Evolution of the Atmospheres of Terrestrial Planets: Focus on Mars and Venus.

Gillmann, C., *Institut de Physique du Globe de Paris, Département de Géophysique Spatiale et Planétaire*

Supervised by Lognonne, P., Institut de Physique du globe de Paris, Chassefière, E., Service d'aeronomie IPSL and van Thienen, P., Utrecht University

A simple atmosphere-interior coupling has been implemented for Venus under stagnant lid convection regime; the atmosphere gains water from the degassing, through a parameterized model of mantle convection, including volatile exchanges between the mantle and the atmosphere (additionally, the mean depth of partial melting is taken into account), and a radiative-convective atmosphere model computes the temperature at the planet's surface. This coupling suggests that a strong links exist between inner (the solid part) and outer layers of the planets.
It also seems essential to study the atmospheric escape which could be a major parameter constraining the surface conditions during the early evolution of terrestrial planets. The initial model of the escape we used is very basic, so we develop this aspect to try and see if realistic results may be obtained with a more complete approach. Thus we use an energy-limited approach to model the escape of hydrogen out of the primitive atmosphere and its entrainment of rare gazes. We compare the evolution of rare gaze depletion and the final (after 4.6 Gy) isotopic ratios to those measured by Venera missions. Results show that depending on the extent of the exosphere and the temperature it reaches, it is possible to explain the present isotopic ratios with only the hydrodynamic escape, especially with hot (such as 500 K to 1500 K) and extended (4 to 8 times the size of the planet) exospheres.
Since hydrodynamic escape mostly takes place during the first hundreds of million years, other processes for atmospheric escape have been considered in order to quantify the loss of volatiles during later periods. Using data from Mars Express and several models such as ones created by Leblanc (2001) or Chassefière, Leblanc and Langlais (2006), a model for the evolution of Martian atmosphere and volatiles has been set up. Crust production rates from a model by Breuer et al. (2006) are taken as input for the mantle degassing and the evolution of the content in water, CO_2 and SO_2 of the atmosphere is studied through different scenarios. We first focused on the present situation as described by available data such as ones from Mars Express in order to study the late evolution of the Martian atmosphere. It appears that a production of at least 0.05 to 0.1 is needed for the atmosphere to be at steady state.
Our second interest is to have a view of possible evolutions of the Martian environment over the whole history of the planet and to try to relate it to specific features discovered, and especially with sulfate formations detected by the OMEGA spectrometer.

Evolution of the Martian Atmosphere: Atmospheric Escape and Degassing.

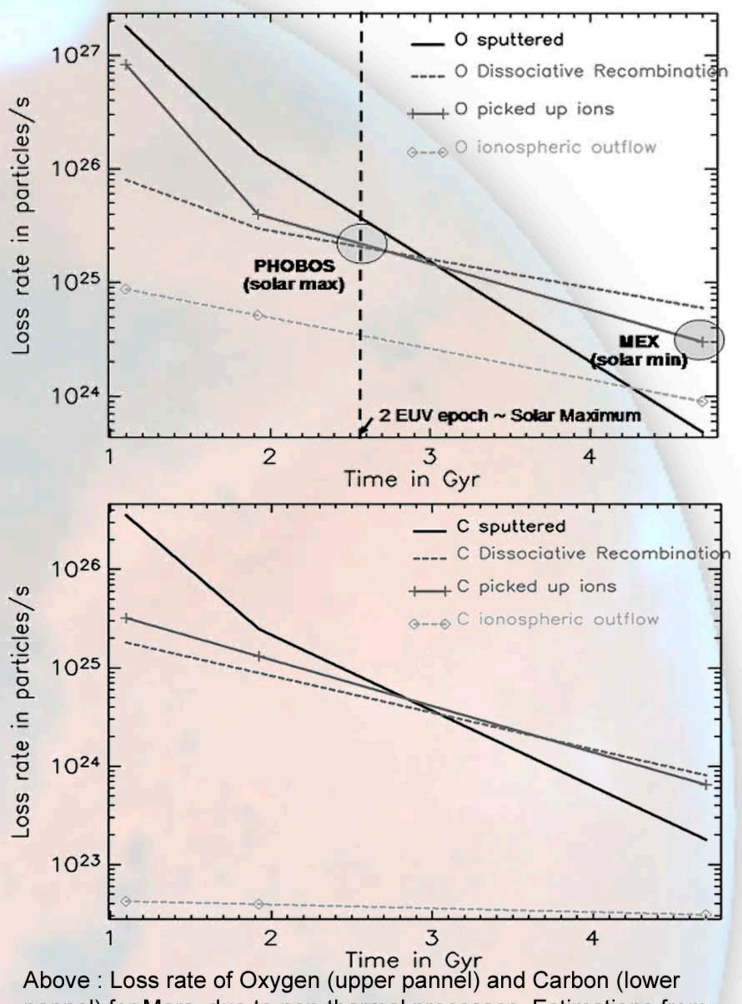
A simple model is used to compute the evolution of different volatiles over time in the atmosphere of Mars during the last 3 Gy. We focus on Water and CO_2. On the first order we consider degassing (input of gases) and escape (which removes the gases from the atmosphere) as the two main processes which control the history of the atmosphere.

