 25, 35$, and 45° . We have set $\omega = 270^{\circ}$ so that solstice occurs at perihelion. Thus, one summer pole receives the maximum possible insolation. We have also performed one calculation in which θ = 45° and $\omega = 0^{\circ}$ so that perihelion is at equinox, producing the maximum equatorial insolation at one equinox and the minimum insolation at the other. The equatorial, equinoctial insolation maxima are independent of θ .

(b) Calculated Temperature and Pressure Cycles, e = 0

Figure 3 illustrates calculations of daily average surface temperatures as a function of latitude and time of year for e = 0 and various values of θ . Mars presently has e =0.09, $\theta = 25^{\circ}$, and $\omega = 250^{\circ}$, so it is intermediate between the case shown in Fig. 3B and that shown in Fig. 6A. In Fig. 3B the summer temperatures are nearly independent of latitude. The polar annual cap sublimes away somewhat after the solar energy peaks because of the finite en-



FIG. 3. The daily averaged ground temperature at various latitudes is illustrated for perihelion occurring at solstice, a circular orbit, and varying obliquity. All the calculations employ a CO₂ albedo A = 0.65 and a CO₂ abundance of 8.9 mbar except for one case in (A) where the calculation for the polar temperatures was run with A = 0.4. The seasonal variation in polar temperature shows that low CO albedos will prevent permanent polar cap formation. Another case was run with A = 0.65 but P = 5 mbar. The constant low polar temperature shows that, if the albedo is high, a permanent polar cap will form even if the abundance of CO₂ is low. Parts (C) and (D) indicate that high polar summer temperatures occur at large obliquity.

ergy required to evaporate the high-albedo CO_2 frost. During most of the year CO_2 deposition maintains polar temperatures constantly near 150°K. The amplitude of the thermal seasonal wave increases with latitude. All these results are in basic agreement with observations of Mars.

At obliquities greater than 35° the summer polar temperatures exceed summer equatorial ones. In fact, for $\theta = 45^{\circ}$, summer polar ground temperatures exceed 273°K throughout the entire day for about 45 Martian days. Of course on Mars under

present conditions low latitudes experience very large excursions of daily temperature, which often have peaks exceeding 273° for portions of the day (Kieffer *et al.*, 1977). As the obliquity increases, the polar CO₂ deposits evaporate earlier and earlier since the sunlight is more intense because of the higher sun angle and the greater length of the summer day at subpolar latitudes. However, polar CO₂ ice deposits form further equatorward during winter since the length of the winter day is lessened at higher obliquity. As the obliquity increases, the magnitude of the annual thermal wave grows at all latitudes.

As the obliquity becomes less than 25°, the poles become colder, the equator becomes warmer, and the annual thermal wave amplitude decreases. The behavior of polar CO₂ depends upon its albedo. In Fig. 3A we present two alternate models: one which inhibits polar cap formation due to very restricted CO₂ abundance (P = 5 mbar and CO₂ frost albedo $A_{CO_2} = 0.65$); and one which inhibits polar cap formation due to enhanced solar heating (P = 9 mbar and $A_{CO_2} = 0.4$). Permanent polar caps form only for the case with higher albedo.

A high polar cap albedo at low obliquity is the more likely case. The permanent north polar cap of Mars, composed of H₂O, has an albedo near 0.4 (Kieffer *et al.*, 1976). There are two reasons for the low albedo. Mars is presently a dusty place, and the thickness of water ice annually sublimed and redeposited on the poles is small ($\ll 1$ cm). Hence the dust-to-ice ratio is large and the ice is not self-cleaning. For CO₂ permanent deposits neither of these effects should occur. A high-albedo cap causes a low atmospheric pressure (Fig. 4) and at low atmospheric pressure dust lifting will not



FIG. 4. The seasonal variation of atmospheric pressure is shown for a circular orbit and for various obliquities. At low obliquity a permanent CO_2 cap may form, decreasing the pressure. Note that the annual mean pressure declines slightly as the obliquity increases above 25° because of the formation of larger annual polar caps.



FIG. 5. The seasonal variation of the CO_2 polar cap boundary as a function of obliquity for a circular orbit. The annual cap is larger for larger obliquity.

occur (Ward *et al.*, 1974; Pollack, 1979). Furthermore, our model with A = 0.65 suggests that 100 cm of polar CO₂ will annually be deposited and removed at $\theta = 15^{\circ}$, cleaning the CO₂ each year.

Figures 4 and 5 show how atmospheric pressure P and the location of the polar cap boundary vary with time of year for various θ . As θ increases the polar cap boundary moves equatorward, the annual mean atmospheric pressure decreases slightly, and the amplitude of the seasonal pressure wave increases. Earlier models with unlimited CO₂ (Ward et al., 1974; Gierasch and Toon, 1973) predicted large increases in P with increasing θ . Ward *et al.* (1974) also found, even with unlimited CO₂, that annual polar caps would extend to lower latitudes when the obliquity increased, as Fig. 5 indicates. However, with limited CO_2 , the seasonal caps contain less CO_2 than with unlimited CO_2 and do not reach latitudes as low. The reason that the atmospheric pressure decreases with increasing obliquity is that greater amounts of CO₂ go into the larger seasonal cap.

If the polar albedo remains high for permanent CO₂ caps, then at 15° obliquity the annual pressure will be quite low, 1 mbar or less. The precise value of atmospheric pressure at low θ is sensitive to the choice of the CO₂ albedo. Ward *et al.* (1974) chose A_{CO_2} = 0.70 and found that at θ = 15° atmospheric pressures were only a few tenths of a millibar.

(c) Calculated Temperature and Pressure Cycles, e = 0.14

Figure 6 illustrates the daily average ground temperature as a function of θ for e = 0.14. The temperatures are similar to those with e = 0 except that the two hemispheres are no longer identical and the amplitude of the seasonal thermal wave increases. With e = 0.14 and $\theta = 35^\circ$, polar ground temperatures in the hemisphere in which perihelion occurs at solstice exceed 273°K. When θ is 45°, temperatures at the pole reach 300°K, exceed 273°K for about 85 days, and the entire region poleward of 50° has daily average midsummer temperatures above 273°. The summer pole in the opposite hemisphere is cooler, and even at $\theta = 45^{\circ}$ ground temperatures there never exceed 260°. Winter temperatures are also altered. The hemisphere with the warmer summer pole is colder in winter and the seasonal cap reaches lower latitudes.

As the equinoxes precess, the two poles



FIG. 6. The seasonal variation of ground temperature at various latitudes is shown for a highly eccentric orbit. Parts (A)-(C) illustrate results for various obliquities when perihelion occurs at southern solstice. The asymmetry between hemispheres is most noticeable at the poles. Part (D) illustrates the seasonal temperature variation at 45° obliquity when perihelion occurs at an equinox. The temperature is most altered at 30° latitude.

switch roles. The intermediate case, $\omega = 0$, is illustrated in Fig. 6D. Here the hemispheres are nearly symmetric except for hysteresis effects. The equatorial temperature shows a strong maximum at the perihelion equinox.

Figures 7 and 8, respectively, illustrate for e = 0.14 the atmospheric pressure changes and the modification of the ice cap edge due to changing θ . These results are similar to those for e = 0 except for an asymmetry between the hemispheres.

(d) Annual Average Temperatures

Seasonal temperature oscillations are quite important for atmospheric pressure changes and for volatiles stored within the top 100 cm of soils, the upper thermal skin depth for the annual thermal wave. For many important processes, however, annual average temperatures are more significant than daily average temperatures. Figure 9 illustrates annual average temperatures for various θ and e = 0.0 and 0.14. Even though the annual average insolation is nearly independent of e, the annual average temperature does depend on e. The greater dependence of temperature on e occurs because the temperature is related in a nonlinear manner to the insolation through σT^4 . Hence, average temperatures are asymmetric between the hemispheres and they vary for fixed θ and different e. The temperature changes in Fig. 9 approximately mimic the annual average insolation variation shown in Fig. 2. As the obliquity increases, solar insolation at high latitudes



FIG. 7. The seasonal atmospheric pressure variation as a function of obliquity for a highly eccentric orbit.



FIG. 8. The seasonal polar cap boundary variation as a function of obliquity for a high eccentric orbit.

increases, as do polar temperatures. The insolation at low latitutdes decreases as do equatorial temperatures, but the changes at low latitudes are smaller than those at high latitudes. Figure 2 shows that the insolation near 35° latitude is nearly independent of θ , but Fig. 9 shows that the temperature is nearly independent of θ near 50° latitude. The latitude shift is due to the nonlinear dependence of T on insolation through σT^4 .

Earlier climate models did not illustrate annual average temperatures except at the poles. Ward et al. (1974) thought polar temperatures would be about 10°K cooler than those shown by Fig. 9 at $\theta = 25$ and 35° because their CO₂ cap never sublimed to expose low-albedo ground, and latent heat kept summer temperatures low. Nevertheless, the temperature change in going from $\theta = 25$ to 35° in the model of Ward *et al.* was about 15°K, which is rather close to the value shown in Fig. 9. Gierasch and Toon (1973), however, pointed out that with unlimited CO₂ polar temperatures would reach 190°K at $\theta = 35^{\circ}$ because of atmospheric heat transport from the equator.

(e) Summary

With limited CO₂ available, the Martian climate for various obliquities and eccen-



FIG. 9. The annual average temperature at different latitudes for various obliquities and eccentricities. Annual average temperatures for all latitudes are given for an eccentricity of 0.14, while for negative latitudes and $\theta = 25^{\circ}$ and $\theta = 45^{\circ}$, they are given for eccentricities of both 0.0 and 0.14. The higher polar temperature at negative latitudes, with $\theta = 25^{\circ}$ and $\theta = 45^{\circ}$, corresponds to e = 0.0. The annual average temperature variation is similar to the insolation variation illustrated in Fig. 2. The temperature is not a linear function of insolation and is a stronger function of eccentricity than is implied by the dependence of annual average insolation on eccentricity. The polar temperatures do not vary greatly between 25 and 35° obliquity. Annual average temperatures are important because most of the regolith responds to annual average temperatures rather than daily average temperatures.

tricities is similar to the present climate. As the obliquity or eccentricity increase, seasonal extremes are exaggerated. The major changes from present-day Martian conditions are that very high (273°K) summertime daily average polar ground temperatures occur at high obliquity and that permanent CO₂ polar caps form at low obliquity if the CO₂ albedo remains high, as seems most likely. The annual average temperatures depend slightly on the orbital parameters. The annual average polar ground temperature may increase by about 10 to 15°K as the obliquity increases from 25 to 35°. Previous Martian climate models erroneously assumed that Mars currently has massive CO₂ permanent polar caps. These early models predicted large pressure and temperature changes that are not found in the present limited CO₂ model.

III. THE SEASONAL CYCLE OF CO₂ INCLUDING CO₂ EXCHANGE WITH THE REGOLITH

In the previous section we discussed Martian climate changes involving the CO_2 in the atmosphere and in the annual polar caps. However, much more CO_2 may lie hidden on Mars. Here we consider whether or not some of this hidden CO_2 could participate in climate changes.

(a) CO_2 Reservoirs

Early analyses of Viking measurements of noble gas concentrations suggest that Mars outgassed 150 to 500 mbar of CO₂ (Anders and Owen, 1977). Since the Pioneer exploration of Venus, theories of the noble gas distributions on the terrestrial planets have been revised, and recent work relating the Viking observations of noble gas inventories to the abundances of other volatiles on Mars implies that 1 to 3 bars of CO₂ may have been outgassed during Martian geologic history (Pollack and Black, 1979). Much of the outgassed CO₂ may now be in the form of carbonates (Toulmin et al., Anders and Owen, 1977). If 4% of the soil mass is CO₂ in a carbonate phase, and if the regolith is 100 m thick, then about 800 g cm^{-2} of regolith could be CO_2 . This is equivalent to about 50 current atmospheres or about 300 mbar pressure. Unfortunately, CO₂ converted to carbonates would be permanently prevented from returning to the atmosphere barring extensive volcanic reprocessing of surface materials.

 CO_2 may also be adsorbed on the regolith, as suggested by Davis (1969) and Fanale and Cannon (1971, 1974, 1978). Direct evidence that heated Martian soil evolves CO_2 was obtained by Viking, but we cannot quantify the fraction desorbed because of the additional possibility of carbonates breaking down (e.g., Biemann *et al.*, 1977).

The Martian topsoil may be composed of a combination of clays and iron oxides (Toon *et al.*, 1976; Toulmin *et al.*, 1977). However, the crust of Mars is most likely

some type of mafic igneous rock such as basalt. Whether most of the regolith is composed of clays, like the surface layer. or basalt, like the crust, is unknown. The Fanale and Cannon (1971, 1974, 1978) experimental studies of CO₂ adsorption on basalt, pulverized to simulate Martian soil, suggest that a 1-km-thick basalt regolith could adsorb 300 to 800 g cm⁻² of CO₂. In a basalt-carbonate regolith, carbonate would contain most of the CO₂. Fanale and Cannon (1978, 1979) suggested that clay minerals may adsorb 10 times more CO₂ than basalts. In such a case 300 to 800 g cm⁻² of CO_2 could be contained in a 100 m clay regolith. A clay-carbonate regolith would have CO₂ equally partitioned between clay and carbonate.

To utilize adsorption data it is convenient to obtain an analytic expression to fit the available data. Figure 10 presents basalt adsorption data of Fanale and Cannon (1971, 1974, 1978, 1979) which were obtained for different basalt powders with different measured surface areas ($5.8 \text{ m}^2 \text{ g}^{-1}$ in 1971 and $4.4 \text{ m}^2 \text{ g}^{-1}$ in 1978). Normally the volume adsorbed is directly proportional to the surface area, so we have normalized the data to $4.4 \text{ m}^2 \text{ g}^{-1}$ which brings the Fanale and Cannon measurements at 195 and 196°K into good agreement.

Unfortunately, there is no universal theory of adsorption. After rejecting several theoretically justifiable analytic forms, we found a good empirical fit to the adsorbed mass per gram of basalt at low temperature using

$$\rho_{a} = \rho_{0}(A_{s}/A_{0})\exp(FT)$$

$$(P/P_{0})^{\chi} \text{ g CO}_{2}/\text{g soil}, \quad (2)$$

where P is the pressure in millimeters of Hg, P_0 is 1 mm, T is the temperature, ρ_0 is 1.5×10^{-2} g CO₂/g soil (equivalent to 7.8 cm³ STP g⁻¹), A_s is the actual adsorption surface area, A_0 is 4.4 m² g⁻¹, χ is 0.275, and F is -1.5×10^{-2} °K. This empirical equation satisfies three important criteria: as P ap-



FIG. 10. The mass of CO₂ absorbed per gram of soil is illustrated as a function of pressure for various temperatures. In (A) the solid curves are simple analytic functional fits to Fanale and Cannon's (1971, 1974, 1978, 1979) data for basalt. In (B) the solid curves are functional fits to Fanale and Cannon's data for clays.

proach 0, ρ_a approaches 0; ρ_a is linear in surface area; and ρ_a depends on only two constant coefficients, F and χ .

Equation (2), as shown in Fig. 10A, overestimates the CO_2 adsorbed by basalt at high temperatures. However, so little CO_2 is then adsorbed and Mars is so much colder than 300°K that this error is not significant for our work. Two important aspects of Fig. 10A are that large temperature changes are needed to remove substantial fractions of CO_2 from the ground and that the curves have a strong pressure dependence, especially below 2 mbar.

Fanale and Cannon (1979) recently performed clay (nontronite) adsorption measurements, which are illustrated in Fig. 10B along with an empirical fit using Eq. (2). In Eq. (2) we employed the same values of χ and F as for basalt. We used 9×10^{-3} g CO_2/g soil for ρ_0 . The value of A_s at 158°K is 78 m² g⁻¹. Fanale and Cannon (1979) point out that the CO₂ adsorption area for clays is temperature dependent because H₂O occupies fewer adsorption sites at lower temperatures leaving more sites available for CO_2 . To simulate this effect we introduced additional temperature dependence by setting $A_{\rm T} = A_{158}(158/{\rm T})^2$. The poor fit to the data in Fig. 10B at 195°K illustrates that the adsorption area has a more complex temperature dependence than this. Further experimental work is necessary to obtain a meaningful improvement in the analytic fit, however. As for basalt, clay adsorption depends strongly on pressure for low pressures. Unfortunately the available data do not extend to low enough pressures to define the trend very well. At low temperatures (160°K) clays release about 10 times more CO₂ than do basalts for the same temperature change. However, near 200°K clays only release a few times more CO₂ than do basalts for a given temperature change.

The depth of the Martian regolith also determines the total CO_2 adsorbed. Fanale (1976) argued that the regolith may be 1 to 2 km deep based upon photographic indications from Mariner 9 and estimated ejecta volumes from craters. However, Fanale and Cannon (1979) suggested that some of the regolith material may have been lithified during Martian geologic history and they therefore adopted a mean regolith depth of a few hundred meters. Anders and Owen (1977) suggested, from comparison of predicted chlorine and sulfur abundances with Viking observations of Cl and S concentrations in the soil, that the regolith is only 70 m deep. However, the new theories of noble gas distributions on the terrestrial planets (Pollack and Black, 1979) would probably considerably raise the predicted abundances of Cl and S and therefore considerably increase the estimated regolith depth. Hence, the regolith is estimated to be between 100 and 1000 m deep with an intermediate value being most likely.

(b) Characteristic Diffusion Times of CO₂ in the Martian Regolith

 CO_2 could be driven between atmosphere and regolith either by temperature or by pressure changes. In Section II we described typical surface pressure and surface temperature changes caused by orbital element variations. However, it is necessary to extrapolate these changes into the ground. A complete solution requires a coupled thermal and vapor diffusion model at several latitudes which is integrated over long time periods. Before proceeding to discuss this problem, we consider the thermal and vapor diffusion time scales, which are often greatly different.

The thermal skin depth (the depth where $\Delta T = e^{-1}$ of ΔT at the surface) is

$$\lambda_{\rm t} = (\gamma/\pi)^{1/2} (I/\rho_{\rm S}C_{\rm H})$$

= 2.2 × 10⁻² cm sec^{-1/2} (y/\pi)^{1/2}. (3)

Here γ is the period of the thermal variation, and we have used typical Martian surface thermal inertia I (6 \times 10⁻³ cal $cm^{-2} sec^{-1/2}$ °K), heat capacity $C_{\rm H}$ (0.17 cal $g^{-1} \circ K^{-1}$), and soil density ρ_s (1.6 g cm⁻³). The thermal inertia, remotely sensed, is relevant for the soil surface, but at depth in the regolith it could increase because of soil compaction. The thermal conductivity also depends on pressure when the gas mean free path is comparable to the pore size in the rocks. Thermal inertia increases approximately as the fourth root of pressure in the range from 1 to 100 mbar for coarse basalt powders (Wechsler and Glaser, 1965). For clays with very small pore sizes the pressure dependence might be much less. Viking observations produced no correlation between surface thermal inertia and elevation except for extreme elevations, suggesting either that the pressure dependence is low or that some other factor such as particle size is more important. Due to these compaction and pressure effects, the thermal skin depth is uncertain by a factor of several. The time for a thermal wave to reach the bottom of a 1-km regolith is about 10^6 years and the time with a 100-m regolith is about 10^4 years. A shallow regolith is optimum for responding to thermal forcing.

The time scale for propagation of a pressure wave depends upon soil porosity, E, tortuosity, q, pore size, r, and adsorption. Viking measurements (Clark *et al.*, 1977) suggest that E is 0.60 ± 0.15 . We adopt a value of 0.5 to represent the regolith column. The tortuosity (which represents the ratio of blind to open paths available to a diffusing molecule) is not known, and we take a value of 5 in agreement with Smoluchowski's ideas (1968). The pore size is the most uncertain parameter.

