Evolution of the atmospheres of terrestrial planets coupled with planetary activity.

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Abstract

A parameterized model of mantle convection, including volatile exchanges between the mantle and the atmosphere, is used to study the thermal history of the Earth, Mars and Venus. Different scaling laws have been chosen to model plate tectonics and stagnant lid convection. Stagnant lid convection does not allow regassing through subduction of hydrated crust. The influence of volatiles is taken into account. The exchanges with the atmosphere (degassing as well as regassing) depend on the spreading rate of the ocean seafloor and thus on the mantle heat flow. Additionally, the mean depth of partial melting is taken into account. The model is run for 4.6Gy. Time series of average mantle temperature, viscosity, total degassing and Rayleigh number are calculated for the three planets. Models show a fast degassing early in their history. The effects of the different parameters are studied. Models show realistic present-day values.
A simple atmosphere-interior coupling has been implemented for Venus under stagnant lid convection regime; the atmosphere gains water from the degassing 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.

Inner Dynamics parameterized convection program.

The model.

We want to model in a simple way the evolution of a planet so we can easily study the main features of its history and test many different scenarios. The main output of the model is the mantle mean temperature and the degazing of volatiles (mainly water). The planet is composed of spherical layers representing the core and the mantle.

The convection is supposed to take place over the whole depth of the mantle.
Heating from the mantle is not taken into account: the main source of heat comes from the radiogenic elements present inside the mantle. The model is influenced by the surface temperature. Besides, the mantle exchanges water (both degassing and regassing) and heat (surface heat flux) with the atmosphere of the planet. Loss of water is due to partial melting located at the ridges (plate tectonics) or volcanism (stagnant lid).
We set a simple temperature and water-content-dependent rheology following the work by Karato and Wu (1993):

$$\dot{\varepsilon} = A\left(\frac{\sigma}{\mu}\right)\left(\frac{b}{d}\right)^m \exp\left[-\frac{(E^* + PV^*)}{RT_i}\right]$$

Where $\dot{\varepsilon}$ is the strain rate, T_i the temperature, P the pressure, σ the stress, d the grain size, A the exponential pre-factor, μ the shear modulus, b the length of Burger's vector, E^* the activation energy, V^* the activation volume and R the perfect gaz constant.
Thus the viscosity is $\eta = \frac{\sigma}{2\dot{\varepsilon}}$.

We use a parameterized approach to model the mantle convection in this program. It is based on the use of scaling laws. Scaling laws differ with the convection mode.
For plate tectonics, such as it is seen on Earth, we take: $Nu = \left(\frac{Ra}{Ra_c}\right)^\beta$ with $\beta = 1/3$ (Christensen, 1985). With Nu the nusselt number and Ra the Rayleigh number.
For stagnant lid convection we take: $Nu = \theta^{\omega}\left(\frac{Ra}{Ra_c}\right)^{\phi}$

(Franck-Kamenetskii, 1969 ; Solomatov and Moresi, 1996) where θ depends from the intern energy and the temperature gradient and ω and ϕ are two fixed parameters.

For each time step, the program solves the simple energy equation.

$$\frac{4}{3}\pi\rho c\left(R_m^3 - R_c^3\right)\frac{\partial T}{\partial t} = -4\pi R_m^2 q + \frac{4}{3}\pi Q\left(R_m^3 - R_c^3\right)$$

Where ρ is the mantle density, c the heat capacity at constant pressure, Q the heat source, q the surface heat flux, R_m the mantle radius and R_c the core radius.

