Dynamics and thermal evolution of terrestrial planets

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Dynamic Planets

- Volcanic and tectonic history
- Magnetic field
- Atmosphere Evolution

- Relate observations to interior dynamics models
Mars
Crustal dichotomy

- Variation in age
- Crustal thickness variation (e.g., Zuber et al., 2000)
- Variation in surface composition (e.g., Christensen et al., 2000)
Volcanic activity

- Strong decrease of volcanic activity with time; first global distribution later concentration in two provinces
- Crustal dichotomy formed in the first few 100 Ma
- Bulk of the large volcanic provinces formed in first 1 Ga
Episodic volcanism but also recent activity in Tharsis and Elysium

Neukum et al, 2004
Venus

- Volcanoes and volcanic lava flows are homogenously distributed at the surface.

- Possibly a recent global resurfacing event (~500-700 Ma) renewed the surface.
Venus: crustal thickness
Mercury

- Crater and scarps cover the surface
  - old surface
  - ~1-3 km decrease of radius caused by thermal contraction since ~4 Ga
Mercury

- Messenger flybys revealed volcanic resurfacing

- Volcanism more widespread than previously expected
Moon

Crustal dichotomy: lunar mare are primarily found on near side
Moon

- Increase of TiO$_2$ with age suggests that source region moved to greater depth.

- Volcanic activity extended, albeit at a small rate, until perhaps 1.5 Ga b.p.
Concept of thermal convection

Thermal buoyancy

\[ \rho = \rho_0 (1 - \alpha T) \]

\[ \alpha = 2 \times 10^{-5} \text{ W/mK} \text{ (thermal expansivity)} \]
Thermal convection (heated from below)

\[ \frac{dT}{dz} \bigg|_{ad} = \frac{\alpha g T}{C_p} \]
Thermal convection (heated from within)

T_0

\[ \text{temperature profile} \]

T_0 \quad \text{T}_1
Do all the terrestrial planets have present-day convection?

A measure for the strength of mantle convection is the Rayleigh number

\[
Ra = \frac{\alpha \rho g \Delta T d^3}{\chi \eta (T,P)} \quad Ra_q = \frac{\alpha \rho g Q d^5}{k \chi \eta (T,P)}
\]

- \(\alpha\): thermal expansivity
- \(g\): gravity
- \(\rho\): mantle density
- \(\Delta T\): temperature difference
- \(d\): mantle thickness
- \(\kappa\): thermal diffusivity
- \(\eta\): mantle viscosity
- \(k\): thermal conductivity
- \(Q\): heat production rate

Critical Rayleigh number
\(~ 10^3 - 10^4\) for onset of convection
Mantle viscosity

\[ \eta = C_1 \exp\left( \frac{E + pV}{RT_m} \right) \quad \text{Newtonian} \]

\[ \eta = C_2 \left( \frac{G}{\sigma} \right)^{n-1} \exp\left( \frac{E + pV}{RT_m} \right) \quad \text{Non-Newtonian} \]

typical values for E and V

\[ E = 300 - 540 \text{ kJ/mol} \]
\[ V = 2 \cdot 10^{-6} - 2 \cdot 10^{-5} \text{ m}^3/\text{mol} \]
# Planetary data

<table>
<thead>
<tr>
<th></th>
<th>Mercury</th>
<th>Venus</th>
<th>Earth</th>
<th>Moon</th>
<th>Mars</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Radius [km]</strong></td>
<td>2439</td>
<td>6052</td>
<td>6378</td>
<td>1737</td>
<td>3394</td>
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<tr>
<td><strong>Mass</strong></td>
<td>0.055</td>
<td>0.815</td>
<td>1.0</td>
<td>0.012</td>
<td>0.107</td>
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<tr>
<td><strong>Density [kg/m³]</strong></td>
<td>5430.</td>
<td>5250.</td>
<td>5515.</td>
<td>3340</td>
<td>3940.</td>
</tr>
<tr>
<td><strong>uncompressed</strong></td>
<td>5300.</td>
<td>4000.</td>
<td>4100.</td>
<td>3400</td>
<td>3800.</td>
</tr>
<tr>
<td><strong>Mol</strong></td>
<td>-</td>
<td>-</td>
<td>0.3335</td>
<td>0.3905</td>
<td>0.3662</td>
</tr>
<tr>
<td><strong>Core Radius/Planet Radius</strong></td>
<td>0.8</td>
<td>0.55</td>
<td>0.546</td>
<td>0.25</td>
<td>0.42 - 0.53</td>
</tr>
<tr>
<td><strong>Rayleigh number</strong></td>
<td>cond. $10^5$</td>
<td>$10^7-10^8$</td>
<td>$10^7-10^8$</td>
<td>$5\cdot10^4 - 10^5$</td>
<td>$10^6-10^7$</td>
</tr>
</tbody>
</table>
Heat transport by convection

