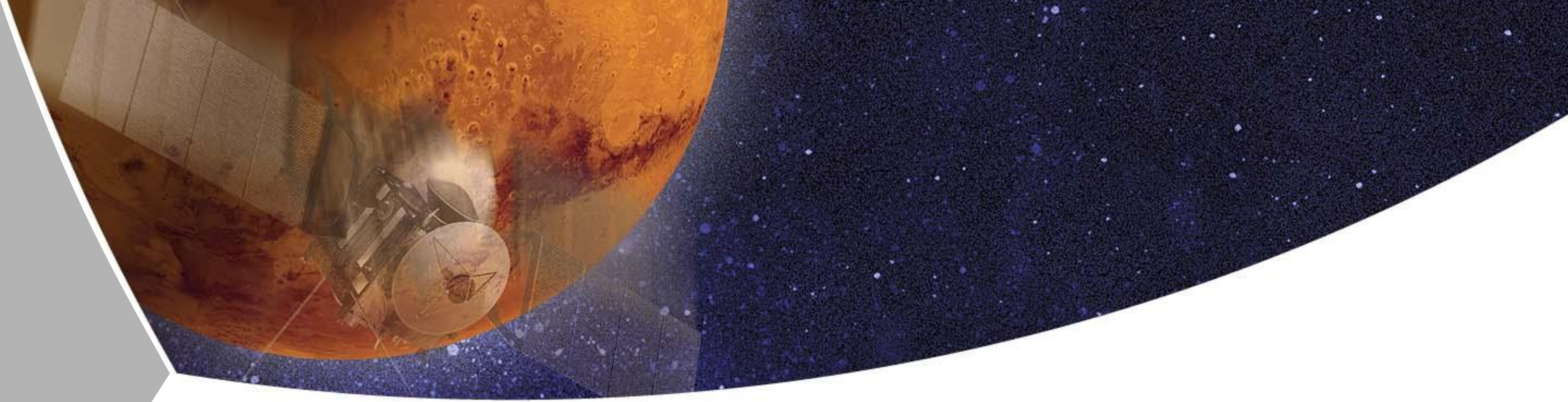
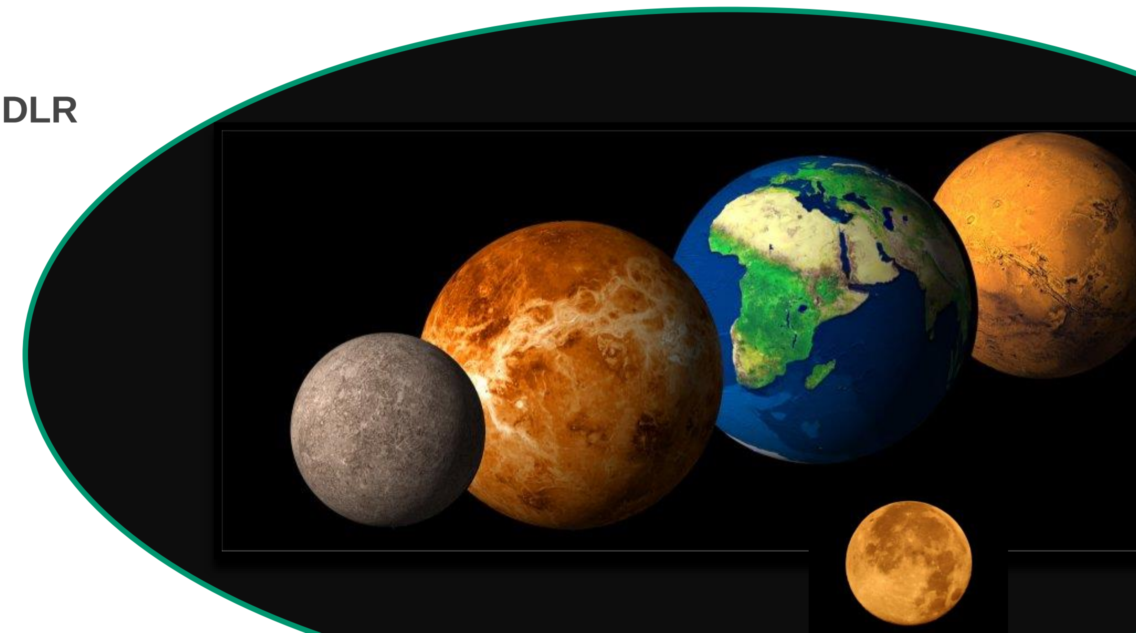


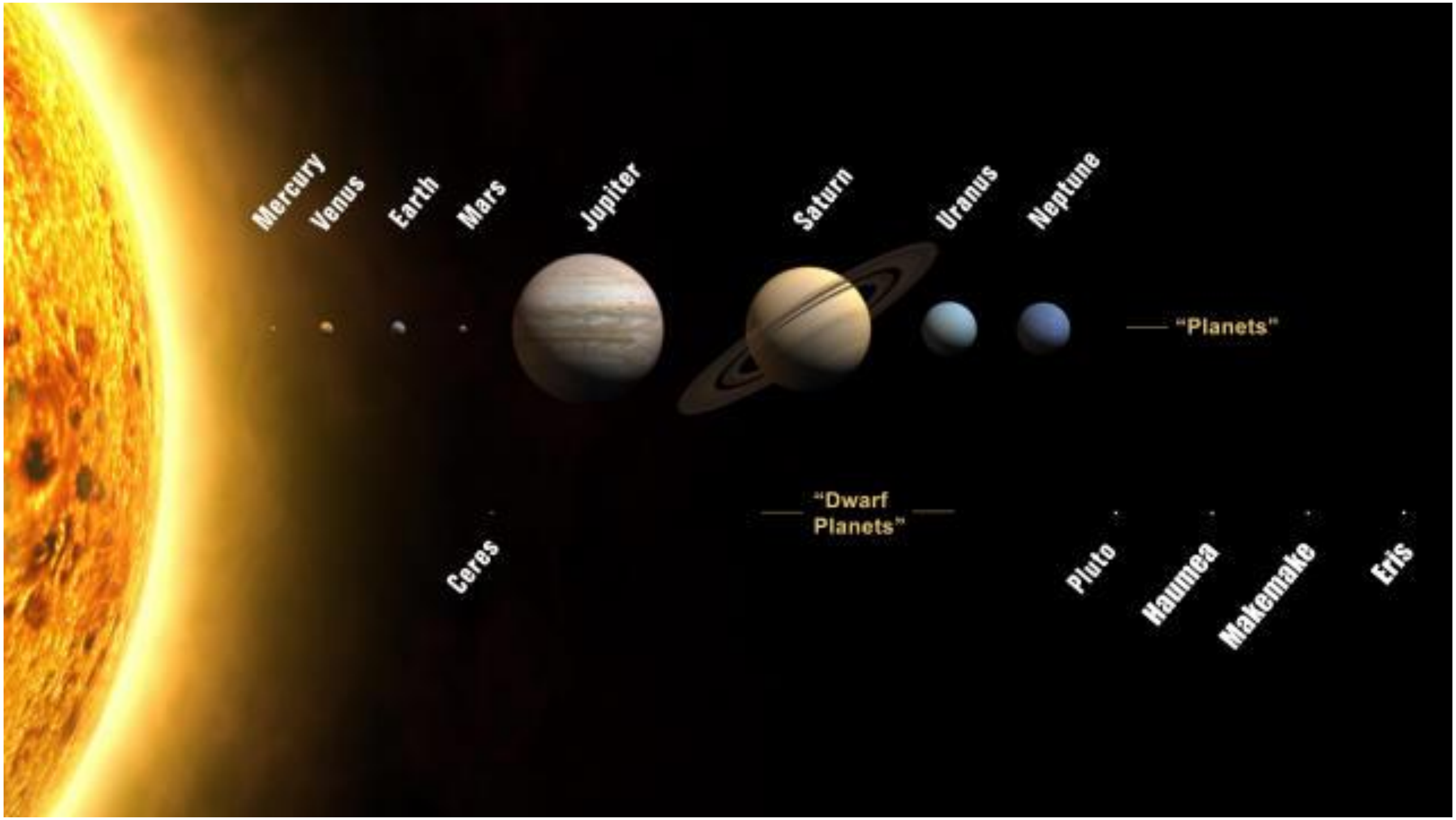
Terrestrial planets and the Moon

D. Breuer

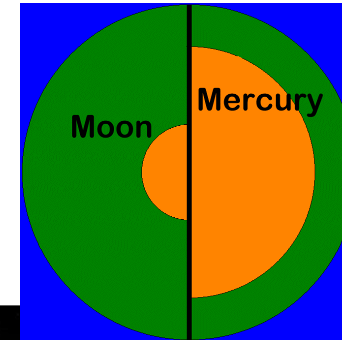
Institute for Planetary Research, DLR

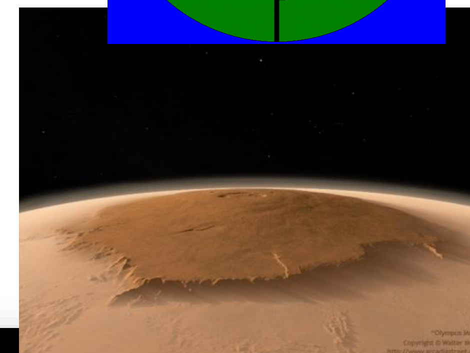


DLR Deutsches Zentrum für Luft- und Raumfahrt e.V.
in der Helmholtz-Gemeinschaft

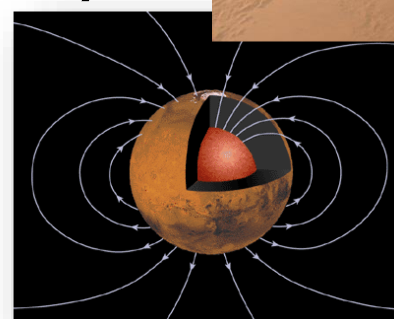

Mercury
Venus
Earth
Mars
Jupiter
Saturn
Uranus
Neptune
"Planets"
"Dwarf Planets"
Ceres
Pluto
Haumea
Makemake
Eris

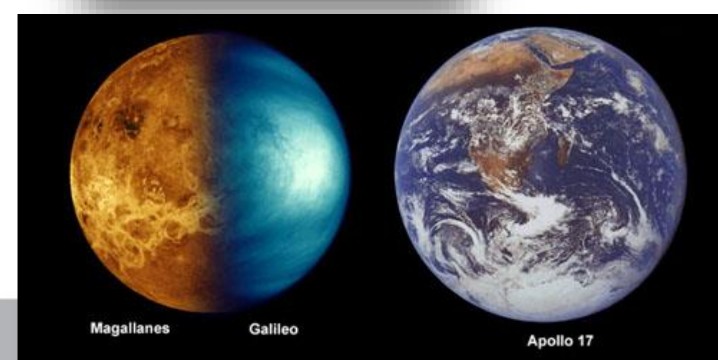
Similarities and Diversities:
Terrestrial Planets and the Moon

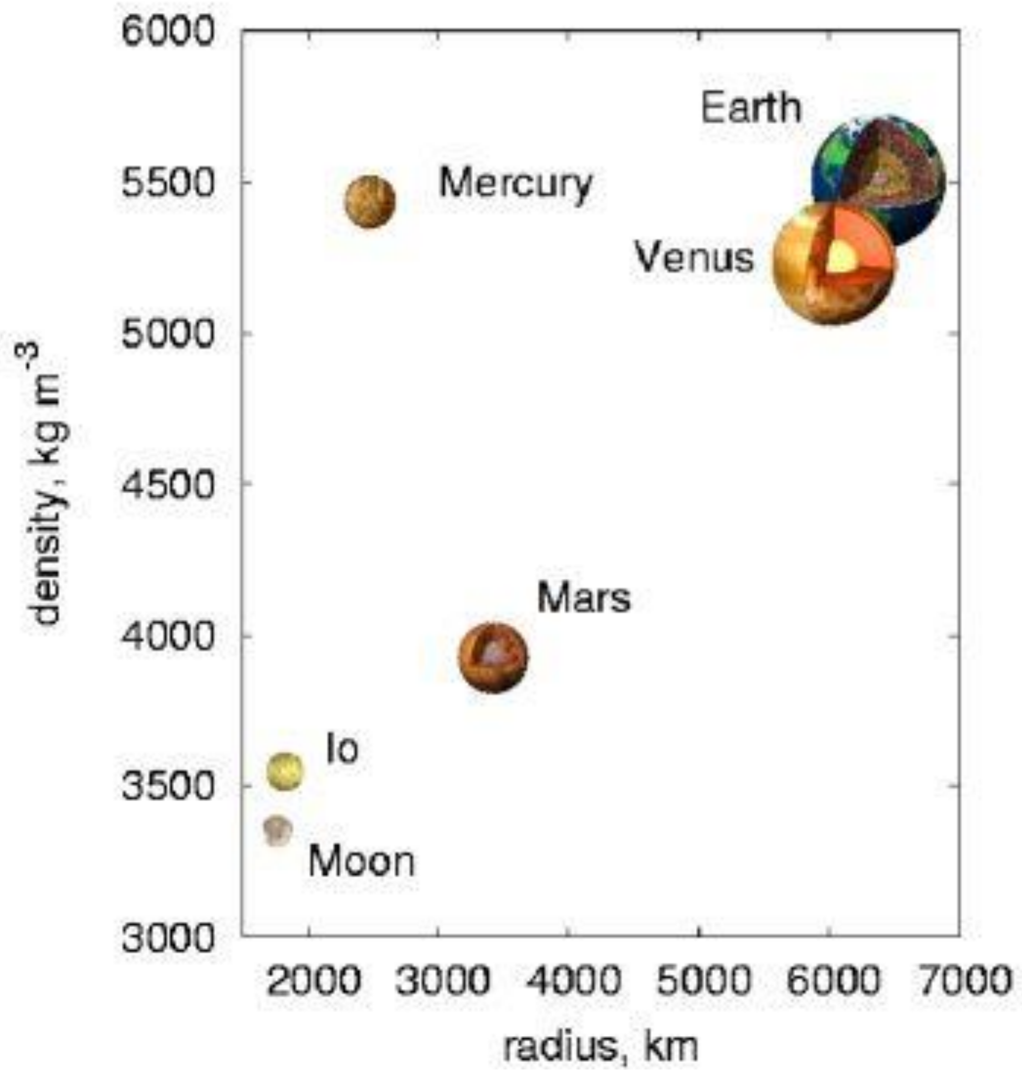




- **Interior structure and composition**
- **Surface structures (volcanic and tectonic)**
- **Magnetic field**
- **Atmosphere**





