History and Escape of the Martian Atmosphere

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Talk based on the paper by Lammer et al. 2012 (submitted to SSR)

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Initial water inventory of growing planet is due to impacts of asteroids and comets

Earth: Water is product of collisions between proto-earth and embryos
Mars: Water added during late stage of its formation

Water delivery based on estimated cumulative collision of asteroids and comets from beyond 2.5 AU with Mars (Lunine et al. 2003):

6 – 27% of Earth ocean (equivalent to 10 – 100 bar)
Thick atmosphere can be created very fast (<10^5 yr) through solidification of a magma ocean with a low initial volatile content

0.05 wt% H2O and 0.01 wt% CO2 can release (initial depth of magma ocean 500 or 2000 km) (Elkins-Tanton 2008)

30 – 70 bar (H2O + CO2)

Surface temperature might have been up to 1500 K due to thermal blanketing and frequent impacts
Early Sun was probably much stronger at shorter wavelengths than today

Saturated EUV flux duration $\sim$100 Myr

Solar nebula evaporation time scale $\sim$10 Myr

Formation time scale of terrestrial protoplanets $< 100$ Myr

Primordial atmosphere was exposed to high EUV radiation for some 10 Myr
Loss of Early Steam Atmosphere

Hydrodynamic blow-off conditions during EUV saturation period

Fast dissociation of molecular species (Tian 2009)

Rapid loss of dense proto-atmosphere

The MEX/OMEGA observations are consistent with an early escape of most of the Mars atmosphere (Bibring et al. 2005)

Early steam proto-atmosphere may have been lost within ~10 Myr
Volcanic Outgassing

Period where outgassing was balanced by escape to space

$$\frac{dM_{i}^{\text{atm}}}{dt} \sim \frac{dM_{i}^{\text{cr}}}{dt} X_{i}^{\text{melt}} \eta_i$$

If martian magmatic CO2 content is comparable to terrestrial content
Hawaiian basalt: $X_{CO2}^{\text{melt}} \sim 0.65$ wt%

Outgassing associated alone with formation of Tharsis bulge: $\sim 1.5$ bar CO2
(Phillips et al. 2001)

CO2 content of martian magma mainly depends on O fugacity, which
was probably overestimated by previous models

For plausible range of P-T conditions, CO2 content is $0.01 - 0.1$ wt% 
(Hirschmann and Withers 2008)

0.04 – 1.5 bar CO2 outgassed since $\sim 3.8$ Gyr
Volcanic outgassing of CO2 and H2O together with volatiles possibly delivered by later impacts are the main sources of volatiles for the secondary martian atmosphere.

It is not clear when the outgassing flux from Mars interior exceeded the expected escape flux so that a secondary CO2 atmosphere could have built up.

By combining a chemical model for CO2 solubility with a parameterized thermal evolution model, the amount of outgassed CO2 can be calculated in a self-consistent way.
Two models:
- global melt channel
- melting in localized mantle plumes

Amount of outgassed CO₂ probably insufficient for a appreciable greenhouse effect

Pressures are however sufficient to allow for transient liquid water on the surface

[Lammer et al., submitted to SSR 2012]
Bulk concentration of water in mantle: 100 ppm

Outgassing efficiency: 0.4

Additional water may have been delivered by comets until the late Noachian or during the LHB

[Lammer et al., submitted to SSR 2012]
Could the martian atmosphere have been eroded by impacts since the late Noachian?

- Simulations by hydrocodes (solving EoM and EoS)
  Difficulties: choice of an appropriate EoS; a proper model of the vapor cloud dynamics; expensive in time and calculations

- Analytical model („tangent plane model“, Melosh and Vickery 1989)
  Vapour plumes created by high speed impacts may eject the entire airmass above the plane tangent to the point of impact

\[
\frac{dM_{\text{atm}}}{dt} = \frac{dM_{\text{del}}}{dt} - \frac{dM_{\text{esc}}}{dt}
\]

\[
\frac{dM_{\text{esc}}}{dt} = \frac{\partial N_{\text{cum}} (> m_{\text{crit}}(t), t)}{\partial t} 4\pi R^2 m_{\text{tan}}(t) f_{\text{vel}} f_{\text{obl}} = k f_{\text{vel}} f_{\text{obl}}
\]

\[
\frac{dM_{\text{del}}}{dt} = k n \left[ \frac{b}{1-b} y_{\text{imp}} f_{\text{vap}} + y_{\text{imp}} (1 - f_{\text{vel}} f_{\text{obl}}) g_{\text{vap}} \right]
\]

\[
n = \frac{m_{\text{crit}}(t)}{m_{\text{tan}}(t)}, \quad m_{\text{tan}} = \frac{m_{\text{atm}} H}{2R}
\]