The pore size could be fixed by the grain size of Martian soil, which is thought to be between 10 and 100 μ m (Kieffer *et al.*, 1977; Shorthill et al., 1976). However, r is most likely to be set by the smallest available particles. Toon et al. (1977) and Pollack et al. (1977) concluded that the size of Martian dust storm particles was about 1 μ m, which suggests a pore radius $r \le 1 \mu$ m. In some materials the pore size distribution is bimodal, so that large pores occur between grains and small pores occur on grain surfaces. Diffusion occurs preferentially in the larger pores, while adsorption occurs mainly in the smaller ones. The radius of these smaller pores can be determined from adsorption measurements as $r = \frac{1}{2}$ (pore volume/surface area). For Fanale and Cannon's basalt this calculation yields $r \simeq 0.05$ μ m. For basalt, it is unlikely that this pore radius is the relevant one for diffusion calculations because it is much smaller than the dimensions of the basalt grains. However, clays are composed of very thin plates which can be closely packed. For clays the adsorption area pore size may be close to the relevant one for diffusion. A clay with 10 times the adsorption area as basalt has $r \simeq 5 \times 10^{-3} \ \mu m$. Hence the possible range of r on Mars is between 5×10^{-3} and $1 \ \mu m$.

The diffusion equation in a porous adsorbing medium is (Barrer, 1967)

$$E \frac{\partial \rho_{\rm g}}{\partial t} = E \frac{\partial}{\partial z} D_{\rm g} \frac{\partial}{\partial z} \rho_{\rm g} - \frac{\partial}{\partial t} \rho_{\rm s} \rho_{\rm a} + \frac{\partial}{\partial z} D_{\rm s} \frac{\partial}{\partial z} \rho_{\rm s} \rho_{\rm a}, \quad (4)$$

where ρ_{g} is the mass density of the gas, D_{g} is the effective volume diffusion coefficient, $\rho_{\rm s}$ is the soil density, $\rho_{\rm a}$ is the adsorbed mass per gram of rock, and D_s is the surface diffusion coefficient. Physically, the second term on the right-hand side of Eq. (4) represents the local conversion from adsorbed CO₂ to gaseous CO₂ and the final term represents a surface flux which occurs on adsorbing materials as the molecules hop along the surface. We shall ignore the surface flux because its diffusion coefficient is extremely complex and cannot be adequately predicted. Furthermore, its neglect yields a smaller error than is already present because of uncertainties about the pore size.

We can convert Eq. (4) to a simpler form by using Eq. (2) and the ideal gas equation

$$\frac{\partial \mathbf{P}}{\partial t} = \frac{1}{C} \frac{\partial}{\partial z} K_{\mathrm{d}} \frac{\partial}{\partial z} P, \qquad (5)$$

where

$$C = E + \rho_{s}\chi\rho_{0}\left(\frac{A}{A_{0}}\right)$$

$$\exp(FT)\left(\frac{P}{P_{0}}\right)^{\chi}\frac{R_{g}T}{P} = E + \frac{\chi P_{adsorbed}}{P},$$

$$K_{d} = \frac{E}{q}\left(\frac{3\pi}{64}\frac{r}{\lambda} + \frac{\pi}{4}\frac{2r/\lambda}{(1+2r/\lambda)} + \frac{1}{(1+2r/\lambda)}\right)\frac{4}{3}r\left(\frac{2R_{g}T}{\pi}\right)^{1/2}.$$
 (6)

Here R_g is the CO₂ gas constant and λ is the mean free path of CO₂ (6 μ m at $P_{ref} = 6$ mm). Here $\lambda = 6 \mu m$ (6 mm/P).

The expression for K_d (Barrer, 1967) includes the fact that when the pore radius is larger than the mean free path for gas molecule collision the CO₂ flows out along streamlines. When r/λ is less than 1 the CO₂ molecules collide with the pore walls and diffuse from the ground. In our case we expect $r/\lambda \ll 1$ (if $r = 1 \ \mu m$, $r/\lambda \approx 0.1$). In the range $0 < r/\lambda < 1.5$, K_d varies from its value when $r/\lambda = 0$ by only 5%. Hence we can extract K_d from the derivative in Eq. (5) and define a vapor diffusion skin depth as

$$\lambda_{\rm d} = (\gamma/\pi)^{1/2} (K_{\rm d}/C)^{1/2} = (\gamma/\pi)^{1/2} L_{\rm d}.$$
 (7)

Values of L_d are given in Fig. 11 for basalt powders having various pore sizes and temperatures. Figure 11 also illustrates the corresponding thermal diffusion term. For clays at low temperatures, the value of C is 10 times larger than for basalt and the diffusion length is $10^{1/2}$ times smaller than those depicted in Fig. 11. Of course, clays are also very likely to have a smaller pore radius than do basalt soils.

Figure 11 shows that except for the small pore sizes typical of clays, the pressure diffusion length exceeds the thermal lengths. For example, during the time that an annual thermal wave penetrates 1 m, a pressure wave will penetrate about 5 m in basalt with $r = 1 \mu m$. If adsorption did not occur the pressure wave could penetrate to 30 m. Adsorption greatly damps diffusion because part of any vapor phase pressure drop is made by desorption, and much more mass resides in the adsorbed phase than in the vapor phase.

It should be recognized that no experimental studies of diffusion of CO_2 through analogs of typical Martian soils are available. Studies of terrestrial analogs would be quite useful to determine more precisely the time required for CO_2 to move through soils.



FIG. 11. The diffusion wavelength is illustrated for various pressures and pore sizes for basalt. For very large pore sizes CO₂ can rapidly flow through the soil, but at small pore sizes CO₂ cannot rapidly diffuse through the soil. For $r = 1 \ \mu m$ we illustrate several cases of interest. We show that diffusion wavelengths for various temperatures do not differ greatly. We also demonstrate a diffusion wavelength curve for a soil that does not adsorb CO₂. The presence of adsorption severely damps the pressure wavelength. Clays are generally 10 times as adsorbing as basalts. Hence the diffusion wavelengths for clavs are roughly 10^{1/2} times smaller than those illustrated here. Also shown is the thermal diffusion wavelength, which is only weakly dependent on pressure. CO2 will normally diffuse faster than heat through basalt unless the pore size is less than about 0.1 μ m. CO₂ will diffuse faster than heat through clay unless the pore size is less than about 0.5 μ m. The quantity depicted is not a proper wavelength or skin depth until it is multiplied by the square root of the forcing period.

(c) Damping of the Annual Pressure Wave by the Regolith

Dzurisin and Ingersoll (1975) suggested that the annual CO_2 pressure wave (recall Figs. 4 and 7) might be damped if the atmosphere were in contact with an isothermal CO_2 reservoir such as the regolith. Before considering exchange of CO_2 with the regolith over time scales relevant for climate, it is first of interest to consider whether or not the regolith now interacts with the annual CO_2 pressure wave. Any such observed interaction could provide experimental guidance for climate problems. Present observations of the seasonal atmospheric pressure wave are in agreement with simple models not involving any regolith absorption (Hess *et al.*, 1977, 1979). However, careful comparisons of models of the annual pressure wave with observations have not been made.

The total amplitude of the annual pressure wave is about 2 mbar. Figure 10 shows that a 2-mbar pressure drop over a basaltic soil will desorb 1 to 2 \times 10^{-4} g of CO_2/g of soil depending on the temperature. Hence, to desorb 1 mbar of CO_2 , the pressure drop must occur over a depth of 75 to 150 m. If a pressure wave is to penetrate 75 m in 300 days, then L_d from Eq. (7) must be about 2 cm sec $^{-1/2}$. Figure 11 requires a pore size considerably in excess of 50 μ m for $L_d = 2$ cm sec $^{-1/2}$. If we assume that the soil is composed of clavs and that the pressure drop of 2 mbar leads to desorption of 1 to 2 \times 10⁻³ g of CO₂/g of soil, then we need to desorb only 7.5 to 15 m of regolith. For such desorption to occur in 300 days a pore size of about 50 μ m would be required.

These results suggest that the annual pressure wave will not be damped by the regolith unless the pore size is much larger than we now suspect or unless surface diffusion is more important than we expect. Experimental studies of terrestrial analogs might be useful to gain further insights into pore sizes and surface diffusion phenomena. The important point, however, is that the lack of observed interaction between the regolith and the atmosphere on an annual time scale is due to the difficulty of diffusion through the regolith and it is not evidence that Martian soil contains no adsorbed CO_2 .

If future observations do indicate regolith-atmosphere interaction on a seasonal time scale, it might prove fruitful to consider an alternate mechanism for removal of CO₂ from the regolith involving the annual thermal wave. The total CO₂ adsorbed on a basalt regolith which is 1 annual thermal skin depth deep is only 2% of the atmospheric pressure. However, a 1-m clay regolith may contain 1.5 mbar of CO₂. During 300 days CO₂ could penetrate 1 m of clay with a pore radius of $1 \mu m$. However, the thermal wave probably would not release the full 1.5 mbar of CO_2 unless r or the adsorption is larger for clays than we expect. A considerable phase lag would exist and, in addition, the amplitude would be damped because adsorption in the winter hemisphere would oppose desorption in the summer hemisphere and because the seasonal thermal wave is significant only at high latitudes.

(d) Atmospheric Pressure Changes Due to Thermal and Pressure Forcing

The characteristic period of the Martian orbital elements is 10⁵ years, during which time a thermal wave could propagate 300 m and a pressure wave in $r = 1 - \mu m$ basalt could penetrate 1.5 km. Hence, a basaltic regolith could desorb CO₂ essentially instantaneously in response to a slow thermal wave. For clays with $r = 0.01 \ \mu m$, a pressure wave could travel only about 40 m during 10⁵ years. However, a 100-m-thick clay regolith would contain such a large amount of CO_2 that desorption over the region affected by the thermal wave would be similar in amount to that for 1 km of basalt. The important point is that pressure diffusion in a basalt or clay regolith on climatic time scales probably does not completely prevent the regolith from responding to climate changes.

We can now estimate the maximum amount of CO_2 exchanged between the atmosphere and regolith caused by the pressure and temperature changes found in Section II by simply assuming that the full pressure or temperature change at the surface propagates through the regolith exchanging the maximum possible adsorbed CO_2 according to Figure 10; and that all the CO_2 exchanged either reaches the atmosphere or is removed from it. We now perform such a calculation for basalt and clay regoliths.

Figure 9 suggests that, if the obliquity changed from 25 to 35° (with e = 0), the ground equatorward of 55° latitude would cool by 2 to 5°K while the ground poleward of 55° latitude would warm by 15°K. We can find the maximum possible CO₂ desorbed by assuming: that the pressure dependence of the adsorption can be ignored; that ΔT occurs over the whole depth of the regolith; that equatorward of 55° latitude T = 220°K and $\Delta T = -2°K$; and that poleward of 55° latitude $\Delta T = 15^{\circ}$ K while T =160°K. Then we find that the equatorward region adsorbs (over 300 m of regolith) about 1.2 g/cm² and the polar regions desorb about 20 g/cm². However, only 20% of the planet's area is above 55° latitude, so the planetary average CO₂ desorbed is about 3 g/cm^2 . Since the present atmospheric pressure on Mars is equivalent to 16 g/cm² the pressure change will be equivalent to only 1.15 mbar. A similar calculation for clays suggests that $\Delta P \simeq 20$ mbar for a 300-m regolith.

If we did not include the equatorial adsorption, then the polar warming could have increased the pressure by about 1.5 mbar for basalt and 25 mbar for clays. The result is linear with regolith depth, so if the thermal wave reaches 1 km rather than 300 m, then 5 mbar could be released from basalt and 82.5 mbar from clays. For clays it is very difficult to justify a 1-km regolith which is diffusively coupled to the surface. In addition, desorbed CO₂ comes from very high latitudes where permafrost is likely to block the pores and further decrease the CO₂ desorption.

Fanale and Cannon (1978) argued that the results of even large (40°C) temperature changes over 10⁵ years will be small (30 mbar) for basalt because the adsorbed mass depends on pressure as well as temperature so that one cannot move along vertical lines in Fig. 10 in studying desorption for varying temperatures. Our analysis, however, shows that realistic temperature changes in equatorial and polar latitude are rather small and partially compensating, so that the net pressure change is very small even when the pressure dependence of the isotherms is ignored.

Fanale and Cannon (1979) have also evaluated clay and basalt desorption for a 15°K polar temperature increase. For clays, they suggested that 50 mbar of CO₂ would be desorbed at high obliquity, but that correcting for the pressure dependence of the isotherms would reduce the pressure to about 29 mbar and correcting for latitudinal dependence of adsorption would further reduce the pressure to 17 mbar. They assumed a larger volume of polar regolith, a higher soil density, and a higher temperature dependence of the isotherms at low temperatures than we did, but their final answer is similar to ours. For basalt, Fanale and Cannon (1979) found $\Delta P = 9$ mbar. However, they considered a 1-km-thick polar regolith extending over a much larger geographic area and a higher soil density (2 g/cm^3 compared with our 1.6 g/cm^3) than we did. The difference between our predicted pressure at high obliquity for basalt (7.15 mbar) and theirs (15 mbar) is a measure of the inherent uncertainty of the calculation.

The preceding discussion has centered on the role of temperature oscillations in desorbing CO₂ from the Martian regolith. The effects are small because: (1) temperature changes from orbital perturbations are small and of opposite sign at different latitudes; (2) the thermal wave cannot propagate very far in 10^{5} years; and (3) desorption is limited by the pressure dependence of the adsorption isotherms.

Orbital element changes cause not only temperature changes but also pressure variations when the permanent polar cap forms at low obliquity. Basalt adsorption is a strong function of pressure when P is less than a few millimeters of Hg. As Fig. 4 shows, the atmospheric pressure can fall to

1 mm or less at low θ . Such low pressures could lead to nearly complete desorption of a basaltic, $r = 1 \,\mu m$, Martian regolith. (The temperature changes due to the altered θ have negligible effects on adsorption by comparison.) For clays with $r \simeq 0.01 \ \mu m$ perhaps only 40 m could be desorbed. For clays we are not certain that the isotherms are strongly sloping with pressure at 1 mm, but a strong slope must develop at some low pressure. In either case, permanent CO_2 caps may form not only from the present atmospheric mass but also from the adsorbed mass. Assuming that the permanent caps occupy 1% of the area of Mars, the thickness of a permanent cap formed from the present atmosphere would be about 10 m. If there is even as little as 30 times as much CO₂ mass in the regolith as in the current atmosphere, the polar caps at low obliquity could reach a depth of 300 m.

It is interesting to speculate concerning possible hysteresis effects in the regolithpolar cap system. For example, suppose declining obliquity reduces the atmospheric pressure to a very low value and slowly desorbs the regolith. If some other process, such as an episode of volcanic activity, occurred to blacken the polar cap, then the CO_2 could be driven into the atmosphere. The CO_2 might not return to the cap if feedback effects maintained high polar temperatures (Gierasch and Toon, 1973) and the CO_2 could not reenter the regolith immediately because of diffusion-time constraints.

(e) Regolith Maintenance of Mean Atmospheric Pressure over Geologic Time Scales

The remaining question concerning regolith adsorbed CO_2 is its role in maintaining the mean Martian atmospheric pressure. Leighton and Murray (1966) noted that the Martian atmospheric pressure is very close to the vapor pressure of CO_2 at reasonable Martian polar temperatures. Ingersoll (1974) pointed out, however, that the polar temperature, and therefore the vapor pressure, is a very strong function of surface albedo, which is not well known, so that the equality of the vapor pressure and the observed pressure is not very exact. Viking results (Kieffer et al., 1976) have now shown that the north polar permanent cap of Mars is not solid CO₂, although the south polar cap may contain some solid CO₂. Since the behavior of the south polar cap during Viking observations seemed anomalous (James et al., 1979), it is likely that the south polar cap does not contain a substantial CO₂ reservoir. Even if the entire cap were CO₂, it would only contain the equivalent of 6 mbar of CO₂ (Fanale and Cannon, 1979). Hence, the atmospheric pressure is probably not the vapor pressure of polar CO₂ deposits. This means that the observed pressure is either a nonequilibrium transient or that it is in equilibrium with the regolith as previously suggested by Fanale and Cannon (1978).

Figure 11 emphasizes that for moderate pore sizes even a 1-km regolith would be in equilibrium with the atmosphere on time scales longer than about 10⁵ years, which is much shorter than reasonable time scales for outgassing, or carbonate formation. In addition, the annual mean pressure is not a strong function of obliquity except for obliquities less than 20°, or greater than 30°, and Ward's (1979) calculations suggest that the Martian obliquity has not been lower than 20° nor higher than 30° for nearly $5 \times$ 10⁵ years. Hence, it is likely that the current pressure is in equilibrium with the regolith. As previously discussed, the thermal wave due to the obliquity oscillation should produce only small pressure changes.

If the atmospheric pressure is controlled by regolith adsorption, then the present atmospheric pressure on Mars is the result of the total CO_2 outgassed, the present regolith depth, and the present conversion of CO_2 into nonadsorbed forms. The pressure is exceedingly sensitive to any modification of these parameters. For example, if the depth of a basalt regolith were increased by 20% the adsorbed CO_2 per gram would decrease by 20% and Fig. 10 indicates that the atmospheric pressure would fall by about 50%. Likewise, CO_2 is probably converted to carbonate on Mars, although we have no direct evidence of the carbonate abundance (Gooding, 1978). If 4% of the regolith is CO₂ tied up in carbonates, then approximately 20 times more CO_2 is stored as carbonate than as adsorbed CO_2 on basalt. Figure 10 shows that increasing the adsorbed CO_2 by a factor of 20 by converting the carbonates to CO₂ would lead to an immense increase in the atmospheric pressure. The precise pressure cannot be calculated because the adsorption at high pressures has not been measured and many adsorption isotherms have greatly increased dependence of adsorption on pressure at higher pressures (e.g., Brunauer et al., 1967). However, for a current regolith containing 50 mbar of adsorbed CO_2 , it would certainly seem possible for the atmospheric pressure to exceed 150 mbar if all the hypothetical carbonates were converted to adsorbed CO₂. If the regolith is composed to clays, then carbonates may contain only 50% of the total CO_2 , and converting the carbonates to CO₂ would only result in a modest pressure increase.

Over geologic history various processes, such as greenhouse warming, might have maintained the Martian surface temperature at much higher temperatures than at present. The curves in Fig. 10 show that at low temperatures basalt adsorption is a strong function of pressure. For example, all the CO_2 desorbed by a sample of 6 mbar, if warmed from 160 to 195°K, would be readsorbed before the pressure reached 25 mbar. However, experimental data suggest that at high temperatures (273°K) the adsorption is a weak function of pressure. If some processes (greenhouse warming in an early atmosphere) considerably warmed Mars for a long period, then little CO_2 would be adsorbed on the regolith.

There may have been some period when Mars had no regolith, had only a weakly adsorbing regolith, or had a regolith that was distributed quite differently than now. Perhaps early in Martian geologic history CO₂ had been outgassed, but no regolith or only a thin basalt regolith had formed. Then CO₂ would remain in the atmosphere-polar cap system. Alternatively, suppose that the regolith on Mars is not distributed uniformly with latitude. At present, extensive dust laminae of total thickness between 1 and 6 km extend poleward of 80° at both Martian poles (Dzurisin and Blasius, 1975; Cutts et al., 1976). In addition, high Martian latitudes are covered by a deep mantle presumably of loose, wind-eroded debris (Soderblom et al., 1973). At present T =170°K for these materials, but suppose that in the past they were all at the equator where $T = 220^{\circ}$ K and that they then covered one-half the planet's surface to a depth of 500 m. The resulting amount of desorbed CO_2 , if the regolith were basalt, would have been about 15 mbar. If the regolith were clay, then 180 mbar of CO₂ could have been desorbed although the pressure dependence of the isotherms would damp somewhat the net CO_2 entering the atmosphere.