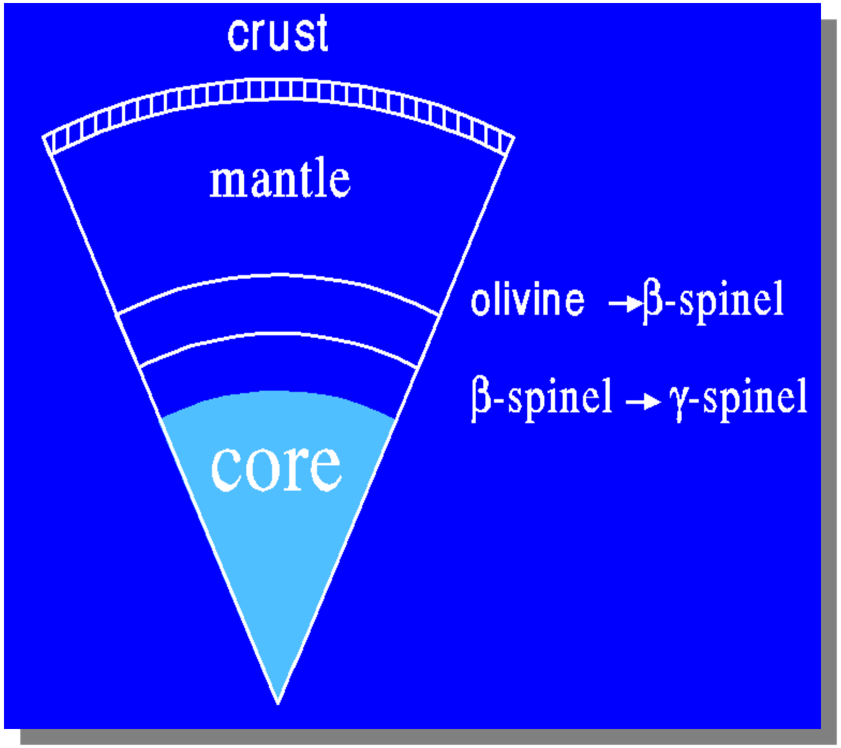


6000
5500
5000
4500
4000
3500
3000
density, kg m-3
Earth
Mercury
Venus
Mars
Io
Moon
2000
3000
4000
5000
6000
7000
radius, km

Ganymede

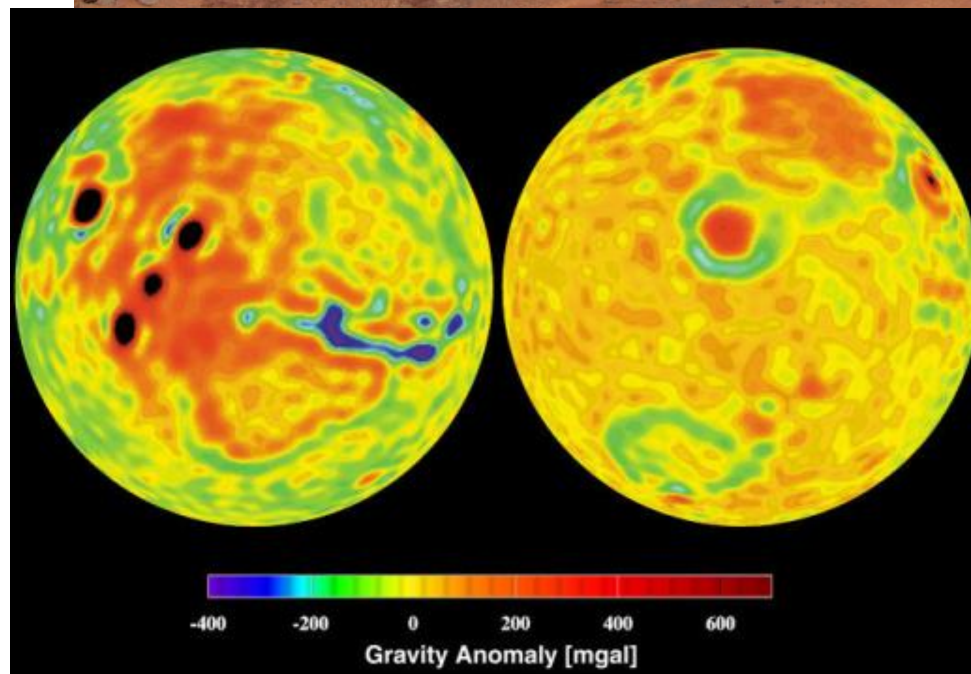
Interior Structure and Composition

- Mass of reservoirs (crust, mantle, core)
- Composition
- Depth of phase transitions and chemical layers
- Variations of pressure, temperature, and density



Data

- Mass
- Gravity field, rotational state, topography
- Chemistry / mineralogy of the surface
- Cosmochemical data (SNC, Moon samples)
- Data from the laboratory
- Heat flow
- Seismology



MGS Gravity Field of Mars

Interior Structure: The Data Set

- **Moon** and **Mars**: Mass, MoI-factor, Samples, Surface Chemistry, Lunar seismology
- **Mercury:** Mass, MoI-factor, C_m/C, Surface chemistry
- **Venus**: Small rotation rate precludes a calculation of the MoI-factor from J_2 under the assumption of hydrostatic equilibrium

Simple Two-Layer Model

$$I = \int x^2 \, dm$$

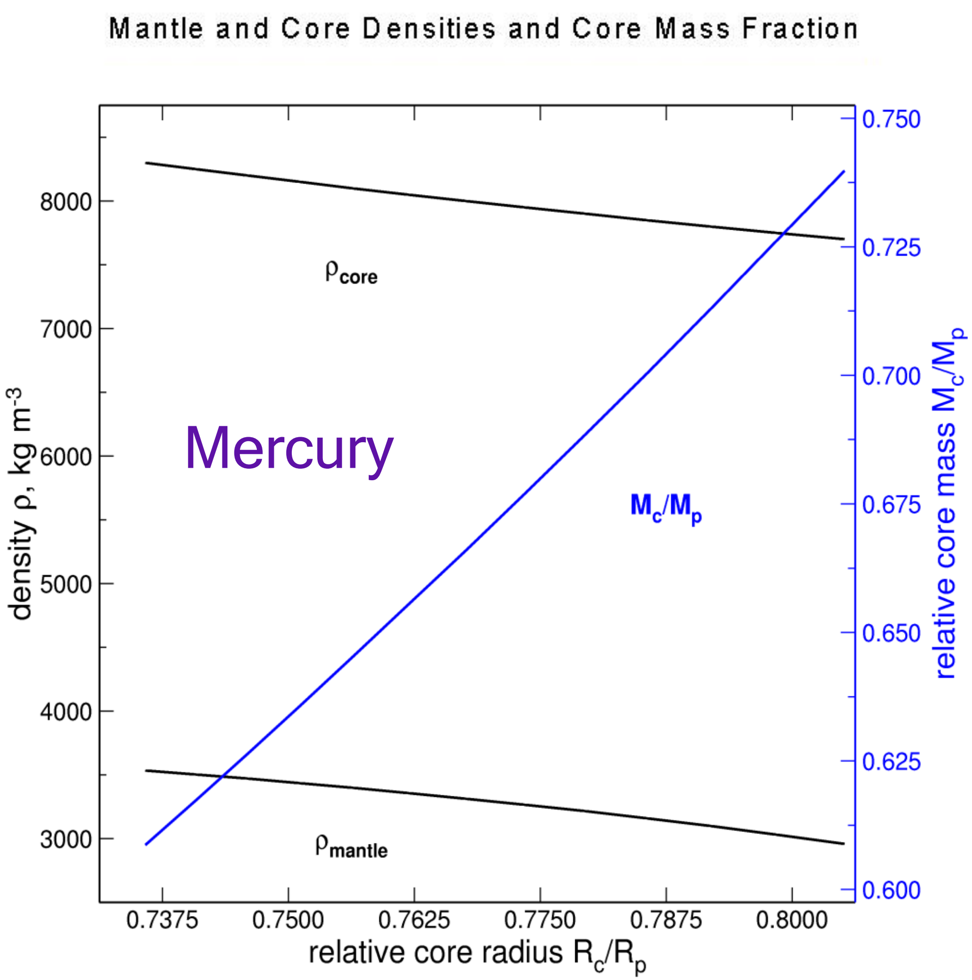
$$M = \frac{4}{3}\pi\left(\left(R_P^3 - R_c^3\right)\rho_m + R_c^3\rho_c\right)$$

$$\frac{I}{MR_P^2} = \frac{2}{5}\left(\left(1-\left(\frac{R_c}{R_P}\right)^5\right)\left(\frac{\rho_m}{\bar{\rho}}\right)+\left(\frac{R_c}{R_P}\right)^5\left(\frac{\rho_c}{\bar{\rho}}\right)\right)$$

Ambiguity: three unknowns, two pieces of data

Two-layered structural models

- Non-uniqueness of even simple interior structure models with $\rho_s = 0$
  - two constraints: mean density, MoI factor
  - three unknowns: mantle and core density, core radius
- Reduce ambiguities by cosmochemistry
  - core densities ranging from pure Fe to eutectic Fe-FeS



Detailed Models of the Interior

Structural Equations

mass, *m*

$$\frac{dm}{dr} = 4\pi r^2 \rho$$

moment of inertia, θ

$$\frac{d\theta}{dr} = \frac{8}{3}\pi r^4 \rho$$

gravity, *g*

$$\frac{dg}{dr} = 4\pi G\rho - 2\frac{g}{r}$$

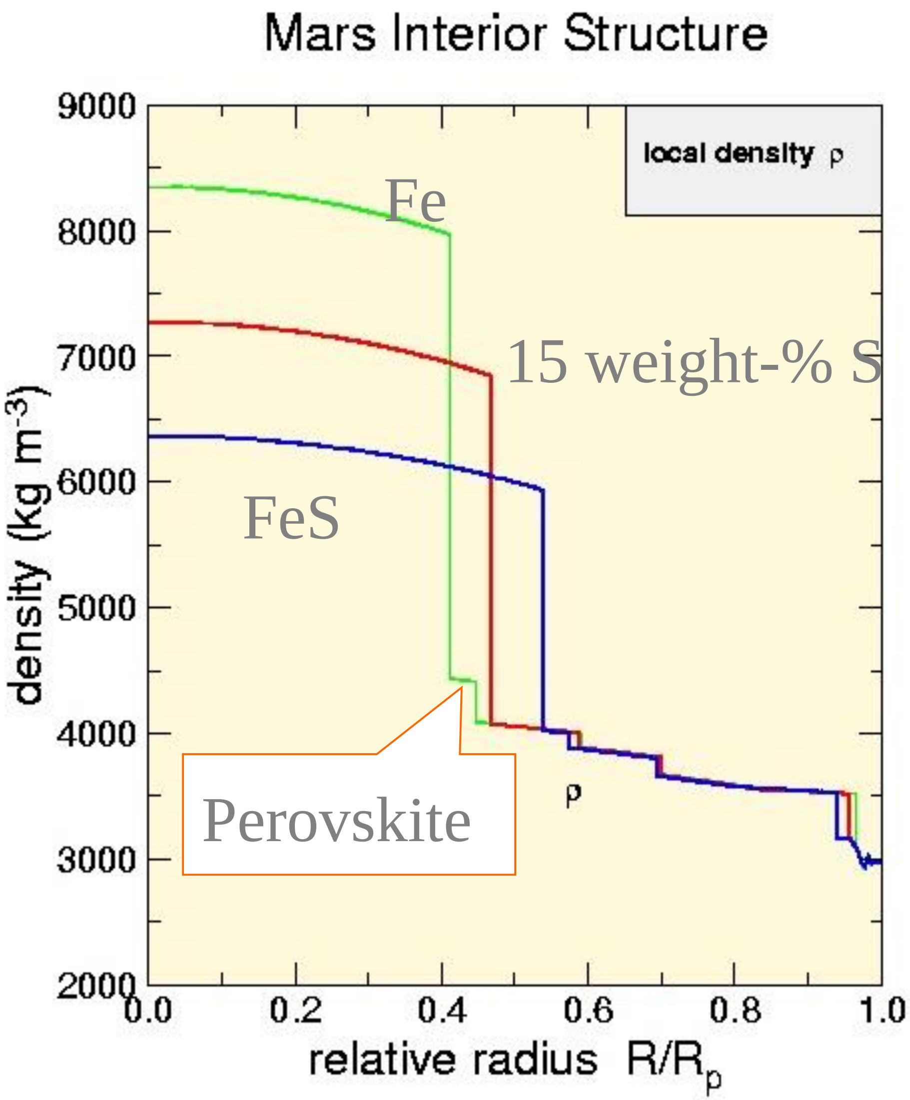
pressure, *p*

$$\frac{dp}{dr} = -g\rho$$

Model assumptions:

