

Evolution of Mars Atmospheric Pressure and Temperature Modeling and Constrains from Meteorites F. Javier Martín-Torres¹, Carles E. Moyano-Camero², and J.M. Trigo-Rodríguez² ¹ Centro de Astrobiología. CSIC-INTA. Madrid, Spain. javiermt@cab.inta-csic.es, ²Institute of Space Sciences (CSIC-IEEC). Campus UAB, Faculty Sciences, C5-p2. Bellaterra, Spain. trigo@ieec.uab.es

Introduction: The geological history of Mars can be split into three primary periods: Noachian (4.5Gyr ago to 3.5Gyr ago), Hesperian (3.5 Gyr ago to 2.9–3.3 Gyr ago), and Amazonian (2.9–3.3 Gyr ago to present). To understand paleoclimate and to correctly interpret the Mars meteorite record we need to understand two types of atmospheric conditions that may have occurred through the history of Mars: a dry, dusty situation similar to the current Mars and a warmer, wetter climate unlike any recent Martian conditions. For the dry, dusty conditions, the key difference between the current and the climate and paleoclimate is the possibility of a much thicker atmosphere. A thicker atmosphere leads to an increase in the atmospheric heat capacity, a change in the transmission of insolation and infrared energy through the atmosphere, and, therefore, a change in the general circulation and behaviour of the atmosphere. For a warmer, wetter climate, the differences are more drastic. To achieve globally “wet” conditions on ancient Mars requires a larger global water inventory. To better understand the two possible paleoclimate states mentioned above we have conducted several studies to examine the evolution of the Mars atmosphere bearing in mind these factors and deriving the expected mass, pressure, and temperature of the Mars atmosphere as a function of time. This evolution will have an impact on the correct interpretation of the Mars meteorite record as will affect the environmental conditions in Mars. For example large atmospheric pressures in early Martian atmosphere would imply the existence of higher temperatures (through greenhouse effect with atmospheric CO₂) and, possibly, the existence of liquid water on the surface. This view agrees with evidences of liquid water in past, like valley networks and outflow channels [1, 2]. This also fits with the only current evidence available of the oldest Martian meteorite known: ALH 84001 with secondary minerals produced by aqueous alteration and even a highly debated hosting of fossils of bacteria [3].

Mass Atmosphere Modeling: The amount of mass in Martian atmosphere at time t , $m(t)$ depends of two variables: the degassed mass, $m_d(t)$, since the beginning of secondary atmosphere (around 3.5 Gyr ago), and the mass lost, $m_l(t)$ by the atmosphere during the same period, i.e.:

$$m(t) = m_d(t) - m_l(t) \quad (1)$$

The rate of degassing was not continuous in time with periods of volcanism and calm but we know that since the end of the accretion period (Noachian epoch) Mars

was cooling. It is expected that the rate of degassing diminished as the planet aged and we assume that the degassing rate could be computed as:

$$m_d(t) = (A/B)[\exp[B(t - t_0)] - 1] \quad (2)$$

where A is the degassing rate at time t_0 (1Gyr) and B is a constant that controls the velocity of degassing.

The mechanisms of atmospheric mass loss are various: drag of ions by the solar wind ($m_i(t)$), dissociative recombination ($m_{dr}(t)$), oxidation of the surface rocks ($m_{ox}(t)$):

$$m_l(t) = m_i(t) + m_{dr}(t) + m_{ox}(t) \quad (3)$$

The three mass terms contributing to the mass loss are computed as follows:

a) Material dragged by solar wind, $m_i(t)$: We assume that the mass loss rate of ions is:

$$dm_i/dt = A \cdot \rho_s \cdot v^2 \cdot L/2, \quad (4)$$

where a is a constant, ρ_s is the solar wind density, L is the solar ultraviolet irradiance normalized to current values, and v is the solar wind velocity. Using current values of the solar ultraviolet irradiance ($L=1$), a value of $v=4 \times 10^5$ m/s, and $\rho_s=7.15 \times 10^{-21}$ kg m⁻³, A should have a value between -4.37×10^8 s m and -1.75×10^9 s m. After [4]:

$$L = C t^m \quad (5)$$

where $C = 1.898 \times 10^{21}$ s^{1.24} and m is an adimensional constant with value -1.24 ; and [5] established that:

$$v = D t^r \quad (6)$$

where $D = 5.27 \times 10^{12}$ m s^{-0.585} and r is an adimensional constant with a value of -0.415 . With equations (5) and (6) it is possible to find the solution of equation (4):

$$m_i(t) = A \cdot r \cdot C \cdot D^2 \cdot (t^{m+2r+1} - t_0^{m+2r+1})/[2(m+2r+1)] \quad (7)$$