Atmospheric Escape.

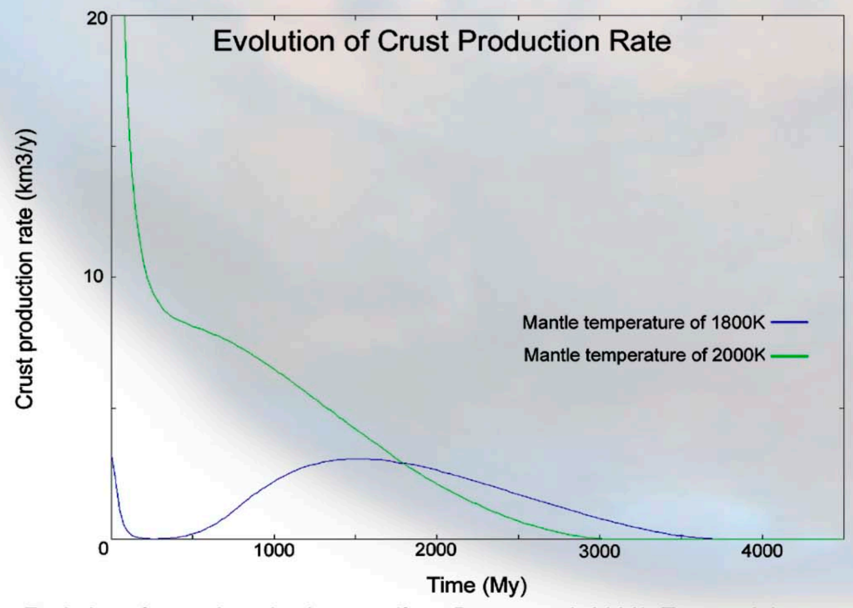
Since our model focuses on the last 3 Gy, the main processes for atmospheric escape are non-thermal. We use data from Chassefière, Leblanc and Langlais (2006) for sputtering, dissociative recombination, ionospheric outflow and ion pick-up to estimate the amount of CO_2 and H_2O lost to space.
-Sputtering corresponds to a mechanism where ions produced in the corona or in the ionosphere can reimpact the neutral atmosphere with enough energy to lead to the ejection of important quantity of neutral atmospheric particles.
-During dissociative recombination, ions produced in the ionosphere by UV photo-ionization recombine with electrons and form in some cases energetic neutrals with enough energy to escape Mars.
-Ion pick-up: ions produced by photo-ionization, electron impact and charge exchange in the Martian exosphere are then dragged along by the moving solar magnetic field lines wrapping the planet.
-Ionospheric outflow: ions are produced within the ionosphere (below the exobase) and can flow in some case up to the ionopause where they are also dragged by the solar wind. Solar UV emission is supposed to be higher in the past than at present time, thus explaining the higher values before 2 Gy.

Degassing.

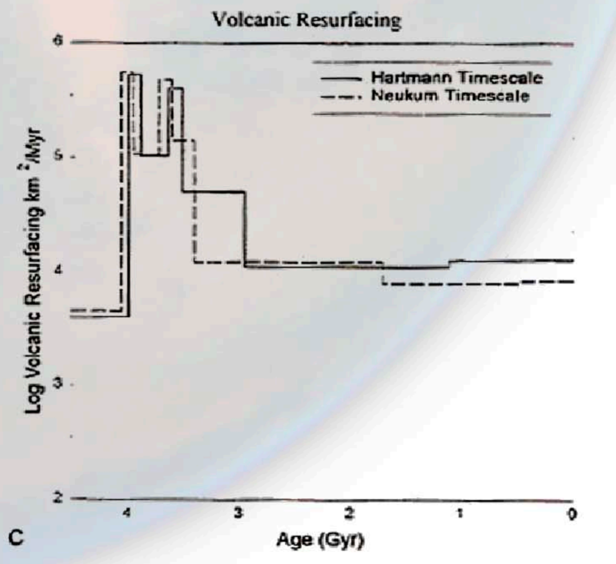
As an input for the amount of volatiles released into the atmosphere, we use data from a numerical study from Breuer and Spohn (2006). They compute the amount of crust produced during the evolution of the Martian mantle. They employed a parameterized model of stagnant lid convection with crust production, core cooling and mantle melting. The model is constrained by both crustal thickness and magnetic field history with help from data from Mars Pathfinder, Mars Global Surveyor and Mars Express. We compared these results with other data from Hartmann and Neukum (2001), obtained by observation of the surface of the planet. And the observation seems to show that activity is not as strong as the numerical seems to imply. We still use input from Breuer et al. but it seems important to point out this difference. We are currently trying to get new estimates of recent volcanic resurfacing in order to better constraint our data.