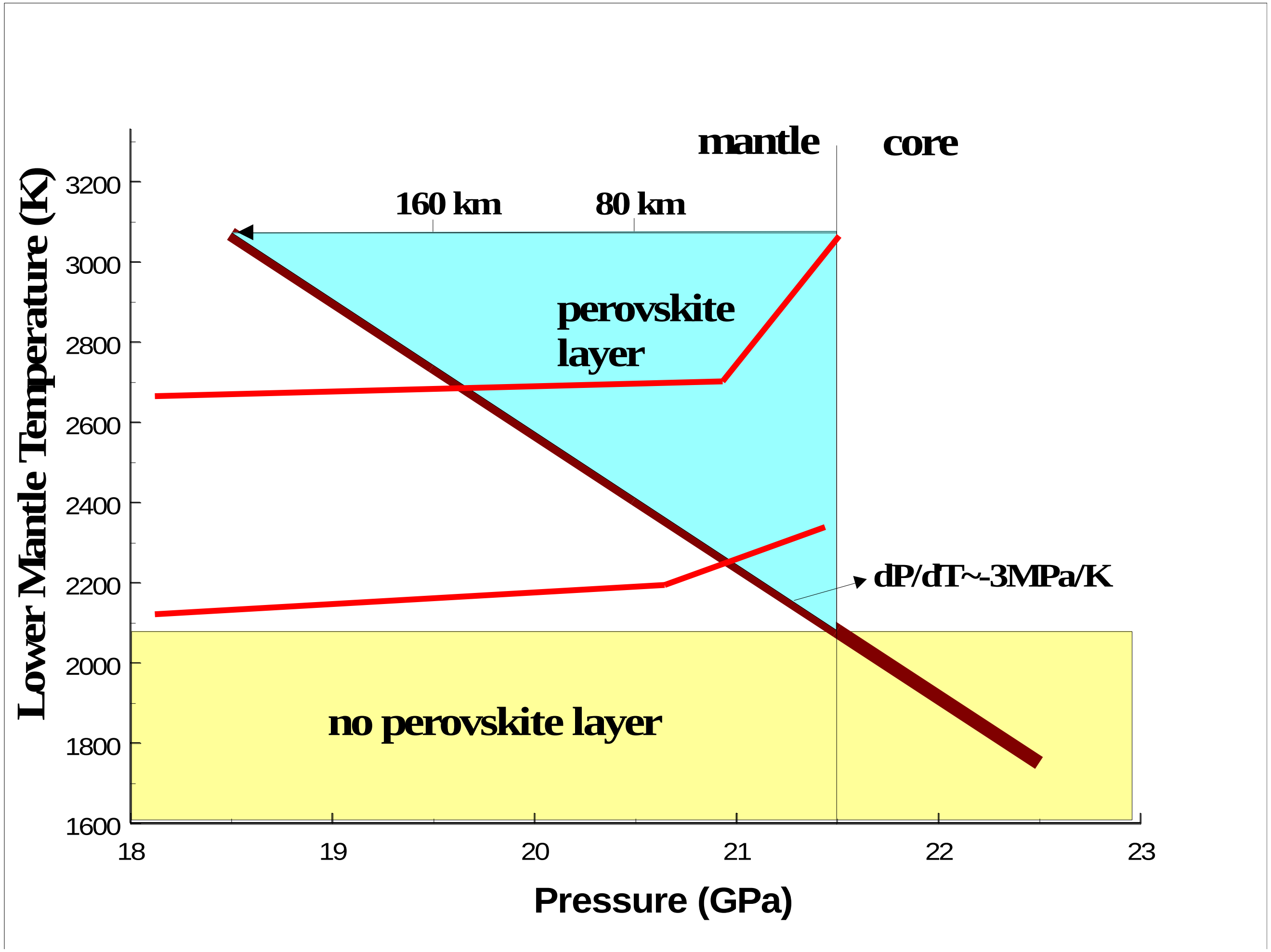
- Spherically symmetric and fully differentiated planets
- Hydrostatic and thermal equilibrium

Interior structure of Mars



Sohl and Spohn, 1997

Perovskite layer thickness near core/mantle boundary dependent on lower mantle temperature



Planetary Data

	Mercury	*Venus*	*Earth*	*Mars*	*Moon*
Radius	0.38	0.95	1.0	0.54	0.27
Mass	0.055	0.815	1.0	0.107	0.012
Density [kg/m^3]	5427.	5243.	5514.	3933.	3344.
MoI	0.346	?	0.3308	0.3662	0.3940
R_c/R_p	0.83	(0.55)	0.546	0.5	0.25

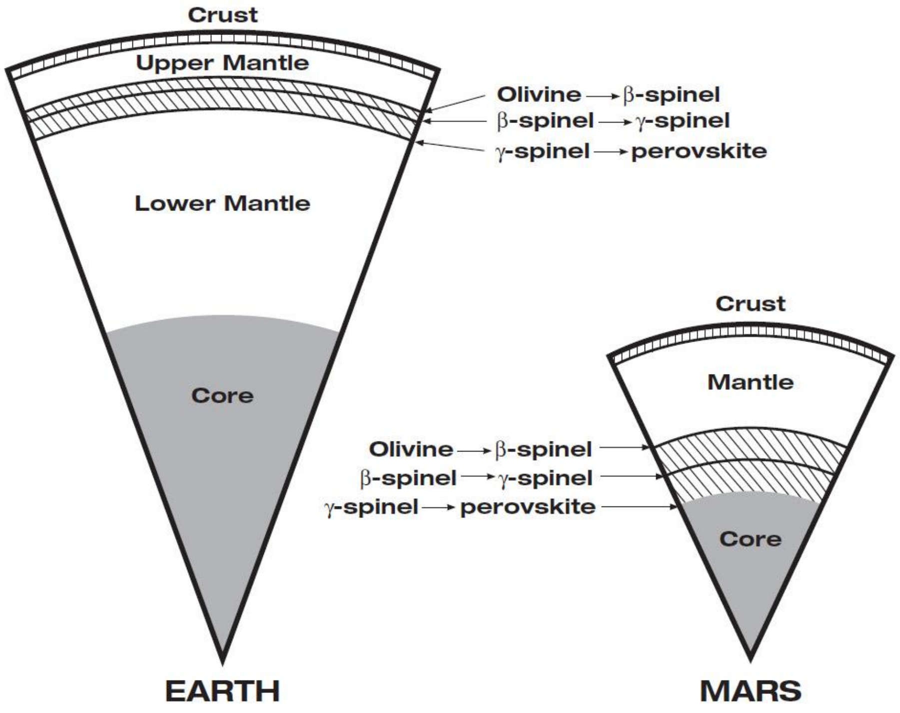
Interior structure: Mars

- Liquid core
  (constraint from tidal Love number k2)

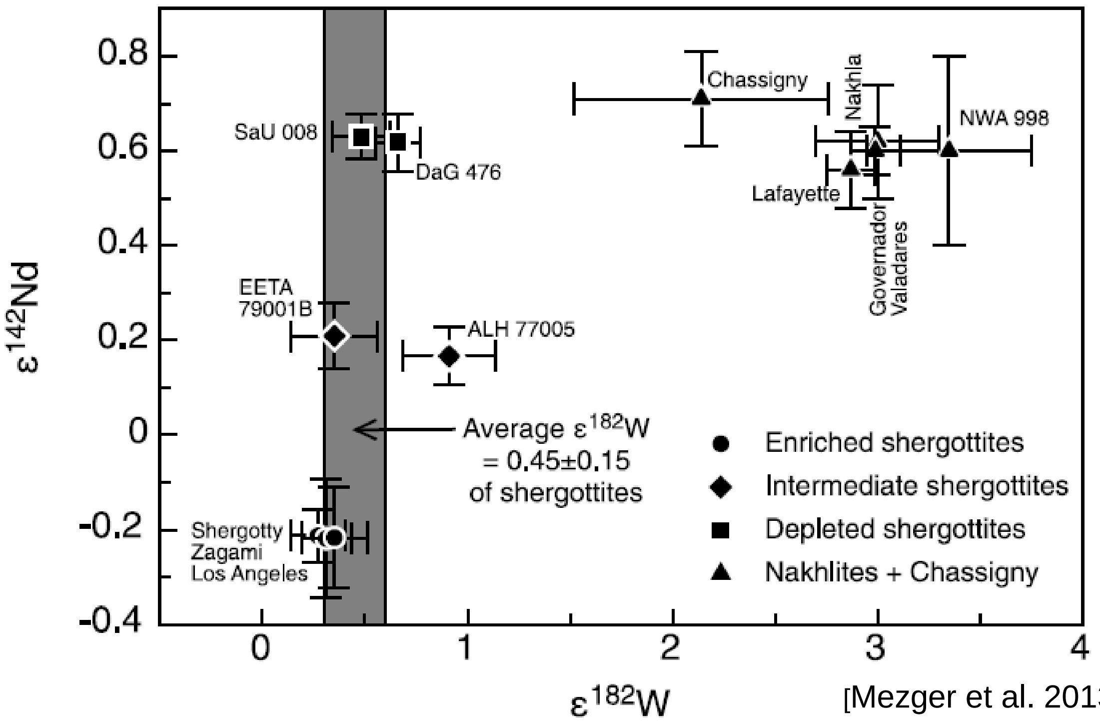
- Iron rich mantle composition
  FeO 15-18% (Earth 8 %)

- Deep phase transition?

- Crustal thickness (57 ± 24 km)
  (constraint from higher degree of gravity and topography - GTR)



Geochemical models

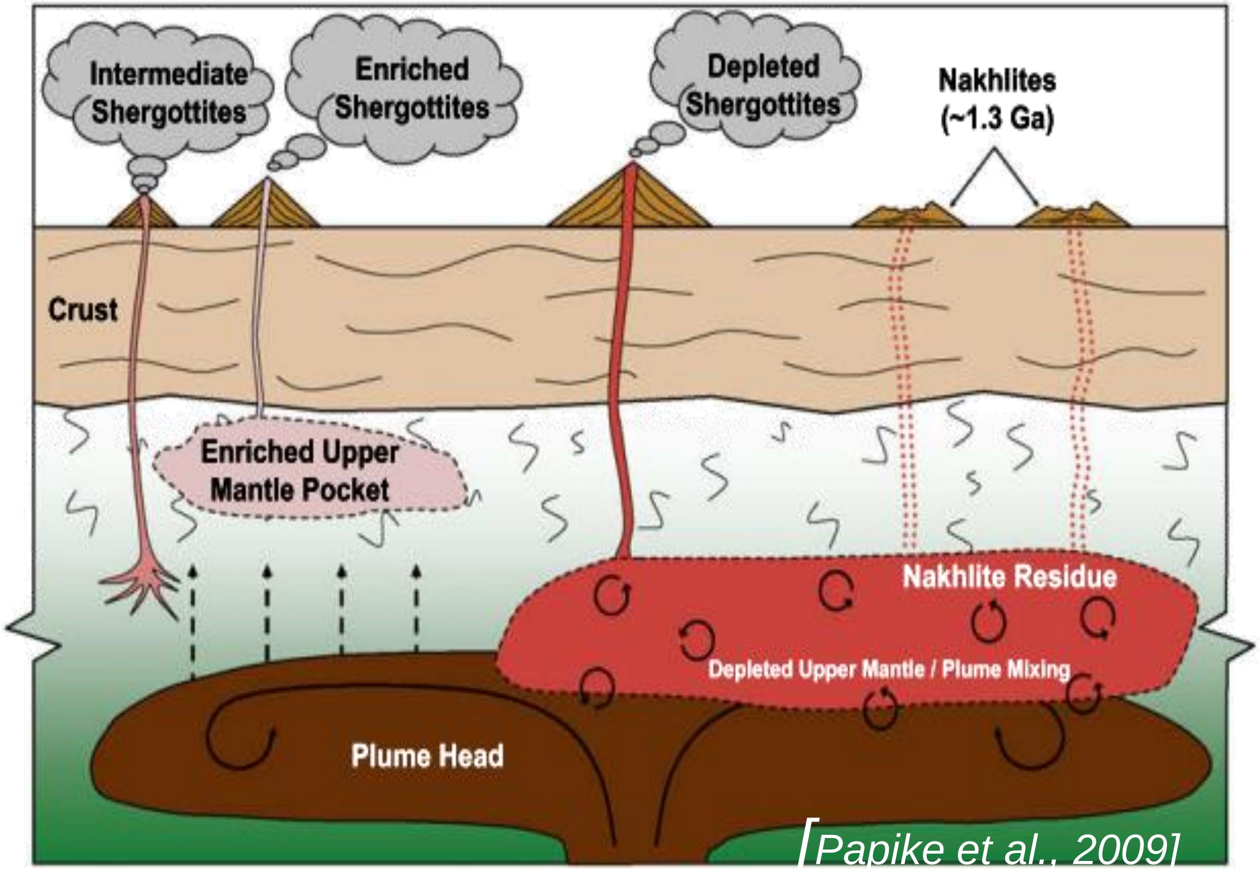


[Mezger et al. 2013]

- Nd anomalies in the SNCs indicate the existence of at least 3 reservoirs, which formed early and did not remix.
- As a comparison, Earth has ε^{142}Nd of 0 to 0.1.

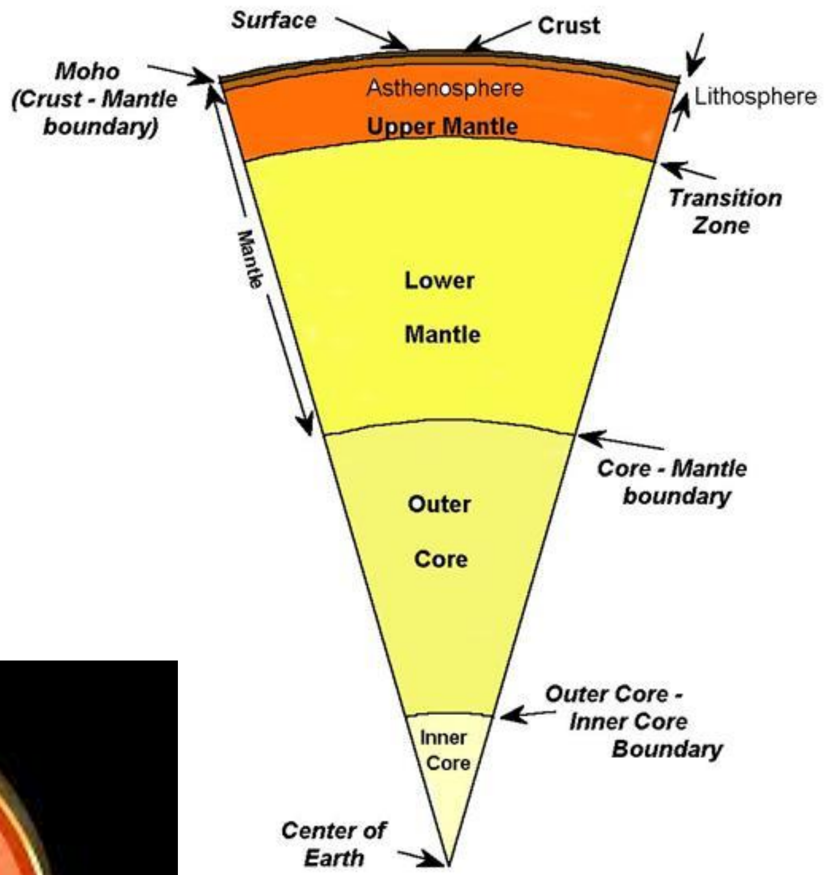
- Large spatial separation and inefficent mantle mixing could account for reservoir preservation.
- This appears to be incompatible with vigorous whole mantle convection.

