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An estimation of the mass dragged by the solar wind from Mars's atmosphere in its geologic history

Héctor Javier Durand-Manterola
Depto. de Física Espacial, Instituto de Geofísica, UNAM, México, D.F. México

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RESUMEN

En el pasado Marte tuvo una atmósfera más densa. Debido a la inexistencia de un campo magnético que protegiera la ionosfera y la exosfera del ataque del viento solar, éste erosionó la atmósfera marciana. Se propone un modelo que describe esta erosión a lo largo del tiempo geológico. Se concluye que: a) La cantidad de volátiles desgasificada en Marte fue del orden de 193.7 TAM (1 TAM = 1 Masa de la atmósfera terrestre = 5.28×10^{18} Kg). b) La cantidad de volátiles arrastrados por el viento solar, si la cronología larga es correcta, está en el intervalo de 0.472 a 1.89 TAM. c) La cantidad de volátiles arrastrados por el viento solar, si la cronología corta es correcta, está en el intervalo de 0.0624 a 0.25 TAM. d) La cantidad de volátiles arrastrada por el viento solar es mucho menor que la masa desgasificada. Por lo tanto, el arrastre ejercido por el viento solar no explica la mayor parte de los volátiles perdidos por Marte.

PALABRAS CLAVE: Marte, viento solar, atmósfera marciana.

ABSTRACT

In the past Mars had a denser atmosphere, but it lacks a magnetic field to protect the ionosphere and exosphere from the solar wind. A model for describing the loss of atmosphere in geologic time is presented. The amount of volatiles degassed from Mars was in the order of 193.7 Terrestrial Atmospheric Masses (TAM). The amount of volatiles dragged by the solar wind, if the large chronology is correct, is in the range of 0.472 to 1.89 TAM. If the short chronology is correct, the loss remains in the range of 0.0624 to 0.25 TAM. The amount of volatiles dragged by the solar wind is far less than the degassed mass; hence the drag exerted by the solar wind does not account for the bulk of volatiles lost by Mars.

KEY WORDS: Mars, solar wind, martian atmosphere.

INTRODUCTION

The Martian atmosphere was significantly more dense in the past than it is today (e.g. Sagan *et al*., 1973; Pollack *et al*., 1987). Evidence of atmospheric erosion is presented by Owen *et al*. (1988). The ratio of HDO to H_2O in Martian water vapor is 3 to 12 times the terrestrial value, implying that substantial amounts of water have been lost during Mars history. Large gas pressures in the early Martian atmosphere imply the existence of higher temperatures due to greenhouse effect with atmospheric CO_2 and H_2O and, possibly, the existence of liquid water on the surface as suggested by valley networks and outflow channels (McElroy *et al*., 1977; Clifford *et al*., 1988; Carr, 1986 and 1987; Cattermole, 1992).

The primitive Martian atmosphere, formed from gas of the primitive nebula (Greenberg, 1989), was probably rich in H_2 (Zhang *et al*., 1993). It was removed during the heavy bombardment (Dreibus and Wänke, 1989) in the first 10^9 years of the planet's evolution mainly by impact erosion (Ahrens *et al*., 1989) and by processes like hydrodynamic escape or "blowoff" (Hunten *et al*., 1989). The secondary atmosphere was formed later by material degassed from the planetary interior (Zhang *et al*., 1993). The current atmosphere of Mars is the residue of this secondary atmosphere. The secondary Martian atmosphere contains primarily CO_2, H_2O, and N_2. This secondary atmosphere has been partly lost through absorption at the planetary surface (Pollack *et al*., 1987; Squyres, 1989) and by escape to space (Pérez de Tejada, 1987, 1992; Zhang *et al*., 1993).

Mechanisms of escape to space may include thermal and nonthermal processes (Hunten, 1982). The thermal mechanism, also known as Jeans escape process (Jeans, 1916; Chamberlain and Hunten, 1987), was first suggested by Waterson in 1846 (Hunten *et al*., 1989). The solar wind does not participate in this kind of escape.