Above : Loss rate of Oxygen (upper pannel) and Carbon (lower pannel) for Mars, due to non-thermal processes. Estimations from Phobos and Mars Express missions are also shown (reproduced Chassefière et al. 2006).



Above : Evolution of crustal production rate (from Breuer et al. 2006). Two models are shown : one with a hot initial mantle (2000K) and one with a lower mantle temperature (1800K).



Above : Evolution of volcanic resurfacing over time. Note that we are only interested in the last 3 Gy here and that the scale shows a surface per Myr which tends to complicate evaluations of melt volume. (Reproduced from Hartmann and Neukum (2001).

Evolution of the Martian Atmosphere: Results.

First models.

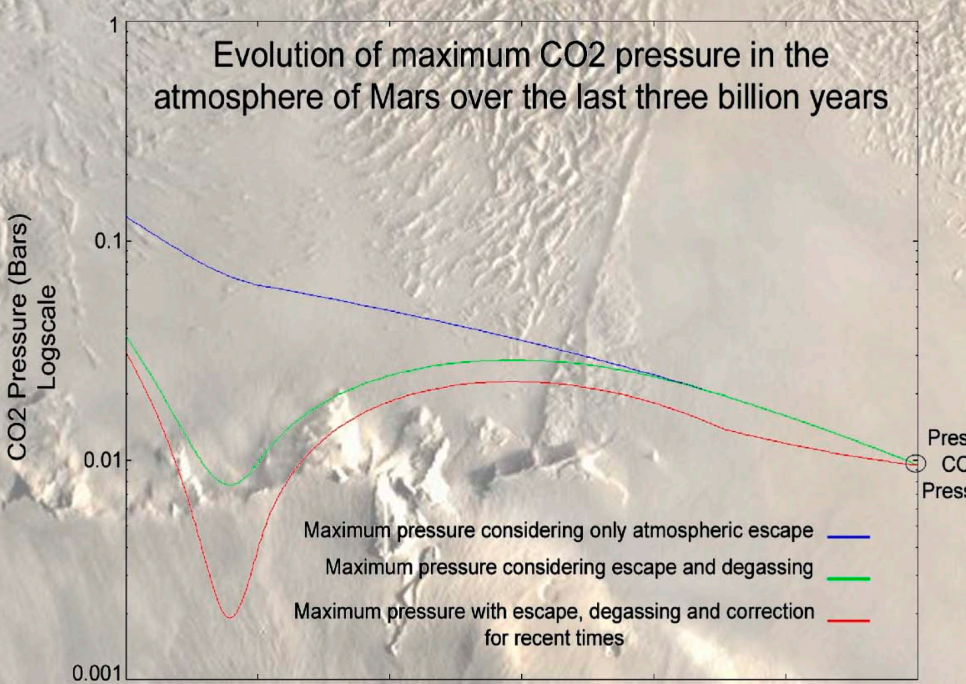
First results are shown to the right on the upper panel. We wanted to have an estimation of the state of the atmosphere at any given time with the main constraint that we obtain the real situation at the present time. For CO_2 that means we enforce present pressure to be just below 10mbars. We then go back in time to find how pressure must have evolved to attain this state. Some study was also done on water pressure but, since many uncertainties still remain concerning present sources (polar ice-caps, ice stored in regolith...), results were still dubious. We did not go back further than 3 Gy. Earlier, many events can occur that are not easily constrained and whose effects are thought as very important such as bombardment, and, very early, hydrodynamic escape.

