The Martian Paleoclimate and Enhanced Atmospheric Carbon Dioxide

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Current evidence indicates that the Martian surface is abundant with water presently in the form of ice, while the atmosphere was at one time more massive with a past surface pressure of as much as 1 atm of CO_2. In an attempt to understand the Martian paleoclimate, we have modeled a past CO_2–H_2O greenhouse and find global temperatures which are consistent with an earlier presence of liquid surface water, a finding which agrees with the extensive evidence for past fluvial erosion. An important aspect of the CO_2–H_2O greenhouse model is the detailed inclusion of CO_2 hot bands. For a surface pressure of 1 atm of CO_2, the present greenhouse model predicts a global mean surface temperature of 294°K, but if the hot bands are excluded, a surface temperature of only 250°K is achieved.

INTRODUCTION

The low mean values of the present atmospheric surface pressure (6 mbar) and surface temperature (219°K) make the existence of liquid water on Mars extremely unlikely (e.g., Ingersoll, 1970). Thus the discovery of sinuous, branching channel systems whose morphology implies water erosion has been recognized for several years as evidence of an extraordinary change in the Martian climate (Masursky, 1973; Sagan *et al.*, 1973). The problem is divided into two parts:

1. Was the Martian atmosphere denser in the past?

2. Did such an atmosphere produce a sufficient greenhouse effect to raise the mean

global temperature to a value consistent with the existence of liquid surface water?

The first question has been given an affirmative answer from two different points of view. Attempts to reconstruct the Martian volatile inventory by studying escape processes (McElroy *et al.*, 1977) or by using a model based on terrestrial and meteoritic volatiles (Anders and Owen, 1977) both lead to the conclusion that the early atmosphere could have been much more massive than the remnant we observe today. A positive answer to the second question has been obtained only by invoking the presence of ammonia as a trace atmospheric constituent in the planet's early history (Sagan and Mullen, 1972; Pollack, 1979). But this is not entirely satisfactory, since the common assumption that the terrestrial planets all began their histories with strongly reducing atmospheres has been undergoing a serious

revision in recent years (Abelson, 1966; Walker, 1976). In particular, we have been able to demonstrate that the greenhouse effect produced by CO_2 and H_2O alone is adequate to compensate for reduced solar luminosity in the case of the early Earth, leading to moderate past global temperatures at a time when CO_2 was more abundant in the Earth's atmosphere (Owen et al., 1979).

We have now extended this approach to a consideration of the paleoclimate of Mars. We shall show in the following discussion that there is reason to believe that a past CO_2–H_2O greenhouse could have been adequate to produce the required warming on this planet as well. Pollack's (1979) results show that a CO_2–H_2O greenhouse would produce liquid surface water for a surface pressure of about 2 bars. But this value of the surface pressure would be larger if Pollack had incorporated the higher planetary albedo resulting from increased atmospheric density. In the following we show that despite this modification, a surface pressure requirement which is lower than Pollack's 2 bars is achieved through use of a more detailed CO_2 radiation model.

MARTIAN CLIMATIC CHANGE

From the present enriched ratio of $^{15}N/^{14}N$ in the Martian atmosphere, McElroy et al. (1977) have calculated that the original partial pressure of nitrogen in the Martian atmosphere must have been on the order of 30 mbar ($\pm$ a factor of 3). Similarly, they noted that the "normalcy" of $^{18}O/^{16}O$ in Martian CO_2 (to $\pm 5\%$) requires the presence of a reservoir of oxygen that can exchange with the atmosphere. This reservoir would be equivalent to a layer of water at least 13 m deep over the entire planet. These are only lower limits and do not specify the *total* volatile content of the early atmosphere, which could be considerably higher than these values imply (Yung et al., 1977; McElroy et al., 1977).

By comparing observed abundances of noble gases and estimates of global potassium from gamma ray data, Anders and Owen (1977) have suggested that Mars is deficient in volatiles compared with Earth, but could still have had an early atmosphere with a total surface pressure in the range of 100 to 400 mbar and a water reservoir equivalent to a 9- to 35-m global ocean. These estimates are based on a model that relies on the relative abundances of volatiles exhibited by carbonaceous meteorites of type $C3V$, which seems to correspond closely with the volatile inventory found on Earth (Ganapathy and Anders, 1974). The volatiles are indexed by the conservative assumption that volatile metals (e.g., thallium) have the same abundance on Mars as on the Moon. This abundance can be increased by a factor 2 to 3 without violating the observed ratio of $^{36}Ar/^{40}Ar$, which provides an estimate of the relative abundance of the most volatile elements to those in a class with potassium. Thus a surface pressure of 0.8 to 1.2 atm of CO_2 is consistent with this model, which would predict a corresponding nitrogen abundance of 8 to 12 mbar and an amount of water equivalent to a layer 54 to 85 m deep, values that are consistent with the lower limits set by considerations of atmospheric escape as described above. Still higher abundances of CO_2 and N_2 are possible if we relax the requirement for a pure $C3V$ model and allow an admixture of material resembling $C1$ and $C2$ meteorites.