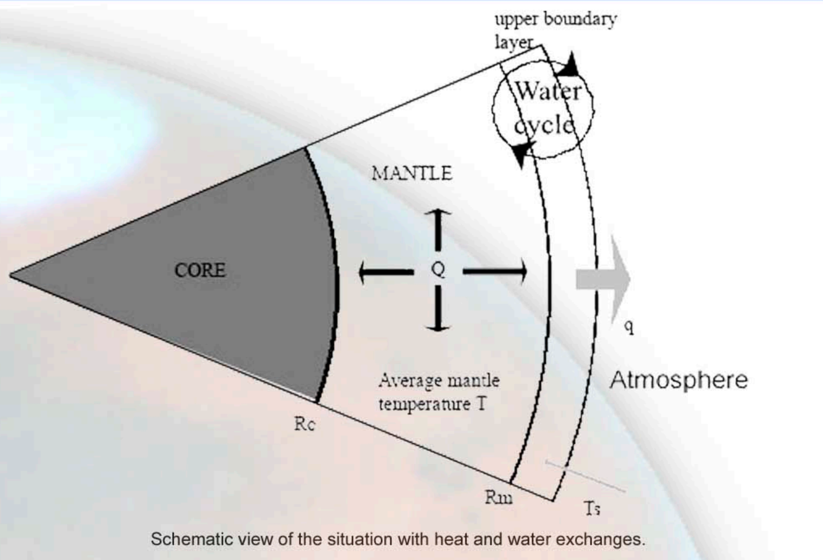
The water exchanges are modelled with a simple approach from (Franck and Bounama, 1995). The water contained in the melted mantle is degassed and a fraction of the water contained in the oceanic crust is reincorporated in the mantle (only when plate tectonics occurs).
It is important to model the amount of mantle material which is melted near the surface. It depends mainly on the speed of the convection near the surface for this speed constrains the amount of mantle going through the "melt zone". This speed is the spreading rate S for the plate tectonics and is given by $q = \frac{S^{1/2}\, 2k(T - T_s)}{[\pi\kappa A_0(t)]^{1/2}}$

according to McGovern and Schubert (1989), with A_0 the total surface of oceanic basins at a given time t.
When using stagnant lid convection, we can't use the same equation which is specifically obtained for a plate tectonics situation. Thus we take this equation:

$$u = 0.05\zeta\frac{\kappa}{d}\theta^{-2n/(n+2)}Ra^{2n/(n+2)}$$

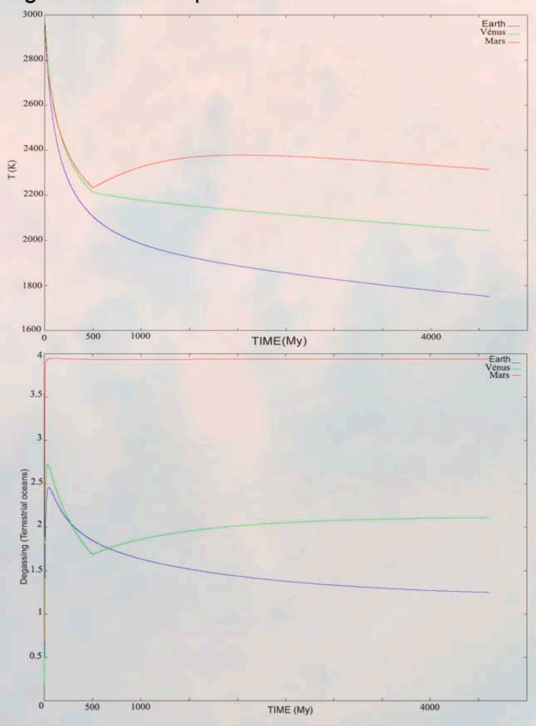
from Reese et al. (1999) for the mantle velocity under the lid. n=3, ζ is related to the pressure-dependent viscosity and the phase transition effects, ζ=0.4.
Tests have been done with this new equation and seem to yield results significantly different from the previous tests.

The program includes the calculation of the melting depth depending on the convection mode. It compares a parameterization of the solidus (Vlaar & Van den Berg,1991 and De Smet et al.,1998) to an adiabatic profile of the mantle temperature of the planet. The difference between plate tectonics and stagnant lid convection (the lid) is taken into account : the lid reduces the melt zone. This is useful to find the quantity of water degassing and might also give insight on the presence of a magma ocean or on the crust production rates.