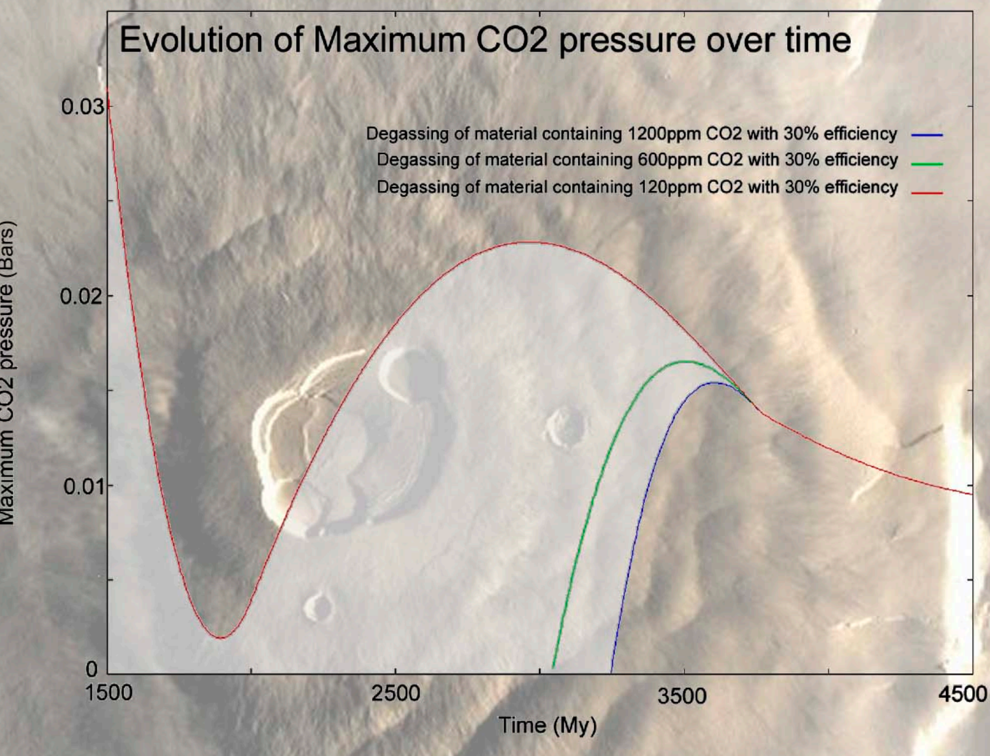
Here we can first see CO_2 maximum pressure when only atmospheric escape is considered. This means that even 3 Gy ago it was unlikely for Mars to have a thick CO_2 atmosphere. This is consistent with studies implying that thick CO_2 atmospheres would first condensate then precipitate at ~1bar. The two other curves take into account degassing based on crustal production rates from Breuer et al. (2006). Values for the maximum CO_2 pressure are significantly lower and stay in the order of several to several tens of millibars. The red one includes a correction used to represent the fact that today Mars is still active (Neukum et al, 2004, shows that volcanic events occurred less than 100My ago with phases of activity as young as 2My). The amount of present volcanism was arbitrarily chosen to balance present atmospheric escape, which might be the case. Further study is being done to get reliable data from the observation and maybe be able to decide if the atmosphere of Mars is at a steady state or is slowly lost into space.
What is really interesting is the minimum around 2Gy. It shows we can obtain today's atmosphere for Mars without needing much primordial atmosphere. In fact most of what we see nowadays might well be a produce of volcanic activity _thus a secondary (and quite recent) atmosphere_ instead of the remnant of a primordial atmosphere.

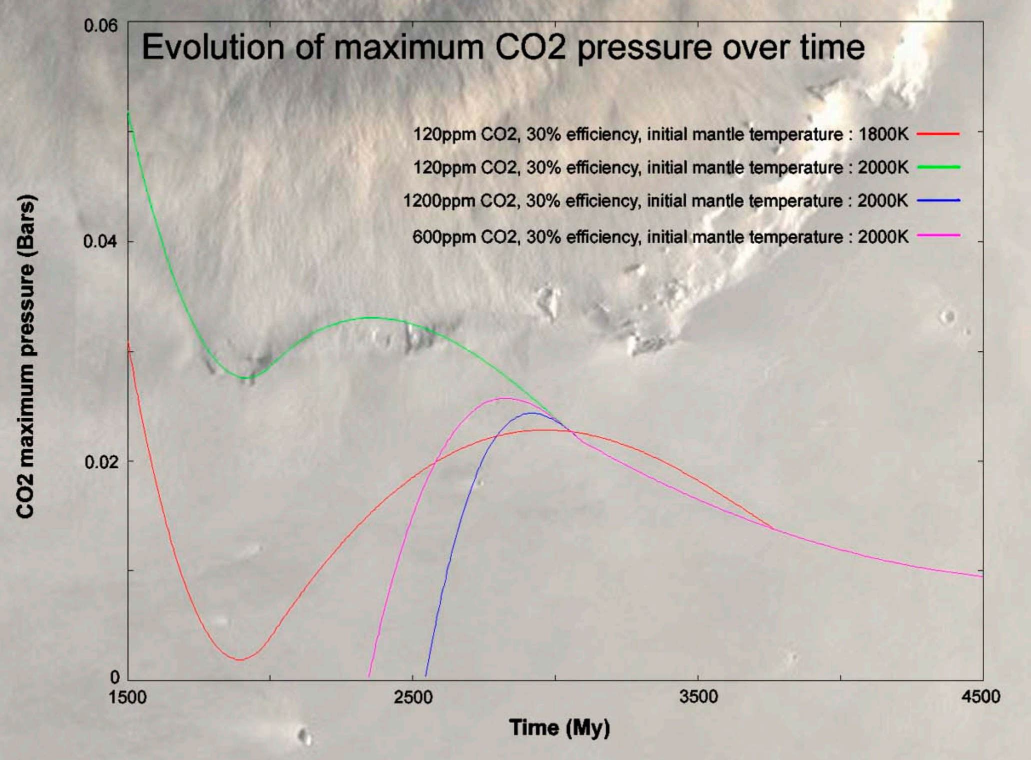
Further study.