Martian Reservois



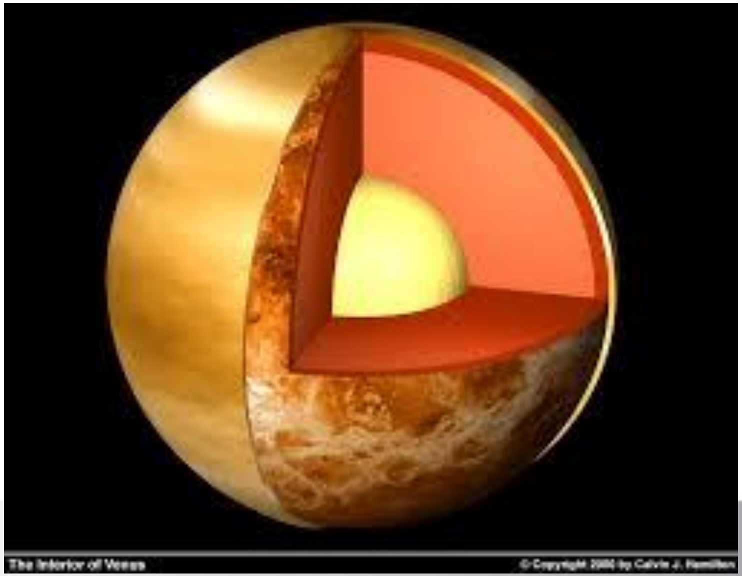
How did they form and remain stable over the entire Martian history?

Interior structure: Venus



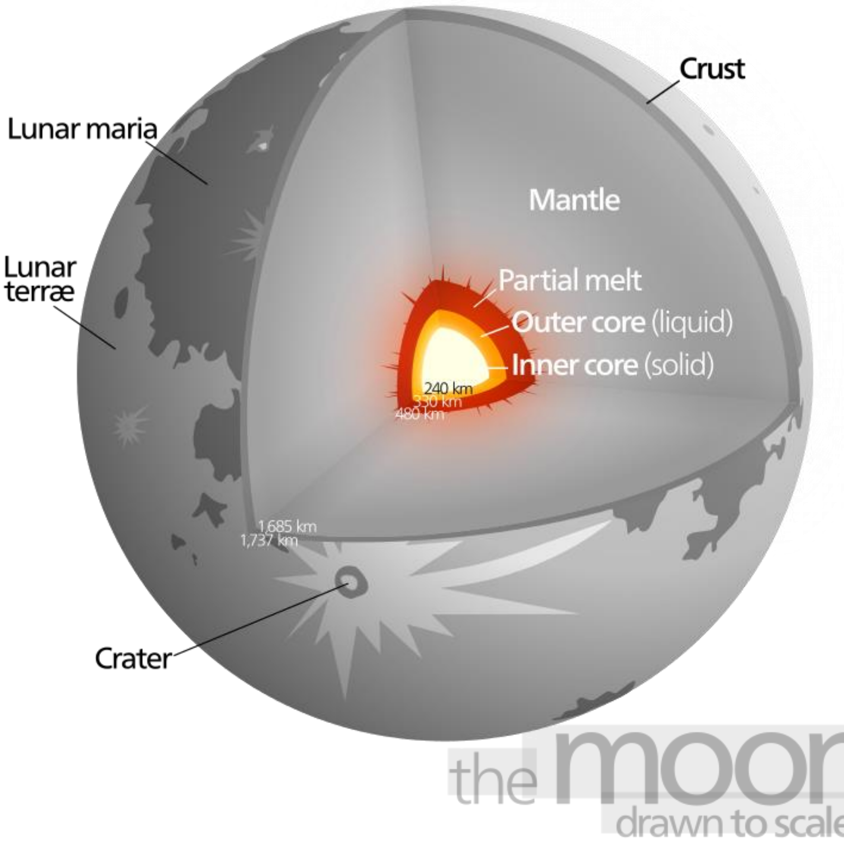
- Fluid core (constraint from tidal Love number k2)

- Crustal thickness: 8 - 30 km



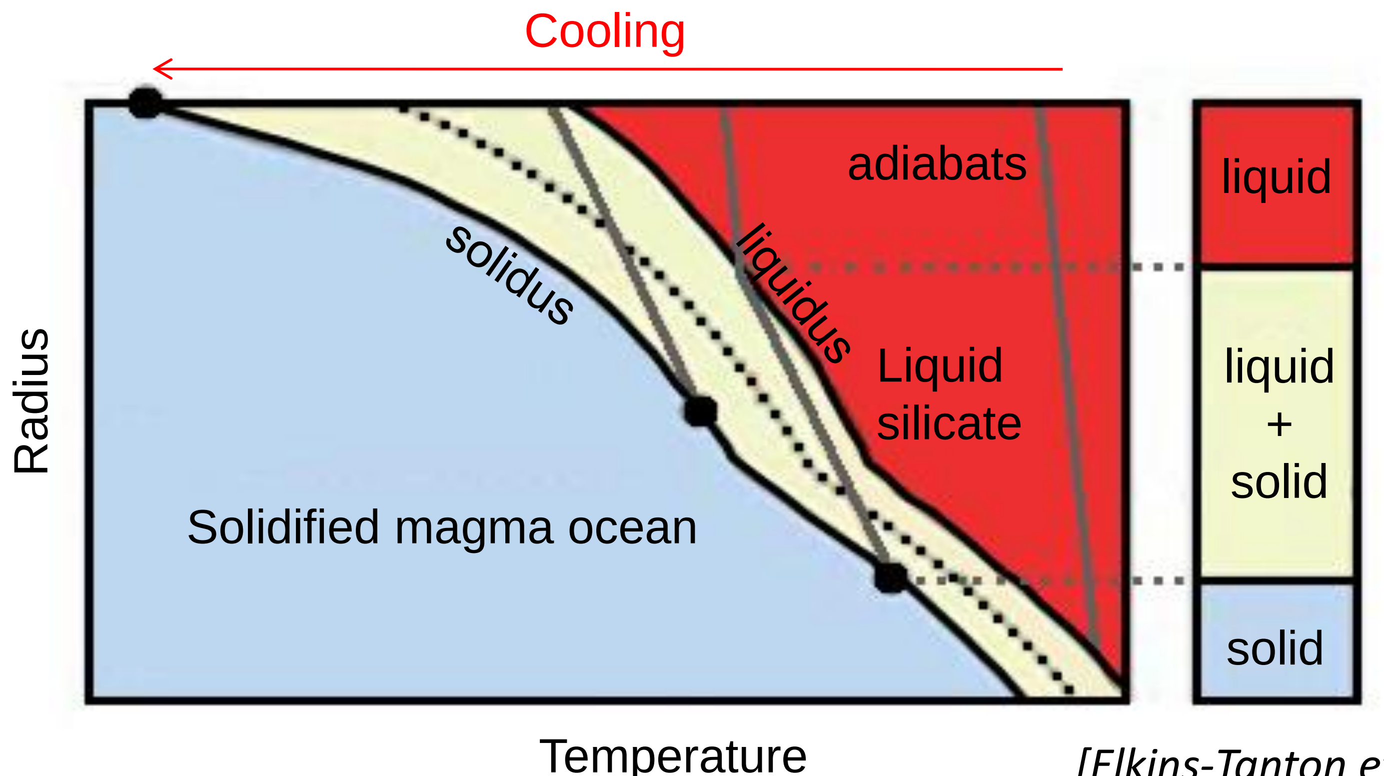
Interior structure: Moon

- Fluid core (constraint from lunar laser ranging)
- Solid inner core? (constraint from seismic data)
- Partial melt zone in deep mantle
- Anorthositic crust (39 ± 5 km)

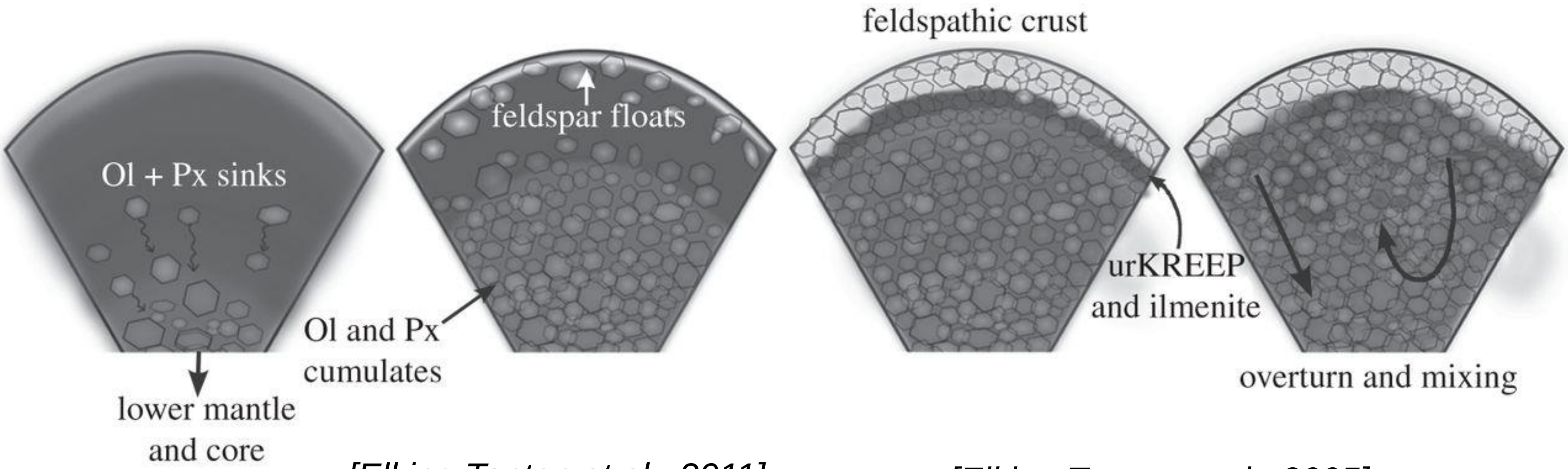


Magma Ocean Crystallization

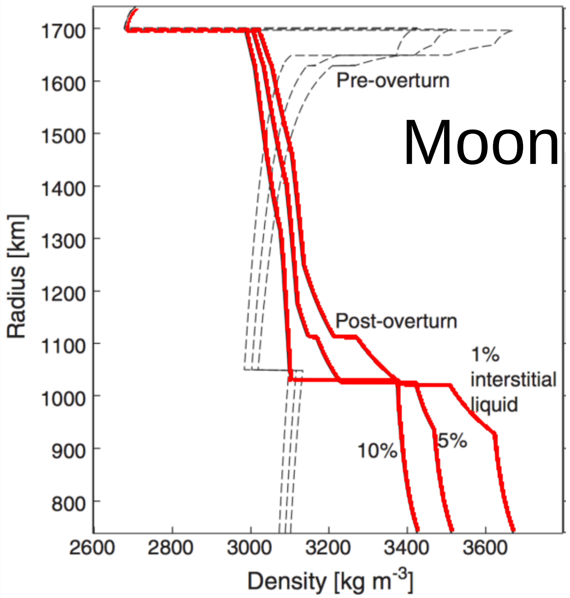
- Fractional crystallization ⇒ unstable density gradient
- Late mantle cumulates enriched in incompatible heat producing elements



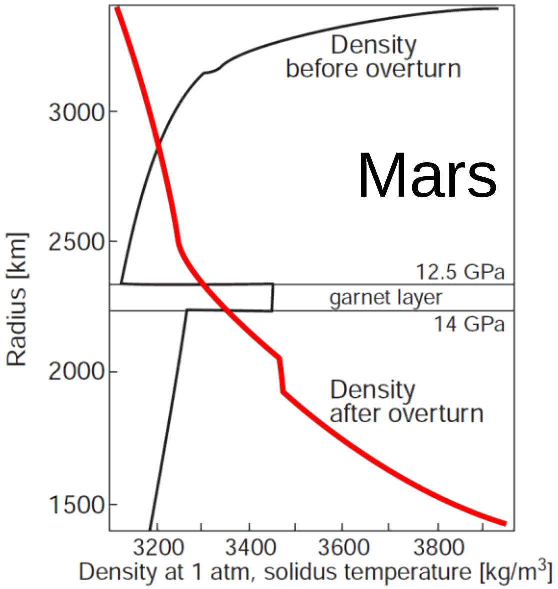
[Elkins-Tanton et al., 2005]



[Elkins-Tanton et al., 2011]

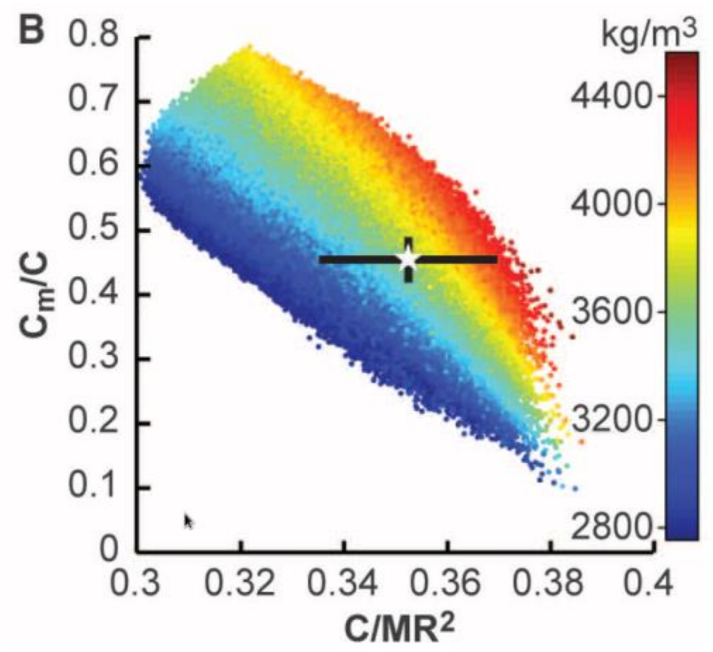


[Elkins-Tanton et al., 2005]



Interior structure: Mercury

- Thin mantle ~ 400 – 450 km
- Liquid core
- Crustal thickness: 35 ± 18 km (large V_c/V_m of 0.1)
- High mantle density could imply the presence of a solid FeS layer
- FeS layer would:
  - constrain T_c to 1600-1700 K.
  - reduce mantle layer to ~300 km



[Smith et al., 2012]

Interior structure: Mercury

- Thin mantle ~ 400 – 450 km
- Liquid core
- Crustal thickness: 35 ± 18 km (large V_c/V_m of 0.1)