Schematic view of the situation with heat and water exchanges.

Three planets are investigated: Mars, The Earth and Venus. In most of the cases, degassing occurs right in the beginning of the evolution and is massive, removing efficiently a good part of the water from the mantle both with plate tectonics and stagnant lid convection. However, stagnant lid convection leads to a drier mantle since it prevents water from getting back in the planet (in this model at least). Thus we see that Mars and Venus seem drier than the earth. Moreover, Mars and Venus are hooter than the Earth because plate tectonics removes heat efficiently from the planet's interior whereas with stagnant lid convection, heat can only cross the lid by diffusion which slows the process and leads to higher mantle temperatures.



These graphics show the evolutions of temperature and water content of the mantle of the three terrestrial, Earth, Mars and Venus. Earth uses plate tectonics whereas Mars and Venus begin with plate tectonics but change to Stagnant lid convection after 500 My.
Here the melting depth is not fixed but rather calculated by the program at each time step. It is very thick at first (several hundred kilometers) but decrease rapidly to values of 0 to 20km after 4.5 Gy. The melting depth is greater for Mars than for Venus or the Earth, this is due to the dominating influence of the gravity.
Venus and Mars are hotter and drier than the Earth.

This basic prameterized model for the history of terrestrial planets seems to be able to give good insight on major large scale processes occuring during the evolution of the planets and leading to their current state.
It would be interesting to use it (and refine it depending on the needs) to study the influence of inner dynamics on the evolution of the surfaces of the planets. We want to couple this program with basic atmospherical models to see if retroactions can be found and if wecan explain somehow how it is the Earth evolved so differently from Mars and Venus.
First, however, we need to study the nature of the influence of different processes leading to the release of volatiles in the atmosphere and try to quantify their effects on the atmosphere.

Quantitative study of different events.

Impact-induced volatile production.

One source of volatiles we did not consider in the convection model is the effect of impacts. They melt some of the material present where the crater forms and release what can be high amounts of water vapour and CO2 in the atmosphere. Reese et al. (2004) try to explain the formation of Tharsis with the impacts of big meteorites. We use their works to find out the effects of these impacts on the release of volatiles.
We use a water density in the mantle of de 4.86kg/m^3, based on the assumption that the amount of water received by Mars during its formation was roughly proportional the one received by the Earth. The density of CO2 in the mantle of Mars is not known. We took 1.64kg/m^3 as a possible value, based on the assumption that carbon was abundant in the mantle of Mars due to possible high quantities of S in its core (Kuramoto, 1997) preventing carbon to enter it by diminishing its solubility in liquid iron.
The amount of water released by impacts such as those studied here corresponds to 0.1 to 1 bar of H20 in the atmosphere of Mars. This corresponds to several tens of meters for a global martian ocean. Maybe such events can be used to explain the evolution of the atmosphere of Mars or some of the water figures found on the planet.

	Case 1			Case 2		
Anomaly radius (km)	800			1500		
Impactor radius (km)	470			880		
Transient crater radius (km)	1000			1600		
Complete melt radius (km)	610			1100		
Crater volume ($10^8 km^3$)	3.8			17		
Retained impact melt volume ($10^8 km^3$)	1.5			11		
Interior viscosity (Pa.s)	10^{23}	10^{22}	10^{21}	10^{23}	10^{22}	10^{21}
Melting duration (Ga)	0.1	0.1	0.1	0.1	0.3	1
Decompression melt volume ($10^8 km^3$)	0.084	0.50	1.9	0.26	3.1	9.8
Mass of degassed water ($10^{17} kg$)	7.71	9.72	16.52	54.72	68.52	101.09
Amount of degassed water (Terrestrial Oceans)	0.00055	0.0007	0.0012	0.0039	0.0049	0.0073
Mass of degassed CO2 ($10^{17} kg$)	2.6	3.28	5.58	18.47	23.12	34.11