(f) Summary

As much as 1 to 3 bars of CO_2 may be adsorbed on the Martian regolith or permanently tied up as carbonates in the regolith. Diffusion of CO₂ through the Martian soil is slow, preventing the adsorbed CO₂ from affecting the annual CO₂ pressure cycle. On climatic time scales diffusion cannot be ignored, but it probably does not prevent the regolith from responding to climate changes in the surface pressure and temperature. In agreement with Fanale and Cannon (1979) we find, even given the most favorable assumptions, that polar ground temperature changes are too small to desorb large quantities of CO₂ at high obliquities. The pressure at $\theta = 35^{\circ}$ is likely to be no greater than 20 mbar for either a clay regolith or a basalt regolith. At low obliquities, the low atmospheric pressure caused by the formation of a permanent CO_2 polar cap may strongly desorb the regolith and large quantities of CO_2 from the regolith may be deposited on the polar caps. The present atmospheric pressure is likely to be in equilibrium with the regolith as suggested by Fanale and Cannon (1978). Over geologic time the atmospheric pressure has probably greatly decreased as carbonates have formed, as highly adsorbing clays have formed, as the regolith has become deeper, and as the regolith has been preferentially deposited in polar regions. During a possible early epoch with high surface temperatures, adsorption would not have been an important sink for CO_2 .

IV. THE SEASONAL CYCLE OF ATMOSPHERIC DUST

Global-scale dust storms, which now occur every Martian year, are an important component of Martian meteorology and climatology. The atmospheric thermal structure is stablized by solar energy deposition in suspended dust grains. Particulate backscattering of solar radiation may significantly reduce the solar energy reaching the summer polar cap and alter its heat budget. Dust grains deposited in the permanent and seasonal ice caps substantially modify the ice albedo which also alters the polar heat budget.

As we have discussed, variations in the orbital elements can modify the atmospheric pressure, the temperature, and the seasonal polar cap boundary. Below we review ideas about how these modifications may alter the dust storm cycle and then we discuss how changes in the dust storm cycle can "feedback" to modify pressure, temperature, and polar cap boundaries.

(a) Effects of Climate Changes on Dust Storm Generation and Decay

The generation of global dust storms on Mars requires local winds at the high velocity end of the Martian wind spectrum. These winds lift dust locally. Then a positive feedback occurs between radiant energy deposition in the suspended dust grains and the global winds; this then enhances dust-lifting winds over a large area of the planet (Pollack *et al.*, 1979). The wind speed required to lift dust depends upon the pressure, while the wind velocities themselves depend upon horizontal thermal gradients.

Wind tunnel studies (Pollack et al., 1976) have shown that the wind speed, u, at the top of the boundary layer required to lift dust varies with pressure as $u\alpha P^{-0.55}$. At present, u is about 95 m/sec, but if P were 1 mbar, u would be 230 m/sec. If P were 25 mbar, u would be about 40 m/sec. Since Martian winds now are barely sufficient to lift dust, it is clear that at lower obliguities and lower atmospheric pressures dust lifting will not occur (Ward et al., 1974; Pollack, 1979). At slightly higher pressures, which could be reached at high obliquity if the regolith is composed of highly adsorbing clays, dust lifting should occur much more easily than it does today (Pollack, 1979).

At the present time, global dust storms occur preferentially during southern hemisphere summer, presumably because wind speeds reach their peak at that time as the solar insolation is maximized at perihelion passage. At slightly higher pressures, dust storm generation may occur during the summers of both hemispheres because of the less restrictive wind speed requirements. Hence, at both larger and smaller obliquities than those of today, dust storm generation may be independent of the eccentricity and precessional cycles.

The mechanisms responsible for creating winds strong enough to lift dust are not known well enough, nor are they simple enough, to allow more than qualitative extrapolations to other climates. Pollack *et al.* (1979) discuss the possible importance of polar cap, tidal, topographical, frontal, and seasonal winds for dust storm generation at present.

Polar cap winds are driven by the thermal contrast between frost-free and frost-covered terrain and by the CO_2 mass flux into and out of the polar caps. Peterfreund and Kieffer (1979) reported that polar cap winds generate many local dust storms. The thermal contrast winds will be unaffected by changes in the orbital elements, but the CO_2 mass flux will be enhanced and the edge of the polar cap will move equatorward at high obliquity because of the larger seasonal exchange of CO_2 as seen in Figs. 4 and 5.

Topographical winds are driven by largescale elevation differences, while tidal winds are driven by the diurnal movement of the subsolar point. The winds should be affected in three ways by orbital element changes. First, both wind systems depend upon the small value of the radiative relaxation time of the Martian atmosphere. At higher obliquities, and hence higher pressures, these wind systems could be damped because the radiative relaxation time will become longer. Second, at higher eccentricities the diurnal temperature gradients will be accentuated near perihelion, and this may increase the amplitude of tidal and topographic winds. Finally, the time of year of perihelion passage and the obliquity may affect the phase, amplitude, and latitude of wind speed maxima. These changes could be important if tidal or topographic winds interact either with each other or with seasonal winds to produce additive effects.

Seasonal and frontal winds will be altered by thermal gradient modifications due to obliquity changes. The frequency of frontal systems may increase and fronts may be concentrated in lower latitudes at higher obliquity because midlatitude temperature gradients will decrease while lower latitude gradients will increase (Figs. 3 and 6). In addition, at high obliquity, summertime temperature gradients as large as those during winter, but of opposite sign, will occur between the summer pole and equator. Strong summertime easterlies would be expected with possible summertime frontal activity.

Global dust storms start to decay as soon as the storms are generated because radiative heating in an optically thick dust layer stabilizes the atmosphere and shuts off the winds that drive the dust-lifting processes (Pollack et al., 1979). For this reason, global dust storms during a single season may not be more frequent at high obliquity than they are at present during southern hemisphere summer. The decay of the dust storms should be nearly independent of any orbital changes. The removal of the dust from the atmosphere apparently occurs preferentially in polar regions as CO₂ "snow" forms around the dust grains and makes them large enough to fall out of the atmosphere (Pollack et al., 1979). Jones et al. (1979) observed from Viking that even nighttime CO₂ frost formation is sufficient to remove dust from the atmosphere. Polar CO₂ deposits will cover a larger geographic area at higher obliquity and more mass will be exchanged between the atmosphere and poles at higher obliquity. Hence, dust may be removed from the atmosphere slightly faster at high obliquity than at low obliquity. Pollack (1979) has pointed out that the polar debris mantles probably mark the limits of the CO₂ annual cap at high obliquity.

(b) Effects of Changes in Dust Storm Activity on the CO₂ Cycle

The CO₂ cycle on Mars is controlled by the polar energy budget which is a strong function of the orbital parameters and of the polar cap albedo. As the obliquity varies from 15 to 35° the annual average solar energy available at the pole changes by nearly a factor of 2. If the CO_2 frost albedo were to change from 0.6 to 0.8, or if the permanent H₂O ice cap albedo were to change from 0.4 to 0.7, the energy budget would also change by a factor of 2. Below we shall show that changes in the dust-toice mixing ratio can cause such changes in the ice albedo so that modifications of dust storm activity due to orbital element change may be just as important to the polar energy budget as are variations in solar insolation caused directly by orbital element variations.

The albedo of the Martian polar caps is

significantly less than unity. The north summer water ice cap is observed to have an albedo near 0.40 (Kieffer et al., 1976). The albedo of the Martian southern permanent ice cap is not known, but Farmer et al. (1979) state that it is >0.55. The albedo of the seasonal CO₂ caps has not been measured in detail, although James et al. (1979) report values of 0.73 in the red and 0.36 in the violet. A spectrally averaged albedo of 0.65 ± 0.05 has long been used in models of the seasonal polar cap recession (e.g., Kieffer et al., 1977; Davies et al., 1977). For comparisons, the albedo of the Earth's Antarctic ice cap is about 0.82 in summer and 0.88 in winter (Hummel and Reck, 1979); terrestrial snows on open tundra have albedos near 0.8.

The Martian ice caps could have low albedos for three reasons. The first, that the ice may be optically thin, is unlikely. Terrestrial observations show that only a few centimeters of snow is enough to completely mask the ground. In addition, simple two-stream radiative transfer calculations of the albedo of CO_2 frost, overlying ground with typical Martian soil albedo, show that only 1 cm of compacted CO_2 is enough to raise the albedo to 95%, Both the CO_2 seasonal caps and the H₂O permanent caps are much thicker than a few centimeters except perhaps in isolated regions, or during the early stages of ice deposition.

A second possible cause of low ice albedos is exposed patches of bare ground. The albedo of broken snow is $A = fA_{ice} + (1 - f) A_{ground}$, where f is the fraction of ice cover. For an ice albedo of unity, f must be 0.53 to obtain an albedo of 0.65, and f must be 0.2 to obtain an albedo of 0.4. Although exposed ground may help lower the polar albedo, the bare ground fraction suggested by these numbers is so large that some other process must be mainly responsible for the low albedo.

The final possible cause of a low polar albedo is that dust is admixed with the ice. Figure 12 shows the spherical albedo of a semi-infinite layer found from the two-



FIG. 12. The spherical albedo of a dirty ice cap varies greatly with the single-scattering albedo of the ice-dust grains composing the ice, but it is only a moderate function of the asymmetry parameter of the grains.

stream approximation (Irvine, 1975) as a function of the single-scattering albedo $\tilde{\omega}_0$ of the ice-dust grains for asymmetry factors g of 0.88 and 0.80. Pollack *et al.* (1979) find that g for Martian dust is about 0.8 and $\tilde{\omega}_0$ is about 0.86. The calculated Martian surface albedo from Fig. 12 is thus about 0.20, which is in agreement with the observed albedo of 0.214 \pm 0.063 (Kieffer *et al.*, 1977), confirming that the two-stream representation gives a reasonable approximation to the albedo.

The surface albedo is a strong function of $\tilde{\omega}_0$, but it is only a weak function of g when in the range of 0.8 to 0.88. According to Hunt (1973), a value of 0.85 to 0.88 for g at visible wavelengths is representative of ice particles for sizes between 4 and 50 μ m. Since the refractive index of CO₂ is close to that of H₂O ice, a similar value of g applies to CO₂ ice (Thompson *et al.*, 1973).

Figure 12 shows that a polar cap albedo of 0.4 requires that $\tilde{\omega}_0 = 0.97$, while an albedo of 0.65 requires that $\bar{\omega}_0 = 0.994$. The single-scattering albedo ($\tilde{\omega}_0$) of a mixture of ice having unit albedo with dust of albedo $\tilde{\omega}_{0d}$ can be related to the volume mixing ratio of dust (V) by assuming that the ice and dust particles are individual grains and that the optical depths of the dust and of the ice grains are given by the geometric cross sections, a good assumption for the particle sizes of interest. Then it is easy to show that

$$V = (N_{\rm i}/N_{\rm d})^{1/2} [(1 - \tilde{\omega}_0)/(\tilde{\omega}_0 - \tilde{\omega}_{0\rm d})]^{3/2} \quad (8)$$

where N_i/N_d is the ratio of the number of ice particles to the number of dust particles.

Pollack *et al.* (1979) suggest that each water ice particle arrives at the polar cap surrounding a dust particle so that $N_i = N_d$. Hence, the dust-to-H₂O ice mixing ratio in the north polar cap, if it has an albedo of 0.4, is about 15%.

We cannot use $N_i/N_d = 1$ for CO₂, however, since it is likely that much more CO_2 is deposited directly onto the cap from the vapor phase than is formed as dustcontaminated snowflakes in the atmosphere. The difference in the behavior of H_2O and CO_2 frost occurs because, unlike H₂O, the vapor abundance of CO₂ is quite high above the pole, so that CO₂ can directly condense onto the surface. An additional factor is that large CO₂ flakes falling rapidly through the atmosphere will preferentially sweep up large numbers of dust particles in their path. A few large CO₂ flakes will therefore be mixed with many smaller dust particles.

From Viking observations of the abundance of dust in the Martian atmosphere Pollack *et al.* (1979) estimated that 2×10^{-2} g cm⁻² year⁻¹ of dust is deposited poleward of 60° in the north polar region of Mars. Hess et al. (1979) estimated from the observed pressure variation on Mars that the mass of CO₂ deposited in the south polar region is about 35 g/cm^2 of CO₂ spread over 45° latitude. The mass of CO₂ in the seasonal north polar cap should be less (Fig. 7). Hence the volume mixing ratio of dust to CO_2 in the north polar cap is approximately 10^{-3} . Using Eq. (8) and a polar albedo of 0.65 we find that for each CO₂ particle there are about 100 dust particles. If the average dust particle radius is $1 \ \mu m$, then the average CO₂ radius is 50 μ m.

If we assume that the CO_2 deposition is constant, we can calculate the polar albedo change caused by variation in the dust mixing ratio by using Eq. (8) scaled from the N_i/N_d values just found. For twice the dust input of Pollack *et al.* (1979), the polar albedo would decrease from 0.65 to 0.55. For one-half this dust input, the albedo would increase to 0.7, and for one-tenth this dust input, the albedo would become 0.85.

Because of the strong sensitivity of albedo to dust content, a strong feedback exists between dust storm frequency, polar albedo, and CO₂ pressure. At high obliquity, the release of CO₂ from clay reservoirs may provide enough additional pressure to increase dust storm frequency. The greater dust storm frequency in turn should cause the CO₂ albedo at both poles to be low and so encourage higher pressures. On the other hand, at low obliquity the lower pressures (Fig. 4) will decrease the dust storm frequency, the polar cap albedo will increase, and still lower pressures should be favored. This feedback should greatly enhance the possibility of the formation of permanent CO₂ polar caps at both poles when the obliquity is lower than at present.

Dust storms under present conditions occur preferentially during southern hemisphere summer. Therefore, the albedo of CO_2 frost should be higher in the southern hemisphere than in the northern hemisphere. If the dust-to-CO₂ mixing ratio is lower in the southern hemisphere than in the northern hemisphere by a factor of 10, the albedo difference could cause the solar energy budgets of the northern and southern poles to differ by a factor of 2. Better observations of the difference in the albedos between the two seasonal caps and the variations of albedo as the caps sublime during the spring might provide insight into the importance of dust to the polar cap albedos.

The effect of embedded dust on the polar ice albedo is very large. Atmospheric dust could also affect the polar energy budget by blocking solar and thermal radiation. Pollack *et al.* (1979) showed that the surface energy budget at low latitudes is not substantially influenced by atmospheric dust

because the increased ground-level infrared radiation from the dust offsets the groundlevel solar energy deficit. Calculations of the radiation balance in the polar regions have been performed by James et al. (1979). They found that dust with a $\tilde{\omega}_0$ of 0.75 would reduce the amount of solar energy absorbed at the polar surface if the polar albedo were 0.5, whereas if the polar albedo were 0.6 the energy absorbed would increase. If $\tilde{\omega}_0$ were 0.85, James *et al.* determined that the dust decreased the amount of energy absorbed at the pole for both polar cap albedos. James et al. ignored the contribution of the dust to the infrared radiation at the ground, which Pollack et al. (1979) found to be important. Suspended dust may either increase or decrease the energy available at the polar surface depending upon the values of polar albedo, single scattering albedo, and the ratio of infrared-to-visible dust extinction. Much further work is required on this difficult problem, but the sensitivity suggests that the net effect is probably not large, especially when it is compared with the factorof-2 change in the energy budget that could result from admixing different amounts of dust and CO_2 in the polar ice.

(c) Summary

In agreement with previous studies (Ward et al., 1974; Pollack et al., 1979; Pollack, 1979), we find that at low obliquity low pressures will prevent dust storms. At high obliquity, slightly higher pressures will enable dust storms to occur throughout the Martian year independently of the precession of the equinoxes. Since dust storms are self-limiting, the intensity of individual storms, their lifetimes, and their frequency of occurrence should be similar to those found now during southern hemisphere summer. At moderate obliquities, dust storms will be most prevalent at large eccentricity when perhelion occurs near solstice. A strong feedback occurs between dust storm frequency and polar cap albedo. At low obliquity the CO₂ seasonal or permanent polar cap albedo should be greater than the seasonal cap albedo now. At high obliquity the CO₂ seasonal caps at both poles should have albedos similar to, or perhaps slightly less than, that of the present seasonal north polar CO₂ cap. At times of moderate θ and high e, such as the present, the two CO₂ seasonal polar caps should have significantly different albedos because of the seasonal asymmetry in dust storm occurrence. The solar energy budget of the two polar caps could easily differ by a factor of 2 because of the difference in the dust content of CO₂ frost. Better observations of the present polar cap albedoes would be valuable.

V. THE SEASONAL CYCLE OF H₂O

Water vapor in the Martian atmosphere undergoes a strong seasonal variation as it condenses on and sublimes from the permanent H_2O polar caps. Over long times, water vapor also interacts with permafrost and with water bound to soil particles. We now consider the water seasonal cycle under different orbital and obliquity conditions with most concern being given to the stability of the polar caps. Then we discuss long term changes in the amount of permafrost and soil water.

(a) H₂O Reservoirs—Atmosphere and Polar Caps

Presently the Martian atmosphere contains seasonally varying water vapor equivalent to 0.3 to 1.3 km³ of ice (Farmer and Doms, 1979; Farmer et al., 1977). Two permanent water ice caps are located at the Martian poles; their combined area is about 10⁶ km², approximately one-half the area of the Earth's Greenland ice cap or one-tenth the area of the Earth's Antarctic ice cap. The thickness of the Martian polar caps is not known, but Cutts et al. (1976) suggest that ice is very thin (≤ 1 cm) in areas of the south pole where the visible ice cover is variable. Dzurisin and Blasius (1975) deduced that the ice cover is typically a few hundred meters or less since ice does not

cover topographic areas with greater height differences. Murray and Malin (1973) identified one polar region in which ice did seem to cover the topography, implying that the thickness in that region may exceed a few hundred meters.

(b) Maximum Summer Ice Cap Temperatures and Sublimation Rates

We have shown by using a simple model (Figs. 3 and 6) that summer polar ground temperatures can be quite high at large obliquities. The thermal history of an ice cap is more complex than that of the soil and, thus, a more detailed model is required to explore its temperature variation and the associated water vapor loss rate. The heat balance equation for an ice cap is

$$(1 - A)\overline{I} + AH - e\sigma T^{4} - K_{ice} \frac{\partial T}{\partial z} - Ec = \rho C_{p} \frac{\partial T}{\partial t}$$
(9)

The initial term is the daily averaged available solar radiation. This term contains an ice albedo A taken as 0.4 in agreement with present values for the north polar cap (Kieffer et al., 1976). The Martian south polar cap has a higher albedo (>0.55, Farmer and Doms, 1979). The term I is the daily average solar insolation. At the pole there is no variation of the insolation during the day. Ice is moderately transparent to solar radiation, but Martian ice contains a large quantity of dust, as evidenced by the low albedo, so we have assumed that all the solar energy is deposited in the top layer (1.5 cm thick) of our ice cap model. Depositing the solar energy in a thicker layer would result in lower temperatures.

The second term in Eq. (9) represents atmospheric heating caused by thermal conduction from the atmosphere to the ground, infrared radiation from CO₂, and infrared radiation from H₂O. In order to evaluate these terms we assume that the summer atmosphere is isothermal and is always warmer than the polar cap by $T_A - T = 10^{\circ}$ C. This assumption is reasonable since bare ground near the pole is warmer than ice at the pole, which will lead to a warm atmosphere over the pole, and since observations show that the atmosphere over the pole is stable (Kieffer *et al.*, 1976). For heat conduction in a stable atmosphere we follow Gierasch and Toon (1973) and use

$$AH1 = \beta PU(T_A - T),$$

where

$$\beta = 2.6 \times 10^{-5} \text{ °K}^{-1}$$

$$U = \text{typical wind speed} = 10^3 \text{ cm sec}^{-1},$$

$$P = \text{ pressure in dynes cm}^{-2}.$$
 (10)

The atmospheric heating due to CO_2 and H_2O are found by using the emissivities of Gierasch and Goody (1968) and Cess (1974):

$$AHCO_2 = 6 \times 10^5 \ln(1 + 1.97 \times 10^{-3}P)$$

$$exp(-960/T_A)ergs cm^{-2} sec^{-1},$$

$$AHH_2O = 0.75 \sigma T_A^4 [1 - exp(-(P_{H_2O}P)^{1/2})]$$

$$9.1 \times 10^{-5})]ergs cm^{-2} sec^{-1}.$$

In calculating the water emissivity $P_{\rm H_2O}$, the water vapor pressure (in dynes cm⁻²) at the surface is used. The water is taken to be uniformly mixed, and it is assumed that the column abundance of water does not exceed that which had been previously sub-limed from the water cap.