We considered several models to test the influence of CO_2 content in the melted material used for the degassing. We assumed that 30% of the volatiles were released into the atmosphere in this model (as shown to the right on the middle panel). CO_2 contents vary from 120ppm to 1200ppm. We observe that for very low CO_2 degassing, a part of the primordial atmosphere remains. However, as soon as the amount of CO_2 available gets higher, the maximum CO_2 pressure drops. This means that with these models, the whole present atmosphere is secondary and originates only from recent volcanism. We can estimate an age for these atmospheres and it can be as young as 1.25 Gy.
These scenarios imply that some process must have efficiently removed (almost) all the primordial atmosphere. Heavy bombardment must have played a role in this, since it is accepted that it could remove up to 99% of the atmosphere, but some other mechanisms might be involved to remove part of the late volcanism-produced atmosphere (before 1.5 Gy). It must be noted that, compared to what we can find on Earth, the CO_2 contents we used are low. It might be realistic to go up to 4000ppm. Moreover, higher degassing efficiency might be used. Thus the Martian atmosphere may be even younger than it is shown on the figure. If studies enable us to have good data concerning recent activity, we could model the late period more accurately. CO_2 content might also have a strong influence on late evolution since it is not sure that late degassing balances present escape. More data is needed to solve this problem.

Other models where used to study the influence of global activity on the evolution of maximal CO_2 pressure. Breuer et al (2006) give several curves for crust production rates depending on the initial temperature of the mantle. We compared results for a 2000K mantle and a 1800K one. (Figure to the right, lower panel).
The red curve corresponds to the cases above and is shown for comparison. We see that for very low CO_2 content, 'higher" pressures can be obtained, however, as soon as we use higher CO_2 contents, the same pattern occurs and it appears that present atmosphere is a secondary volcanism-produced one.
The atmospheres obtained with these scenarios are a little older than in the precedent case due to the late period of lower volcanic activity (the last 1.5 Gy instead of the last 750 My).







Atmosphere : Coupling and Escape.

Basic coupling.

We first used a simple convective-radiative atmosphere model based on a work by Phillips et al (2001) in the case of Venus. The convection parameterized model provides the amount of volatiles released by the mantle into the atmosphere. The atmosphere model calculates the green house effect and gives a surface temperature to the convection model. The program compares a radiative temperature (in the upper part of the atmosphere, due to the green house effect) given by :

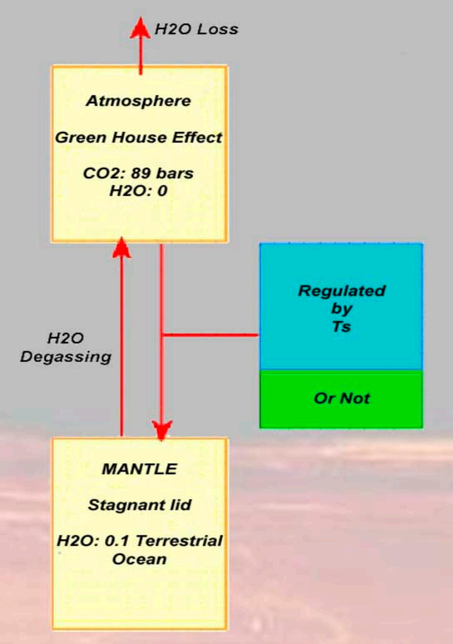
$$T_r(z) = T_e\left(\tau(z) + \frac{1}{2}\right)^{1/4}$$

(where T_e is the effective radiative temperature of the atmosphere (323 K) and $\tau(z)$ is the total opacity in the infrared due to green house gases)
with a convective temperature (in the lower part of the atmosphere, found with an adiabatic gradient) and given by :

$$T_c(z) = T_s - \Gamma z$$

where Γ is the adiabatic gradient and T_s the previous surface temperature.

We also chose to allow the atmosphere to escape into space. In this experiment this was done in a very basic way with a simple exponential decrease with a time constant of about 160 My (Grinspoon 1993). This should represent the hydrodynamic escape of Hydrogen due to Extreme UV.



Schematic figure representing the links between different parts of the model.

We used the program in two ways: one which is fully coupled and one where the new surface temperature is calculated but not used by the convection model. On the figure on the right, we can thus see the influence of this parameter and see how important it is to link the different layers when studying planets.