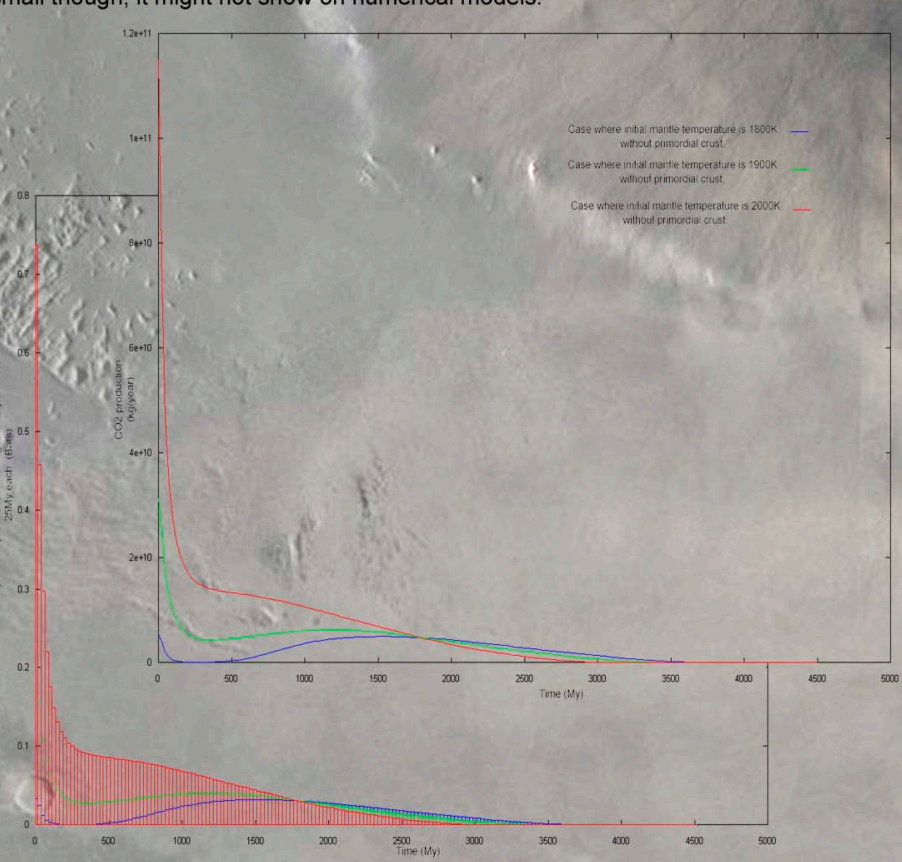
Table : Results from Reese et al., 2004 about impact induced melting. Our own calculations of corresponding volatile production have been added.

Planetary activities.

One of the problems that retained our attention was the question of the equilibrium of CO2 cycles on Mars and the Earth which is directly linked to their activities. On Earth, direct measures are not possible due to the presence of high quantities of CO2 in the atmosphere and ocean (Coltice et al., 2004). It is not possible either to affirm CO2 cycle on Earth is at steady state even if we assume it is not far from it (Javoy et al., 1982). On Mars (on the contrary to the Earth) there is atmospheric escape at the present day. Since there is still several meters CO2 ice left on the south pole, it can be hypothesised that CO2 ice sublimation compensates for this atmospheric escape. This case however seems quite surprising for we would be witnessing the "last days" of CO2 ice on Mars: given the relatively small amounts of CO2 ice present, the process cannot last long. Maybe there are other processes that could replenish the CO2 lost into space.
We tried to compare the efficiencies of atmosphere replenishment. On Mars, at present time, it takes around 200My to escape 10mb of atmosphere. Earth is not subject to atmospheric escape but another mechanism exists. CO2 precipitates and is extracted from the atmosphere then carbonates are trapped in subducting oceanic crust (Javoy et al., 1982). This process extracts around 1bar every 50My. A quick calculation gives us the ratio of the speeds of extraction: 1/400. If we assume the atmosphere of Mars and the Earth are near steady state (that they do not move from this state over geological timescales), it comes that the activity of Mars should be around 400 times less important than what occurs on the Earth. Considering Earth's crust production to be around 20km^3/year (we can compare to the activity of Venus: 0.4km^3/year according to Nimmo and McKenzie, 1997), we find that Mars' crust production should be around 0.05km^3/year (around 10 times less than Venus). This number is really small but seems to be in agreement with numerical studies such as those from Breuer and Spohn (2003 and 2006) that indicates very small mean crustal production over the last 500My. These considerations are valid if the density of CO2 in the mantles of Mars and the Earth are of the same order. If it differs widely, this must be taken into account. For exemple, if Mars' mantle if richer in carbon, it can be even less active and still be able to sustain its present atmosphere.
If we consider a reasonable sized caldera of about 100 meters deep and a radius of 10km, we get a melt volume of around 30km^3 .With the numbers used above, this caldera can sustain the atmosphere for about 600years.
We did the same for an impact crater the size of Argyre Basin (600km wide). Using equations proposed by Schmidt and Housen (1987) and Melosh (1989), we find a melt volume of around $10^6 km^3$ which is enough (using the same calculations as above) to sustain the atmosphere for 10My for present conditions (at the time when Argyre was formed, escape was more vigorous).