In Mars the Venus-like interaction with the solar wind (Cloutier *et al*., 1999) is an important factor in the loss of atmospheric gases. Nonthermal escape mechanisms include charge exchange, dissociative recombination, impact dissociation, photodissociation, ion-neutral reaction, sputtering by solar wind proton incidence, solar wind pick-up, and ion escape (Hunten, 1982; Chamberlain and Hunten, 1987; Hunten *et al*., 1989). On Mars Crider *et al*. (2000) show that electron impact ionization is a source of ions. When ions reach the exosphere they are swept off by the convective electric field of the solar wind. Nonthermal processes dominate the present atmospheric escape flux from Mars and are responsible for a 75% enrichment of the $^{15}N/^{14}N$ ratio (Hunten *et al*., 1989).

Different estimates for the erosion rate of the Martian atmospheres have been proposed (Sagan *et al*., 1973; Pollack *et al*., 1987; Kar, 1996; Pérez de Tejada 1987).

In this work we develope a model to describe the solar wind erosion for the Mars atmosphere over geologic time.

EROSION OF THE PLANETARY ATMOSPHERE BY SOLAR WIND

I assume that the rate at which the planet loses mass is proportional to the rate at which the solar wind supplies energy. The kinetic power can be expressed as:

$$P_{sw} = \varepsilon v \sigma , \tag{1}$$

where ε is the kinetic energy by unit mass of solar wind, v is the velocity of the wind, and σ is the transverse section of the planetary exosphere.

From observations conduced onboard the Phobos II spacecraft Lundin *et al*. (1990) find that the rate of mass loss is $3x10^{25}$ ions/s. A similar estimate was obtained by Pérez de Tejada (1987) from a calculation of the transport of solar wind momentum to the Martian ionosphere across the velocity boundary layer. Brace *et al*. (1982) have reported, for Venus from Pioneer Venus Orbiter observations, a mass loss rate of $7x10^{26}$ ions/s. This amount is 23.3 times larger than the reported Martian mass loss rate but it is not comparable to the ratio of the atmospheric masses of Venus to Mars, which is about 13 500. This shows that the mass loss rate is not proportional to the total atmospheric mass and not dependent on it.

The solar irradiance L is an essential factor in the mass loading process. The amount of mass transported by the solar wind cannot be larger than the ionized mass. The amount of ions must be proportional to the amount of ultraviolet photons if $n_i << n$ where n_i is the ion density and n is the neutral density. The condition $n_i << n$ obtains in the case of planetary atmospheres.

Thus we may assume that the mass loss rate of Mars and Venus is proportional to the solar ultraviolet irradiance and to the kinetic power of the solar wind. The heliocentric distance of Mars is 2.108 times the heliocentric distance of Venus. Thus in Venus we may expect 4.443 times the Mars values of ultraviolet radiation and kinetic power, as both quantities vary with the inverse of the square of the distance to Sun. Venus should have lost 19.74 times more material than Mars. This figure is close to the actual one. Thus, the assumption that the ionic mass loss rate is proportional to the ultraviolet irradiance, and to the solar wind kinetic power is a reasonable one.

The mass loss rate *dM/dt* can be taken as proportional to the rate of injection of energy in the system *dE/dt* and to the rate of injection of ions in the system dn_i/dt :

$$\frac{dM}{dt} = A \frac{dE}{dt} \frac{dn_i}{dt} , \tag{2}$$

where M is the atmospheric mass, t is time, and A is a constant.

The energy required to drag the ions is supplied by kinetic energy of the solar wind. Then the rate of injection of energy equals the kinetic power of the solar wind. From equation (1)

$$\frac{dE}{dt} = \varepsilon v \sigma , \tag{3}$$

where v is the solar wind speed, ε is the kinetic energy density ($= \rho v^2/2$, where ρ is the solar wind density), and σ is the transversal section of the exosphere of the planet ($= \pi(r_p+h)$, where r_p is the planetary radius and h is the height of the exosphere in the terminator.