b) Dissociative recombination, $m_{dr}(t)$: Dissociative recombination is a mechanism by which an ionized molecule is recombined with an electron and then dissociated, producing fragments that may acquire velocities larger than planetary escape velocity. We assume that the dissociative recombination rate is proportional to the magnitude of the flux of incident solar ultraviolet radiation [6]:

$$dM_{dr}/dt = b \cdot L \quad (8)$$

so that with equation (5) we obtain:

$$m_{dr}(t) = b \cdot C \cdot (t^{m+1} - t_0^{m+1}) / (m + 1) \quad (9)$$

c) Surface oxidation: As argued in [5] the rate (dm_{ox}/dt) at which oxygen atoms may be immobilized at the surface could be as large as 1.3×10^{12} atoms $m^{-2} s^{-1}$, this is equivalent to a total mass loss of 5 kg/s around the planet. Assuming this rate as constant in time we get that:

$$m_{ox}(t) = 5(t - t_0) \quad (10)$$

Total mass: From previous equations (1), (3), (7), (9) and (10) the equation that describes the evolution of total atmospheric mass in time is:

$$M(t) = (A/B)[\exp[B(t - t_0)] - 1] - A \cdot r \cdot C \cdot D^2 \cdot (t^{m+2r+1} - t_0^{m+2r+1}) / [2(m + 2r + 1)] - bC(t^{m+1} - t_0^{m+1}) / (m + 1) - 5(t - t_0) \quad (11)$$

This equation depends on parameters A and B. Taking $t = t_p$, in which $t_p = 4.6$ Ga (present time), we can find A in function of B:

$$A = B \{ M(t_p) + A \cdot r \cdot C \cdot D^2 (t_p^{m+2r+1} - t_0^{m+2r+1}) / [2(m + 2r + 1)] + bC(t_p^{m+1} - t_0^{m+1}) / (m + 1) + 5(t_p - t_0) / \exp[B(t_p - t_0)] - 1 \} \quad (12)$$

where B (parameter controlling the velocity of degassing of the planet in time) is a free parameter of the model.

Near Surface Temperature Modeling: At the planetary surface the temperature produced by the greenhouse effect is:

$$T^4(t) = T_e^4 [1 + 3 \cdot \tau / 4] \quad (13)$$

(Chamberlain and Hunten, 1987) where T_e is the mean planetary black body temperature and τ is the atmosphere optical thickness. After [7], [8], and [9], T_e can be written as:

$$T_e(t) = T_p [1 + 0.4(1 - t/t_p)]^{-1/4} \quad (14)$$

where T_p is the present value of T_e , t is time, and t_p is the age of the Sun. After the value of the total mass obtained in (11) the atmospheric temperature at the surface becomes:

$$T(t) = T_p [1 + 3t_p M(t) / 4M_p]^{1/4} [1 + 0.4(1 - t/t_p)]^{-1/4} \quad (15)$$

Surface Pressure Modeling: The atmospheric pressure P as function of time through the perfect gas equa-

tion normalized with respect to the present values (P_p, M_p, T_a) is given by

$$P(t) = M(t)T(t)P_p / M_p T_p \quad (16)$$

Using equations (11), and (16) we evaluate P(t).

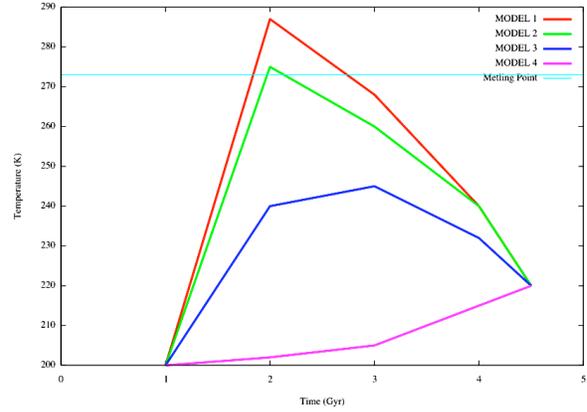


Figure 1. Near surface T as a function of time for different set of free parameters in the model. The melting point temperature is plotted for reference.

Results and discussion: We have developed a 1D model of the evolution of Martian mass, near surface temperature and pressure considering the main production and loss processes of Mars atmosphere and the radiative conditions on Mars. As concluded in [10] SNC meteorites are martian rocks that provide valuable information about the atmospheric composition of Mars over time. They provide a complementary tool for atmospheric modeling and a straightforward method for model validation. Our next step will be validate our model with martian meteorites and tune the free parameters in our model.

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