Influence of Crustal production.

Based on work by Breuer and Spohn (2006) we tried to calculate an estimation of the contribution of crustal production to the atmosphere within the framework of numerical models. We have calculated the annual production of CO2 due to crustal production and the CO2 pressure in the atmosphere over periods of 25My. The Order of these results is about the same as those presented above. However in this calculation no more crust is produced in the most recent times. Since the amount needed is very small though, it might not show on numerical models.



Tharsis.

Tharsis is a major feature of Mars' surface. Many people have theorized about it's formation whether by assuming a simple hot plume rising from the core/mantle boundary (e.g. Kiefer 2003), a passive upwelling driven by downwellings or even major impacts (Reese et al. 2004). We have just started a project to model the possible creation of Tharsis by a hot plume using a 3D-numerical model. But it is also interesting to study the implications of Tharsis existence and activity.

It has been proposed that Tharsis played a major role in the climate of Mars either due to its mass and influence on obliquity or by the amount of volatiles released into the atmosphere during its creation. We study the second hypothesis. It is important to look at some of the implications of the presence of such a volcanic feature. Keeping in mind what we discussed above about planetary activity, it is also good to keep in mind that signs of recent activity have been observed (Hartmann et al. 1999) only 40 to 100 million years ago.

A rough estimation of the volume of Tharsis gives a result around 300 millions km. When using the same numbers as in the previous section, we find it means about 10 meters water in a global martian ocean. We can make the same kind of calculation for the CO2 budget and we find (again with the values previously used) 0.1 bars of CO2. Given that the numbers are subject to debate, it can be argued that it is reasonable to expect values between these results and one more order. That is to say Tharsis would be a source of 10 to 120m water and 0.1 to 1.5 Bars CO2.

We can look at another aspect of Tharsis: the production of SO2. If we take the Earth-like value of around 122 Tons of SO2 for 15km (eruption of the Laki), we get 2.4 millions GTons of SO2, that is to say 12 Tons/m .We can calculate the pH of the water produced by Tharsis under these conditions. We simply find that the pH should thus be between around 1 and 2.5.
What we need to address is the question of the availability of SO2: we have to compare the timescales of the reactions of SO2 with the surface and of the formation of Tharsis. This might decrease the amount of SO2 available to "acidify" water and might lead to high pH. What would be also interesting is to compare this fast calculation to the conditions of formation of sulphates that seem to have been found on the surface of Mars with the OMEGA spectrometer (Bibring et al., 2005).

Atmosphere : Coupling and Escape.

Basic coupling.