Subsequent to the submission of this paper, Pollack and Black (1979) have presented another model for predicting planetary volatile inventories. They estimate 1 to 3 bars for the amount of outgassed CO_2 on Mars.

Thus there appears to be no difficulty in identifying a complement of volatiles that could have provided sufficient water and an adequate surface pressure (many times that of the triple point) to permit fluvial erosion during the early history of Mars. The transformation of this primordial environment to the thin, dry, cold atmosphere we observe today can be understood by invoking a

combination of escape, chemical deposition, and storage in current reservoirs. Among the latter we can list the regolith (Fanale and Cannon, 1978), the residual polar caps (Farmer et al., 1977), and subsurface permafrost (Smoluchowski, 1968). But high atmospheric pressure is not a sufficient condition to permit extensive fluvial erosion. The next step is to demonstrate that this early atmosphere, by means of a CO_2–H_2O greenhouse, would also have caused a sufficient increase in the mean global temperature of the planet to allow the abundant water to exist in the liquid state.

Pollack (1979) has modeled a CO_2–H_2O atmosphere; he concludes that a CO_2 surface pressure of about 2 bars would be required to produce the warming necessary for the presence of liquid water on the Martian surface. With a surface pressure of 1 atm his model yields a global mean surface temperature of 252°K. Recently, however, Owen, et al. (1979), in modeling an enhanced carbon dioxide greenhouse for the early Earth, have illustrated the importance of detailed modeling of several hot bands of CO_2. In the following section we adopt a climate model which reduces Pollack's estimate of CO_2 surface pressure necessary for the presence of liquid surface water. The model additionally incorporates the modifying influence of increased planetary albedo due to Rayleigh scattering within a dense CO_2 atmosphere.

CLIMATE MODEL

The global climate model which we employ is a modification of the terrestrial radiative–convective model of Augustsson and Ramanathan (1977), retaining atmospheric opacity sources due only to carbon dioxide and water vapor. This model provides a detailed incorporation of the CO_2 hot bands located in the 9- to 10-μm region and at 12 and 18 μm, in addition to the continuum absorption by water vapor. We have adopted the continuum H_2O opacity model of Ramanathan (1976) with the following

minor modification. Following the suggestion of Roberts et al. (1976) we reduced the H_2O–air broadening coefficient by a factor of 5 from that given in Ramanathan (1976). We do not include the 7.6-μm pressure-induced band of CO_2 within the modified Augustsson–Ramanathan model, since we estimate that this band will be of no importance for surface pressures considered here.

Minor modifications of the model consist of simply replacing terrestrial input parameters with those appropriate to Mars; these are summarized in Table I. Our use of the present solar constant, despite the fact that a warmer Mars most likely coincided with a colder Sun, is discussed in the following section within the context of seasonal changes. A surface albedo of 0.225 was chosen to produce a planetary albedo close to the reported value of 0.214 for present Martian conditions (Kieffer, et al., 1977). The surface emissivity of 0.9 coincides with that for terrestrial deserts (Sellers, 1965), and this should be reasonably representative of Mars. As in Pollack's (1979) model, the critical lapse rate within the troposphere is 70% of the dry adiabatic lapse rate (5°K km^{-1}) to crudely simulate processes which could cause a reduction from the dry adiabatic value. For lack of a more rational choice, we have employed the terrestrial value of surface relative humidity (0.77), assuming as did Pollack that the relative humidity is constant throughout the troposphere. The stratospheric temperature is computed by assuming radiative equilibrium, and in this respect the present model

TABLE I

MODEL INPUT PARAMETERS FOR MARS

Quantity	Value
Gravitational acceleration	373 cm sec^{-2}
Solar constant	583 W m^{-2}
Surface albedo	0.225
Surface emissivity	0.90
Lapse rate	3.5°K km^{-1}
Surface relative humidity	0.77

differs from Pollack's model, in which stratospheric temperature is arbitrarily fixed.

A further modification of the climate model is necessary to incorporate CO_2-broadened rotational lines, as opposed to air-broadening within the terrestrial model. For this purpose we employ the self-broadened CO_2 mean line halfwidth of Cess and Ramanathan (1972), while the broadening pressure for water vapor bands (except for the continuum e-type bands) is doubled to account approximately for CO_2 broadening versus air broadening (Varanasi, 1971; Izatt *et al.*, 1969).