[Pham et al. 2011]
Atmospheric Mass Evolution

Effects of impactor velocity, vaporization and impact angle included

Original model

Different hydrocode simulations

\[ 10 < n < 2400 \]

Exponentially decaying impactor flux

<table>
<thead>
<tr>
<th>Factor</th>
<th>Asteroids</th>
<th>Comets</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>SP comets</td>
<td>LP comets</td>
</tr>
<tr>
<td>( f_{vel} )</td>
<td>0.08</td>
<td>0.83</td>
</tr>
<tr>
<td>( f_{obi} )</td>
<td>7.54</td>
<td>2.16</td>
</tr>
<tr>
<td>( y_{imp} )</td>
<td>0.01</td>
<td>0.3</td>
</tr>
<tr>
<td>( f_{vap} )</td>
<td>0.34</td>
<td>1</td>
</tr>
<tr>
<td>( g_{vap} )</td>
<td>0.21</td>
<td>1</td>
</tr>
</tbody>
</table>
Effect of LHB

Best erosion scenario

Delivery scenario

Hydrocode simulation suggests $n > 40$ for LHB (Svetsov 2007)

With impactors having 1 wt% of CO2, total delivery of CO2 could be ~300 mbar and up to several bar (~100 m GEL) of H2O

Large impacts during LHB may have triggered transient wet and warm conditions on the martian surface

Size of impactors during LHB for optimum erosion case

[Lammer et al., submitted to SSR 2012]
Atmospheric Loss Processes

Impact erosion (maybe relevant on early Mars)
→ light and heavy species (H, O, N) are eroded from the atmosphere hit by asteroids or comets

Hydrodynamic expansion and thermal (Jeans) outflow (relevant on early Mars)
→ the EUV flux of the young Sun heats the upper atmosphere, leading to hydrodynamic expansion and escape

Photo-chemical reactions (relevant on Mars all the history)
→ light and heavy species (H, O, C, N, etc.)
→ light species (H, H\textsubscript{2}, D)

Ion pick up (relevant if magnetosphere is not shielding)
→ plasma flow

Sputtering (non-magnetized → reduced on early Mars)
→ light and heavy species of the upper atmosphere can be sputtered by solar wind plasma if the planet has no or a weak magnetic field

Plasma instabilities (non-magnetized → reduced on early Mars)
→ all ion species at the magnetopause/ionopause-transition layer

Cool ion outflow (non-magnetized → reduced on early Mars)
→ light and heavy ions with energies larger than the escape energy
Detached vortices can carry away ionospheric particles (Brace et al. 1982)

Martian ionopause appears rather stable with regard to KH (Möstl et al. 2011)

⇒ Loss is not significant

Fig. 12 Nonlinear evolution of the Kelvin-Helmholtz instability. The time series of the mass density is shown, from an MHD simulation with periodic boundary conditions in the x-direction. The mass density changes from the upper to the lower plasma layer and exhibits an increase of up to ten times (see the color code: blue: low density, red: high density). In the upper layer, the plasma flows from left to right. In the lower layer, the plasma is at rest. Initially small perturbations of the boundary layer separating the two plasma layers evolve into a KH vortex.
Ion Loss Rates

[Barabash et al 2007]

The biggest puzzles in martian planetology. We have measured the current loss rate due to the solar wind interaction for different species: $Q(O^+) = 1.6 \cdot 10^{23}$ per second = 4 grams per second (g s$^{-1}$), $Q(O_2^+) = 1.5 \cdot 10^{23}$ s$^{-1} = 8$ g s$^{-1}$, and $Q(CO_2^+) = 8 \cdot 10^{22}$ s$^{-1} = 6$ g s$^{-1}$ in the energy range of 30 to 30,000 electron volts per charge. These rates can be propagated backward over a period of 3.5 billion years, resulting in the total removal of 0.2 to 4 millibar of carbon dioxide and a few centimeters of water. The escape rate is low, and thus one has to continue searching for water.