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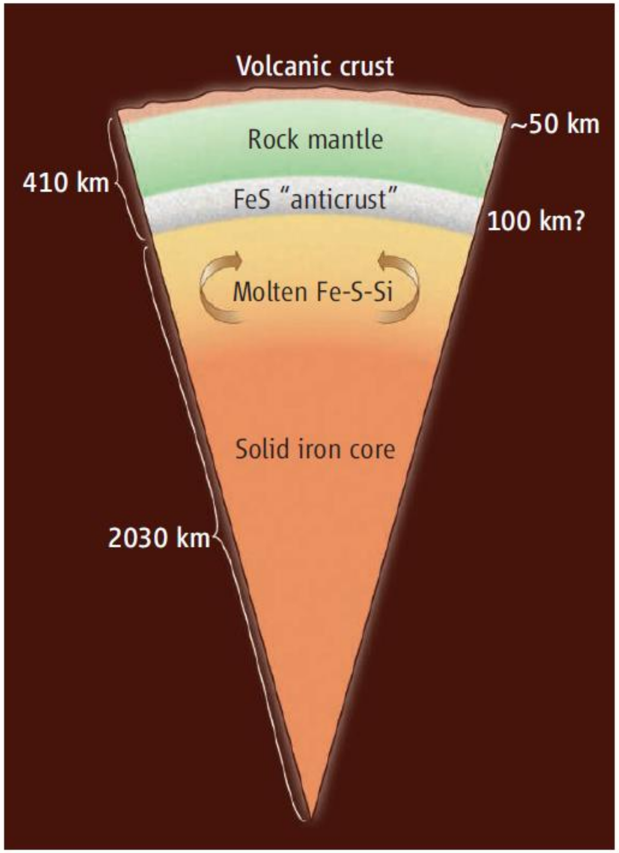
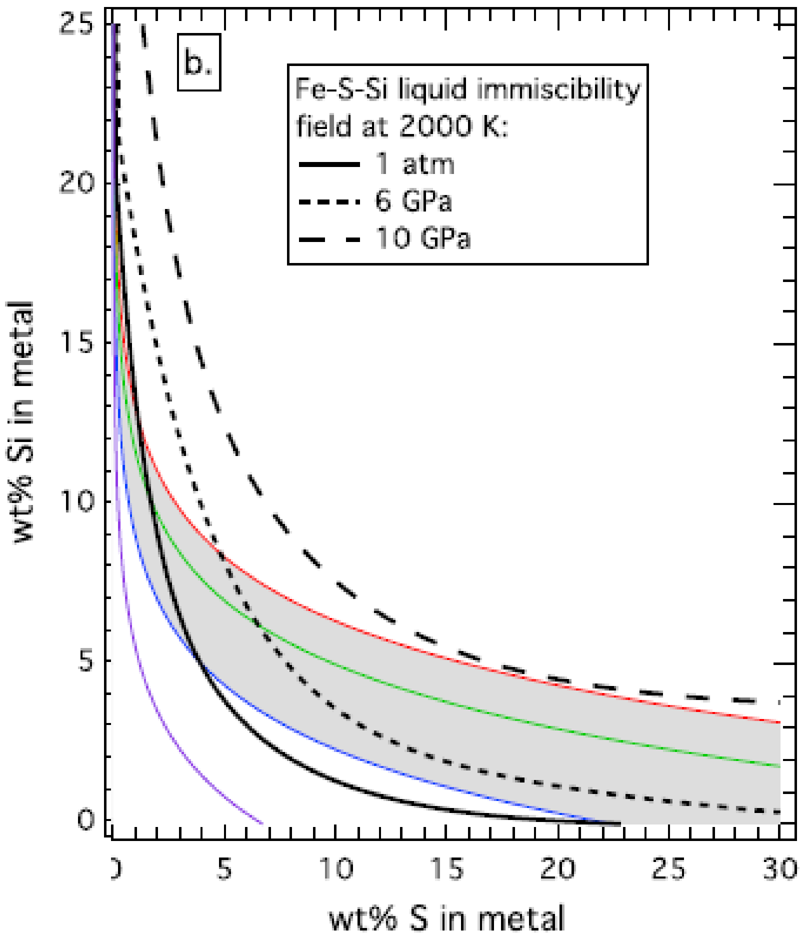


[KcKinnon 2012]

Core composition

- Elevated S content at surface → low $fO2$
- Si will partition in the core under reducing conditions → Fe-S-Si liquids
- Fe-S-Si liquids immiscible at P in upper core and solid FeS less dense than residual liquid

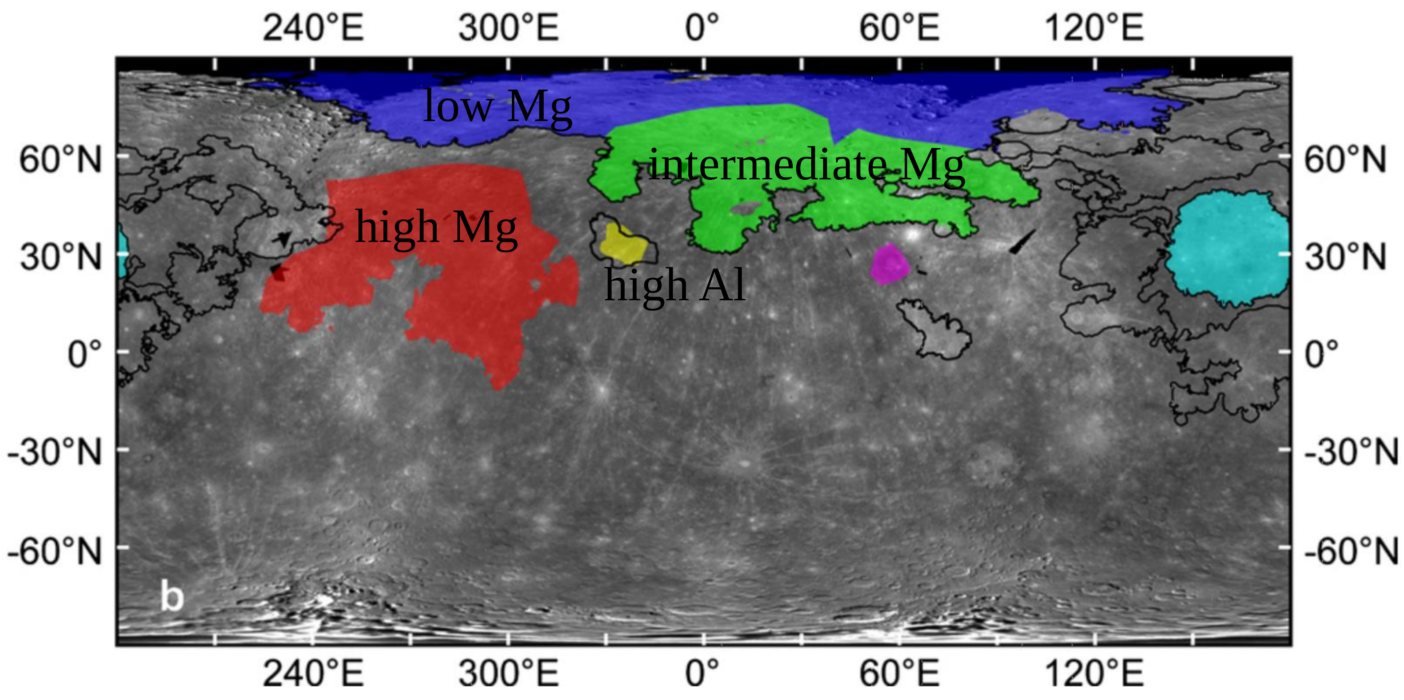


Chabot et al. 2014

Mantle reservoirs in Mercury

- X-Ray Spectrometer (XRS) and Gamma-Ray Spectrometer (GRS) :

The differences in composition of various terranes indicate that Mercury has a chemically heterogeneous mantle



[Weider et al., 2015]

Similarity and Diversity of Terrestrial Planets and the Moon

- **Interior structure and composition**
- **Surface structures (volcanic and tectonic)**
- **Magnetic field**
- **Atmosphere**

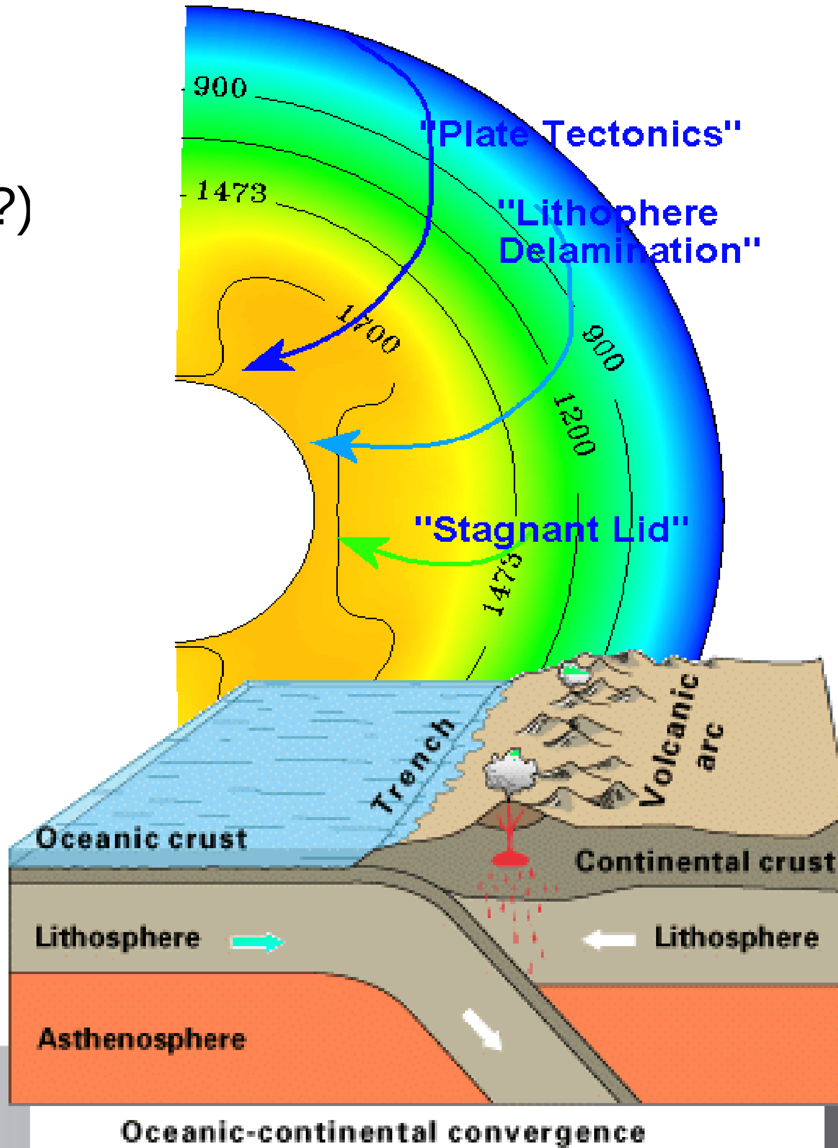
Heat Transport Mechanisms

- Plate tectonics
  (Earth, early Mars?, early Venus?)

- Stagnant lid convection
  (Mercury, Venus?, Mars, Moon)

- Lithosphere delamination
  (Venus?, early Earth)

Magma transport (volcanism)



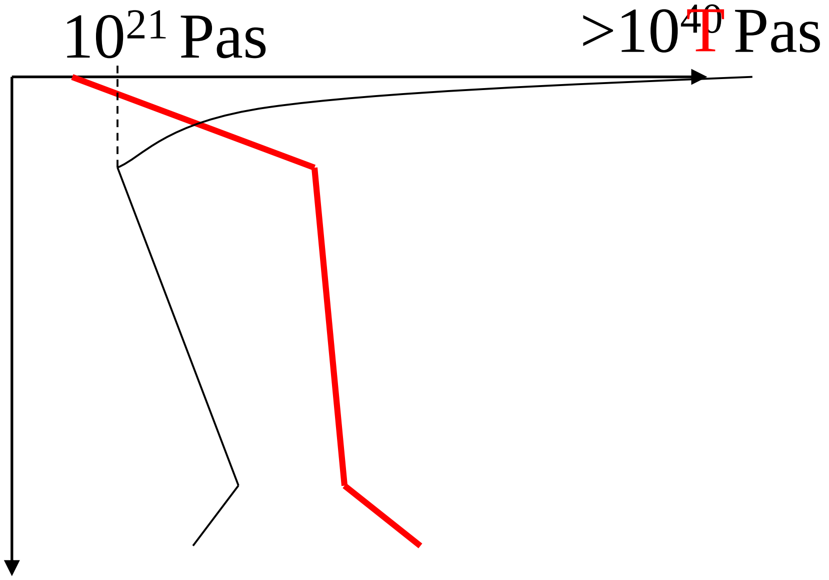
Mantle viscosity

$$\eta = C_1 \exp\left(\frac{E+pV}{RT_m}\right)$$

typical values for E and V

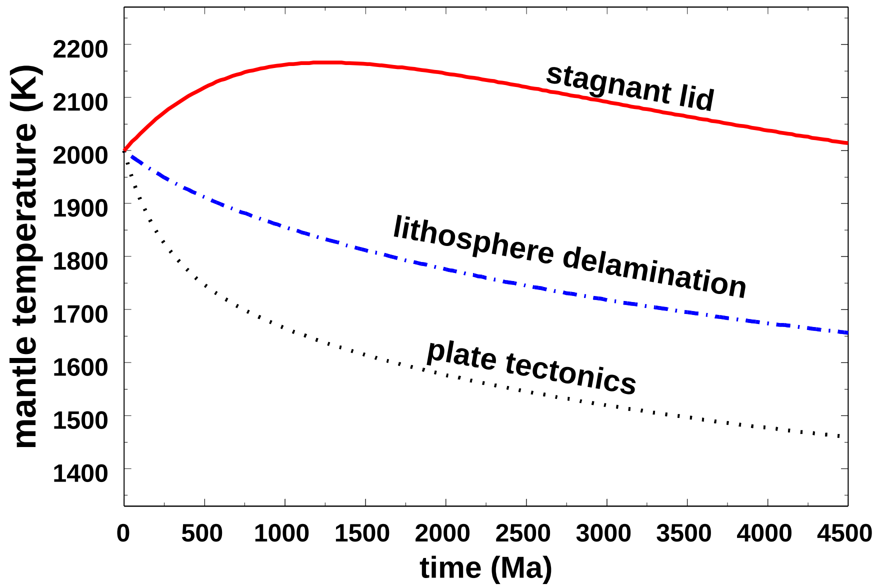
E = 300 – 540 kJ/mol

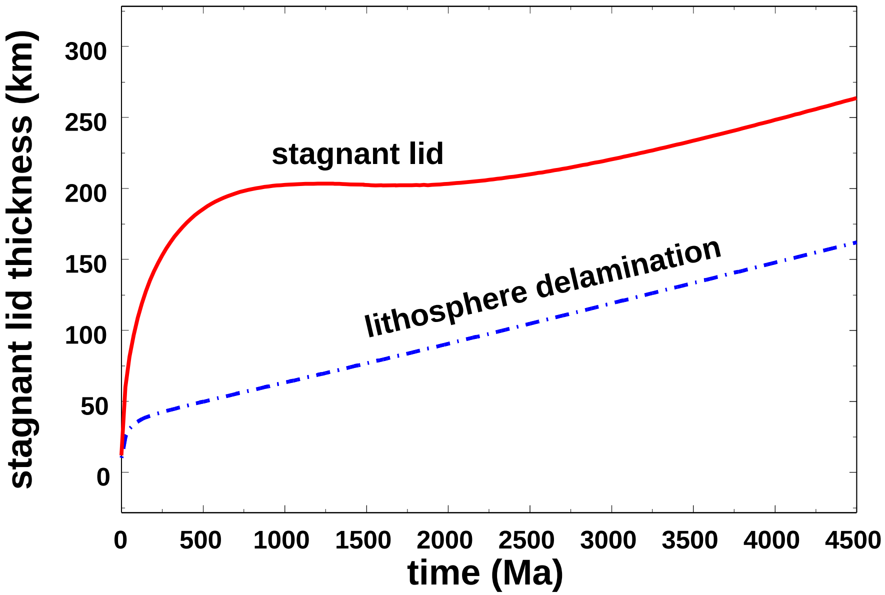
V = $2\cdot10^{-6}$ – $2\cdot10^{-5}$ m^3/mol