The energy loss terms in Eq. (9) include ground radiation losses (we take an emissivity of 0.95) and conduction into the ice. The conductivity of pure ice at 273°K is about 5 \times 10⁻³ cal cm⁻¹ sec⁻¹ °K⁻¹ and at 150°K it is about 10^{-2} cal cm⁻¹ sec⁻¹ °K⁻¹ (Hobbs 1974). Snow at 273°K has a density dependent K of 6.8 \times 10⁻³ $\rho_{\rm snow}^2$. Dense arctic snow with $\rho = 0.33$ g cm⁻³ has K = 7 \times 10⁻⁴ cal cm⁻¹ sec⁻¹ °K⁻¹ (Maykut and Untersteiner, 1971). Permafrost with 20% soil content can have a 50% lower K than does pure ice (Clark, 1966). To encompass the possible range of conductivities we perform calculations with $K = 8.6 \times 10^{-3}$ (Davies et al., 1977) and with $K = 1 \times 10^{-3}$ cal cm⁻¹ sec⁻¹ $^{\circ}$ K⁻¹.

The final energy loss term is evaporative cooling. Two evaporative mechanisms are at work. A turbulent water vapor flux occurs as winds blow across the ice (Ingersoll, 1974):

$$EC1 = 0.002\rho_{w}(0)UL,$$

$$L = \text{latent heat of ice}$$

$$= 2.8 \times 10^{10} \text{ergs g}^{-1},$$

$$\rho_{w}(0) = \text{water vapor mass density.} (11)$$

Also convective water vapor evaporation occurs because water is lighter than CO_2 (Ingersoll, 1970)

$$EC2 = 0.17 \rho_{\rm w}(0) D \left[(\Delta \rho / \rho) (g / \nu^2) \right]^{1/3} L$$

ergs cm⁻² sec⁻¹,

where

$$\begin{split} &\Delta \rho / \rho = 26 P_{\rm H_{2}O} / (44P - 26P_{\rm H_{2}O}), \\ &\nu = 3.75 \times 10^{-7} (T^{5/2}/T + 269) (R/P) \rm cm^2 \\ &\rm sec^{-1} = viscosity of CO_2, \\ &D = 32T^{3/2} / P ~\rm cm^2 ~\rm sec^{-1} = diffusion \\ &\rm coefficient of ~H_2O ~\rm in ~CO_2, \\ &R = \rm gas ~\rm constant. \end{split}$$

Here we assume that there is negligible water vapor except just above the ice.

The final term in Eq. (9) is the temperature change of the ice. We used an ice heat capacity of 1.4×10^7 ergs g⁻¹ °K⁻¹ which is appropriate for 200°K (Hobbs, 1974). We also considered the energy in the vapor phase molecules caused by vapor pressure changes, but this effect is negligibly small.

To solve Eq. (9) we used the same thermal model as previously employed for the ground except that we took 20 layers of 1.5cm thickness and 20 layers of 30-cm thickness. We began the calculation at the time of complete CO_2 sublimation as found from our earlier calculations (Figs. 3 and 6) and set the beginning ice temperature at all depths in the ice at the annual mean temperature from our earlier calculations.

Figure 13 illustrates ice surface temperatures for various values of obliquity with the maximum possible eccentricity. Also noted is the total thickness of ice sublimed from each cap during the summer season.



FIG. 13. The seasonal dependence of the temperature of a water ice cap is a strong function of the obliquity. Given in parentheses are the thicknesses of ice lost from each polar cap during the summer. The greater thickness is lost from the pole reaching higher temperature. Substantial water loss rates occur at high obliquity. As discussed in the text, the loss rate is very sensitive to ice albedo and thermal conductivity.

Both caps are assumed to have an albedo of 0.4 and a K of 10^{-3} cal cm⁻¹ sec⁻¹ °K⁻¹. The solar insolation integrated over the year is identical at both poles because the pole with the larger insolation at solstice has the shorter summer. However, sublimation of ice at the two poles is not the same because it depends nonlinearly on the insolation. Most evaporation occurs when the ice is warmest.

An evaluation of the sensitivity of the results presented in Fig. 13 to the processes included in Eq. (9) suggests that the albedo, thermal conductivity, and evaporative losses are the major uncertainties. Table I presents the results from a number of calculations exploring the variation of the water mass loss rate caused by changes in these parameters.

Higher water loss rates could occur if the polar albedo were less than 0.4 but, because the northern hemisphere winter is now as dusty as is possible, we do not believe that lower albedos are likely. A calculation with $K_{ice} = 10^{-4}$ cal cm⁻¹ sec⁻¹ °K⁻¹ produced negligible differences from the results given in Fig. 13. Davies *et al.*

Fixed parameter	Variable parameter	Obliquity	Loss rate from hot pole	Loss rate from cool pole	
	A = 0.4	25	1.5	0.2	
		35	20	5	
		45	45	20	
$K = 1 \times 10^{-3} \text{ cal cm}^{-1} \text{ sec}^{-1} ^{\circ}\text{K}^{-1}$					
P = 5 mbar	A = 0.5	25	0.5	0.05	
$U = 10 \text{ m sec}^{-1}$		35	10	2	
		45	30	10	
	A = 0.6	25	0.1	0.01	
		35	5	0.3	
		45	15	2	
	A = 0.4	25	0.1	0.02	
$K = 8.6 \times 10^{-3} \text{ cal cm}^{-1} \text{ sec}^{-1} ^{\circ}\text{K}^{-1}$		35	7	1	
$P = 5 \text{ mbar}$ $U = 10 \text{ m sec}^{-1}$	}	45	25	8	
	A = 0.6	25	0.005	0.001	
	,	35	0.1	0.05	
	P = 5 mbar	45	43		
$K = 1 \times 10^{-3} \text{ cal cm}^{-1} \text{ sec}^{-1} ^{\circ}\text{K}$	P = 25 mbar	35	40		
$U = 10 \text{ m sec}^{-1}$	}	45	78/0.3 melt		
	P = 50 mbar	35	52/77 melt		
	-	45	69/282 melt		
	A = 0.4; P = 5 mbar	25	0.7		
$U = 10 \text{ m sec}^{-1}$	A = 0.4; P = 50 mbar	35	22		
$\mathbf{K} = 0.0 \times 10^{\circ}$ cal cm \cdot sec $\cdot \cdot \mathbf{K}$	A = 0.6; P = 5 or 50 mbar	35	0.1		
$U = 10 \text{ m sec}^{-1}$ $K = 1 \times 10^{-4} \text{ cal cm}^{-1} \text{ sec}^{-1} \text{ °K}$	A = 0.4; P = 5 mbar	45	53		

TABLE I

POLAR CAP WATER ICE LOSS RATES (g cm⁻² year⁻¹) FOR ORBIT OF MAXIMUM ECCENTRICITY

(1977) argue that such low values of K would lead to inconsistencies with presently observed water vapor abundances over the north polar ice cap. Suppressing evaporation by shutting off the wind has a small effect because convective evaporation limits the temperature.

Ingersoll (1970) pointed out that the convective evaporative flux declines as the atmospheric pressure increases. The flux is inversely dependent upon atmospheric pressure because it is driven by density differences between the atmosphere just above the ice and the atmosphere at higher altitudes. With larger pressures, adding a fixed water vapor amount causes a smaller fractional density change than occurs at smaller CO₂ pressures. Without wind, but with an atmospheric pressure of 25 mbar, we find that for e = 0.14 and $\theta = 45^{\circ}$ ice temperatures at the warmer pole reach 273°K, 78 cm of ice sublime, and 0.3-cm melt. If the atmospheric pressure reaches 50 mbar, then the melting point of ice is reached and maintained for about 60 Martian days, 70 cm of ice sublimes, and 280 cm of ice melts. Ice melting also occurs with P = 50 mbar when $\theta = 35^{\circ}$.

Now let us consider the factors that may reduce the water vapor loss rates. At low obliquity, suppressing evaporation reduces the water loss rate, as can be seen in Table I by comparing the $\theta = 25^{\circ}$ cases with and without wind. At high obliquity, evaporation is the temperature limiting process. In order to balance the energy, approximately the same physical mass must be lost even if the evaporation rate is reduced. This occurs by raising the temperature, which in turn increases the incoming energy from atmospheric infrared radiation and hence causes greater mass loss.

The most effective way to limit the mass loss is to increase the thermal conductivity or the albedo. Table I shows that, at large obliquity, large water loss rates (several centimers/year) do not occur if the albedo and the conductivity are both high. Ice melting will not occur at $\theta = 35^\circ$, even at 50mbar pressure, if the conductivity is high.

At a moderate θ , such as 25°, the albedo and thermal conductivity also strongly control the water loss rate. The present south polar cap of Mars may have an albedo of >0.5 (Farmer *et al.*, 1979), even though the northern cap has an albedo of 0.4. Table I shows that the southern cap with an albedo of 0.6 would lose less water than the northern cap with an albedo of 0.4, as is observed. In addition, Table I shows that changing the thermal conductivity in the range from 1×10^{-3} to 8×10^{-3} alters the water loss by a very large factor.

Comparing the hot- and cold-pole loss rates in Table I with the loss rates for changes in various parameters shows that the water loss rate dependence on eccentricity is no stronger than its dependence upon obliquity, thermal conductivity, or albedo. Hence asymmetries between the permanent polar caps could be caused equally well by differences in albedo, thermal conductivity, or position in the precessional cycle, and the differences may well be frozen in from an earlier phase of the obliquity cycle.

Figure 13 and Table I do not show water loss rates from the polar regions for obliquities less than 25° . The loss rates are likely to decline radically as θ decreases because the polar ice albedo and conductivity will rise at low obliquity as dust storm frequency drops because of declining CO₂ pressure. More importantly, if permanent CO₂ polar caps form, the water loss rate will drop to an extremely small value. For example, at a temperature of 150°K the water vapor loss rate is about 10⁻⁶ g cm⁻² summer⁻¹ with a wind speed of 10 m/sec at 5 mbar of pressure. Ingersoll (1974) and Leighton and Murray (1966) previously pointed out the great importance of permanent CO₂ deposits as traps for water vapor.

(c) Water Ice Deposition Rates at the Poles and Exchange between Polar and Nonpolar Ice Reservoirs

Water ice not only sublimes from the summer poles, but it also is transported toward the winter pole. Some ice may reach the winter permanent pole to replace ice lost during the previous summer, other ice may be deposited in the annual CO_2 polar cap, and yet more ice may exchange with mineral surfaces and permafrost.

The available evidence does not allow us to estimate the net gain or loss rates for the current permanent polar caps. The only numerical data on water return to the poles are those of Farmer et al. (1979) who observed that the globally integrated atmospheric water vapor declined during the northern hemisphere winter by an amount equivalent to 4×10^{-3} g cm⁻² spread poleward at 60°N. The latitude was chosen to agree with the boundary of the seasonal north polar cap. Pollack et al. (1977, 1979) noted that dust particles probably serve as the carriers of water into the polar regions and that the mass of ice per dust particle is about the same as the mass of dust, suggesting a water vapor deposition rate of about 2×10^{-2} g cm⁻² year⁻¹ over the area of the north annual polar cap. The number is approximately equal to the deposition each year of all the water vapor in a hemisphere containing 10 precipitable micrometers of water. Hence this value is close to

an upper limit for the possible current ice deposition rate.

These water vapor return rates are less by 1 to 2 orders of magnitude than the water loss rates shown by Fig. 13 for $\theta = 25^{\circ}$. The implication is that both Martian polar caps are currently being destroyed or (Table I) that the ice conductivity is close to pure ice for both caps and additionally that the south cap has an albedo near 0.6. The highest temperature reached in the south cap calculation for A = 0.6 and $K = 8.6 \times$ 10^{-3} cal⁻¹ g⁻¹ °K⁻¹ was 193°K, which is much higher than the CO₂ frost point on present-day Mars.

There is no reason to believe that polar caps on Mars are permanent features. At other times in Martian history the caps may have been larger, smaller, or even nonexistent. A similar physical situation exists on Earth, and the evidence shows that the Earth's permanent polar caps were much larger 2×10^5 years ago and were absent 4×10^7 years ago. Large ice sheets have probably existed on Earth for a total of less than 10% of the Earth's geologic history (Pollack, 1979).

Since Mars has no oceans, its polar caps can undergo long-term alterations only by exchanging H₂O with the regolith. The Martian permanent polar caps occupy 1% of the planet. If the regolith were 10% H₂O by volume, then only 10 to 100 m of globally distributed regolith is needed to accommodate all the ice in polar caps 100 to 1000 m thick. Carr and Schaber (1977) suggest that the ice content of some Martian soils may be as high as 50%. If permafrost with 50% ice content spreads from 45 to 40° latitude, then the ice in a 100-m ice cap could be stored in only 30 m of soil. The available data (Farmer et al., 1977, Farmer and Doms, 1979) do not show any obvious indication of a seasonal regolith source or sink of water, but it is likely that water is currently stored reversibly in the regolith as permafrost, adsorbed water, and water of hydration. We shall consider each of these reservoirs in turn.

Clark (1978) shows that montmorillonite clays strongly adsorb H₂O and that sulfates, such as MgSO₄, undergo rapid hydration at temperatures near 210°K with present Martian water vapor abundances. A soil dominated by clays, but also containing about 10% magnesium sulfate, would contain 30, 15, or 5% of its dry weight as water for temperatures of 195, 215, and 235°K, respectively, for a column water content of 10 pr μ m of uniformly mixed water. At 215° about half this H₂O would be attached to the sulfates. By contrast, a basalt regolith at 215°K along with a water pressure of 10 pr μ m would only adsorb about 0.3% of its weight as H₂O (Fanale and Cannon, 1974).

The storage of H_2O in clays and sulfates is likely to be a strong function of latitude, depth, and season. Because of the strong temperature dependence of the adsorption and because of hysteresis effects, the diurnal thermal wave probably permanently dessicates the equatorial and midlatitude soil within a few centimeters of the surface, while the upper meter of soil in the polar regions could exchange water because of the seasonal thermal wave (see Clark, 1978).

Figure 14 illustrates the adsorbed water on smectite clays for various temperatures and pressures (Anderson et al., 1978). For an average water vapor amount of 10 pr μ m, soil colder than 195°K will be in vapor equilibrium with permafrost rather than with adsorbed water. On the other hand, soil only a few degrees warmer than 195° will be strongly desorbed. In polar regions, where substantial adsorption occurs, the major exchange will be between permafrost and atmospheric water vapor as long as permafrost is present. Of course, the contact between the permafrost and atmosphere will be indirect because even in the polar regions the annual thermal wave will seasonally remove the permafrost in the upper soil layers. At slightly lower latitudes, just beyond the permafrost margin, substantial clay regolith adsorption will oc-







FIG. 14. The mass of water adsorbed per gram of clay is a very strong function of the water pressure and of the temperature. Adsorption will precede permafrost formation, but when enough water is available to fill completely the adsorption sites so as to make permafrost, then permafrost formation will be the dominant physical process occurring. Because of the large quantity of water probably available, permafrost formation is likely to be more important for Martian climate than adsorption. The precipitable water column abundance was found by assuming that the water was uniformly mixed.

cur. At still lower latitudes only a few percent of the regolith could be adsorbed H₂O. The low-latitude (high-temperature) adsorption is a weak function of temperature but a modest function of pressure. For example, if the H₂O pressure fell from 10 to 1 pr μ m, soil at 205° would lose about 2% of its weight in water. In this way a 10²-m thick regolith covering half the planet might lose about 4 × 10² g cm⁻² of H₂O, which is sufficient to add about 200 m of ice to the poles.

Permafrost has been considered by several authors. Carr and Schaber (1977) discussed geologic evidence for permafrost and suggested that permafrost was widespread with H₂O composing 50% of the soil volume in some regions. Fanale (1976) and Farmer and Doms (1979) have shown that permafrost could be in equilibrium with the present atmospheric water content at latitudes above 40°. However, Smoluchowski (1968) has pointed out that only a few tens of meters of fine-grained regolith are sufficient to prevent permafrost from subliming from equatorial regions over the age of Mars.

The edge of the equilibrium permafrost boundary depends upon temperature and atmospheric water content. Presently, ice is in equilibrium at a temperature near 200°K. A factor-of-10 change in water vapor amount would change the equilibrium temperature boundary by about 10 to 15°K, which is equivalent to a latitude shift of about 10° for lower water amounts and 20° for higher water amounts. Likewise, for fixed H₂O abundance the boundary may shift if the mean temperature shifts. At 35° obliquity the permafrost should extend 5° further equatorward than it does now. At 15° obliquity the permafrost should extend about 3° further poleward than at present.

In the above discussion we have suggested that substantial amounts of permafrost and adsorbed water could move into and out of the soil as P_w and T vary. We have already indicated that the temperature wave can penetrate 100 m in only 10⁴ years. We must also consider the diffusion of H₂O into and out of the soil. This problem has all the complexity of the CO_2 diffusion problem discussed earlier.

The H₂O diffusion problem differs from that of CO₂ in several ways. First, relative to the atmospheric abundance much more H_2O is adsorbed on soil than is CO_2 . In solving the CO₂ problem we showed that, even if all the CO₂ in one pressure wavelength were desorbed, it would contribute only a small amount of CO₂ to the atmosphere and, therefore, we did not actually solve the diffusion problem. For H₂O, however, we must indeed solve the diffusion equation because of the large water abundance. A single molecule of H₂O can diffuse to a great depth based upon diffusion lengths such as those shown in Fig. 11. However, the gradient of the vapor density caused by diffusion greatly limits the actual mass flux. Second, because the H₂O in the atmosphere is a very tiny fraction of that adsorbed or present as permafrost, damping of pore diffusion by adsorption is very important and surface diffusion along the adsorbing surfaces could also be significant. Finally, the diffusion mean free path for H₂O depends on both the pore radius and the CO₂ pressure. For small pore sizes ($r < 1.0 \ \mu$ m) H₂O molecules collide mainly with pore walls in attempting to escape, while for larger pore sizes H₂O molecules collide with CO₂ molecules. Hence, H₂O diffusion may depend upon CO₂ pressure changes.

As a guide to the diffusion problem we can ignore the adsorption processes and calculate at time t the distance Z to the top of an evaporating permafrost layer by assuming that the diffusion occurs essentially in steady state. Then

$$Z = [2Dt((\rho_{\rm w}(Z) - \rho_{\rm w}(0))/\rho_{\rm ice}]^{1/2} \quad (12)$$

while the mass lost during t is $E\rho_{ice}Z$, where E is the porosity of the soil. This calculation, and the results, are essentially identical to those of Smoluchowski (1968) for the same temperature and diffusion coefficient.

Table II gives the values of Z for $t = 5 \times$

10⁴ years, for temperatures [which control the water vapor pressure $\rho_w(Z)$] from 200 to 200°K, for $\rho_w(0) = 0$, and for diffusion coefficients *D* between 0.01 and 10 cm² sec⁻¹. Diffusion coefficients *D* cannot be much larger than 10 cm² sec⁻¹ because that limit is set by the present CO₂ pressure. For pore sizes of a few microns or less *D* in centimeters squared per second is approximately numerically equal to the pore size in microns.

It is obvious from Table II that only a thin layer of permafrost can be lost or gained during a 5×10^4 -year period. There are two physical limitations expressed by these numbers. Most important, the mass in the vapor phase over the ice is very low. For example, at 210°K uniform mixing would allow only about 50 pr μ m of vapor phase H₂O above exposed ice. Diffusion greatly limits the amount of water in the column to a lesser value at any given time and also prevents rapid replacement of any vapor removed from the column. Suppose, for example, that exposed ice at 210°K is allowed to replenish by diffusion the column of air above it with H₂O 10 times during a year. Then during 5×10^4 years only 25 m of ice could be lost.