The rate of injection of ions is proportional to the amount of ultraviolet photons from the Sun:

$$\frac{dn_i}{dt} = B \sigma L , \tag{4}$$

where B is a constant and L is the solar irradiance at the distance of the planet. Anbar *et al*. (1993) show that photo-dissociation of CO_2 is temperature-dependent, i.e., the number of ions produced by this process depends on irradiance and on the temperature of the photodissociated gas. However, since the photodissociation rate coefficient of CO_2 does not change by an order of magnitude over the range of Martian temperatures, we may take it as constant.

From equations (2), (3), and (4) we obtain

$$\frac{dM}{dt} = \mathrm{A} B \sigma^2 \varepsilon v L . \tag{5}$$

Substituting the expression for ε:

$$\frac{dM}{dt} = \frac{\mathrm{A} B \sigma^2 \rho}{2} v^3 L . \tag{6}$$

Assuming σ and ρ to be constant in time we can take $AB\sigma^2\rho/2 = \alpha$ as a constant. Thus

$$\frac{dM}{dt} = \alpha v^3 L \ . \tag{7}$$

This differential equation controls the drag process.

With the ion flux from the Phobos probe, and assuming that solar wind removes atoms from the Mars atmosphere transferring momentum to the ionospheric ions produced by the ultraviolet radiation of the Sun, Lundin *et al*. (1990) proposed a total flux of 0.5 to 1.0 kg s^{-1}. Lammer and Bauer (1991) found that the dominant mechanism to remove mass is the drag of ions by the solar wind, which they estimated as 0.25 kg/s. Pérez de Tejada (1998), assuming viscous drag by the solar wind, proposed a flux of 0.5 kg/s. I assume for the purpose of this work that the mass loss rate of ions is between 0.25 kg/s and 1 kg/s. We may evaluate the coefficient α in equation (7) with these values and using the current value of the solar ultraviolet irradiance in Mars, 50.55 J $m^{-2}s^{-1}$, and the solar wind velocity 4×10^5 m/s. This yields a value of α in the range of -7.7275×10^{-20} J^{-1} m^{-2} s^3 to -3.091×10^{-19} J^{-1} m^{-2} s^3.

From theoretical studies of solar evolution we know that solar ultraviolet luminosity is a decreasing power of time (Zahnle and Walker, 1982), then it is:

$$L = L_a C\, t^m \ , \tag{8}$$

where L_a is the current ultraviolet irradiance at Mars (50.55 J m^{-2} s^{-1}), $C = 1.898 \times 10^{21}$ $s^{1.24}$ and m is a constant exponent with value -1.24.

The solar wind speed is also a power of time (Newkirk, 1980):

$$v = D\, t^r \ , \tag{9}$$

where $D = 5.27 \times 10^{12}$ m $s^{-0.585}$ and the exponent r is -0.415.

With equations (8) and (9) we may find the solution of equation (7) as follows:

$$M(t) - M(t_a) = -\frac{\alpha L_a C D^3}{(m+3r+1)}\left[t_a^{m+3r+1} - t^{m+3r+1}\right]. \tag{10}$$

This expression represents the amount of volatile mass lost by Mars from time t to the present time t_a.

THE MASS DEGASSED FROM MARS IN ITS GEOLOGICAL HISTORY

Let us estimate the amount of volatiles degassed in the geological history of Mars in terms of units of terrestrial atmospheric mass (TAM = 5.28×10^{18} kg). The current volatile inventory at Earth is 0.77 TAM of N_2, 0.21 TAM of O_2, 60 TAM of CO_2 (mostly in carbonate rocks), and 600 TAM of H_2O (261 TAM in the oceans and the rest in rocks). Altogether this makes 661 TAM (Pollack, 1981). We may assume that all the volatiles degassed in Earth's history are present today because the drag by solar wind was inhibited by the terrestrial magnetic field, and the loss from other mechanisms is low.