In addition, it is necessary to replace the Rayleigh scattering parameterization employed in the terrestrial model by one appropriate to a CO_2 atmosphere with variable surface pressure. This was accomplished by scaling the Rayleigh optical depths of Penndorf (1955), incorporating the differences in the refractive index and depolarization factor between CO_2 and air. The monochromatic Rayleigh albedo at six representative wavelengths, applicable for global mean conditions, was then evaluated by employing the scattering formulation of Wang (1972). From this the surface-atmosphere albedo was evaluated at wavelengths relevant for Rayleigh scattering. At larger wavelengths the model incorporates solar absorption by near-infrared bands of CO_2 and H_2O.

RESULTS AND DISCUSSION

Calculated global-mean surface temperatures are illustrated in Fig. 1 as a function of surface pressure for two values of tropospheric lapse rate. These results start at the present Martian surface pressure (0.006 atm), for which the model-predicted global surface temperature is 219°K. As surface pressure increases, the increase in surface temperature is at first rather gradual, since here the primary increase in atmospheric opacity is caused by the 15-μm CO_2 ground-state band. But as the surface pressure is increased beyond about 0.3 atm,
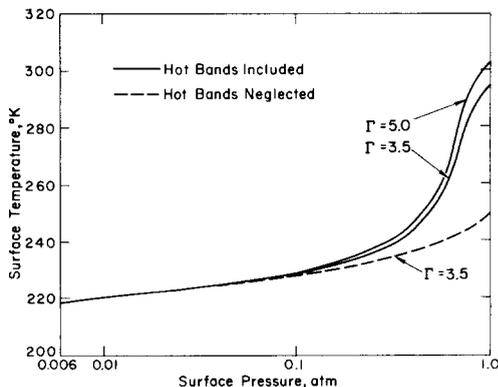

FIG. 1. Global annual surface temperature as a function of surface pressure. The inclusion or deletion of hot bands refers to the CO_2 bands. Γ is the tropospheric lapse rate.

the corresponding temperature increase becomes dramatic. This is produced by a rapid increase in atmospheric opacity resulting from the strong temperature dependence of the CO_2 hot bands, together with a corresponding enhancement in the atmospheric water vapor opacity caused by the coupling between H_2O abundance and surface temperature. An important ingredient in the temperature dependence of the H_2O opacity is the continuum e-type absorption, which varies as the square of the H_2O partial pressure. For surface pressures greater than about 0.3 atm, there is a rapid increase in H_2O partial pressure with temperature. To give an example, as the surface pressure increases from 0.6 to 1.0 atm, the H_2O abundance increases by a factor of 24.

From the comparison in Fig. 1, the CO_2 hot bands clearly trigger the strong opacity–temperature coupling. For a surface pressure of 1 atm, the hot bands cause a 44°K increase in surface temperature. Again, the sensitivity of the model surface temperature to inclusion of hot bands would have been considerably smaller had the model not included H_2O e-type absorption.

The increase in surface temperature would be even greater if there were not a partially compensating increase, within our model, of the planetary albedo. This is illus-

trated in Fig. 2, which shows the variation of planetary albedo with surface pressure for $\Gamma = 3.5°K$ km^{-1} and with the hot bands included. For present conditions (0.006 atm) the planetary albedo is 0.215, less than the surface albedo (0.225) because of solar absorption by near-infrared bands of CO_2. As surface pressure increases, CO_2 as well as H_2O solar absorption will further decrease the planetary albedo relative to the surface albedo. But this effect is more than compensated by an increase in planetary albedo due to Rayleigh scattering, as is illustrated in Fig. 2. For surface pressures greater than 0.05 atm, Rayleigh scattering predominates over CO_2–H_2O solar absorption. The planetary albedo variation between 0.6 and 1.0 atm is caused by a very rapid increase in H_2O solar absorption from 0.6 to 0.8 atm, which reduces the albedo, after which the H_2O bands are saturated, with Rayleigh scattering again causing an albedo increase. If we ignore the atmospheric modification of planetary albedo and simply employ the presently observed value of 0.214, then for a surface pressure of 1 atm the model yields a surface temperature of 303°K, which is 9°K higher than when atmosphere-albedo coupling is included.

An important feature illustrated in Fig. 1 is the extreme sensitivity of global surface temperature to changes in surface pressure for pressures near the upper limits discussed in the second section. In the vicinity
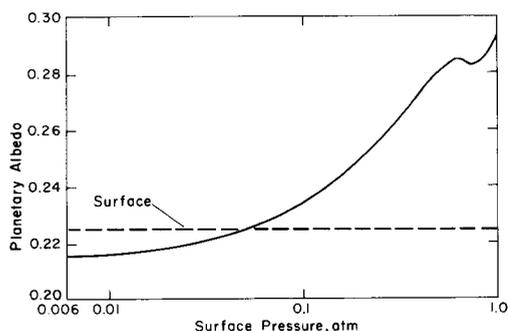

FIG. 2. Planetary albedo as a function of surface pressure for $\Gamma = 3.5°K$ km^{-1} and with hot bands included. The value of the surface albedo is 0.225.

of 0.6 atm, a 10% change in surface pressure causes the surface temperature to change by nearly 8°K. Because of this extreme sensitivity, the model results should be interpreted only in a qualitative context, since the model will be similarly sensitive to other changes.