As we pointed out earlier, the ice-loss rates from exposed polar ice can be quite large at high obliquity but quite small at low obliquity. At low obliquity the atmospheric H_2O pressure will drop and temperatures at latitudes equatorward of 60° latitude will *increase*. Both factors will push both the edge of the equilibrium permafrost bound-

TABLE II

DEPTH OF PERMAFROST (CM) AFTE	r 5	X	104	YEARS
-------------------------------	-----	---	-----	-------

Diffusion coefficient $(cm^2 sec^{-1})$	Temperature (°K			
(em see)	200	210	220	
0.01	10	20	40	
0.1	35	65	130	
1	110	200	410	
10	340	650	1300	

ary and the edge of the regolith adsorption boundary poleward at low obliquity. Hence, the regolith will attempt to supply ice to the pole. The actual amount of ice supplied is very difficult to determine. Because of the great temperature dependence for the vapor pressure of ice, most of the supply of vapor will come from the present permafrost region whose annual mean temperature is about 200°K. Table II suggests that this region could lose about 0.5 m of ice over 5×10^4 years. If this permafrost region of Mars occupies 10% of the planet's area, and the permanent polar cap occupies 1%, then perhaps 5 m of ice would be added to the ice cap. It is conceivable that ten times as much as this could be added if the upper few meters of regolith had a very large pore size. In that case the low CO_2 pressure at low obliquity could greatly increase the H₂O diffusion rate. Another possibility is that geologic processes, such as slumping on slopes, could be important and might continually keep fresh permafrost exposed. It is unlikely, however, that winds would expose fresh permafrost because of the inability of winds at low atmospheric pressure to move dust. The seasonal thermal wave may also play a significant role in allowing permafrost to escape when the pressure is low. The region poleward of 60° latitude has an annual temperature of 175°K. At such a low temperatures only a few centimeters of permafrost could be desorbed. However, summertime temperatures would increase the temperature of the upper meter of soil so that diffusive losses would increase. If the upper meter of permafrost over onethird of the planet is lost, then 30 m of ice would be added to the caps.

At high obliquity, the poles will lose more water during the summer season than now, and temperatures equatorward of 55° latitude will drop. These processes will move the equilibrium permafrost boundary to lower latitudes. For example, if the mean H₂O amount increased to 60 pr μ m at 35°

obliquity, then the permfrost boundary could move to about 25° latitude. The vapor pressure above ice at about 210°K is 60 pr μ m, and about 20% of the planet's area is between 25 and 40° latitude. Table II suggests that under these circumstances the regolith might consume 20 m of ice from the caps. The amount of ice consumed could increase if higher water vapor abundances occur, or if the regolith has a very large pore size in its upper ten meters and the CO₂ pressure does not increase dramatically at high obliquity. Wind-blown dust might also speed the processes by hydrating in suspension and then becoming buried, or by simply covering exposed surface ice. High ice cap loss rates may force increased permafrost. If the polar cap loss rates exceed the gain rate, and if the permafrost cannot take up the difference, then the high atmospheric vapor pressures will occur. These high pressures will drive ice into the regolith and they will force the polar cap gain rate to increase until balance is achieved.

The reader may note that several different processes control the polar cap budget. If the two polar caps are out of balance with each other the feedbacks reinforce the imbalance by lowering the albedo of the cap losing net water. If both caps are out of balance with the atmosphere-regolith system, the feedbacks depend on the sign of the imbalance. At low obliquity, permanent CO₂ caps will have net H₂O gain rates, and the atmospheric water will be reduced until the ice cap gain balances the loss by the permafrost. At high obliquity, when large polar-ice-loss rates occur, the loss will continue if permafrost can adsorb the loss. If not, atmospheric water will build up, increasing the polar cap gain rate. A high water gain will increase polar cap albedos and decrease the ice loss rate. Hence at high obliquity the polar caps may balance their gain and loss rates so that the difference is equal to the regolith removal rate.

The movement of water vapor through adsorbing soil is difficult to calculate theo-

retically. Experimental studies of this process, using terrestrial analogs, could be quite valuable for improving our estimates of permafrost transfer rates.

(d) Formation of Ice Sheets on the Polar Laminae

The polar laminae appear to be dusty ice deposits extending slightly beyond the north polar cap to 80°N latitude and well beyond the south polar cap to 70°S (Cutts 1973). Pollack (1979) has proposed that the permanent polar cap, the polar laminae, and the polar debris mantle form a continuous sequence of geologic formations differing, according to their latitude, in their ability to retain water ice. Specifically, the debris mantels originated from atmospheric dust that was deposited in the annual CO₂ caps but was unable to retain any attached ice during the summer. This process was observed by Viking (Jones et al., 1979). The polar caps retain water ice for very long periods. The laminae are the intermediate case of dust sometimes able to retain

ice and sometimes not, so that the ice-todust mixing ratio varies with depth. Ice as permafrost is stable at the latitudes of the laminae, but the laminae are too thick to have formed by vapor diffusion of ice into soil. A dusty ice sheet covered with a meter of dust would be stable at the latitudes of the laminae because the dust would prevent high summertime surface temperatures from reaching the ice and causing it to sublime. Buried ice would be at the annual mean temperature and would be cold enough to be in equilibrium with atmosphere water vapor amounts or more likely to grow by picking up atmospheric water. Therefore the laminae ice would not diffuse out through a thin overlying dust layer. Here we attempt to identify the conditions that might have led to ice being permanently deposited in the area of the polar laminae before a dust layer covered it or before a thick, high-albedo ice sheet was present to stabilize the ice.

Table III presents water loss rates at 80 and 70° latitudes for an albedo of 0.25,

Fixed parameter	Variable parameter	Obliquity	Loss rate from hot pole	Loss rate from cool pole
)	P = 5 mbar	15	0.5	0.05
	lat 80°	20	5	0.7
4 = 0.25		25	15	5
$U = 10 \text{ m sec}^{-1}$	P = 1 mbar	15	0.5	0.03
$K = 1 \times 10^{-4} \text{ cal cm}^{-1} \text{ sec}^{-1} \text{ °K}^{-1}$	lat 80°	20	5	0.5
) // 10	P = 1 mbar	15	4	0.2
	lat 70°	20	10	0.7
$J = 10 \text{ m sec}^{-1}$				
$K = 1 \times 10^{-4} \text{ cal cm}^{-1} \text{ sec}^{-1} \text{ °K}^{-1}$	80° lat	20	2	0.1
P = 5 mbar $ A = 0.4$	70° lat	20	5	0.2
$U = 10 \text{ m sec}^{-1}$				
$K = 1 \times 10^{-3} \text{ cal cm}^{-1} \text{ sec}^{-1} \text{ °K}$	80° lat	20	0.2	0.005
P = 5 mbar $A = 0.4$	70° lat	20	1	0.01

 TABLE III

 LAMINAE WATER LOSS RATE (g cm⁻² year⁻¹)

typical soil conductivity, and for obliquities of 15, 20, and 25°. In performing these calculations we used the daily averaged solar insolation, even though there is a diurnal cycle at both 70 and 80°. It may be noted that the lower CO₂ pressures at $\theta =$ 15° play only a small role in the water budget by reducing the CO₂ greenhouse warming slightly.

These calculations suggest that, at very low obliquity, ice can begin to accumulate if the rate of supply is on the order of 0.1 to $0.5 \text{ g cm}^{-2} \text{ year}^{-1}$. This is a very high rate of supply, as we pointed out previously, and it suggests that the ice component of the laminae does not form simply as an extended thin dusty ice sheet.

It is most likely that the laminae originate at low obliquity when a large permanent polar cap of CO₂ forms from the present atmosphere and from CO₂ desorbed out of the regolith. Such a permanent cap of CO_2 would trap water released from permafrost reservoirs very effectively. The laminae may also form at moderate obliquity by growth of permanent ice sheets if one pole has much lower ice loss rates than the other. Then the preexisting permanent ice sheet at the low-loss-rate pole may grow at the expense of the other polar cap. Of course, alternatives exist. Perhaps wind deposits H₂O ice into small piles, or perhaps shadowing by surface relief allows some ice to remain each summer. These patches of ice might then be able to grow into a more extensive, stable ice sheet. Large-scale slopes could also play a role by preventing solar energy from directly reaching the ice. Once enough ice (probably a few centimeters) is present to raise the albedo of the ice well above that of the ground, the loss rate of water ice without the permanent CO₂ might be low enough to allow frost to accumulate.

Table III shows some typical loss rates from established polar ice sheets at 20° obliquity. With $K = 10^3$ cal cm⁻¹ sec⁻¹ °K⁻¹ and high albedo (0.4), the water ice loss rates at 80° for the cooler pole are very low. In the dust-free conditions at low obliquity, an even higher albedo may be anticipated.

At 70° latitude the calculated loss rates of ice are larger than those at 80°, but the differences are not always striking. Several additional factors not included in our calculations may contribute to the laminae being confined to high latitudes. Permanent CO₂ deposits are unlikely to have ever extended much beyond 80°, so large H₂O ice sheets at low latitudes could not have been established. In addition, the calculations we performed ignore the diurnal cycle. We have already shown (Figs. 3 and 6) that the daily average summer temperature is not a strong function of latitude at low obliquity, which is why the loss rates differ only slightly between 70 and 80°. Water loss rates are likely to be controlled even at 70° by diurnal temperature cycles. A much more complex model, including an atmospheric boundary layer, diurnal condensation and evaporation cycles, and diurnal transport out of the boundary layer, is needed to examine fully the water loss rates under these conditions. Finally, at low latitudes, exposed ice would not be stable at high obliquities, and the ice component of the low-latitude laminae might therefore have been eroded, whereas the CO₂ annual caps partly protect the laminae at higher latitudes.

The reader may note that there is no clear distinction in observed thickness, structure, or geographical extent between the present north polar cap and a layer of northern laminae with a thin cap of dust. Ice surrounding dust grains never seems to be stable on bare polar ground according to our calculations of ice loss rates. Rather, it seems more likely that the laminae are remnant, buried, permanent H_2O ice caps that formed at both poles during periods of low obliquity or at the colder pole during periods of moderate obliquity and high eccentricity.

(e) Summary

At low obliquity, CO₂ permanent caps

will effectively trap water vapor. Permanent H₂O ice caps will form partly from atmospheric water vapor, but mainly through water supplied from permafrost. The thickness of such caps could be tens of meters, but slow diffusion of water vapor through the upper several meters of regolith limits the cap thickness. At intermediate obliquities, water loss rates from the polar caps will be very large unless the conductivity of the polar ice is close to that of pure ice. At high eccentricity one polar cap will have a significantly larger water ice loss rate than the other if both caps have the same albedo. However, the cap with the higher insolation will probably have a higher albedo than the other cap because dust storms will preferentially occur during the winter season of the low insolation cap. The albedo difference equalizes the water loss rate, or perhaps causes the high insolation pole to lose less water than the low insolation pole. This situation probably occurs on present-day Mars. At high obliquity both poles experience large ice loss rates. moving the permafrost boundary equatorward. At high obliquity it is possible that a small amount of ice could melt in windsheltered regions, if the atmospheric pressure increased substantially. Ice is stable in the region of the polar laminae only when thick, high-albedo ice sheets are present. No obvious mechanism exists to incorporate ice into forming dusty laminae. It is most likely that the laminae are buried, remnant, permanent polar caps. Hence over long time periods the latitude boundary of the polar debris mantle marks the furthest equatorward boundary of the annual CO₂ caps and the boundary of the laminae marks the furthest equatorward boundary of the permanent water ice caps.

VI. THE RELATION BETWEEN ORBITAL VARIATIONS AND H₂O-CO₂-CLOUD GREENHOUSES

Numerous dessicated channels cover the surface of Mars. Their origin is unknown, but an early Martian climate, warmer and wetter than now, may have played a role in their formation. Pollack (1979), Sagan and Mullen (1972), and Sagan (1977) pointed out that reduced atmospheres composed of CH₄ and NH₃ could maintain high surface temperatures through a greenhouse effect with small (100 mbar) pressures of CH₄. Such atmospheres are consistent with many models of outgassing for Mars and Earth (Fanale, 1971), with theories of the origin of life on Earth, with ideas on the nature of the early terrestrial atmosphere, and with concepts about the maintenance of the early Earth's high surface temperatures despite low solar luminosity. The great disadvantage of reduced atmospheres is that they are unstable with respect to hydrogen escape following photodissociation and may therefore have very short lifetimes.

Another possibility is that high surface temperatures were maintained by a greenhouse effect in a CO₂-H₂O-cloud atmosphere (Pollack, 1979; Cess et al., 1980). Such an atmosphere has the advantages of having a relatively long lifetime and of being chemically identical to the presently observed atmosphere. Many investigators (e.g., Walker, 1976) believe that the Earth's early atmosphere were dominated by CO₂ rather than CH₄. Owen et al. (1979) show that the temperature of the early Earth could have been maintained above freezing despite lower solar luminosity by a plausible CO₉-H₉O atmosphere. Fanale (1971)pointed out that the outgassed atmosphere of Mars would be dominated by CO₂ and H₂ assuming the H₂O to be cold trapped. The formation of CH₄ occurred only if the escape of H_2 was not rapid compared with the outgassing rate. He adopted a catastrophic, accretional degassing model so that H₂ had a relatively long enough lifetime to convert CO₂ to CH₄. In 1971 it was believed that Mars was not geologically active, but now the observed prevalence of large volcanoes suggests continual geologic activity. Moreover, most evidence (Anders and Owen, 1977; Pollack and Black, 1979) suggests

that Mars is not fully degassed, which may imply a more gradual degassing epoch. The question of degassing history and the CO₂/CH₄ ratio needs to be reanalyzed based upon the data obtained since 1971. The disadvantage of a CO₂ atmosphere is that relatively large amounts of CO₂ are required to maintain high temperatures. Early estimates by Anders and Owen (1977) suggested that less than 500 mbar of CO₂ had been outgassed on Mars. However, more recent estimates (Pollack and Black, 1979) that consider revised theories of noble gas abundances on the terrestrial planets suggest that 1 to 3 bars of CO_2 may have been outgassed. These outgassing estimates make an early CO₂-H₂O greenhouse much more plausible.

Below we explore in greater detail the nature of CO_2-H_2O -cloud greenhouses. First, we discuss the maintenance of substantial quantities of atmospheric gases. Then we elucidate some difficulties in determining precisely the pressures needed for keeping high surface temperatures. It should be noted that the climate considered represents a substantial departure from that of present-day Mars. A complete analysis of such climates is beyond the scope of the present work.

(a) Maintenance of CO_2 and H_2O in the Atmosphere

The present atmospheric pressure on Mars is far too low to produce an effective greenhouse. High CO₂ pressures are required to produce appreciable greenhouse warming by the 15- μ m CO₂ bands and other weaker CO₂ bands, and also to pressure broaden H₂O bands. Since atmospheric H₂O is restricted by the atmospheric temperature, an increased CO₂ greenhouse effect is also needed to raise temperatures to allow more H₂O into the atmosphere. As we have shown, changes in the orbital elements of Mars are unable to drive more than a few tens of millibars of CO₂ out of the regolith and into the atmosphere-polar cap system. Hence to achieve significant

 CO_2 pressures and a significant CO_2-H_2O greenhouse, we must return to a period before the formation of an extensive clay-carbonate regolith.

The processes forming the Martian regolith are uncertain. Fanale (1976) argued that meteoroid bombardment produced a 2-km thick basaltic megaregolith. As pointed out earlier, basalt is not strongly adsorbing at high surface temperatures so, if the CO_2 atmosphere appeared before the formation of a basalt regolith, a high-surface-temperature climate could be stable. However, if the CO₂ outgassed slowly as the regolith formed, it would be incorporated into the regolith. Likewise, if at low obliquity the CO₂ pressure were reduced and temperatures fell, then CO₂ would be permanently lost to the regolith. Higher obliquity could not return the CO₂ to the cold atmosphere. Channel features are found only on heavily cratered terrain, so a scenario of early CO₂ outgassing, gradual regolith production by meteoroid impacts, carbonate formation, and loss of an atmosphere is consistent with the timing of the end of channel formation.

Another plausible scenario that might lead to high surface pressures is that meteoroid bombardment produced a basaltic regolith only 100 m thick which could not adsorb all the available CO2 as it was outgassed. However, geochemical processes operating at high surface temperatures may have slowly produced more adsorbing clays from the basalts. Some CO₂ may have been converted into carbonates, and clays as well as basalt may have been transported to polar regions where low temperatures increased adsorption. The outgassing rate, the rate of formation of clays and carbonates, and the rate of transport of regolith to the poles would have determined the longevity of the high-temperature state. These geochemical processes may cause any hightemperature climate to be self-destructive since the rate of carbonate and clay formation is greatly enhanced by high temperatures and by the availability of liquid water. Future studies of the rate of loss of atmospheric pressure due to carbonate and clay formation would be quite interesting. F. Fanale (private communication, 1980) suggests that a high-CO₂-pressure, high-temperature climate might not last more than 10^6 years. On Earth, volcanism associated with continental drift recycles clays and carbonates into CO₂ and other volatiles in the atmosphere, but continental drift has not occurred on Mars.

During the period before the loss of CO_2 to geologic reservoirs, it is necessary to maintain the CO_2 in the atmosphere, and this requires high polar temperatures. As we shall discuss in the next section, a 0.5to 2-bar CO_2 atmospheric pressure can maintain global average temperatures above 273°K. It is likely that heat transport by such an atmosphere, in a manner analogous to that on the Earth, could maintain polar temperatures above 230° K. Hence CO₂ once established at 0.5- to 2-bars pressure would not tend to form permanent polar caps since CO₂ would be far below its vapor pressure at polar temperatures.

The process stabilizing the atmosphere in the above example is atmospheric heat transport to the poles. Gierasch and Toon (1973) examined the importance of atmospheric heat transport to the stability of the atmospheric pressure on Mars against changes in the polar heat budget. Their conclusions are illustrated in Fig. 15 where the solid lines represent solutions of a simplified polar heat balance equation for various levels of annual average available



FIG. 15. Because of the strong dependence of atmospheric heat transport on atmospheric pressure, the polar temperature increases rapidly with atmospheric pressure. The various solid lines illustrate this dependence for different amounts of solar energy at the pole. The dashed curves are the vapor pressure relations for CO₂ and for CO₂-H₂O clathrate. Single intersections of solid and dashed curves represent stable climates. When multiple intersections occur, both the higher- and lower-pressure intersections are stable. As the solar energy available at the pole increases, more CO₂ is added to the atmosphere. Because of feedback, the pressure is very sensitive to solar insolation. ϵ and β are variable parameters entering the heat balance equation. (See Gierasch and Toon, 1973.)

solar insolation, S. The heat balance equation determines the temperature of the polar cap consistent with the atmospheric heat transport, which in turn is controlled by the pressure of the atmosphere. An additional constraint when permanent CO_2 polar caps are present is that the atmospheric pressure is controlled by the polar cap temperature through the vapor pressure relation. This relation is illustrated in Fig. 15 by dashed lines for CO₂ and CO₂-H₂O clathrate. Stable climatic states are represented by intersections of the dashed and solid lines. For example, the S curve labeled 1.8×10^4 ergs cm⁻² sec⁻¹ intersects the CO₂ vapor pressure curve at a polar temperature of about 144°K and an atmospheric pressure of 6 mbar. If Mars had permanent CO₂ caps with albedos of 0.77, then S would now have this value and the model would represent present-day Mars. If the obliquity increased to 35° , then S would increase to about 2.4 (for fixed albedo) and the only stable climate would have a pressure above 1 bar.

The climate illustrated by Fig. 15 has multiple solutions. As a result of the multiple solutions, if there are permanent CO_2 polar caps, then, once initiated, a highpressure climate is stable until the solar energy input to the pole drops well below the solar input needed to initiate the hightemperature state. Gierasch and Toon (1973) did not include the greenhouse warming caused by increased CO_2 , so the high-pressure climate is even more stable than Fig. 15 suggests. Of course, the production of clays and carbonates would eventually destroy the high-pressure climate.