- Effectivity of heat transport is measured by the Nusselt number

\[
Nu = \frac{q_{\text{cond}} + q_{\text{con}}}{q_{\text{con}}}
\]

- Nusselt number is proportional to the Rayleigh number

\[
Nu = aRa^\beta
\]
Mantle dynamics

- Mantle material flows like a viscous fluid for high temperature and stress
- Convection transports heat: the driving force is the thermal bouyancy due to the change of density with temperature
- Viscosity is strongly temperature dependent
Thermostat effect

- Strong temperature dependence of the viscosity
  - Low temperature $\rightarrow$ high viscosity $\rightarrow$ low Ra $\rightarrow$ inefficient cooling (low Nu) $\rightarrow$ temperature increase
  - High temperature $\rightarrow$ low viscosity $\rightarrow$ high Ra $\rightarrow$ efficient cooling (high Nu) $\rightarrow$ temperature decrease

![Graph showing temperature over time with two curves, one for low and one for high temperature, indicating temperature decrease over time for both.]
Heat Sources

- **Primordial Energy**
  - Energy of accretion
  - Energy of differentiation (core formation)

- **Radiogenic Heat**
  - Decay of $^{238}\text{U}$, $^{235}\text{U}$, $^{232}\text{Th}$ and $^{40}\text{K}$
**Accretion**
Gravitational potential energy is converted to kinetic energy

Kinetic energy is converted to thermal energy

**Differentiation**
Light material rises to the surface

Dense material falls to the core converting gravitational potential energy to thermal energy

**Radioactive Decay**
Post-accretional temperature distribution

\[ T_a(r) = h \frac{GM(r)}{C_p r} \left(1 + \frac{ru^2}{2GM(r)}\right) + T_e + \Delta T_{a,ad} \]
Post-core formation temperature
Accretion and Core Formation

- Isotope data \(^{182}\text{Hf}^{182}\text{W}\) suggests early and rapid core formation
  - Earth < 60 Ma
  - Mars < 20 Ma

What are the initial thermal conditions after core formation?
## Radioactive heat sources in the primitive mantle

<table>
<thead>
<tr>
<th>Planet</th>
<th>Concentration</th>
<th>$H_0$ [pW/kg]</th>
<th>$H_\rho$ [pW/kg]</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>U [ppb]</td>
<td>Th [ppb]</td>
<td>K [ppm]</td>
</tr>
<tr>
<td>Mercury</td>
<td>30</td>
<td>120</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>8</td>
<td>30</td>
<td>550</td>
</tr>
<tr>
<td></td>
<td>0</td>
<td>400</td>
<td>0</td>
</tr>
<tr>
<td>Venus</td>
<td>22</td>
<td>79</td>
<td>220</td>
</tr>
<tr>
<td>Mars</td>
<td>16</td>
<td>64</td>
<td>160</td>
</tr>
<tr>
<td></td>
<td>28</td>
<td>101</td>
<td>62</td>
</tr>
<tr>
<td></td>
<td>16</td>
<td>56</td>
<td>305</td>
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<tr>
<td></td>
<td>16</td>
<td>55</td>
<td>920</td>
</tr>
<tr>
<td>Earth</td>
<td>20.3</td>
<td>79.5</td>
<td>240</td>
</tr>
<tr>
<td></td>
<td>22</td>
<td>78.2</td>
<td>232.4</td>
</tr>
<tr>
<td></td>
<td>21</td>
<td>84.1</td>
<td>240</td>
</tr>
<tr>
<td></td>
<td>18</td>
<td>64</td>
<td>180</td>
</tr>
<tr>
<td>Moon</td>
<td>60.9</td>
<td>223</td>
<td>178</td>
</tr>
<tr>
<td></td>
<td>33</td>
<td>125</td>
<td>83</td>
</tr>
<tr>
<td></td>
<td>62.8</td>
<td>224</td>
<td>102</td>
</tr>
</tbody>
</table>
Radioactive heat sources