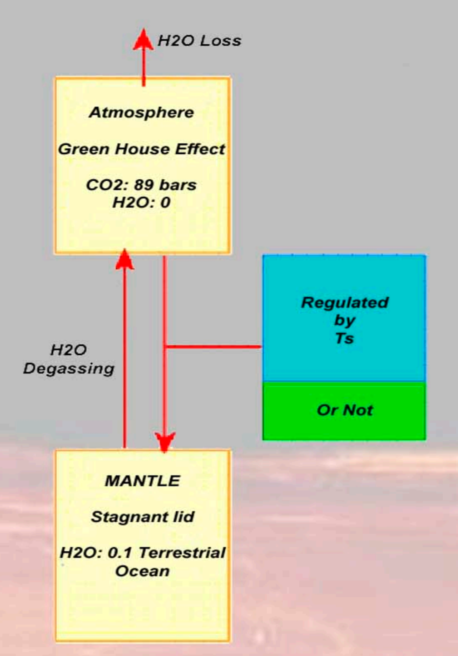
We first used a simple convective-radiative atmosphere model based on a work by Phillips et al (2001) in the case of Venus. The convection parameterized model provides the amount of volatiles released by the mantle into the atmosphere. The atmosphere model calculates the green house effect and gives 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 H2O 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.

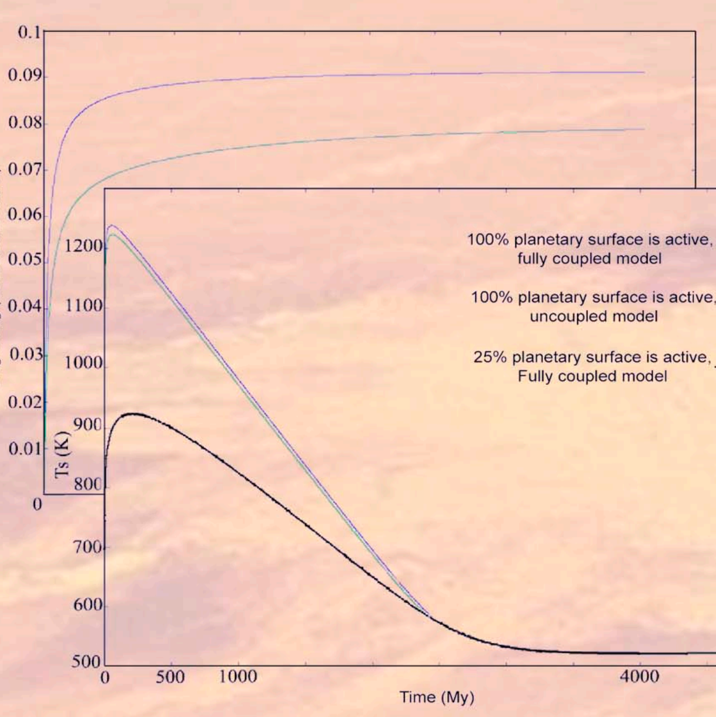


Figure showing the surface temperature and degassing of Venus for coupled (blue) and uncoupled (green) models and the influence of the active surface of the planet (black , 25% of surface is active). Initial water content is 0.1 Terrestrial Ocean, Initial mantle temperature is 2500K.

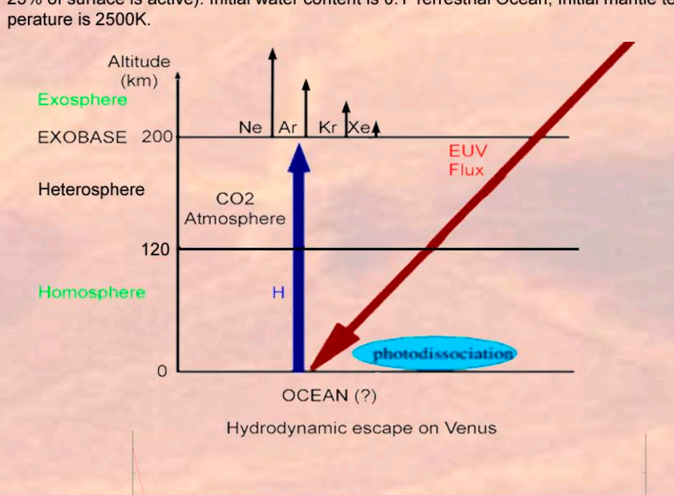
Atmospheric escape.