Another very likely possibility is that CO_2-H_2O clathrate might dominate the polar cap (Dobrovolskis and Ingersoll, 1975; Miller and Smythe, 1970). Clathrate is thermodynamically favored over CO_2 at Martian polar temperatures and, since H_2O ice is brought into the caps as finely divided grains in good contact with the atmosphere, it is likely that clathrate formation would readily occur. Figure 15 indicates that CO_2 clathrate greatly limits the Martian atmospheric pressure, and, for the range of values of S shown, atmospheric pressures greater than 20 mbar are difficult to obtain. It is possible that increased greenhouse warming would increase the slope of the thermal balance solution (the solid lines in Fig. 15) and thereby allow much higher pressures. The value of S may also be much larger than the values indicated in Fig. 15.

Figure 15 shows that a high-pressure climate must either be initiated by, or maintained by, an increase in available solar insolation. Figure 2 indicates that obliquity variations are capable of considerably modulating the polar heat budget. The solar insolation is also very sensitive to the polar cap albedo. Normalizing to a CO_2 albedo of 0.6 and on obliquity of 25°, we find

$$S(A, \theta)/S(0.6, 25^{\circ}) = 5.91(1 - A) \sin \theta.$$
 (13)

This ratio is 0.6, 1.35, and 1.67 for obliquities of 15, 35, and 45°, respectively, with A fixed at 0.6. For an obliquity of 25° it takes values of 1.25 and 0.75 for polar cap albedos of 0.5 and 0.7, respectively. With A = 0.6 and $\theta = 25^\circ$, S = 3.1.

Hence, if the CO₂ albedo is now 0.6, then Fig. 15 suggests that if CO₂ were available, a runaway condition could occur at present. If, earlier in Martian history, the solar luminosity were 0.7 of its present value, S at $\theta = 25^{\circ}$ would be 2.0, even for an albedo of 0.6.

The polar energy would be greatly increased by lowering the albedo or increasing the obliquity. Ward et al. (1979) suggested that during its early history Mars may have had a mean obliquity near 32° with excursions to 45 and 25°. Such an obliquity would make a high-pressure climate easier to achieve and maintain. For example, with $\theta = 45^{\circ}$, A = 0.5, and current luminosity, S could reach a value of 6×10^4 ergs cm^{-2} sec⁻¹. Hence we can conclude that reasonable modifications of the polar albedo obliquity and are probably

sufficient to initiate a high-pressure climate for either CO_2 or its clathrate.

(b) CO_2 -H₂O-Cloud Greenhouse

Pollack (1979) computed surface temperatures for various CO₂ pressures, relative humidities, and tropospheric lapse rates, and he found that global average temperatures above 273°K occurred when the CO₂ surface pressure was about 2 bars. Cess et al. (1980) performed similar calculations but they concluded that a pressure of only 1 bar would produce a temperature above 273°K. Both pressure estimates greatly exceed early predictions of the Martian CO₂ inventory (Anders and Owen, 1977). However, new predictions of outgassed CO₂ exceed the pressures required to obtain a surface temperature greater than 273°K by either of these radiative calculations (Pollack and Black, 1979). The dependence of surface temperature upon surface pressure increases rapidly and becomes highly nonlinear as the surface temperature approaches 273°K. Hence predictions of the pressure needed to obtain a surface temperature of 273° K are quite sensitive to the details of the radiative model. The major differences between the models of Pollack (1979) and Cess *et al.* (1980) are that the latter authors employ a greater pressurebroadening coefficient than Pollack and that the details of the treatment of the CO₂ bands are slightly different. Here we shall point out that clouds and a possible low surface albedo might have been important to climate early in Martian history. Very low surface pressures are found to produce high surface temperatures.

Table IV presents radiative-convective calculations of surface temperatures for various pressures of CO_2 , a planetary albedo of 0.25, and for various values of an insolation factor, f, which is the ratio of the available solar energy to that presently available with globally averaged conditions.

Most of the cloud-free values given in Table IV were taken from the calculations of Pollack (1979), who assumed a fixedtemperature profile with a lapse rate, Γ , of 3.5° K km⁻¹ and a relative humidity $R_{\rm h}$ of

	Insolation factor, f	Insolation Surface pressure (mbar)					ibar)	
		6	100	300	600	1000	1500	
No cloud								
Global average	1.0	217 (218)	228	233	(248; $\Gamma = 3.5$) (250; $\Gamma = 5$)	252 (259; $\Gamma = 3.5$) (263; $\Gamma = 5.0$) (254; $R_{\rm h} = 0.5$)	(272)	
Global average with $A = 0.1$ or equator with A = 0.25	1.25	232	244	255	$(284; \Gamma = 5)$			
Equator with 1 A = 0.1 2	1.5	242	258	273				
	2.0	265	295					
Cloud								
Global average	1.0 1.25		260	298	285			
Equator with $A = 0.1$	1.5		280					

TABLE IV Results of Greenhouse Calculations

unity. He obtained the surface temperature by balancing the outgoing thermal flux against the available solar energy. Pollack showed that his results were not very sensitive to the lapse rate, nor the relative humidity.

Also presented in parentheses in Table IV are new results for cloud-free conditions. These calculations were performed with a radiative-convective scheme that converges to a radiative equilibrium thermal profile in regions where the profile is subadiabatic. The solar energy is assumed to be deposited only at the surface. The calculations also include a pressure broadening coefficient appropriate for CO₂ that Pollack (1979) inadvertently overlooked. However, the difference in the surface temperatures calculated with the two methods is only a few degrees except at high pressures. The new calculations show that only 1.5 bars of CO₂ are needed to obtain a 273°K mean temperature.

The basic conclusion to be drawn from the numbers in Table IV is that CO_2 pressures near 1.5 bars are required under cloud-free conditions in this model to achieve global average surface temperatures above the freezing point of water. The temperatures are not very sensitive to relative humidity nor to lapse rate, as is apparent from the 1-bar values in Table IV.

Table IV also suggests several means by which lower pressures might yield high temperatures. For example, the present albedo of Mars is quite high, $A \sim 0.25$, because of the presence of abundant iron oxides. As we have already pointed out large atmospheric pressures could only be achieved before extensive regolith modification by geochemical processes. At this stage in Martian history, iron oxides may not have been formed. Kieffer et al. (1977) report that Mars presently has a bimodal surface albedo distribution, suggesting that the Martian surface is composed of one material with an albedo near 0.1 and another, much brighter material. Hence, it is reasonable to assume that the

planetary albedo was near 0.1 in early Martian history. The insolation factor for this albedo is 1.2, and CO₂ pressures near 600 mbar could then yield global temperatures above freezing!

Global average temperatures of 273° K may not be the relevant criteria for fluvial features. For example, if one only desires equatorial average temperatures of 273° K, then for an obliquity of 25° , f = 1.22 on a yearly mean and P = 600 mbar is needed. If the albedo is 0.1 and only high equatorial temperatures are needed, then an f of about 1.5 and 300 mbar of CO₂ will suffice. At high surface pressures, atmospheric heat transport will lower equatorial temperatures and raise polar ones. Hence precise estimates of the CO₂ pressures needed to achieve high equatorial temperatures are difficult to make.

The calculations tabulated in Table IV consider only the greenhouse warming due to CO_2 and H_2O . Perhaps other trace gases that are effective at producing a greenhouse might have been present in the early Martian atmosphere. For example, Oyama *et al.* (1979) suggest that SO_2 may be important for the Venus greenhouse. We plan to consider further the importance of trace gases in future work.

Another possibility for enhanced greenhouse effects is that high clouds may form. The advantage of clouds is that pressure broadening is not needed to make them efficient infrared radiators. The disadvantage of clouds as greenhouse constituents is that they reflect solar insolation. Whether clouds warm or cool depends upon their optical depth in the visible and infrared, their area, their altitude, and their composition (Schneider, 1972).

Several cloud materials might contribute to a greenhouse. Dust now dominates the thermal balance on Mars. However, it is unlikely that dust could form more extensive clouds than at present because solar energy deposition in the dust stabilizes the atmosphere and stops the dust lifting processes. CO_2 ice has very narrow infrared bands and would not be an efficient greenhouse material. Our radiative-convective calculations confirm the speculation of Gierasch and Toon (1973) that a high pressure CO_2 thermal profile has its upper atmosphere at temperatures below the CO_2 frost point. This situation occurs in calculations for present-day Mars, which is corrected by appealing to efficient large-scale vertical heat transport through dynamics (Pollack *et al.*, 1979). A similar process, and perhaps solar energy deposition in the gas as well, is needed to prevent the formation of CO_2 clouds over most of Mars at high pressures.

Below we consider only water ice clouds because they have greater infrared emission than CO₂ clouds and, unlike dust, water cloud particles do not have to be lifted from the surface. Table IV presents a series of cloud greenhouse calculations performed with a radiative-convective 1-D model. As we have noted, the effect of clouds on the radiation budget depends upon various cloud properties. Rather than attempt to explore the entire range of possibilities, we have studied only those clouds that should be among the most effective at increasing the surface temperature. This deliberate bias is intended to determine the maximum additional warming that clouds might plausibly contribute, and is not meant to represent the most likely case. Hence we chose a water cloud fractional coverage of 75%, a particle size distribution with a mean size of 4 μ m, a cloud visible optical depth of about 5, and a cloud top pressure of 10% of the surface pressure. We also used a relative humidity of unity and an adiabatic lapse rate of 5°K km⁻¹.

If the clouds are effective as greenhouse warmers, then using as large an area of coverage as possible enhances their impact. However, it did not seem reasonable to expect 100% cloud cover since the water vapor sources would be localized to the equatorial and summer polar regions and restricted to a small fraction of the planet's surface since no oceans are anticipated. Particles of sizes near 4 μ m do not have substantial fall velocities, so they could remain in the atmosphere for long periods of time. Particles of such size have large visible asymmetry factors, so the clouds have relatively low visible albedos, but the clouds' drops are large enough so that the clouds effectively block infrared radiation. A large visible optical depth insures that the clouds will have infrared optical depths that are large (1 to 3). Finally, optically thick clouds radiate effectively at the ambient temperature. The higher the ambient temperature, the more energy they radiate to space and the less effective they are at creating a greenhouse. Therefore, high-altitude clouds are the best greenhouse candidates, but once they reach an altitude where the temperature varies slowly with height, further increases in altitude do not substantially improve the ability of the clouds to create a greenhouse. We chose a cloud top pressure altitude of 10% of the surface pressure since it was the lowest altitude at which the thermal profile became nearly constant. Typical terrestrial radiative-convective models place clouds at 0.4 to 0.5 of the surface pressure.

We assumed in our cloud calculations that all the available solar energy was deposited at the ground. The solar energy was reduced over 75% of the planet by reflection at the cloud top surface calculated with a two-stream approximation for a conservatively scattering cloud (Irvine, 1975), including multiple ground reflections.

For particles of the size of interest the asymmetry factor, g, is about 0.85. With ground albedo A = 0.25 and cloud optical depth of 5, the planetary albedo is about 0.43. This case represents f = 1. If the surface albedo were 0.1, the planetary albedo would be 0.35 and f = 1.15.

Table IV shows that favorably placed clouds may raise surface temperatures over a cloud-free case by about 20 to 40° K. Global average temperatures above 273° K then require less than 600 mbar of pressure. If the surface albedo is 0.1, a global surface

temperature of 273°K occurs at pressures near 300 mbar. A high equatorial temperature occurs near 300 mbar for an albedo of 0.25 and, if the surface albedo is low, then equatorial temperatures can be 273°K for pressures near 100 mbar.

(c) Summary

New estimates of the CO₂ outgassed over Martian geologic history suggest that more than enough CO₂ is available to produce a significant greenhouse effect (Pollack and Black, 1979). A high-obliquity state might be required to initiate a high-pressure climate, but once initiated, greenhouse warming and atmospheric heat transport might prevent polar cap formation even at low obliquity. The atmosphere would be destroyed as carbonates and adsorbing clays form and as the regolith is transported to polar regions. If CO₂ had been slowly outgassed after the formation of a thick, cold, adsorbing regolith, the atmospheric pressure could never have built up. New greenhouse calculations by Cess et al. (1980) suggest that only 1 bar of CO₂ is needed to achieve global average temperatures above 273°K with the current Mars albedo and solar luminosity. We studied a less sensitive radiation model that predicts that 1.5 bars are needed to achieve 273°K. We found with this less sensitive model that if the surface albedo of Mars were lower in its early history, or if equatorial conditions are more relevant than global averages ones, or if extensive high clouds form, then only 500 mbar of CO₂ are needed to obtain temperatures of 273°K. A great many details of a high-pressure CO₂ climate need to be considered further. In particular, the upper atmosphere needs to be warmed to prevent CO₂ frost formation. Atmospheric heat transport to polar regions needs to be taken into account, for it may significantly depress equatorial temperatures and raise polar ones.

VII. DISCUSSION

In the preceding sections we considered

variations in the CO_2 , dust, and H_2O cycles caused by changes in the orbit and obliquity on Mars. We also discussed the implications of enhanced CO_2 pressures for greenhouse warming. Here we would like to unify the ideas developed by describing how the climate might differ between the present orbital configuration and various alternatives. Then we will mention the observational evidence for climate change and finally propose models for the formation of the polar laminae and of the fluvial features.

At present, the Martian obliquity is about 25°, but periodically in the recent past the obliquity has reached values less than 15° (Ward, 1979). As the obliquity decreases, the poles receive less solar energy and, if, the polar cap albedo is not too low, as seems likely, permanent CO₂ polar caps will form. The formation of permanent polar caps will reduce the atmospheric pressure to values of 1 mbar or less. The precise value of the pressure depends on the albedo; higher albedos will cause lower pressures. The reduced atmospheric CO₂ pressure will strongly desorb CO₂ from the regolith augmenting the permanent CO₂ caps with the limit being set by the ability of CO₂ to diffuse from the regolith and also by the amount of CO₂ presently adsorbed on the regolith. The regolith could contribute a several-hundred-meter thickness of CO₂ over the area of the present caps, while the present atmospheric pressure corresponds to only about 10 m.

Permanent CO₂ caps will be strong water vapor sinks and thus atmospheric water vapor will decline at low obliquity. In addition, equatorial and midlatitude annual mean temperatures will be several degrees higher than current values. Lessening amount of atmospheric water vapor and increased midlatitude temperatures will force the equilibrium permafrost boundary to a higher latitude than now. Water will be supplied to the permanent polar caps, but the water supply to the caps will be strongly limited by diffusion so that during a half obliquity cycle of 5×10^4 years a few

meters of H₂O might be added over the area of the present permanent polar caps. Other processes assisting diffusion, such as slumping that exposes fresh ice, or the presence of large pore sizes or cracks near the surface, could substantially augment the H₂O supplied to the poles. The seasonal thermal wave might drive high latitude permafrost to the poles, thus adding a few tens of meters of ice to the caps. If permanent CO₂ polar caps cover a larger geographic area than the present permanent H_2O caps, or if local slopes and other thermal balance factors are favorable, extensive water ice sheets may develop over the area now covered by the polar laminae. If these sheets reach great enough depths to have high albedos and high thermal conductivities they will be stable at higher obliquities than those at which they formed.

As the CO₂ pressure drops, dust-storm activity will cease, because at low pressure only unreasonable wind speeds, in excess of the speed of sound, would be able to lift dust. Since no dust will be available, both the CO₂ and H₂O caps would have much higher albedos than the present H₂O and CO₂ caps. Hence the decline of dust storms strongly augments the trend of the CO₂ and H₂O cycles.

The atmospheric circulation at low obliquity will be quite different than at present. Since there will be no dust storms, the vertical thermal structure will be under greater dynamic control. Feedback between tidal winds and local dust storms will not occur. Large temperature gradients, perhaps similar to those at the present southern summer pole, will drive violent winds at the edge of the summer permanent polar caps. Larger meridional winds will be caused by the necessity of supplying the mass of the annual CO₂ polar caps with a less massive atmosphere. The radiative time constant will be less than at present and the ground temperature will be under even greater radiative control than it is now.

The orbital eccentricity and longitude of perihelion will probably play no significant role in determining the basic climate at low obliquity.

Ward (1979) shows that periodically the Martian obliquity may reach values as high as 35° . Ward *et al.* (1979) suggest that the obliquity could have had a mean value of 32° and reached an extreme near 45° prior to the formation of the Tharsis ridge. As the obliquity increases the poles receive more solar energy so that temperatures of the ground and ice will increase.

Polar ground temperatures during summer can exceed 260°K at $\theta = 35^{\circ}$ when e = 0.14. However, at $\theta = 35^{\circ}$, annual average polar temperatures will exceed present temperatures by only about 10°K whereas equatorward of 60° latitude annual average temperatures will be several degrees cooler than at present.

The small changes in annual average temperature will slightly decrease adsorption in the polar regolith and increase adsorption in the equatorial regolith. A globally distributed basalt regolith would desorb enough CO_2 to provide an increase of only a few millibars of pressure. Larger increases might occur if the regolith were concentrated in polar regions or if the regolith were composed of clays. A clay regolith could potentially release several tens of millibars of CO_2 to the atmosphere. Such small pressure increases are significant for dust storm activity, but not for greenhouse warming.

High temperatures for water ice can occur at large obliquity. In wind-sheltered regions, at fairly high atmospheric pressures (50 mbar), low-albedo water ice could melt at $\theta = 35^{\circ}$. However, it is more likely that thermal conduction and evaporative heat losses will prevent water ice from reaching the melting point. The water vapor loss rate from the polar cap is a strong function of temperature and therefore of albedo and thermal conductivity; nevertheless, it is likely that large water loss rates will occur. Dusty water ice would be highly unstable in the polar laminae area. Atmospheric water vapor amounts will increase, and since equatorial temperatures will decline at high obliquity, the equilibrium permafrost boundary will move to lower latitudes. The polar water ice will attempt to transfer to the regolith, but regolith diffusion will limit the amount of new permafrost formed. Only a few tens of meters of polar ice might find its way into the regolith unless atmospheric water vapor builds up to high levels or soil pore sizes are very large near the Martian surface. The polar caps will disappear at high obliquity if permafrost formation is rapid enough.

The dust storm cycle will be strongly modified at high obliquity if a few tens of millibars of CO₂ are desorbed from regolith reservoirs. In that case wind speeds needed to lift dust will decline significantly. Dust storms will occur during the summers in both hemispheres. However, it is unlikely that dust storms will be more prevalent or intense than they are today during southern hemisphere summer. Atmospheric heating due to solar energy deposition in the dust will stabilize the atmosphere and will prevent substantially more dust storm activity than at present. Since dust storms will occur during both summers, the rate of supply of dust to the two polar regions will be more nearly comparable than is the case now. Because more water will be mobilized in summer and then deposited with the dust in winter than now, it is likely that both permanent polar caps, if they exist, will have a slightly higher albedo than the present northern H₂O polar cap but a lower albedo than the present southern H₂O polar cap.

The atmospheric circulation at high obliquity will be characterized by strong summertime easterlies because summer polar temperatures are higher than equatorial ones. Unlike the present Martian summer, which is relatively free of baroclinic disturbances, a high-obliquity summer may have frontal activity. If the atmospheric pressure increases, the radiative time constant will increase proportionately so that tidal and topographic winds will be damped.

The orbital eccentricity and the longitude of perihelion may be less important at high obliquity than it is now. The dust-storm cycle will be unaffected by the eccentricity (if atmospheric pressures increase). However, the water balance between the poles may be controlled by the eccentric orbit. If permanent water ice polar caps are found at high obliquity, they will be much more likely to transfer between the poles with the precession of the equinoxes than they are now. Then the pole with summer containing perihelion will have the smaller polar cap.

Ward (1974, 1979) and Murray et al. (1973) show that the eccentricity of Mars varies from 0.0 to 0.14 and that the longitude of perihelion slowly varies. The eccentricity barely affects the annual average temperatures. Therefore, permafrost boundaries (now located near $\pm 40^{\circ}$ latitude) and regolith adsorption processes cannot be affected substantially by changes in e or ω . Eccentricity and ω do not significantly affect the climate at very low θ , but they might make significant changes in the water cycle at very large θ . Ward (1979) shows that during a modest fraction of Martian history, θ oscillations about the mean are fairly small. During these periods, variations in e and θ play important roles by creating hemispherical asymmetries.