- It is generally assumed that there are no heat sources in the core

- Urey ratio =
  heat generated by radioactive heat sources / surface heat flow

- U is around 0.45 for the Earth
Melt segregation results in a redistribution of radioactive elements

- Radioactive elements are incompatible and enriched in the melt
- Basaltic melt is enriched by a factor of 5 relative to the mantle material
- Depletion of the mantle in radioactive elements with time
Heat Transport Mechanisms

- Plate tectonics  
  (Earth, early Mars?, early Venus?)
- Stagnant lid convection  
  (Mercury, Venus?, Mars, Moon)
- Lithosphere delamination  
  (Venus?)

Magma transport (volcanism)
Stagnant Lid

Lithosphere Delamination

Plate Tectonics

\[ T_1 = T_m - 2.21 \eta (d\eta/dT)^{-1} \]

\[ T_1 = 1073 \text{ K} \]
‘Driving’ Temperature Contrast

![Graph showing relationships between mantle temperature and delta T for plate tectonics, lithosphere delamination, and stagnant lid models. The graph includes lines for different values of A: A=3.5E4 and A=6.5E4.]
Mantle Temperature

![Graph showing mantle temperature over time with labels for stagnant lid, lithosphere delamination, and plate tectonics.](image)
Lid Thickness

![Graph showing the relationship between stagnant lid thickness and time (Ma). The graph includes a line for stagnant lid thickness and a dashed line for lithosphere delamination.]
Stagnant Lid Convection

- The figure shows the thermal evolution of a lunar model according to Spohn et al. (2000)

- The planet cools by thickening its lithosphere while the deep interior stays warm
Cumulative Energy Loss

![Graph showing cumulative energy loss over time (Ma) with curves for plate tectonics, lithosphere delamination, and stagnant lid.]
Surface Heat Flow

- Plate tectonics
- Lithosphere delamination
- Stagnant lid
Urey Ratio

type of the heat produced by radioactive decay to the total surface heat flow
Transition to the stagnant lid regime

- Different regimes occur depending on the viscosity contrast: Mobile and SL
- Mobility criterion: $M_1 = \delta_{rh} \text{Nu}$
2D and 3D Convection Models

- Full set of hydrodynamic equations
- Local parameters (e.g. temperature, velocity field)
- Mantle flow pattern
2D and 3D Convection Models

- basic laws in Boussinesq approximation (density constant except in buoyancy term)

  - conservation of mass
    \[ \nabla \cdot \tilde{u} = 0 \n\]

  - conservation of momentum
    \[ \frac{\partial \tilde{u}}{\partial t} + \tilde{u} \cdot \nabla \tilde{u} = -\nabla p + \rho g \alpha T \zeta + \nabla (\eta \nabla \tilde{u}) \]

  - conservation of energy
    \[ \frac{\partial T}{\partial t} + \tilde{u} \cdot \nabla T = \kappa \nabla^2 T + R \]
2D and 3D Convection Models

- basic laws in Boussinesq approximation (density constant except in buoyancy term)

  - conservation of mass
    \[ \nabla \cdot \vec{u} = 0 \n\]

  - conservation of momentum
    \[ 0 = -\nabla p + RaT\hat{z} + \nabla (\eta \nabla \vec{u}) \]

  - conservation of energy
    \[ \frac{\partial T}{\partial t} + \vec{u} \cdot \nabla T = \kappa \nabla^2 T + \frac{Ra_q}{Ra} \]
Parameterized Models

- Simple scaling laws (e.g. $\text{Nu} \sim \text{Ra}$)
- Global parameters as function of time (e.g. mean temperature, heat flow)
Parameterized Convection Including Core Cooling and Crustal Growth

- Energy equation: mantle and core
- Equation for lithosphere growth
- Equation for crustal growth
\[
\rho_m C_m V_m \frac{dT_m}{dt} = -q_m A_m + q_c A_c + V_m Q_m,
\]
\[
\rho_c C_c V_c \frac{dT_c}{dt} = -q_c A_c
\]
\[ q_m = k \frac{T_m - T_s}{s} \]

\[ \delta \approx \text{const.} \cdot \text{Ra}^{-\beta} \]

\[ q_c = k \frac{T_c - sT_m}{\delta} \]

\[ \delta_c = \left( \frac{\chi g \text{Ra} \alpha}{\alpha g \Delta \gamma} \right)^{1/3} \]
\[
\rho_m C_m (T_m - T_l) \frac{dl}{dt} = -(q_m) + k \frac{\partial T}{\partial z}_{z=z_l}
\]