Several tests have been run and it appears that they all seem to show a first event characterized by a high surface temperature. This corresponds to the early massive degassing by the mantle in our convection models. After this event, atmospheric hydrogen escapes and surface temperature decrease over the first three billion years to reach values which are lower than those we can measure on Venus at present time. This value corresponds to a dry atmosphere.
Even if the early high temperature is not alarming (Kasting proposed this possibility in 1986) and could be an argument in favour of a primordial magma ocean, the late "low" surface temperature is annoying. The difference between the results and reality could be explained by a recent activity on Venus that was not modelized such as an hypothetic global resurfacing but we need to find out if with minor changes, our model could lead to more realistic present surface temperatures.
Reduction of the "active" surface of the planet to 25% of the total surface reduces sensibly the temperature reached during the first millions years but fails to lead to more permanent change in the surface temperature. Even with this approach, not enough water is degassed in the later period.
Even the late degassing due to the change of convection regime from plate tectonics to stagnant lid (see curves above, in the first part) fails to induce a late increase of surface temperature.
This is due to the atmospheric escape which doesn't allow high quantities of H_2O in the atmosphere in more recent times. Since the model we used for this part of the program is really crude, it seems interesting to investigate further atmospheric escape and try and model it more accurately.
In the present state, this model can still illustrate the need to study the coupling between atmosphere and inner dynamics. The difference between the fully coupled model and the one where atmosphere has no influence on mantle evolution is plain to see.

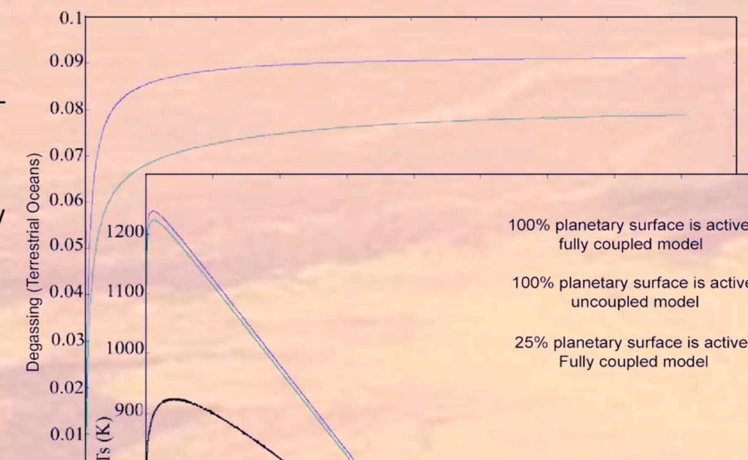


Figure showing the surface temperature and degassing of Venus for coupled (blue) and uncoupled (green) models and the influence of the active surface of the planet (black : 25% of surface is active). Initial water content is 0.1 Terrestrial Ocean, Initial mantle temperature is 2500K.

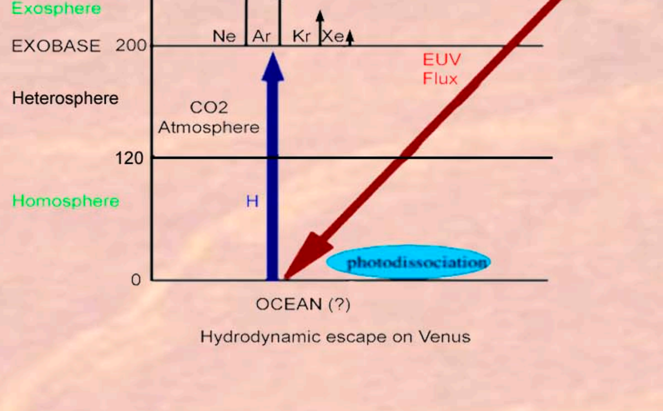
Atmospheric escape.

We have realized a simple model for the atmospheric escape from Venus. The program calculates the amount of hydrogen lost into space due to hydrodynamic escape. We do not take into account the effect of sputtering, impacts or solar winds. We can also find out the amount of noble gases dragged along during the outflow of hydrogen and compare the results with data from the Venera missions. Two aspects of this work are interesting. First, we could use this parameterization of the atmospheric escape in the coupling, secondly this study can give good insight on whether hydrodynamic escape has a major influence on the evolution of planets and the need to take it (or other mechanisms) into account to explain the current state of the atmosphere.
This approach is based on work by Zanhle and Kasting (1987),Kasting and Pollack (1983), Hunten, Pepin and Walker (1987) and Chassefière (1996).