When the Martian orbit is circular, the hemispheres have comparable energy budgets. It is likely that global scale dust storms do not occur during such periods at modest θ , but our knowledge of Martian dust-storm generation is not sufficient to determine whether dust storms occur at e =0. If dust storms do not occur, then the albedos of both polar caps should be large during such periods, while if dust storms do occur, both poles would have albedos like that of the present northern water ice cap. Detailed stratigraphy of the polar caps may some day answer this question.

At large eccentricities, hemispherical

asymmetry is accentuated, especially when perihelion occurs near a solstice, as at present. Then global dust storms will occur preferentially during the summer in which perihelion occurs, the albedo of that summer cap will be higher than the albedo of the other cap, and the water ice cap in that hemisphere will be smaller than the other because of larger water loss rates. This simple picture can be upset by feedbacks between the dust cycle and the water and CO₂ cycles. For example, the present smaller southern water ice cap, found at the pole receiving greater summer insolation as expected, lost less water vapor than the larger northern cap during the year of Viking observations. Presumably this contradiction is caused by a recent change in the heat balance caused by dust storms creating a lower ice albedo at the northern pole.

In order to verify or improve upon these ideas concerning climate changes due to orbital element forcing, we need the following basic knowledge: the permanent polar cap albedo in the southern hemisphere; the thermal conductivity of the permanent polar caps and polar laminae; the geographical distribution, depth, pore size, and composition of the regolith; the amount of CO_2 and H₂O adsorbed on the regolith; the amount of CO_2 in the southern permanent polar cap; and the thickness of the water ice in the polar caps. Experimental studies of the diffusion of CO₂ and H₂O through porous, adsorbing soils would be quite useful. We also need better theoretical models of the mechanisms generating dust storms, of regolith desorption, of the transport for H₂O and dust, and of the polar heat budget. This information would improve our understanding of the processes causing climatic change.

Evidence for climatic change is also required. It is important to note that variations in eccentricity and longitude of perihelion will be recorded as hemispherical asymmetries, while changes in the obliquity cycle will be recorded symmetrically. Unfortunately, since both cycles occur, it will not be easy to separate symmetric and asymmetric variations without precise dating.

The most obvious evidence that the orbit affects the climate is that global dust storms now occur only during southern hemisphere summer. This asymmetry most likely is due to higher winds occurring when solar insolation is maximized near perihelion. As perihelion precesses, the time of year when dust storms occur will change.

A second suggestion of the importance of eccentricity and perihelion passage, long used as evidence for strong orbital insolation control over the water budget, is that the area of the southern polar cap is much less than that of the northern polar cap. Unfortunately, there are several unresolved difficulties. The observed water budget of Farmer et al. (1979) showed less water loss from the south pole than the northern one, and James et al. (1979) suggest that the south pole might have a residual CO₂ cap. These observations imply that the orbit does not control the polar heat budget directly through the insolation. Probably the dust-storm asymmetry has created different albedos at the north and south poles so that orbital insolation variations control the heat budget indirectly through the dust storms. Another possibility is that the difference in the water loss rates between the caps is so small at present obliquities that only a small fraction of the polar cap mass can be transferred between the caps during a precessional cycle. In that case the present areal coverage of the two caps is a fossil of a higher obliquity state. A final possibility is that geological factors such as regional slope differences may control the polar heat budget. The importance of regional topography is strongly suggested by the fact that neither polar cap is symmetric about the rotational axis. Detailed models of the polar cap heat balance are necessary to determine the importance of topography to the heat balance.

Another asymmetric geological feature, possibly related to climate, is the vast (10^5 km^2) , low-albedo, dune field partially encircling the north pole in the region poleward of 80° latitude which does not contain laminae or permanent ice. Cutts *et al.* (1976) find that the dunes have a complex relation to the polar ice, sometimes overlying the ice and sometimes underlying it, suggesting that the dunes and ice are interactive.

It is possible that the north polar dunes are the remnant of some easily erodable geologic structure found only in the north polar region and therefore have no climatic significance. However, the mechanism of supply of dust and water to the polar cap region can naturally explain them. Individual dust grains entering polar latitudes become coated with H₂O ice and eventually fall out onto the annual caps when CO₂ condenses on the particles. During the following summer, at most latitudes, the dust grains are freed when the CO₂ and ice sublime away, and a general polar debris mantle, rather than dunes, forms. However, the dust which falls on the permanent polar caps does not completely lose its coat of water ice, and so is trapped on the polar cap. The dunes are found within the same latitude belt as the permanent cap, but because of their lower albedo and presumably lower thermal conductivity, the dunes are less able to retain water ice than the polar caps, although they are better able to keep water ice than the lower latitude dust deposits. The small amount of ice left around the dust grains at least until summer serves to cement many of them together into large saltating particles. The single-scattering albedo of a large particle is lower than that of a small particle composed of the same material. Therefore, dunes composed of clumps of fine dust particles could have lower spherical albedo. Under this scenario the dunes would not occur in the southern hemisphere because ice is less stable, due to the higher summertime solar insolation. Haberle et al. (1979) also pointed out that an ice cement could form saltating grains from dust. They suggested that the greater dust supply to the north pole favored dunes there. While we agree that the dust supply is significant, we feel that the greater longevity of the ice at northern latitudes is also of great importance.

Although the southern polar laminae are larger in area than the northern ones, the northern laminae are thicker than the southern ones. By mass the laminae are symmetric about the equator. The structure of the laminae has been described in detail by Cutts et al. (1973, 1976, 1979). They are layers of dust and possibly ice exposed in places as fine terraces of approximately equal thickness (≈ 30 m) that extend uniformly over hundreds of kilometers (Fig. 1). The scarps along which the terraces are exposed were originally thought to be edges of a series of stacked plates, each composed of many terraces. More recent observations (Cutts et al., 1979) indicate that the scarps, which spiral about the poles are primarily the edges of troughs cut into the laminae, although isolated escarpments also exist. Cutts et al. (1979) have also identified an undulating character to the flat regions separating the troughs. The nearly complete absence of craters in the laminae suggests that they are fairly youthful or that they are continually being resurfaced.

We require mechanisms for creating three properties of the laminae. First, it is necessary to explain why dust preferentially collects in the polar areas. Second, we must elucidate the formation of the terraces. Finally, it is necessary to account for the troughs and large-scale escarpments. All three of these observations can be explained through the use of a model having a variable dust-to-ice mixing ratio. The variation is driven by the orbital element changes that alter the polar heat budget and the dust storm cycle.

Both poles are probably permanent net dust sinks because dust storms do not preferentially remove dust from the polar regions, and because particles from dust storms are preferentially deposited in the polar regions. Global dust storms now originate in the southern hemisphere and spread northward. Hence the dust is not preferentially obtained from the north polar region. We do not know the source regions of dust in the southern hemisphere, but there is no reason to believe that the pole is a preferred source region. Dust and ice are codeposited as small particles in the north annual cap and to a lesser extent in the south annual cap because CO_2 snow formation is a strong atmospheric clearing mechanism (Pollack *et al.*, 1979).

Cutts et al. (1973, 1979) suggested that dust collects at the poles because erosion is prevented in the polar regions by water ice cementing the grains. Such a process is probably significant in the permanent cap region. However, extensive debris mantles surrounding both poles to fairly low latitudes (35°), but not covering equatorial regions (Soderblom, et al., 1973), suggest that ice formation is not the dominant process keeping dust in the polar regions. We are inclined to believe that the poles are a net dust sink because of enhanced deposition there by CO₂ snow, since global dust storms do not originate there, and because the polar dust grains cannot be easily lifted by the wind. Water ice formation about dust particles increases their sizes by only small amounts. Winds are probably able to suspend such ice-dust particles nearly as easily as single dust particles. However, dust particles of such size are not directly lifted by the wind but are injected by larger saltating particles. Such large particles may be generally absent in the polar regions since they are not carried there by winds and this missing ingredient suppresses wind erosion.

No observational evidence is available to strongly limit theories concerning the reasons that the polar ice and dust form a terraced structure in the laminae. Evidently erosion-resistive soil occurs at regular depths, but whether the erosion-resistive segment is caused by chemical cementing

grains, variations in particle size, of changes in particle composition, or some other property is not known. A reasonable hypothesis is that the terracing is caused by variations in the dust-to-ice mixing ratio through the laminae. The more resistive flats are primarily compacted dust, while the more easily eroded segments are largely water ice with a minor component of unconsolidated dust. When the terrace edges are exposed to the atmosphere the water ice sublimes and the surrounding ground loses its strength and therefore slumps. Strong evidence for this mechanism is that debris mantles are not terraced at lower latitudes, where surface ice is probably never stable. The cause of the varying dustto-ice ratio may partly be changes in the relative deposition rate of dust and ice as suggested by Pollack (1979). In addition, and perhaps more importantly, water ice can be preferentially removed as water vapor sublimes from a shrinking polar cap. As water vapor sublimes, dust imbedded in the ice will become concentrated until eventually a thin layer of nearly pure dust caps the ice so as to prevent further ice loss.

There are two types of changes that will create variations in the dust-to-ice mixing ratio: changes in the longitude of perihelion, and changes in the obliquity. For an eccentric orbit one pole is favored to receive both dust and polar ice. For example, the north polar cap of Mars now covers most of the laminae in the northern hemisphere, and dust storms lead to high dust deposition rates there. As perihelion precesses, the north cap will begin to lose water to the south cap. As the north cap shrinks, the dust at the top of the cap will become concentrated into a layer. Eventually, the dust layer will become thick enough to protect the underlying ice from the seasonal temperature cycle. If the present dust-to-ice mixing ratio is 0.15 (as implied from the present polar albedo), then to obtain a 1.5-m-thick layer of dust, 8.5 m of ice must be lost. This is about all the ice that could be lost from the pole unless the upper dust layer is heavily eroded. Hence one layer will be formed during half a precessional cycle at one pole, and one layer will be formed during the other half cycle at the other pole. The strongest evidence that such a mechanism occurs is that the north polar cap now covers most of the northern laminae.

One problem with this model is that large amounts of ice are being lost from the permanent polar cap system and transferred to the laminae. In order to account for the present H₂O cap, the continued formation of laminae, and their nearly uniform thickness, one must assume that the rate of supply of water to the larger polar cap from nonpolar cap reservoirs is very constant in time and is equal to the rate of loss of water caused by burial in the laminae. As we shall discuss, the formation of scarps in the laminae may supply the needed H₂O. An added problem is that the water loss rate must be substantial for the present obliquity so that water can be driven between the poles. The permanent polar cap thickness must be moderately close to the laminae thickness plus 10 m, or about 40 m.

The second terracing mechanism is obliquity oscillations. At low obliquities duststorm activity will cease, and H₂O ice sheets will form within permanent CO₂ caps over a large polar area from water supplied by permafrost. Once formed, these sheets will be stable as the obliquity increases. At moderate obliquities dust storms will occur so that both ice and dust will be deposited. At larger obliquities any exposed ice will be driven off the permanent caps and back into regolith reservoirs. A layer of dust will cap the underlying ice and stablize it against the seasonal temperature wave. Permafrost reservoirs must be able to supply the difference between the permanent cap water and that lost to the laminae during each obliquity cycle. This loss of water to the laminae need not imply a huge permafrost source of water. The formation of scarps in the laminae may also recycle the needed water. This mechanism does not require large polar water loss rates at the present obliquity. It does, however, imply that a permanent polar-cap thickness on the order of 40 m is needed to form a 30-m layer of laminae, since at high obliquity only a single 10-m layer of ice needs to be lost to create a dust cap from dust embedded in the ice.

In reality, both eccentricity and obliquity variations may be forming laminated terrain. Hence a complex record of formation may eventually be found in the laminae structure.

A problem with forming the terraces in either of these ways is that one 30-m layer is created approximately every 10^5 years. The supply of water and dust to the north cap as estimated by Pollack *et al.* (1979) is sufficient to build a 30-m layer during 10^5 years. However, the total thickness of the laminae is only a few kilometers (Dzurisin and Blasius, 1975). Hence only 10^7 years is represented in the laminae if each layer is conserved and represents 10^5 years. Although the lack of craters suggets a possible youthful age for the laminae, it is difficult to imagine why they should have begun to be formed only 10^7 years ago.

One possibility is that the laminae are not conserved. Cutts et al. (1976) note that the laminae crater age can be explained by a crater filling rate of 1 km/10⁶ years. Such a time is consistent with the time for orbital motions to create 1 km of laminae. The most plausible mechanism for destroying 1 km of laminae every 10⁶ years is horizontal propagation of the scarps that incise the laminae. As we shall discuss, and as Howard (1978) has pointed out, laminae composed mainly of water ice could propagate. The troughs are spaced at intervals of about 50 km and do have depths on the order of 1 km. The laminae will be obliterated as they form if the troughs migrate at the modest rate of 5 cm year⁻¹ and if the dust released is removed gradually by winds. This process may be the major one regulating the thickness of the individual laminae.

Another mechanism that could increase the estimated age of the laminae is plastic thinning due to outward flow caused by the overburden load, followed by sublimation at the edges as suggested for low-latitude permafrost features by Squyres (1978). If the laminae are 90% ice, simply removing the ice at the bottom but retaining the dust extends the age of the laminae to 10^8 years. A reasonable hiatus in dust storm and ice deposition activity does not suffice to extend greatly the age of the laminae. For example, large eccentricities only occur during about half of Martian history; if dust storms only occur at large eccentricity, then the age of the laminae could be increased by a mere factor of 2.

The final property of the laminae to explain, and the one that has attracted most attention, is the deep troughs that dissect the laminae. Earlier these troughs were thought to be the edges of sets of young laminae stacked uniformly on top of older laminae. Such a structure suggested a doubly periodic constructional mechanism to Murray *et al.* (1973). The present concept of the structure of the laminae does not suggest a series of steps of progressively younger age.

More recent work has revolved around the idea that both erosional processes and constructional processes play an important role (Howard, 1978; Cutts et al., 1979). Howard assumed (incorrectly) that the laminae were part of a stepped topography. He thought that the steps were all of the same age and believed that the higher insolation on the equatorward-facing slopes preferentially removed ice from the slopes and led to its deposition on the high-albedo flat areas separating the slopes. Winds carried away dust from the slopes leading to a poleward migration and formation of steep faces. Howard favored a model in which the layers forming the terraces of the laminae are being formed from the material composing the slopes as they erode. Hence he believed that the laminae in the north are forming under present conditions, while

those in the south, where there are not albedo differences between flats and scarps, are not forming at present. Why the slopes are terraced is not clear in Howard's model. An advantage of Howard's hypothesis is that the laminae continually recycle material and do not require water to be supplied from the regolith. Howard thought that the laminae are composed mainly of water ice.

Cutts et al. (1979) agreed with Howard that insolation control is important in forming the slopes, but they believed that the laminae are mostly dust and that the ice serves mainly as a dust trap. Hence, Cutts et al. have no mechanism to extend the age of the laminae by recycling material. According to their model the equatorwardfacing slopes of the troughs do not retain ice because they receive higher insolation than flats. Since the slopes do not retain ice, they do not retain dust either. Cutts et al. pointed out the existence of an undulating character for the flats separating the troughs. They thought that these undulations are caused by a gradually advancing or receding permanent polar cap and a cyclic dust deposition cycle that causes the dust retention region to gradually shift in a wavelike manner. Such a mechanism implies different time scales or phases for dust cycles and water cycles.

We propose a model of the laminae that is closely related to Howard's (1978) model. As we pointed out, we believe that approximately 40-m-thick layers composed of ice and dust are deposited during obliguity and/or eccentricity cycles as the remnants of the permanent polar caps of these eras. As the obliquity and eccentricity change, the ice is forced off the polar caps. The dust at the upper surface is concentrated until eventually a thin layer of insulating dust covers the dusty ice and prevents further water loss through the upper surface. Hence, we believe that orbitally forced insolation variations are responsible for the terraced structure. However, we believe, like Howard, that regional slope differences create insolation differences unconnected with orbital variations. These regional slopes cause water ice to be unstable on equatorward-facing slopes. Hence these are defrosted zones. Since the laminae are largely water ice, the terraced structure is exposed and eroded along these slopes. The slopes propagate poleward and the propagation eventually leads to the observed defrosted bands of rather uniform horizontal structure. Water ice is unstable on these slopes at all eccentricities and most obliquities, so they cut through numerous terraces.

If equatorward-facing slopes are unstable, it follows that poleward-facing slopes are more stable than flat terrain. Hence in addition to equatorward-facing slopes propagating poleward by erosion, one also expects poleward-facing slopes moving poleward by deposition. We propose that such propagating slopes form the alternate sides of the troughs There is no reason for the deposition and erosion processes to proceed at identical rates. Hence one may see isolated escarpments in various regions. The undulating terrain may represent such propagating, isolated features. Conceivably, the interaction of these structures may lead to growth or decay of their amplitude.

This model has several advantages. The terraces are formed by orbital variations. In Howard's (1978) simplest model the terraces were formed concurrently with the troughs. Therefore, the south pole presented a problem because it now has no high-albedo flats that could grow. The escarpments and troughs in our model are formed through insolation controlled by slopes. We believe that both poleward and equatorward slopes will propagate respectively by deposition and erosion. Hence troughs can form, as well as undulating terrain. Unlike Cutts et al. (1979), we feel that the laminae are composed mainly of water ice so that the scarps can move. Propagation allows the development of long uniform scarps from original small nonuniformities such as craters, as pointed out by

Howard (1978). Propagation releases ice at the laminae edges and recycles the polar laminae, both filling and destroying craters and reducing the necessity for continual resupply of the laminae with permafrost water ice. This recycling allows the laminae to be much older than 10^7 years, despite the fact that one 30-m layer forms every 10^5 years.

The final features thought to suggest Martian climate change are the channels. There are many different types of these features, implying different origins. (See Sharp and Malin, 1975). Some channels, mostly found in chaotic terrain, seem to have been formed during a single episode following the catastrophic collapse of a volatile-rich terrain. Although the morphology of other features, such as channels on crater rims and channel networks covering mountain systems, suggests rainfall, it does not require rainfall. Many geologists believe that the channels were fed primarily from ground water sources. A much better knowledge of the requirements for the formation of the channels, such as duration of flow, number of repeated episodes of flow. water source, and whether or not the flow was capped by ice, is needed before any reliable climatological inferences can be drawn.

The geographic distribution of the channels is also not well known. Sagan et al. (1973) presented data showing that the channels are concentrated in the equatorial regions, but the distribution may represent observational bias rather than the true distribution. Sagan et al. (1973) pointed out that, if channels are more abundant in the equatorial regions, a climate-controlled water origin would be favored. Pieri (1976) noted that small Martian channels (which are the most numerous channels and are typically less than 150 km long and 1 to 10 km wide) are restricted to low-albedo, highly cratered terrains. The former restriction is possibly observational since in highalbedo terrains the lack of contrast makes it more difficult to see the channels, and since high-albedo terrains tend to be covered

with a debris mantle that may hide the stream beds.

The restriction of the small channels to highly cratered terrain may be a significant clue to their age. Most geologists believe that the small channels, which are the most suggestive of a climatic origin, are older than the larger channels that appear to have been formed catastrophically. This belief is based on crater counts (Pieri, 1976; Sharp and Malin, 1975). The age of both channel types, according to these workers, seems to be billions of years. Channels are absent in younger geologic unit of the planet, such as most volcanic areas. An opposite point of view was expressed by Masursky et al. (1977), who asserted that cratering counts show that channels have formed during all periods from 3×10^9 until 5×10^8 years ago.

Many Martian channels probably form as the result of geological processes, such as volcanic activity or other geothermal events which melt subsurface ice. The time history of such channels depends on the geological thermal history of Mars. Most Martian channels do not have any obvious geothermal region at their source. Since these channels are geographically widespread, a global heat source is suggested to supply liquid H_2O . A plausible heat source is increased surface temperatures caused by climatic warming.