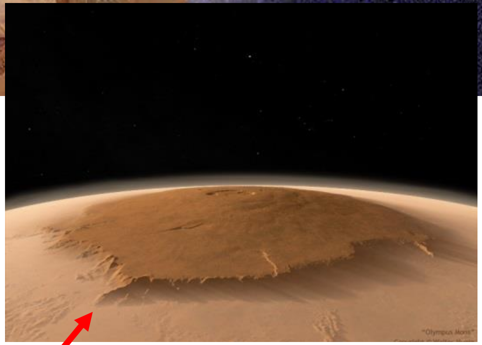
Difference between plate tectonics and one-plate tectonics

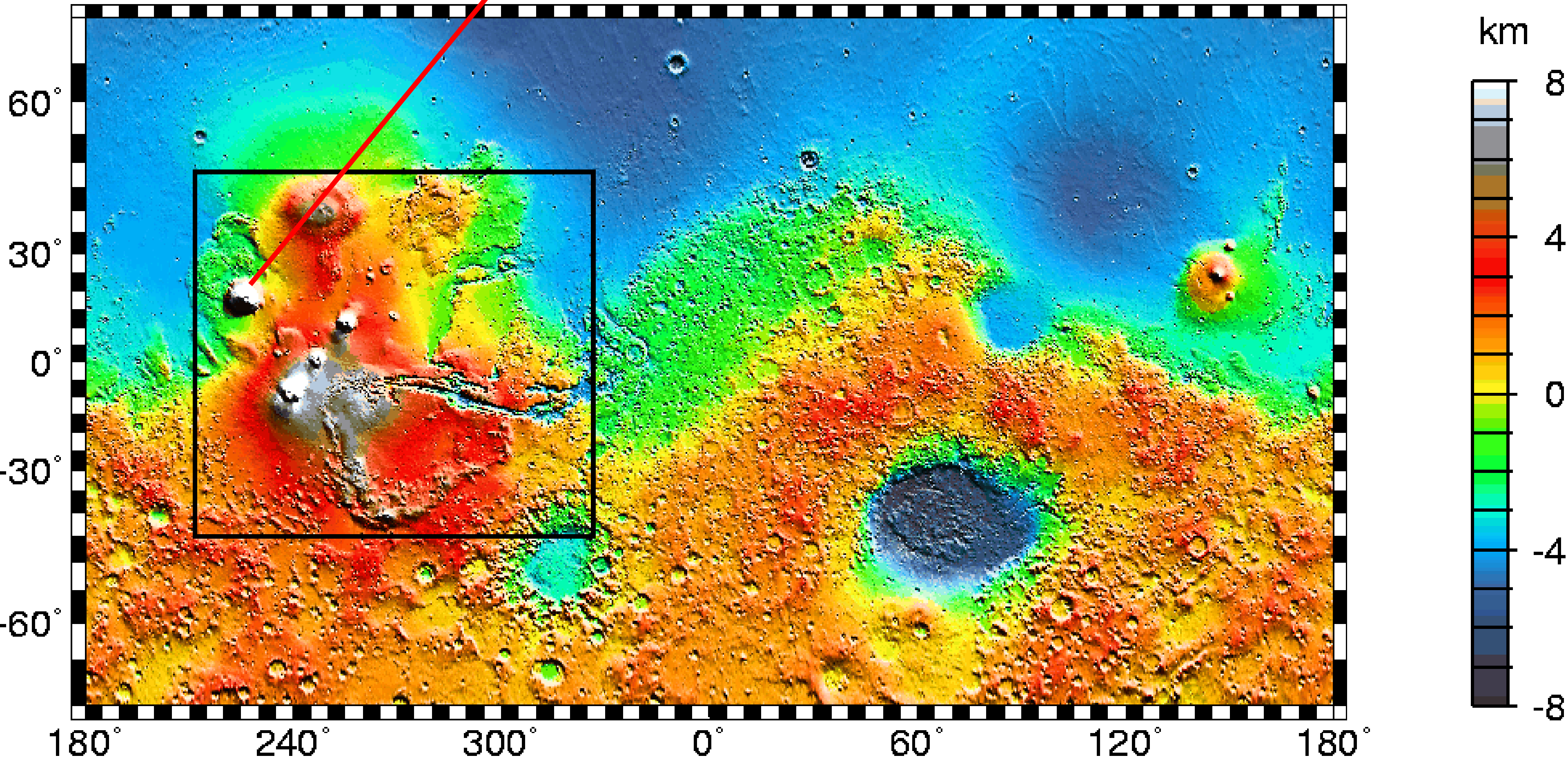
- Plate tectonics: Efficient cooling of the interior
- One-plate tectonics: Inefficient cooling of the deep interior but ‘thick‘ cold lithosphere



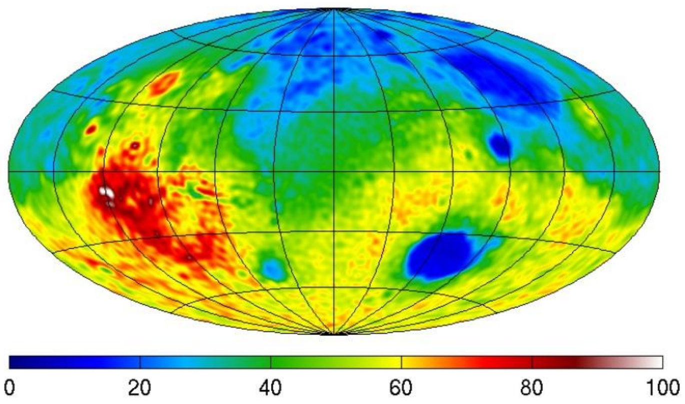


Mars

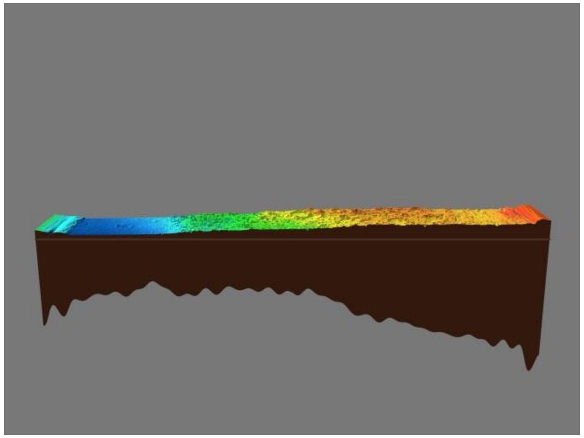




Crustal dichotomy

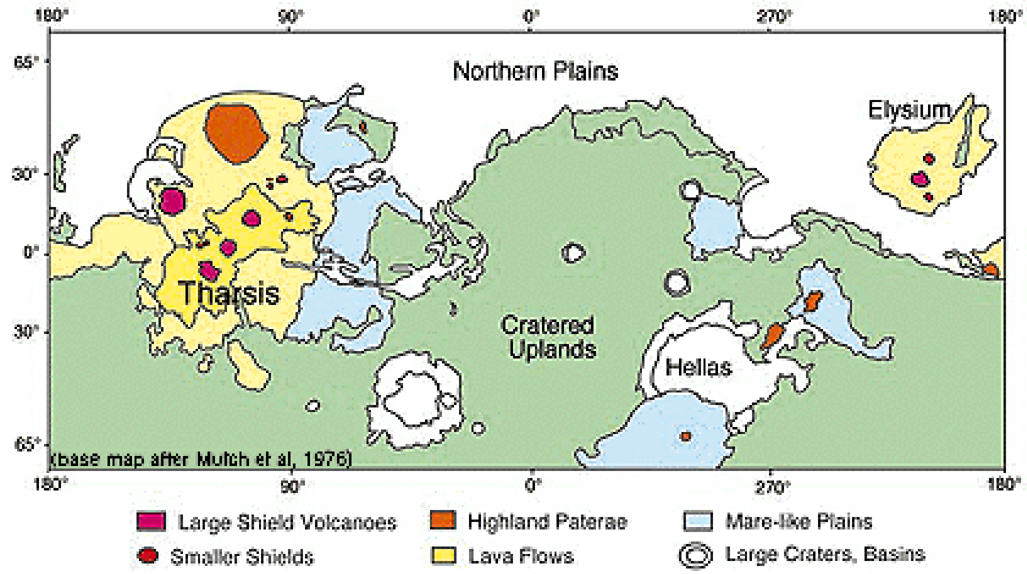


- Variation in surface age
- Crustal thickness variation
  (e.g., Zuber et al., 2000)
- Variation in surface composition (e.g., Christensen et al., 2000)
- Crustal dichotomy formed in the first few 100 Ma

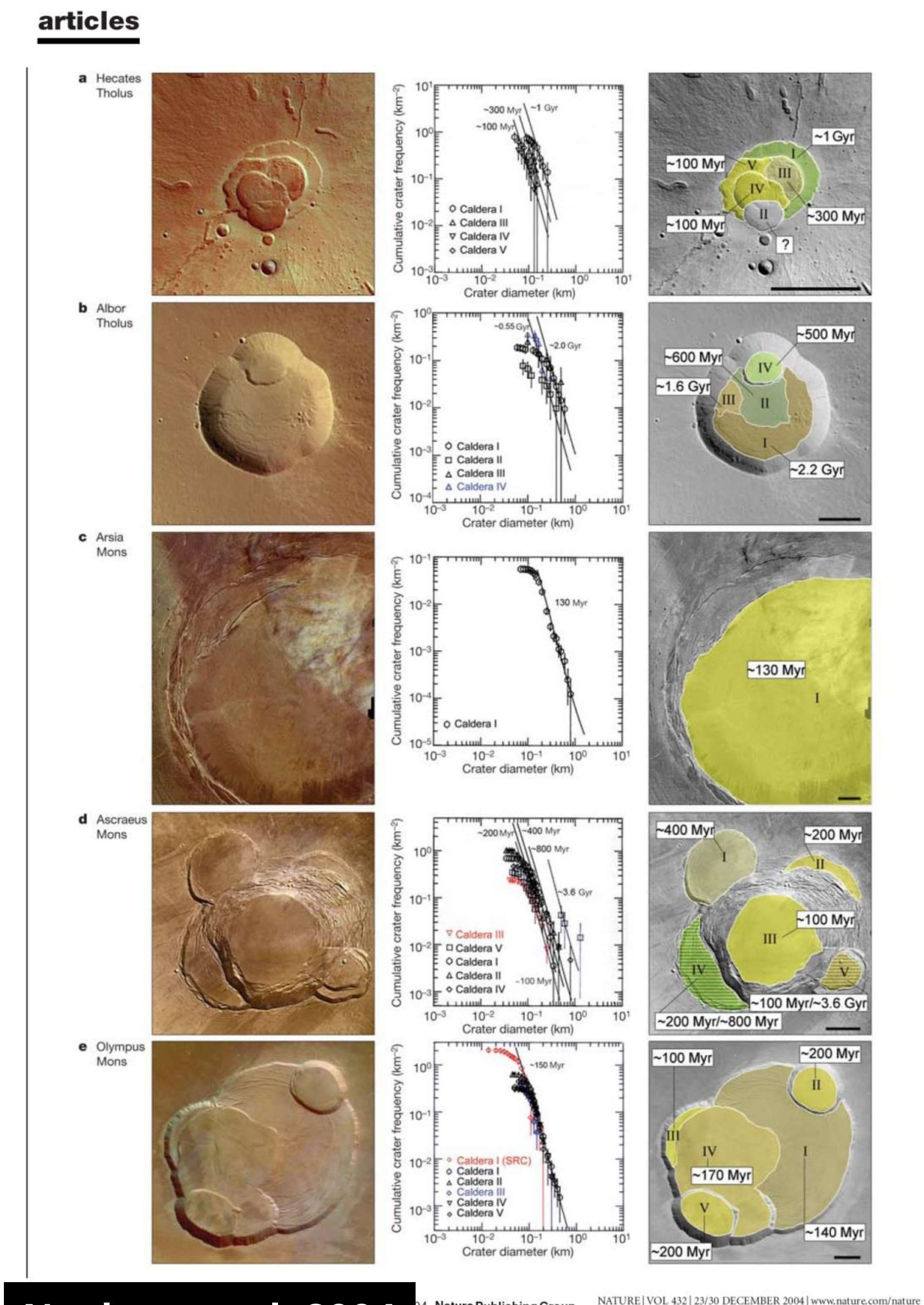


Volcanic activity

- Strong decrease of volcanic activity with time
- Early global distribution later concentration in two provinces (Tharsis and Elysium)
- Bulk of the large volcanic provinces formed in first 1 Ga
- Average crust thickness:
  33 – 81 km