For comparative purposes we have also calculated the predictions of our model with the same input quantities employed by Pollack (1979), namely: (1) a tropospheric relative humidity of 1.0, (2) a surface emissivity of 1.0, (3) a planetary albedo of 0.24, and (4) use of pressure-broadened line halfwidths appropriate to air rather than CO_2. For a surface pressure of 1 atm, our model yields a surface temperature of 290°K, as compared with Pollack's value of 252°K. Clearly there is a substantial difference between the two models, the cause of which we are unable to specifically identify. We simply note that our radiative–convective model contains a much more detailed radiative transfer formulation than does Pollack's model.

Although our model calculations employ the current solar constant for Mars, a warmer Mars in the distant past would have undoubtedly had to contend with a colder Sun, since models for the evolution of the Sun require an increase in solar luminosity by about 25% since the formation of the solar system (Newman and Rood, 1977). But a colder Sun would, on a seasonal basis, be more than compensated by the Martian elliptic orbit, since the maximum eccentricity is 0.14 (Murray, *et al.*, 1973), such that global insolation at perihelion can be as much as 30% greater than the annual mean insolation. Depending upon longitude of perihelion, summer hemispheric insolation would be even larger. It would, of course, be advantageous to have a more quantitative estimate of this effect from, say, seasonal climate modeling. Finally, there is ample margin for an even denser CO_2 atmosphere (and hence a greater greenhouse) in the inventory of Martian volatiles reviewed in the second section.

To test the sensitivity of our model results to input parameters, we have replaced our previously chosen values for the surface albedo (0.225) and surface emissivity (0.9) by 0.18 and 1.0, respectively, which might be more appropriate to a moist Martian surface, while arbitrarily reducing the surface relative humidity to 0.5. These revisions have little effect upon the model results, reducing surface temperature by 5°K for a surface pressure of 1 atm.

One point which we have so far ignored concerns the probability that a warm and wet Mars should have clouds within its atmosphere which, in turn, would influence surface temperature. To illustrate the possible climatological effects of Martian clouds, consider first the terrestrial case. Satellite measurements of the outgoing infrared radiation from Earth (Ellis, 1978) suggest that the presence of clouds reduces the outgoing infrared radiation by about 20 W m^{-2} which would, by itself, lead to global warming. But this effect is countered by an enhanced planetary albedo resulting from clouds. To estimate the corresponding reduction in absorbed solar radiation, ΔQ_a, it follows that if we assume 50% cloud cover

$$\Delta Q_a = 0.5(S/4)(\alpha_c - \alpha_s), \qquad (1)$$

where $S = 1360$ W m^{-2} is the solar constant, while α_c is the cloudy sky albedo and α_s that for clear skies. Choosing $\alpha_c = 0.43$ and $\alpha_s = 0.18$ (Cess, 1976), then $\Delta Q_a = 43$ W m^{-2}, which is more than twice the infrared reduction, implying that the presence of clouds cools the Earth.[2]

But one cannot directly extrapolate this conclusion to a warm Mars. To give an example as to why this cannot be done, let us naively assume that clouds would exert the same effect upon the Martian radiation budget as they do for Earth. Thus with $S = 583$ W m^{-2} and choosing $\alpha_s = 0.28$ as

[2] This implication that clouds cool the Earth should not be confused with the climatological effect resulting from a small change in the present cloud cover fraction or "cloud-amount feedback," which is quite a different matter (e.g., see Cess and Ramanathan, 1978).

representative of a warm Mars (see Fig. 2), it follows from (1) that $\Delta Q_a = 11$ W m^{-2}, which is half the infrared reduction, such that within this context the presence of clouds on Mars would lead to surface warming.

CONCLUSIONS

Our present modeling endeavors suggest that past amounts of atmospheric CO_2 (≈ 1 atm), consistent with limits discussed in the second section, could, through a CO_2–H_2O greenhouse, lead to Martian surface temperatures which would permit liquid water to persist on the planet's surface, particularly when seasonal effects are taken into account. But in view of the numerous complexities in attempting to model the past Martian climate, it would be presumptuous to attempt to draw any quantitative conclusions concerning the Martian paleoclimate through use of our simplistic global annual radiative–convective climate model. What we *can* conclude is that the CO_2 hot bands must properly be included in any such model or grossly incorrect results will be obtained. When such models are done in detail, the possibility *cannot* be ruled out that a past warm Mars, with liquid surface water, was a consequence solely of a CO_2–H_2O greenhouse resulting from a *reasonable* enhancement of atmospheric CO_2.

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