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).

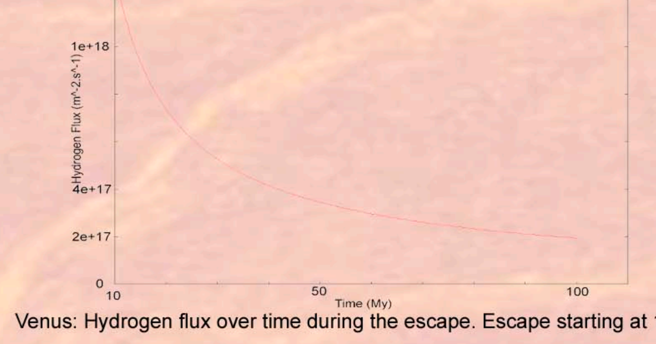
The model is energy-limited. The EUV Flux on top of the atmosphere gives the energy needed to create a thermal escape flux which will drag along elements and molecules whose mass is lower than the critical mass m_c.

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

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



Venus: Hydrogen flux over time during the escape. Escape starting at 10 My.

Data.

The Venera and Pioneer missions have provided us with some data about noble gases on Venus, however it is still scarce and limited. We use some of these to get some constraints on our models.
We'll use mainly the isotopic ratios of Neon (isotopes 20 and 22) and Argon (Isotopes 36 and 38). All models whose results are in the domain obtained thanks to these measures are the most realistic cases for they lead to present observations. Present data on noble gases abundance also enable us to find an estimation of primordial abundances of these gases in the atmosphere of Venus. Xenon is taken as a reference for it is the most heavy noble gas we have information on so it seems the least susceptible to have escaped or been subject to major changes. The initial composition of the atmosphere of Venus we have calculated is closer to solar composition.

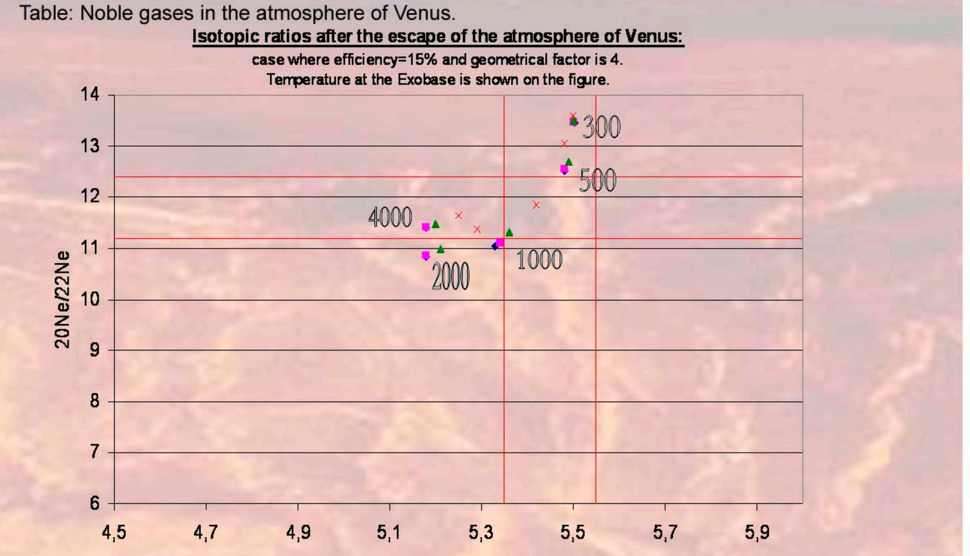
Non-Radiogenic Isotopes, Mixing Ratios.		
20Ne	4.3-13 ppm	Donahue and Russell, 1997
36Ar	20-50 ppm	Donahue and Russell, 1997
84Kr	7-38 ppb	Donahue and Russell, 1997
132Xe	< 10 ppb	Donahue and Russell, 1997
Isotopic Ratios		
20Ne/22Ne	11.2-12.4	Istomin et al., 1983
21Ne/22Ne	<0.067	Istomin et al., 1983
36Ar/28Ar	5.35-5.55	Istomin et al., 1983
40Ar/36Ar	1.09-1.13	Istomin et al., 1983