The model is energy-limited. The EUV Flux on top of the atmosphere gives the energy needed to create a thermal escape flux which will drag along elements and molecules whose mass is lower than the critical mass m_c.

$$m_c = m_1 + \frac{k_b T F_1}{b g X_1} \qquad F_2 = \frac{m_c - m_2}{m_c - m_1}\frac{N_2}{N_1}F_1$$

Where 1 refers to H and 2 the element that is dragged along. N is the quantity of the element, m its mass, F its flux out of the atmosphere. T is the temperature, g the gravitational acceleration, b the diffusion parameter, and k the boltzmann constant.

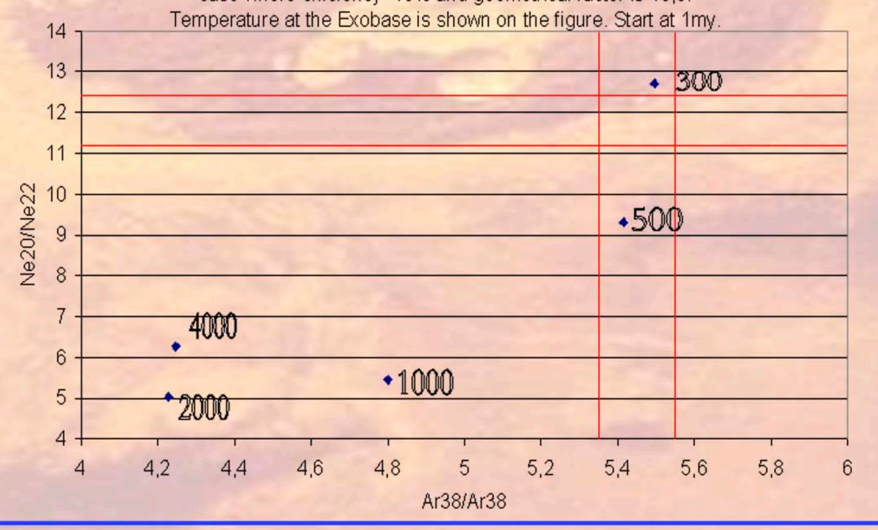


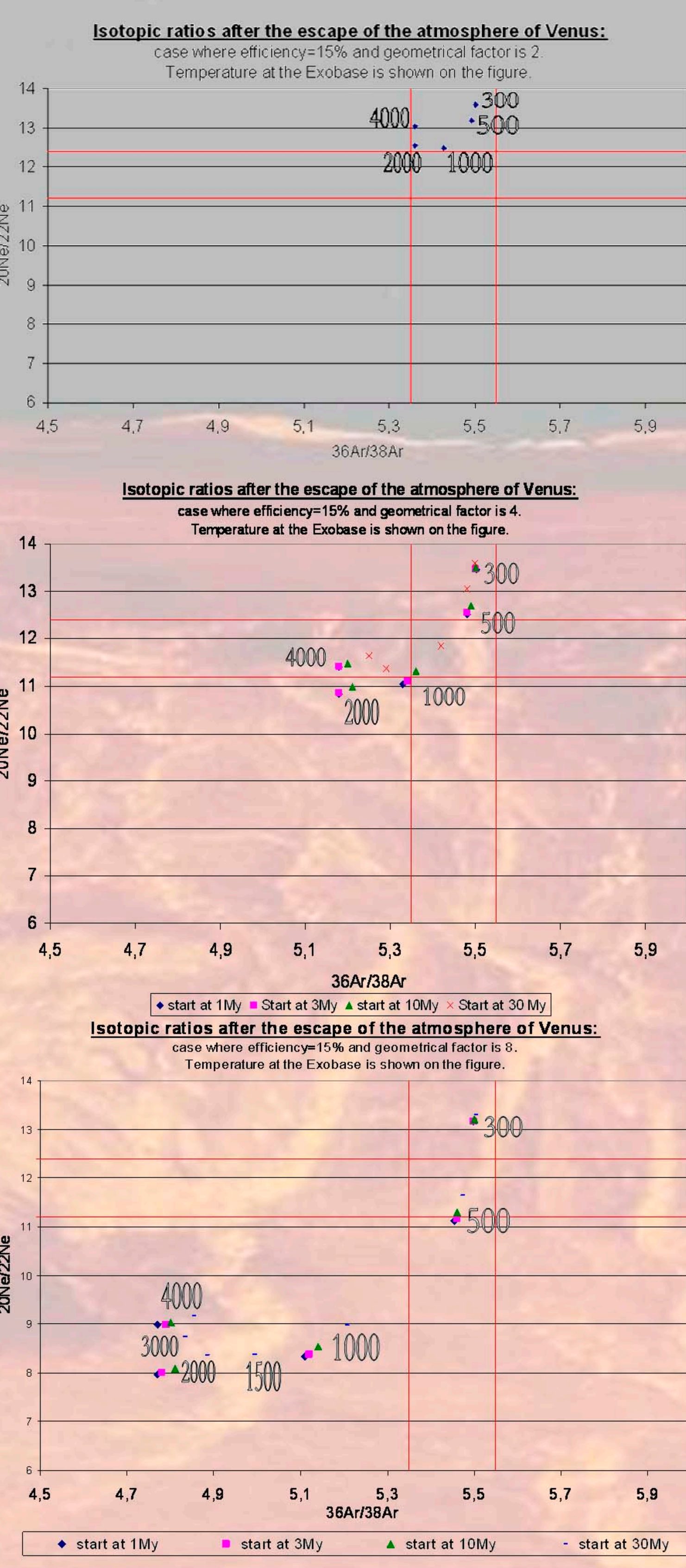
Hydrodynamic escape on Venus

Results.

Several parameters are investigated such as the time when the escape starts, the geometrical factor (directly linked to the amount of energy available for the escape, it is the radius of the zone where energy from the Sun is absorbed by the atmosphere. Its usual value is 2 planetary radiuses but when the temperature at the exobase is high or when the atmosphere expends, it can reach higher values) or the temperature at the exobase. The figures to the right show some of the results obtained for these models. They display calculations for different values of the precited parameters and show the area (red square) where results are expected to stay if they are to fit with in-situ measures.
We can see that isotopic ratios decrease when exobase temperature increases up to a value of 2000K. After this value, isotopic ratios increase. This is due to the competition of two different effects affecting fractionation. One is the escape and the other is the gravity. At high temperatures, the atmosphere is sufficiently hot to bring heavy elements in high amounts to the upper levels of atmosphere. Thus the effect due to gravity is removed and the fractionation is decreased.
When observing the difference between results at different starting dates, we see that the case starting after 30My presents a real difference. This leads us to think that after 30My a significant part of the escape has already been done in the other cases. Escape is most active during the first tens million years; after that it is a lot less important. This is supported by previous figure of hydrogen flux out of the atmosphere.
It should be noted that we usually fall into the area of "realistic values" for temperatures between 500K and 1000K. For a geometrical factor of 2, we need even higher temperatures. This does not agree with the hypothesis from Kasting and Pollack (1983) that temperature at the exobase of Venus should be around 250K to 300K, which are surprising low values. To get results that fit the present data with temperatures of 300K, we need to have a geometrical factor of more than 12 planetary radiuses. This could be the case but it seems quite extreme.