To estimate the amount of volatiles degassed from Mars over its geologic history we may use the two-component accretion model by Ringwood (1977, 1979) and by Wänke (1981). This model assumes that the planets were formed by a highly reduced volatile-free component A and an oxidized volatile-containing component B. The mixing ratio of A to B for Mars is 60:40 and for Earth 85:15 (Dreibus and Wänke, 1989). Thus if Mars has the same mass as Earth its amount of volatiles should be 40/15= 2.666 times the volatiles of Earth. But the mass of Mars is 0.11 times that of Earth (Marov, 1985). Then according to the two-component model the amount of volatiles on Mars should be 2.666 x 0.11 = 0.293 times that on Earth, i.e., 0.293 x 661 TAM = 193.7 TAM.

Figure 1 shows equation (10) for the two extreme values of the constant α. Assuming the lower value of α, Mars lost, over the past 4.48 Ga, one TAM. For the larger value of α the same amount of mass was lost in the past 4.3 Ga. If the drag of the solar wind acted since the birth of Mars, 4.6 Ga ago, the amount of atmosphere lost by Mars is in the range from 3.73 TAM to 14.9 TAM.

DISCUSSION AND CONCLUSIONS

The results are smaller than the estimation of the mass degassed in all the history of Mars. Recently the Mars Global Surveyor Mission discovered a remanent magnetic field of Mars that suggests the presence of a dynamo in the past (Acuña *et al*., 1998).). If so the solar wind could have been stopped by the Martian magnetosphere. Thus the mass eroded by the solar wind could actually be smaller than the earlier values.

Recent results on the remanent magnetism in the Martian crust shows that the Martian dynamo existed in Early and Middle Noachian (Barosio and Acuña, 2000). Solar wind erosion would have started at the end of Middle Noachian or the beginning of Late Noachian. Absolute geologic times are not available for Mars, but Neukum and Wise (1976) and Hartmann *et al*. (1981) develop two chronologies. From the Neukum and Wise model (short chronology). Late Noachian began 3.85 Ga ago, when Mars had an age of 0.75 Ga. From the chronology by Hartmann *et al*. (large chronology), Late Noachian began 4.4 Ga ago, when Mars had an age of 0.2