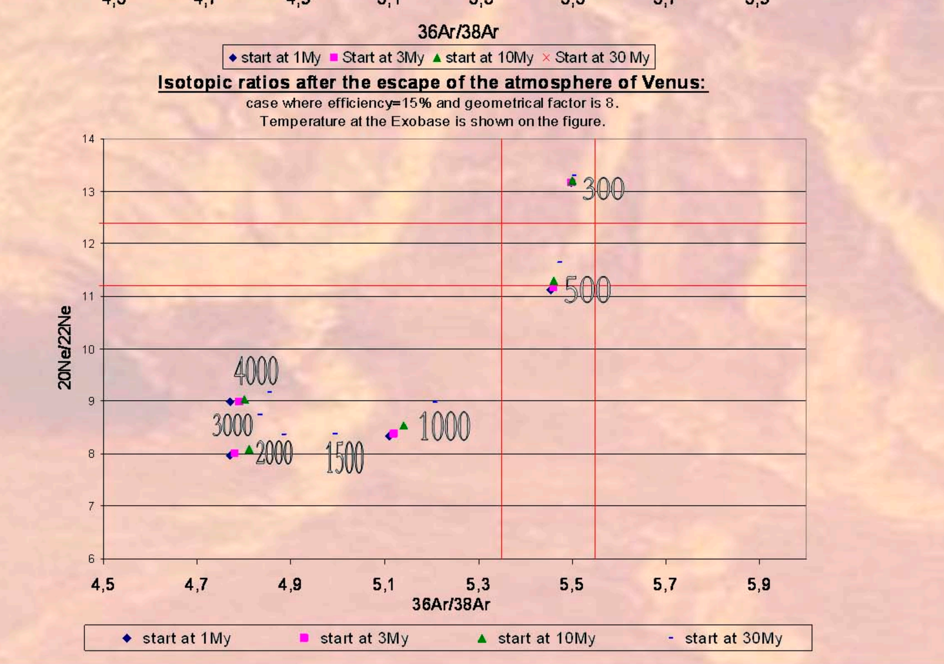
Table: Noble gases in the atmosphere of Venus.

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.





Conclusion/perspectives.

Our aim is to provide a broad overview of the different conditions that might be necessary to make a planet reach a state where it can sustain life on the long term. Thus we investigate links between the atmosphere and the mantle of the most known planets: Mars, Venus and the Earth. We hope to be able to distinguish major actors of the evolution of planets and how process can lead to the present state of terrestrial planets we study.
This work shows a first attempt to quantify the effects of some of the processes that are thought to be dominant. It shows the strong links between different layers both solid and fluid, of the terrestrial planets and the necessity to take these links into account when studying them.

this is but the start of the study and many things can be improved in the modelling. Still it looks promising.
First, the convection model should be tweaked to yield crustal production rates, which could be compared to those we used here. It could also be compared to volcanic activity. Concerning the problem of activity on Mars, it might be useful to try and set up an history of degassing based on the age of the surface of the planet. It could be a nice way to have a good estimation of the volatile input for the atmosphere.
Another way that need further investigation is the question of the Sulphates discovered on Mars and their possible link with Tharsis and the pH of the water outgassed by Tharsis' formation. Maybe we can find a way to explain the presence of these Sulphates. But even if it can't be explained this way, we would get some more constraints that would prove useful for the rest of the study.
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).

Finally, We are working on a 3D-model to investigate what happened in the Tharsis region. This is not clearly linked to the work presented here but will clearly give insight on some interesting aspects presented here, such as the volatile input by Tharsis.