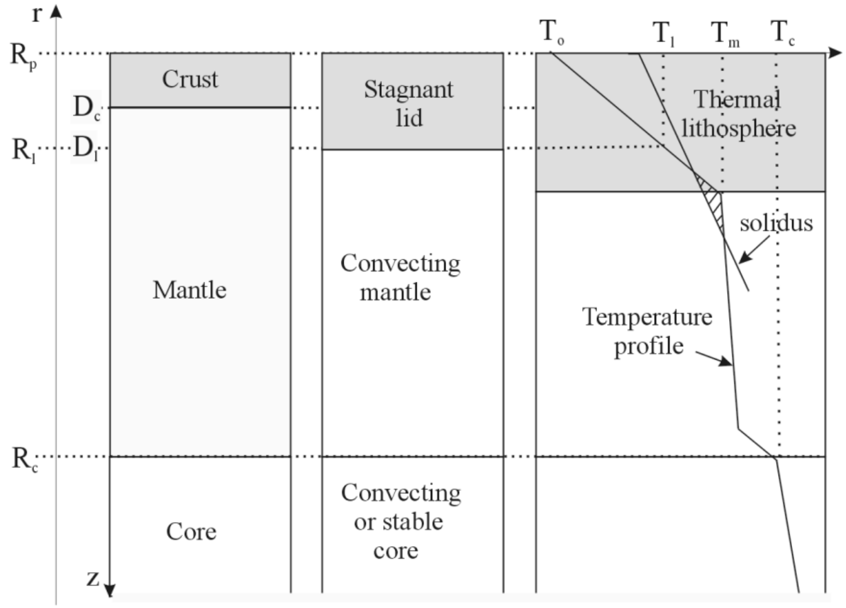
Episodic volcanism but also recent activity in Tharsis and Elysium

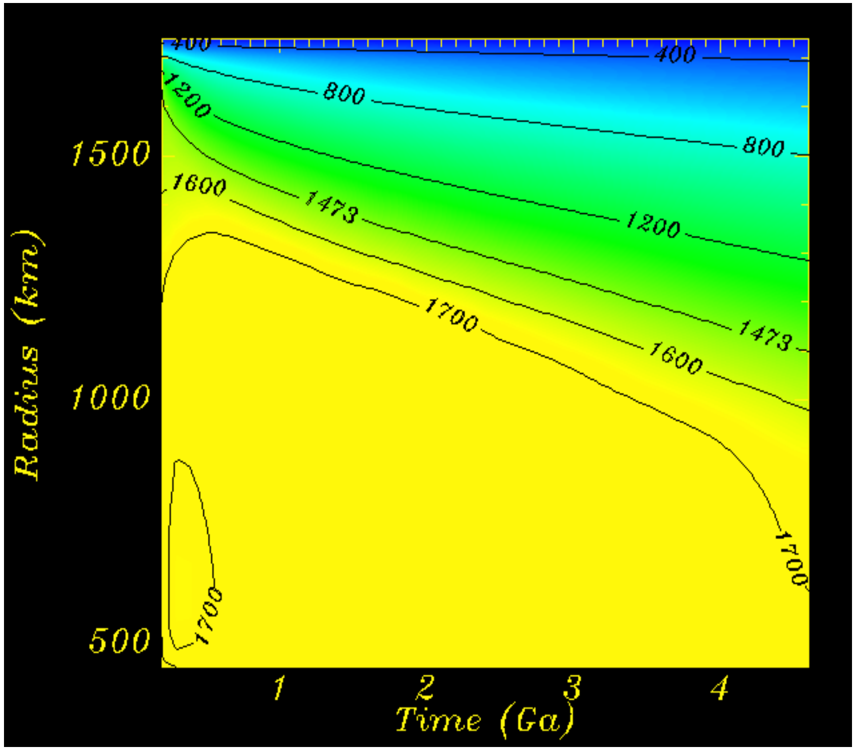


Neukum et al, 2004

Crust Formation in a One-Plate Planet

- Melt production underneath the stagnant lid

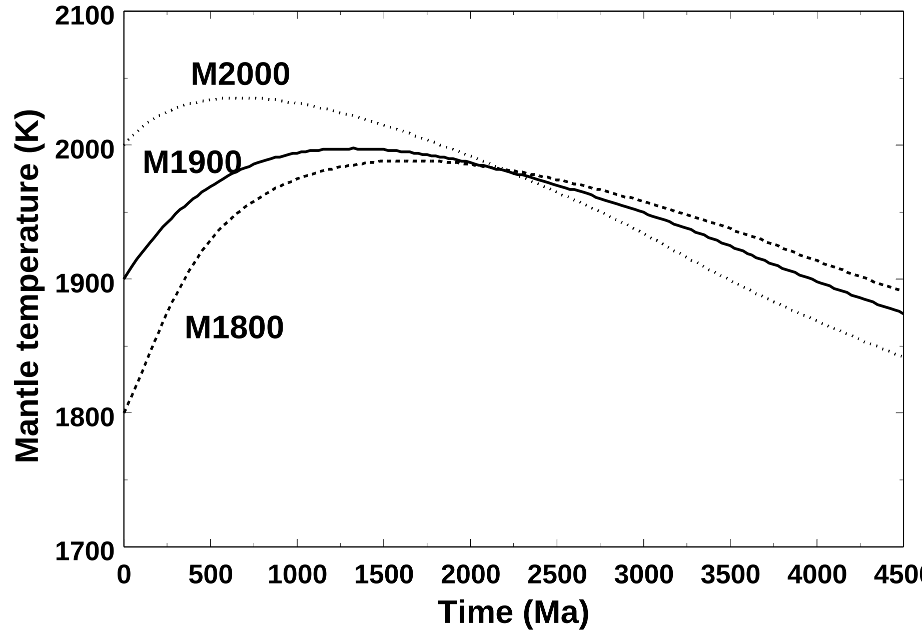


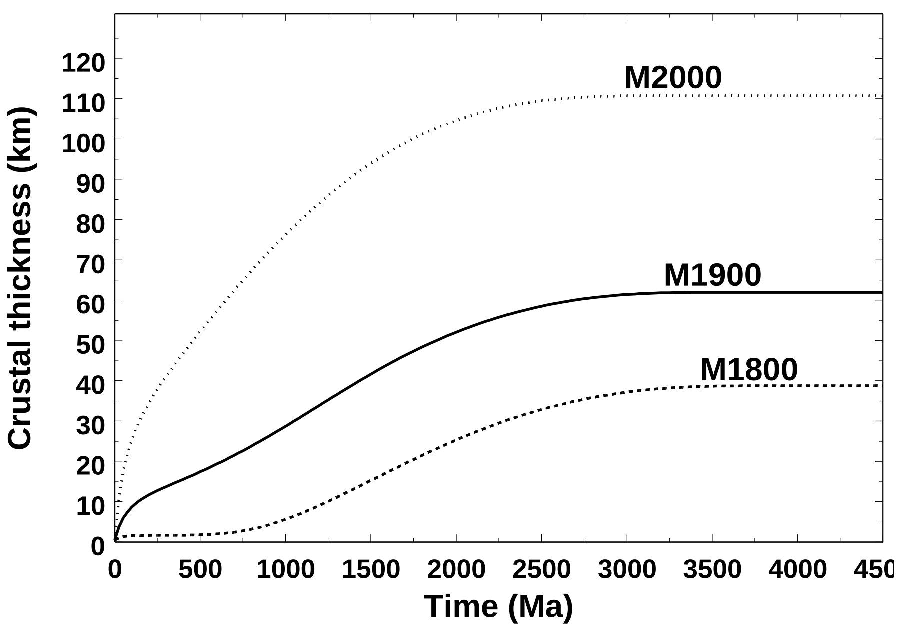


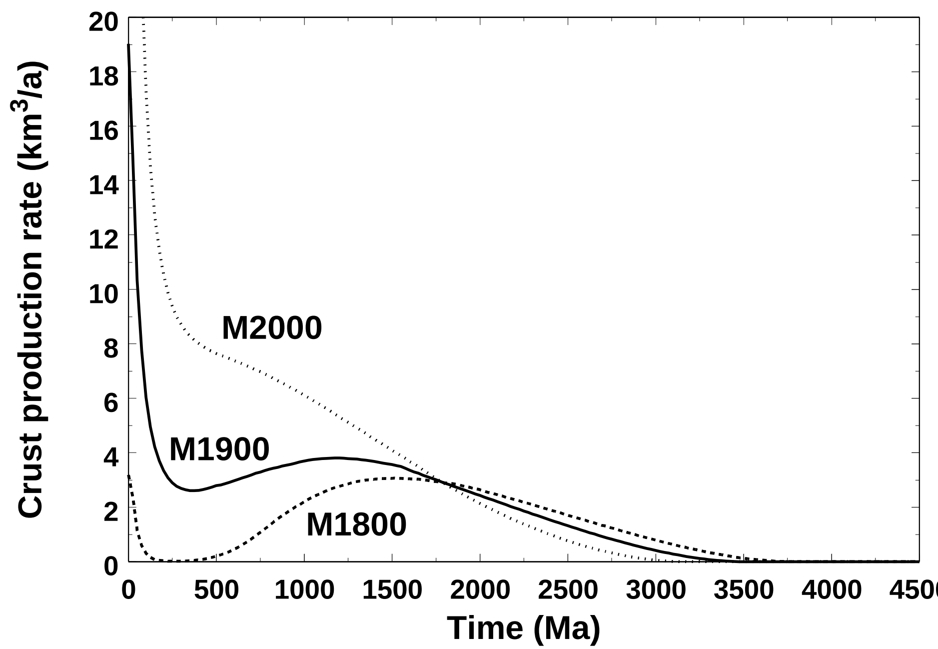
Mantle temperature

Crustal thickness

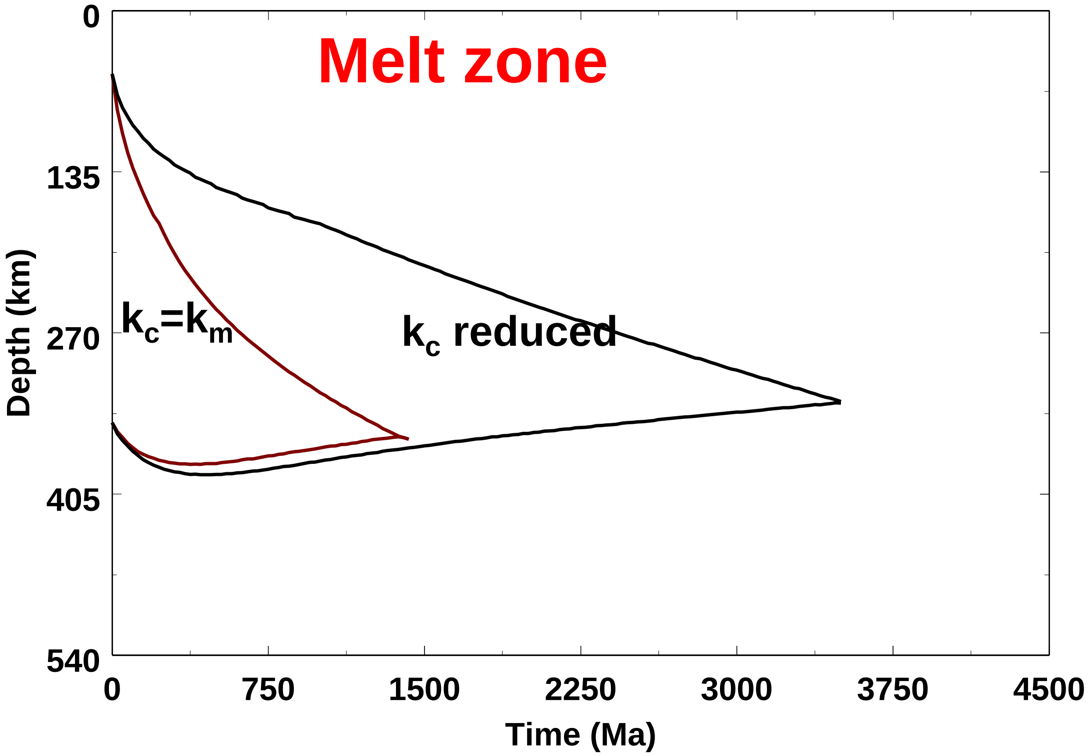
Crustal growth rate

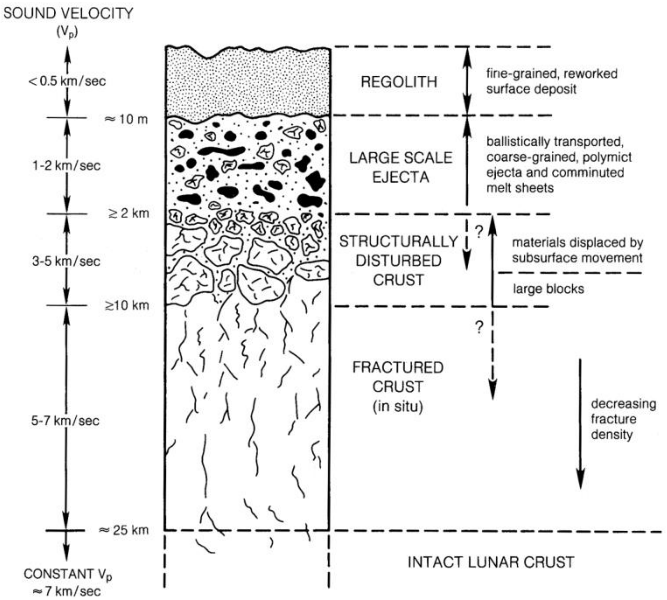






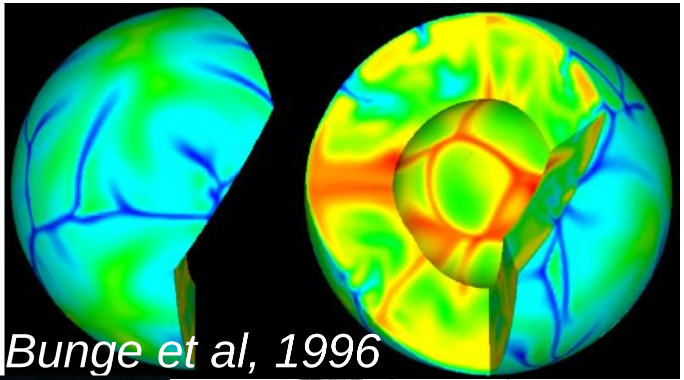
Influence of crustal thermal conductivity