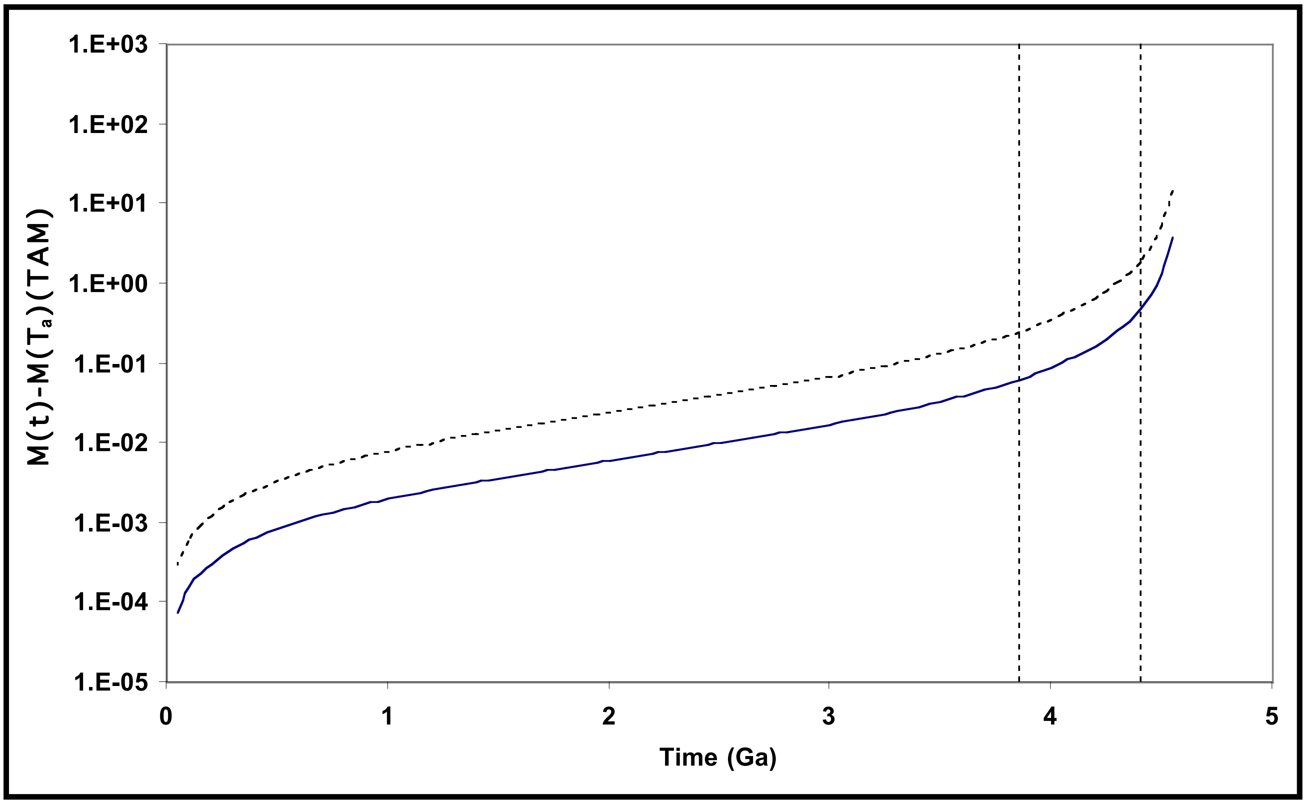

Fig. 1. Mass in TAMs dragged by the solar wind from t years ago to present time. Zero in time scale is the present. The solid line is equation (10) with the minimum value of α and the dotted line is for the maximum value of α. Vertical lines show the time at which the Late Noachian begins, 3.85 Ga ago for the Neukum and Wise (1976) model (short chronology) and 4.4 Ga ago for the Hartmann *et al*. (1981) model (large chronology).

Ga. These limits yield for the large chronology a loss of mass in the range from 0.472 TAM to 1.89 TAM, and for the short chronology from 0.0624 TAM to 0.25 TAM with my model.

These numbers agree roughly with Pérez de Tejada (1992), who estimates 0.82 TAM to the loss of mass in 4.6 Ga. Luhmann *et al*., (1992) estimates a loss of 1.087 TAM in the past 3.5 Ga. My model predicts a dragged mass from 0.033 TAM to 0.133 TAM for this period of time.

The amount of lost water cannot be estimated from the model because we don't know the ratio of atmospheric gases to water in the early Mars materials; but one may assume that it was sufficient to produce an ocean of some meters depth.

In conclusion: (a) the amount of volatiles degassed from Mars was in the order of 193.7 TAM, (b) The amount of volatiles dragged by the solar wind if the large chronology is correct is in the range of 0.472 to 1.89 TAM, (c) The amount of volatiles dragged by the solar wind if the short chronology is correct remains in the range of 0.0624 to 0.25 TAM, (d) The amount of volatiles dragged by the solar wind is much less than the degassed mass. Therefore the drag exerted by the solar wind does not explain the bulk of the volatiles lost by Mars.

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Héctor Javier Durand-Manterola

Depto. de Física Espacial, Instituto de Geofísica,
UNAM, 04510 México D. F., México
Telephone: (525) 56 22 41 41
Fax: (525) 55 50 24 86
Email: hdurand@fis-esp.igeofcu.unam.mx