[Manning et al 2011]

We develop a parametric fit to the results of a detailed magnetohydrodynamic (MHD) study of the response of ion escape rates ($O^+$, $O_2^+$ and $CO_2^+$) to strongly varied solar forcing factors, as a way to efficiently extend the fits with ion mass. We integrate the $CO_2^+$ and oxygen ion escape rates over time, and find that in the last 3.85 Gyr, Mars would have lost about $25^{+85}_{-19}$ mbars of $CO_2$ and $0.64^{+0.62}_{-0.34}$ meters of water (from $O^+$ and $O_2^+$) from ion escape.
given for three different ratios of atoms to molecules at the exobase and are used to obtain a better estimate of the total loss of atmosphere due to the pickup ion sputtering. Using the pickup ion fluxes of Zhang et al. [1993], \( \sim 120 \) mbar of O and \( \sim 60 \) mbar of CO\(_2\) are lost. The loss of O if associated with H\(_2\)O would be equivalent to \( \sim 4 \) m averaged over the Mars surface.
Extrapolation of present values to earlier epochs by assuming an exponential dependency of escape rates with time
C\(^+\) Pick Up Escape Rates

Subsolar magnetopause standoff distance related to a martian magnetic moment decreasing from 0.1 \(M_E\) to 0.01 \(M_E\) during the first Gyr for three different SW densities.

Loss rates:

- \(3\) EUV (2.5 Gyr ago):
  \(~1 \times 10^{26} \text{ s}^{-1}\)
- \(10\) EUV (3.9 Gyr ago):
  \(~3 \times 10^{26} \text{ s}^{-1}\)
- \(20\) EUV (4.2 Gyr ago):
  \(~5 \times 10^{30} \text{ s}^{-1}\)

[Tian et al. GRL, 2009]
Sources of hot particles

O: Dissociative recombination of \( \text{O}_2^+ \)
\[
\text{O}_2^+ + e \rightarrow \text{O}(^3P,^1D) + \text{O}(^3P,^1D,^1S) + \Delta E
\]

C: Dissociative recombination of \( \text{CO}^+ \)
\[
\text{CO}^+ + e \rightarrow \text{C}(^3P,^1D) + \text{O}(^3P,^1D,^1S) + \Delta E
\]

Photodissociation of \( \text{CO} \)
\[
\text{CO} + h\nu \rightarrow \text{C}(^3P) + \text{O}(^3P)
\]

N: Dissociative recombination of \( \text{N}_2^+ \)
\[
\text{N}_2^+ + e \rightarrow \text{N}(^4S,^2D) + \text{N}(^4S,^2D,^2P) + \Delta E
\]

Dissociative recombination of \( \text{NO}^+ \)
\[
\text{NO}^+ + e \rightarrow \text{N}(^4S,^2D) + \text{O}(^3P,^1D) + \Delta E
\]

Photodissociation of \( \text{N}_2 \)
\[
\text{N}_2 + h\nu \rightarrow \text{N}(^2D,^2P) + \text{N}(^2D,^2P)
\]
Table 1. Used Total and Differential Collision Cross Sections, $\sigma$ and $d\sigma$, for Elastic, Inelastic, and Quenching Collisions as well as the Energy Loss Depending on the Colliding Partner