Plate tectonics

\[T_l = T_s = 220K\]

Stagnant lid convection

\[T_l = T_m - \Delta T_e\]

Stagnant Lid

Mantle

Core

Heat conduction equation

- \[T_s\]
- \[T_l\]
- \[T_m\]
- \[Q_m\]
- \[q_m\]
- \[q_c\]
- \[q_s\]
\[
\frac{dD_c}{dt} = \frac{D_{pot} - D_c}{D_m} u m_a \frac{V_a}{V_m}
\]
Generation of melt tends to buffer temperature variations due to consumption and release of latent heat

- Energy available for a temperature increase of 100 K results in \( \Delta T \) of ~ 50 K due to consumption of latent heat

\[
\Delta T = \frac{L}{C_p} \frac{V_{\text{melt}}}{V}
\]
Example Mars
Observed Crustal Evolution

- Average crust thickness: 50 – 100 km
- Primordial crust 20 - 45 km
- Strong decrease of crustal productivity since the Noachian
- Recent volcanism
Earth: Crust Formation in plate tectonic regime

- Efficient crust formation at the divergent plate boundaries (pressure released melting close to the surface) \(~ 17 \text{ km}^3/\text{a}\)
- Two-stage crust formation
- Possible strong plume volcanism in early evolution
Crust Formation in a One-Plate Planet

- Melt production underneath the stagnant lid
Mantle temperature

![Mantle temperature graph with time (Ma) on the x-axis and mantle temperature (K) on the y-axis. The graph shows three curves labeled M1800, M1900, and M2000, representing different mantle temperatures over time.](image)
Crustal thickness and growth rate

![Graph showing crustal thickness and production rate over time](image)
Present-day crustal thickness as function of initial mantle temperature and viscosity

![Graph showing the relationship between initial mantle temperature and crustal thickness with different viscosities represented.

The graph has the x-axis labeled "Initial mantle temperature (K)" ranging from 1700 to 2100 K, and the y-axis labeled "Crustal thickness (km)" ranging from 0 to 220 km.

There are four curves on the graph, each representing a different viscosity:
- \( \eta = 10^{19} \text{ Pas} \)
- \( \eta = 10^{20} \text{ Pas} \)
- \( \eta = 10^{21} \text{ Pas} \)
- \( \eta = 10^{22} \text{ Pas} \)

The curves show how crustal thickness increases as the initial mantle temperature increases, with higher viscosities resulting in thicker crusts.

The graph illustrates the impact of initial mantle temperature and viscosity on the present-day crustal thickness.
Conclusions Mars I

To explain the observed crustal growth

- Trade off between reference viscosity and initial mantle temperature:
  - the higher the reference viscosity the higher the initial mantle temperature
  - e.g. \( \eta_{\text{ref}} = 10^{19} \text{ Pas} \quad T_m(t=0) = 1700 - 1800 \text{ K} \)
  - \( \eta_{\text{ref}} = 10^{21} \text{ Pas} \quad T_m(t=0) = 1900 - 2100 \text{ K} \)
Conclusions Mars II

- Crustal evolution with early plate tectonics shows either a peak in crustal activity 2 Ga ago or no crust formation after early stage of plate tectonics; inconsistent with observations

- Stagnant lid convection consistent with crustal evolution
Plume Volcanism on Mars?

- Early global volcanic activity reduces during the evolution in mainly one or two regions: Tharsis & Elysium

- Significant signal in the gravity field
Martian mantle convection without phase transitions

1.6 GA

2.5 GA

3 GA

3.4 GA

in der Helmholtz-Gemeinschaft
Core is cooling !!!!

Thermal boundary layer is necessary
  e.g. at the bottom of mantle

Cooling of the core ➔ rapid disappearance
  of thermal boundary layer
Conditions required to sustain plumes

- Perovskite layer
  - With a perovskite layer close at the core-mantle boundary it might be possible to generate a large plume early in the evolution but this will be very weak if existant after about 1 Ga.

- Heated core
- Depth dependence of viscosity
- Chemical layering