One possible cause of climatic warming is that the solar luminosity may have been higher at some past time. Although the possibility cannot be excluded, the current theory of solar evolution does not suggest that the solar luminosity was ever substantially higher in the past. In fact, the opposite is most likely. In addition, a solarluminosity increase large enough to raise Martian temperatures substantially would also lead to very much higher terrestrial temperatures, possibly high enough to cause a runaway greenhouse on Earth (Pollack, 1979).

The most likely cause of warmer climate is an early, more massive atmosphere capable of producing a substantial greenhouse. Pollack (1979) has pointed out that only modest total pressures in CH_4 , NH_3 atmospheres are needed to raise global average temperatures above freezing. Unfortunately, the photochemical stability of such atmospheres is unknown. If one assumes that during the climatic epoch in question the temperature of the top 100 m of previously frozen soil must increase substantially, then the atmosphere must maintain high temperatures for 10^4 years, the thermal penetration time.

A massive CO₂ atmosphere has several advantages over a reduced one. CO₂ is presently observed in great abundance on Mars, Venus, and Earth. Some theories of the evolution of the atmospheres of the terrestrial planets (Walker, 1976) suggest that early CO₂-dominated atmospheres, containing very little hydrogen, are more likely than highly reduced, hydrogen-rich atmospheres. A CO₂ atmosphere has a long lifetime, being eventually destroyed by geochemical processes that form carbonates and a clay regolith. About 1 bar of CO₂ is needed to yield global average Martian surface temperatures above freezing with present solar luminosity and Martian albedo (Cess et al., 1980). However, we have shown that much lower CO₂ pressure—as low as 100 mbar in the extreme, but 500 mbar being more reasonable—could yield sizable greenhouse effects. In particular, if the Martian surface albedo were lower prior to the formation of abundant iron oxides, then only about 500 mbar of CO₂ would be needed to yield a global average temperature at the freezing point of water. Abundant high H₂O clouds could also contribute to a greenhouse, and, furthermore, it is not clear the global average temperatures are the relevant ones for creating channels. Equatorial temperatures will be substantially higher than global average ones. Another likely possibility is that other trace gases, such as SO₂, may augment the CO₂ greenhouse.

The present polar temperatures on Mars are so cold that if the abundance of CO_2 were increased it would freeze out at the

poles or enter the regolith. Ammonia would also freeze out at present-day temperatures. Variations in the Martian orbit and, in particular, higher obliquities at an early period (before the formation of carbonates or an adsorbing regolith) may have been an important trigger in moving frozen volatiles into the atmosphere so that a substantial greenhouse could occur. It is not certain whether or not a greenhouse, once initiated, would be stable at low obliquity. Certainly large climate changes, even during a greenhouse epoch, would occur because of obliquity and orbital variations. The suggestion by Ward et al. (1979) that the early Martian obliquity may have always been as large, and usually larger than, the present one may be quite important. However, our knowledge of the dynamics of a climate so substantially different from the present Martian climate is too scanty to determine clearly the significance of changes in orbit and obliquity.

Our understanding of the possible causes of climate change on Mars is rapidly increasing as we learn more about the dynamics of the present Martian climate. However, a great many uncertainties persist as to the nature of the climate changes that have actually occurred. Our current stratigraphy is very crude, so the timing and duration of most events are poorly known. In order to understand the formation mechanisms of the channels and of the polar laminae we must know more about these structures, such as the water-to-ice ratio in the laminae, the nature of the physical change across terraces in the laminae, and the relation of channels to their fluid reservoirs. We must rely on the future exploration of Mars to resolve many of these uncertainties.

ACKNOWLEDGMENTS

We greatly appreciate discussions with C. Leovy, A. Ingersoll, A. Seiff, J. Cuzzi, and F. Fanale, and preprints from J. Cutts, F. Fanale, and B. Farmer. The work by Toon was partly supported by a NASA joint research interchange with Cornell University (NCA2-OR175-702).

REFERENCES

- ANDERS, E., AND OWEN, T. (1977). Mars and Earth: Origin and abundance of volatiles. *Science* 198, 453– 465.
- ANDERSON, D. M., SCHWARZ, M. J., AND RICE, A. R. (1978). Water vapor adsorption by sodium montmorillonite at -5°C. *Icarus* 34, 638-644.
- BARRER, R. M. (1967). Surface and volume flow in porous media. In *The Solid–Gas Interface 2* (E. A. Flood, Ed.), Vol. 2, pp. 557–609. Dekker, New York.
- BERGER, A. L. (1976). Obliquity and precession for the last 5,000,000 years. Astron. Astrophys. 51, 127-135.
- BIEMANN, K., ORO, J., TOULMIN, P., III, ORGEL, L.
 E., NIER, A. O., ANDERSON, D. M., SIMMONDS, P.
 G., FLORY, D., DIAZ, A. V., RUSHNECK, D. R.,
 BILLER, J. B., AND LAFLEUR, A. G. (1977). The search for organic substances and inorganic volatile compounds in the surface of Mars. J. Geophys. Res.
 82, 4641-4658.
- BRIGGS, G. A. (1974). The nature of the residual Martian polar caps. *Icarus* 23, 167–191.
- BRUNAUER, S., COPELAND, L. E., AND KANTRO, D. L. (1967). The Langmuir and BET Theories. In *The Solid-Gas Interface* (E. A. Flood, Ed.), Vol. 1, pp. 77-105. Dekker, New York.
- CARR, M. H., AND SCHABER, G. G. (1977). Martian permafrost features. J. Geophys. Res. 82, 4039– 4054.
- CESS, R. D. (1974). Radiative transfer due to atmospheric water vapor. Global considerations of the Earth's energy balance. J. Quant. Spectrosc. Radiat. Transfer 14, 861-871.
- CESS, R. D., RAMANATHAN, V., AND OWEN, T. (1980). The Martian paleoclimate and enhanced atmospheric carbon dioxide. *Icarus* 41, 159-165.
- CLARK, B. C., III (1978). Implication of abundant hygroscopic minerals in the Martian regolith. *Icarus* 34, 645–665.
- CLARK, B. C.., III, BAIRD, A. K., ROSE, H. J., JR., TOULMIN, P., III, CHRISTIAN, R., KELLIHER, W. C., CASTRO, A. J., ROWE, C. D., KEIL, K., AND HUSS, G. R. (1977). The Viking X-ray fluorescence experiment, analytical methods and early results. J. Geophys. Res. 82, 4577-4594.
- CLARK, S. P., JR. (1966). Handbook of Physical Constants, Geological Society of America Memoirs, Vol. 97, p. 478. Geol. Soc. Amer., New York.
- CUTTS, J. A. (1973). Nature and origin of layered deposits of the Martian polar regions. J. Geophys. Res. 78, 4231-4249.
- CUTTS, J. A., BLASIUS, K. R., CARR, M. H., GREELEY, R., AND MASURSKY, H. (1976). North polar regions of Mars: Imaging results from Viking 2. Science 194, 1329–1337.
- CUTTS, J. A., BLASIUS, K. R., AND ROBERTS, W. J.

(1979). Evolution of Martian polar landscapes: Interplay of long-term variations in perennial ice cover and dust storm intensity. J. Geophys. Res. 84, 2975– 2994.

- CUZZI, J. N., AND MUHLEMAN, D. O. (1972). The microwave spectrum and nature of the subsurface of Mars. *Icarus* 17, 548-560.
- DAVIES, D. W., FARMER, C. B., AND LAPORTE, D. D. (1977). Behavior of volatiles in Mars' polar areas: A model incorporating new experimental data. J. Geophys. Res. 82, 3815-3822.
- DAVIS, B. W. (1969). Some speculations on adsorption and desorption of CO_2 in Martian bright areas. *Icarus* 11, 155–158.
- DOBROVOLSKIS, A., AND INGERSOLL, A. P. (1975). Carbon dioxide-water clathrate as a reservoir of CO₂ on Mars. *Icarus* 26, 353-357.
- DZURISIN, D., AND BLASIUS, K. R. (1975). Topography of the polar layered deposits of Mars. J. *Geophys. Res.* 80, 3286-3306.
- DZURISIN, D., AND INGERSOLL, A. P. (1975). Seasonal buffering of atmospheric pressure on Mars. *Icarus* 26, 437-440.
- FANALE, F. P. (1971). History of Martian volatiles: Implications for organic synthesis. *Icarus* 15, 279– 303.
- FANALE, F. P. (1976). Martian volatiles: Their degassing history and geochemical fate. *Icarus* 28, 179– 202.
- FANALE, F. P., AND CANNON, W. A. (1971). Adsorption on the Martian regolith. *Nature* 230, 502-504.
- FANALE, F. P., AND CANNON, W. A. (1974). Exchange of adsorbed H_2O and CO_2 between the regolith and atmosphere of Mars caused by changes in the surface insolation. J. Geophys. Res. 79, 3397–3402.
- FANALE, F. P., AND CANNON, W. A. (1978). Mars: The role of the regolith in determining atmospheric pressure and response to insolation changes. J. Geophys. Res. 83, 2321-2325.
- FANALE, F. P., AND CANNON, W. A. (1979). Mars: CO₂ Adsorption and capillary condensation on clays—significance for volatile storage and atmospheric history. J. Geophys. Res. 84, 8404– 8414.
- FARMER, C. B., DAVIES, D. W., HOLLAND, A. L., LAPORTE, D. D., AND DOMS, P. E. (1977). Mars: Water vapor observations from the Viking orbiters. J. Geophys. Res. 82, 4225-4248.
- FARMER, C. B., AND DOMS, P. E. (1979). Global seasonal variation of water vapor on Mars and the implications for permafrost. J. Geophys. Res. 84, 2881-2888.
- GIERASCH, P. J., AND GOODY, R. M. (1968). A study of the thermal and dynamical structure of the Martian lower atmosphere. *Planetary Space Sci.* 16, 615-646.
- GIERASCH, P. J., AND TOON, O. B. (1973). Atmo-

spheric pressure variation and the climate of Mars. J. Atmos. Sci. 30, 1502-1508.

- GOODING, J. L. (1978). Chemical weathering on Mars. *Icarus* 33, 483-513.
- HABERLE, R. M., LEOVY, C. B., AND POLLACK, J. B. (1979). A numerical model of the Martian polar cap winds. *Icarus* 39, 151–183.
- HAYS, J. D., IMBRIE, J., AND SHACKLETON, N. J. (1976). Variations in the Earth's orbit: Pacemaker of the ice ages. *Science* 194, 1121-1132.
- HESS, S. L., HENRY, R. M., LEOVY, C. B., RYAN, J. A., AND TILLMAN, J. E. (1977). Meteorological results from the surface of Mars: Viking 1 and 2. J. Geophys. Res. 82, 4559–4574.
- HESS, S. L., HENRY, R. M., AND TILLMAN, J. E. (1979). The seasonal variation of atmospheric pressure on Mars as affected by the south polar cap. J. Geophys. Res. 84, 2923-2927.
- HOBBS, P. V. (1974). *Ice Physics*, p. 359. Oxford Univ. Press (Clarendon), London/New York.
- HOWARD, A. D. (1978). Origin of the stepped topography of the Martian poles. *Icarus* 34, 581-599.
- HUMMEL, J. R., AND RECK, R. (1979). A global surface albedo model. J. Appl. Meteor. 18, 239-253.
- HUNT, G. E. (1973). Radiative properties of terrestrial clouds at visible and infra-red thermal window wavelengths. *Quart. J. Roy. Meteor. Soc.* 99, 346– 369.
- INGERSOLL, A. P. (1970). Occurrence of liquid water. *Science* 168, 972–973.
- INGERSOLL, A. P. (1974). Mars: The case against permanent CO₂ frost caps. J. Geophys. Res. 79, 3403-3410.
- IRVINE, W. M. (1975). Multiple scattering in planetary atmospheres. *Icarus* 25, 175–204.
- JAMES, P. B., BRIGGS, G., BARNES, J., AND SPRUCK, A. (1979). Seasonal recession of Mars' South Polar cap as seen by Viking. J. Geophys. Res. 84, 2889– 2922.
- JONES, K. L., ARVIDSON, R. E., GUINNESS, I. A., BRAGG, S. L., WALL. S. D., CARLSTON, C. E., AND PICK, D. G. (1979). One Mars year: Viking Lander imaging observations of sediment transport and H₂O-condensates. Science 204, 799-806.
- KIEFFER, H. H., CHASE, S. C., JR., MARTIN, T. Z., MINER, E. D., AND PALLUCONI, F. D. (1976). Martian north pole summer temperatures: Dirty water ice. Science 194, 1341–1344.
- KIEFFER, H. H., MARTIN, T. Z., PETERFREUND, A. R., JAKOSKY, B. M., MINER, E. D., AND PALLU-CONI, F. D. (1977). Thermal and albedo mapping of Mars during the Viking primary mission. J. Geophys. Res. 82, 4249-4291.
- LEIGHTON, R. R., AND MURRAY, B. C. (1966). Behavior of carbon dioxide and other volatiles on Mars. *Science* 153, 136-144.
- MASON, B. J. (1976). Towards the undertanding and prediction of climate variations. *Quart. J. Roy. Meteor. Soc.* 102, 473-498.

- MASURSKY, H., BOYCE, J. M., DIAL, A. L. SCHABER, G. G., AND STROBELL, M. E. (1977). Classification and time of formation of Martian channels based on Viking data. J. Geophys. Res. 82, 4016–4038.
- MAYKUT, G. A., AND UNTERSTEINER, N. (1971). Some results from a time dependent thermodynamic model of sea ice. J. Geophys. Res. 76, 1550-1575.
- MCELROY, M. B., KONG, T. Y., AND YUNG, Y. L. (1977). Photochemistry and evolution of Mars' atmosphere: A Viking perspective. J. Geophys. Res. 82, 4379-4388.
- MILANKOVITCH, M. (1930). Mathematische Klemalehie und astronomische theorie der klemaschqankungen. *Handbuch der Klimatologie* (W. Koppen and R. Geiger, Eds.), Vol. I, Teil A. Berlin.
- MILLER, S. L., AND SMYTHE, W. D. (1970). Carbon dioxide clathrate in the Martian ice cap. Science 170, 531-533.
- MURRAY, B. C., AND MALIN, M. C. (1973a). Polar volatiles on Mars—Theory versus observations. *Science* 182, 437–442.
- MURRAY, B. C., WARD, W. R., AND YEUNG, S. C. (1973b). Periodic insolation variations on Mars. *Science* 180, 638–640.
- OWEN, T., BIEMANN, K., RUSHNECK, D. R., BILLER, J. B., HOWARTH, D. W., AND LAFLEUR, A. L. (1977). The composition of the atmosphere at the surface at Mars. J. Geophys. Res. 82, 4635– 4639.
- OWEN, T., CESS, R., AND RAMANATHAN, V. (1979). Infrared CO₂ greenhouse to compensate for reduced solar luminosity on early Earth. *Nature* 277, 640– 642.
- OYAMA, V. I., CARLE, G. C., WOELLER, F., AND POLLACK, J. B. (1979). Venus lower atmosphere composition: Analysis by gas chromatography. *Science* 203, 802–805.
- PETERFREUND, A. R., AND KIEFFER, H. H. (1979). Thermal infrared properties of the Martian atmosphere. 3. Local dust clouds. J. Geophys. Res. 84, 2853-2863.
- PIERI, D. (1976). Martian channels. Distribution of small channels on the Martian surface. *Icarus* 27, 25-50.
- POLLACK, J. B. (1979). Climatic change on the terrestrial planets. *Icarus* 37, 479-553.
- POLLACK, J. B., AND BLACK, D. C. (1979). Implications of the gas compositional measurements of Pioneer Venus for the origin of planetary atmospheres. Science 205, 56-59.
- POLLACK, J. B., COLBURN, D. S., FLASAR, F. M., KAHN, R., CARLSTON, C. E., AND PIDEK, D. (1979). Properties and effects of dust particles suspended in the Martian atmosphere. J. Geophys. Res. 84, 2929-2945.
- POLLACK, J. B., COLBURN, D., KAHN, R., HUNTER, J., VAN CAMP, W., CARLSTON, C. E., AND WOLF, M. R. (1977). Properties of aerosols in the Martian

atmosphere, as inferred from Viking lander imaging data. J. Geophys. Res. 82, 4479-4496.

- POLLACK, J. B., HABERLE, R., GREELEY, R., AND IVERSEN, J. (1976). Estimates of wind speeds required for particle motion on Mars. *Icarus* 29, 395– 417.
- POLLARD, D. (1978). An investigation of the astronomical theory of the ice ages using a simple climate-ice sheet model. *Nature* 272, 233-235.
- SAGAN, C. (1977). Reducing greenhouses and the temperature history of Earth and Mars. *Nature* 269, 224–226.
- SAGAN, C., AND MULLEN G. (1972). Earth and Mars: Evolution of atmospheres and surface temperatures. Science 177, 52-56.
- SAGAN, C., TOON, O. B., AND GIERASCH, P. J. (1973). Climatic change on Mars. *Science* 181, 1045–1049.
- SCHNEIDER, S. H. (1972). Cloudiness as a global climatic feedback mechanism: The effects on the radiation balance and surface temperature of variations in cloudiness. J. Atmos. Sci. 29, 1413-1422.
- SHARP, R. P., AND MALIN, M. C. (1975). Channels on Mars. Geol. Soc. Amer. Bull. 86, 593–609.
- SHORTHILL, R. W., MOORE, H. J., HULTON, R. B., SCOTT, R. F., AND SPITZER, C. R. (1976). The environs of the Viking 2 lander. *Science* 194, 1309– 1318.
- SMOLUCHOWSKI, R. (1968). Mars: Retention of ice. *Science* **159**, 1348–1350.
- SODERBLOM, L. S., KREIDLER, T. J., AND MASURSKY H. (1973). Latitudinal distribution of a debris mantle on the Martian surface. J. Geophys. Res. 78, 4117– 4122.
- SQUYRES, S. W. (1978). Mars: Flow of erosional debris. *Icarus* 34, 600-613.
- THOMPSON, S. B., ARNOLD, F., AND SANDERSON, R. (1973). Optical effects of cryodeposits on low scatter mirrors. Astronaut. Aeronaut. 35, 229–235.
- TOON, O. B., POLLACK, J. B., AND SAGAN, C. (1977). Physical properties of the particles composing the Martian dust storm of 1971-1972. *Icarus* 30, 663-696.
- TOULMIN, P., III, BAIRD, A. K., CLARK, B. C., KEIL, K., ROSE, H. J., JR., CHRISTIAN, R. P., EVANS, P. H., AND KELLIHER, W. C. (1977). Geochemical and mineralogical interpretation of the Viking inorganic chemical results. J. Geophys. Res. 82, 4625– 4634.
- WALKER, J. C. G. (1976). Implications for atmospheric evolution of the inhomogeneous accretion model of the origin of the Earth. In *The Early History of Earth* (B. G. Windley, Ed.), pp. 537-546. Wiley, New York.
- WARD, W. R. (1974). Climatic variations on Mars, 1, Astronomical theory of insolation. J. Geophys. Res. 79, 3375-3386.
- WARD, W. R. (1973). Large-scale variations in the obliquity of Mars. Science 181, 260-262.
- WARD, W. R. (1979). Present obliquity oscillations of

Mars: Fourth order accuracy in orbital e and I. J. Geophys. Res. 84, 237-241.

WARD, W. R., BURNS, J. A., AND TOON, O. B. (1979).
2, Past obliquity oscillations of Mars: Role of the Tharsis uplift. J. Geophys. Res. 84, 243-259.

WARD, W. R., MURRAY, B. C., AND MALIN, M. C.

(1974). Climatic variations on Mars. 2. Evolution of carbon dioxide atmosphere and polar caps. J. Geophys. Res. 79, 3387-3395.

WECHSLER, A. E., AND GLASER, P. E. (1965). Pressure effects on postulated lunar materials. *Icarus* 4, 335-352.