<table>
<thead>
<tr>
<th>Hot Particle</th>
<th>Collision Type</th>
<th>$\sigma$</th>
<th>$d\sigma$</th>
<th>Collision Partner</th>
<th>$\sigma$</th>
<th>$d\sigma$</th>
<th>$E_{\text{loss}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>O($^3P$)</td>
<td>elastic</td>
<td>$Tully$ and $Johnson$ [2001]</td>
<td>1–10 eV (Figure 4)</td>
<td>$Balakrishnan$ et al. [1998a]</td>
<td>0–3 eV (Figure 1)</td>
<td>not available</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>$Kharchenko$ et al. [2000]</td>
<td>0–5 eV (Figure 2)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>inelastic</td>
<td>-</td>
<td></td>
<td>$Balakrishnan$ et al. [1998a]</td>
<td>0–3 eV (Figure 2)</td>
<td>0–3 eV (Figure 8)</td>
<td></td>
</tr>
<tr>
<td>O($^1D$)</td>
<td>elastic</td>
<td>like $O(^3P)$</td>
<td></td>
<td>$Balakrishnan$ et al. [1999]</td>
<td>0–2 eV (Figure 1)</td>
<td>not available</td>
<td></td>
</tr>
<tr>
<td></td>
<td>inelastic</td>
<td>-</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>quenching</td>
<td>$Yee$ et al. [1990]</td>
<td>0.01–10 eV (Figure 4)</td>
<td>$Matsumi$ and Sarwaruddin Chowdhury [1996]</td>
<td>0–0.9 eV (Figure 9)</td>
<td>$Zahr$ et al. [1975], $Tully$ [1974]</td>
<td></td>
</tr>
<tr>
<td>O($^1S$)</td>
<td>elastic</td>
<td>$Yee$ and $Dalgarino$ [1985]</td>
<td>0–5 eV (Figure 4)</td>
<td>like $O(^1D)$</td>
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<td></td>
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<td></td>
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<td></td>
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</tbody>
</table>

[Gröller et al. 2010]
Figure 2: Energy distribution function (EDF) for high (solid lines) and low (dashed lines) solar activities at 240 km altitude. Panel (a) shows the EDF for hot O and hot C atoms and panel (b) for hot N atoms.
Figure 3: Exosphere densities for high (solid lines) and low (dashed lines) solar activities. Panel (a) shows the densities for hot O and hot C atoms and panel (b) for hot N atoms.
Cumulated escape of C due to DR of CO\(^+\) based on model of Tian et al. 2009

<table>
<thead>
<tr>
<th>Time [Gyr ago]</th>
<th>Surface Pressure [mbar]</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>7</td>
</tr>
<tr>
<td>2.5</td>
<td>26</td>
</tr>
<tr>
<td>3.7</td>
<td>48</td>
</tr>
<tr>
<td>3.9</td>
<td>56</td>
</tr>
</tbody>
</table>
**Table 3** Estimated min. and max. outgassed (VO: volcanic outgassing) CO$_2$ in units of bar ~4 Gyr ago and the expected min. and max. range of impact eroded (IE) or delivered (ID) atmosphere during the late heavy bombardment (LHB) period. CO$_2$ escape of various atmospheric loss processes (IL: ion loss; KH: Kelvin Helmholtz instability triggered ionospheric detached plasma clouds; SP: sputtering; DR: dissociative recombination; PD: photo dissociation) from observations and models integrated since that time.

<table>
<thead>
<tr>
<th>Sources and loss processes</th>
<th>CO$_2$ [bar]</th>
</tr>
</thead>
<tbody>
<tr>
<td>VO ~4 Gyr ago, max.</td>
<td>~0.2–0.5 bar</td>
</tr>
<tr>
<td>VO ~4 Gyr ago, min.</td>
<td>~0.05 bar</td>
</tr>
<tr>
<td>ID ~ LHB</td>
<td>~0–0.3 bar</td>
</tr>
<tr>
<td>IE ~ LHB</td>
<td>~0–0.15 bar</td>
</tr>
<tr>
<td>IL since ~4 Gyr ago</td>
<td>~0.001–0.1 bar</td>
</tr>
<tr>
<td>KH since ~4 Gyr ago</td>
<td>~0.001 bar</td>
</tr>
<tr>
<td>SP since ~4 Gyr ago</td>
<td>~0.001–0.1 bar</td>
</tr>
<tr>
<td>DR since ~4 Gyr ago</td>
<td>≤0.005–0.05 bar</td>
</tr>
<tr>
<td>PD since ~4 Gyr ago</td>
<td>≤0.005–0.05 bar</td>
</tr>
</tbody>
</table>

[Lammer et al., submitted to SSR 2012]
Martian Atmosphere Evolution

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