For almost every case we tried, isotopic mixing ratios where of the right order. That is to say they fitted the present values. This however was not true for the Neon which stayed at high abundances no matter what case was chosen.

We also investigated the amount of water lost in each case. It varies from around 3 to 7 "Terrestrial Oceans" (depending on the temperature at the exobase of the atmosphere of Venus) for a geometrical factor of 2 to 10 to 12 for 4 planetary radiuses and to 30 to 40 for a geometrical factor of 8. These quantities seem huge but it was already known that when attempting to explain present isotopic ratios only by Hydrodynamic escape would lead to the loss of many "terrestrial oceans". Maybe an estimation of the amount of water lost by the planet can help to constraint the conditions of the escape of the atmosphere of Venus.





Conclusions and perspectives.

Our purpose is to better understand what kind of mechanism can occur in the atmospheres of other terrestrial planets. Here, with studies of Mars and the late evolution of its CO_2 pressure and Venus and its early hydrodynamic escape, we can better understand what led to the present state of the planets we study.
The simple numerical models we used to compute the late evolution of Mars' atmosphere tends to imply that the present atmosphere we observe on the planet might not be the remains of a primordial one but, on the contrary, only a secondary atmosphere produced mainly (if not entirely) by recent volcanic activity. The atmosphere might indeed be as young as 1Gy. Moreover, it seems that it never was thick and dense during the past 3 Gy, based on the evolution of CO_2 pressures.
Still, we need more data on recent volcanic activity (i.e. during the last 500My or 1Gy) to better constrain the degassing.
The CO_2 content of melted materials is also an important parameter and stronger constraints are needed to be able to get more precise and more significant results.
Once this is done, a geochemical study of the contents of the atmosphere during each scenario and its comparison with available data would allow us to get more constraints on the evolution of the atmosphere. It might be particularly interesting to study Argon and to have a look at isotopic fractionation of Carbon and Nitrogen. Depending on it, we could narrow the window for the age of the atmosphere, assuming that it is as young as this study implies.

The atmospheric escape is the most technical part of what needs to be improved. With our very simple approach, we left aside two main features of atmospheric escape: the cold trop at the boundary between stratosphere and troposphere which would limit the amount of hydrogen available in the upper layers of the atmosphere (it could only get there by diffusion through the cold trap) and the effect of oxygen in the escape (oxygen can be dragged along by hydrogen and would slow the flux toward space).