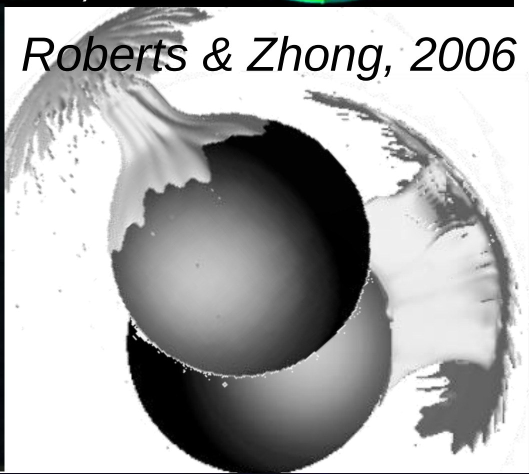


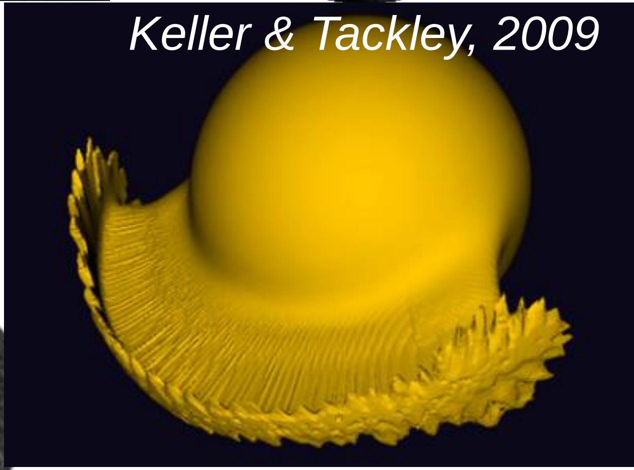
Mars Convection Patterns

- Depth-dependent viscosity produces long-wavelength convection patterns
- Phase-transitions may even induce a degree-1 structure
- A viscosity jump in the mid-mantle also produces long-wavelength convection patterns


Bunge et al, 1996


Breuer et al, 1998


Roberts & Zhong, 2006


Keller & Tackley, 2009


Harder & Christensen, 1996

Without endo-thermic phase transition:

t=1.0 Ga

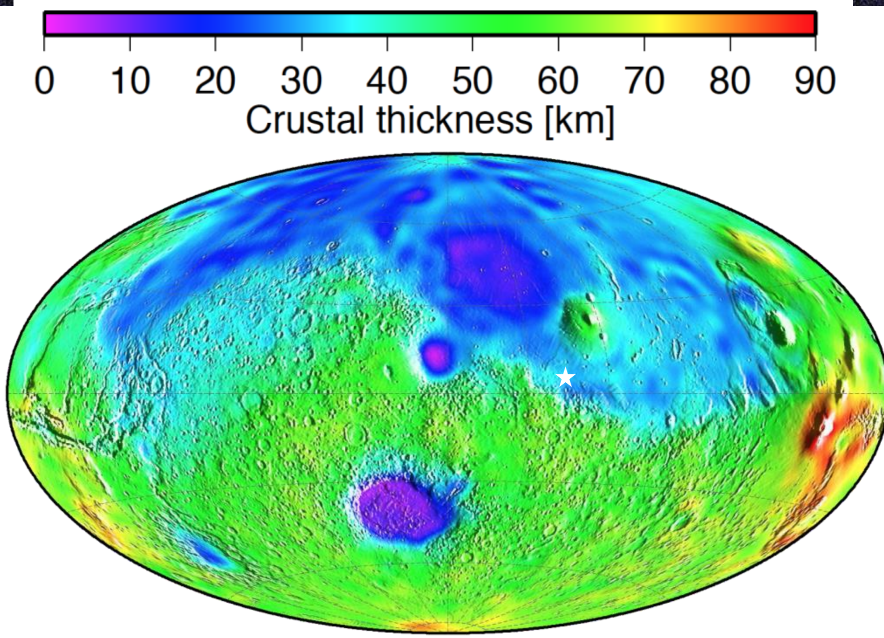
t=4.5 Ga

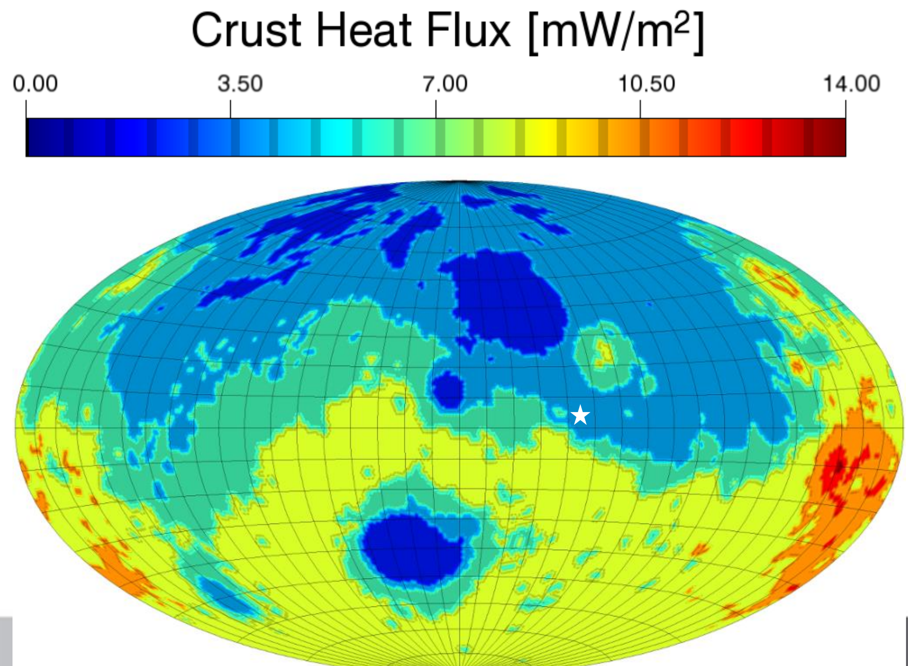
Including endo-thermic phase transition:

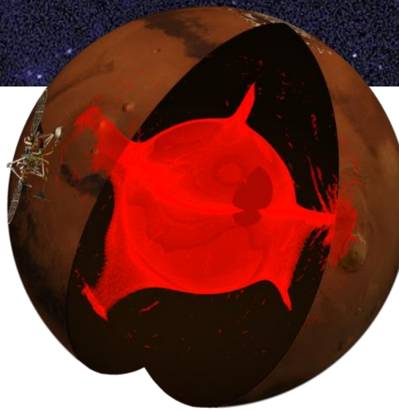
Buske, 2006

Investigate the role of insulating crust and depth-dependent viscosity on convection pattern

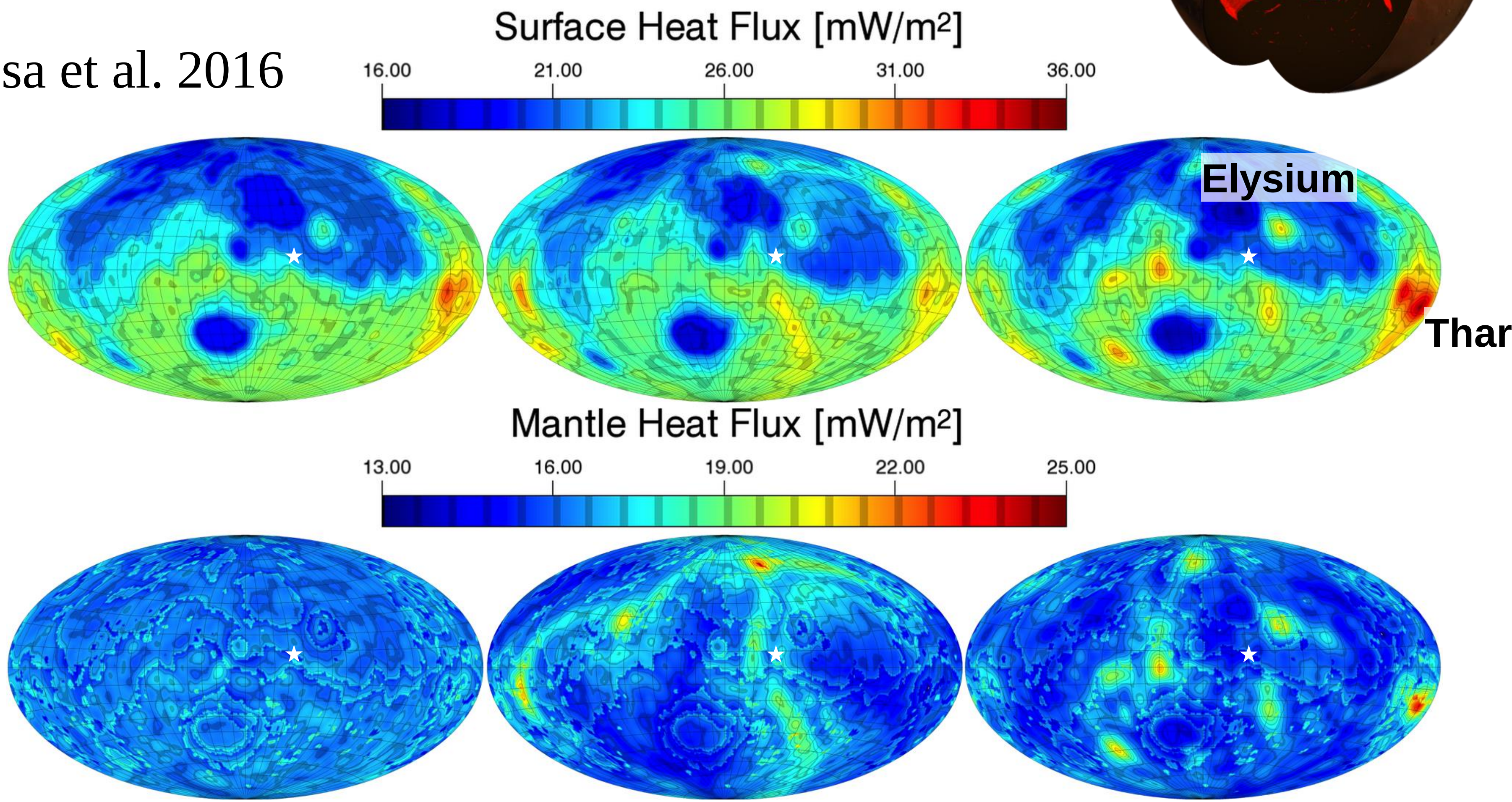
- 3D fully dynamical simulations of interior thermal evolution
- Assuming a crust
  - using the model of *Neumann et al. [2004]*
  - with low conductivity (3 W/m K)
  - enriched in HPE *[Hahn et al., 2011]*







Plesa et al. 2016



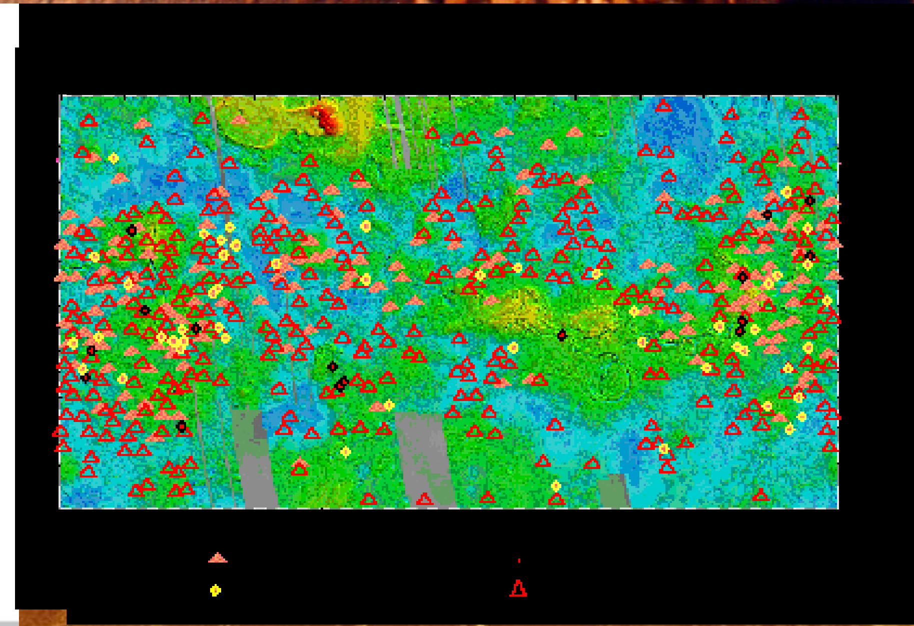
V = 6 cm^3/mol V = 6 cm^3/mol V = 10 cm^3/mol

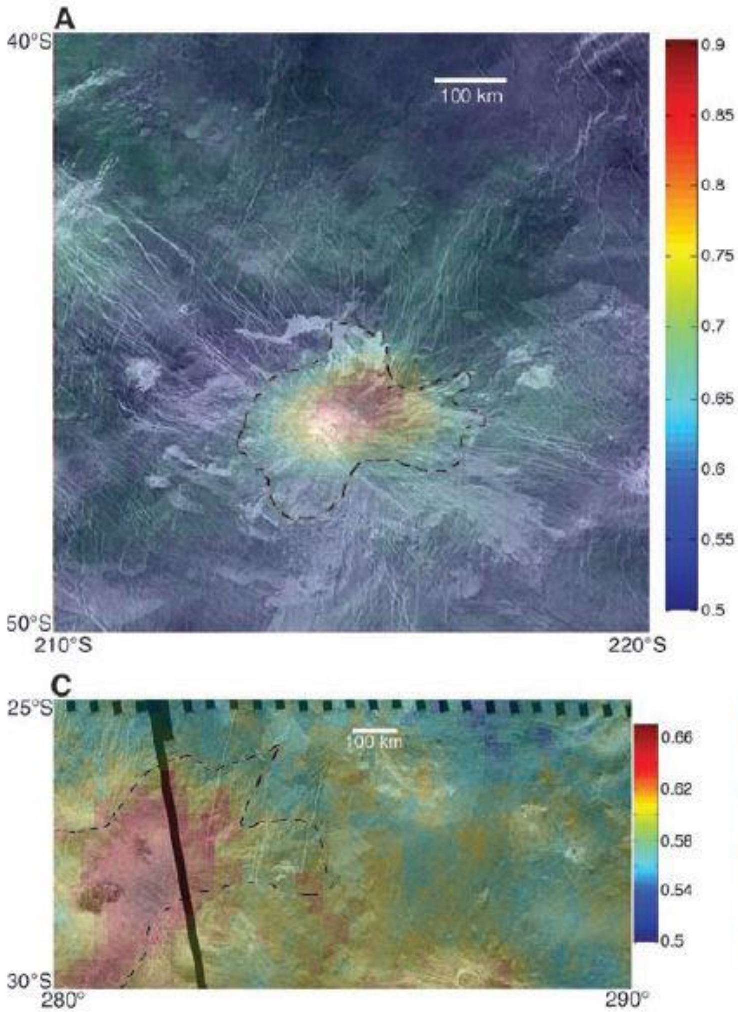
Viscosity jump = 50

Venus

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

- Resurfacing of the surface
  - Catastrophic resurfacing (~ 500-700 Ma) renewed the surface
  - Equilibrium resurfacing





Thermal emissivity of the surface: Nine – hotspots - with volcanism, broad topographic rises and positive gravity anomalies have been identified

Smrekar et al. 2010

Venus:

Terra:

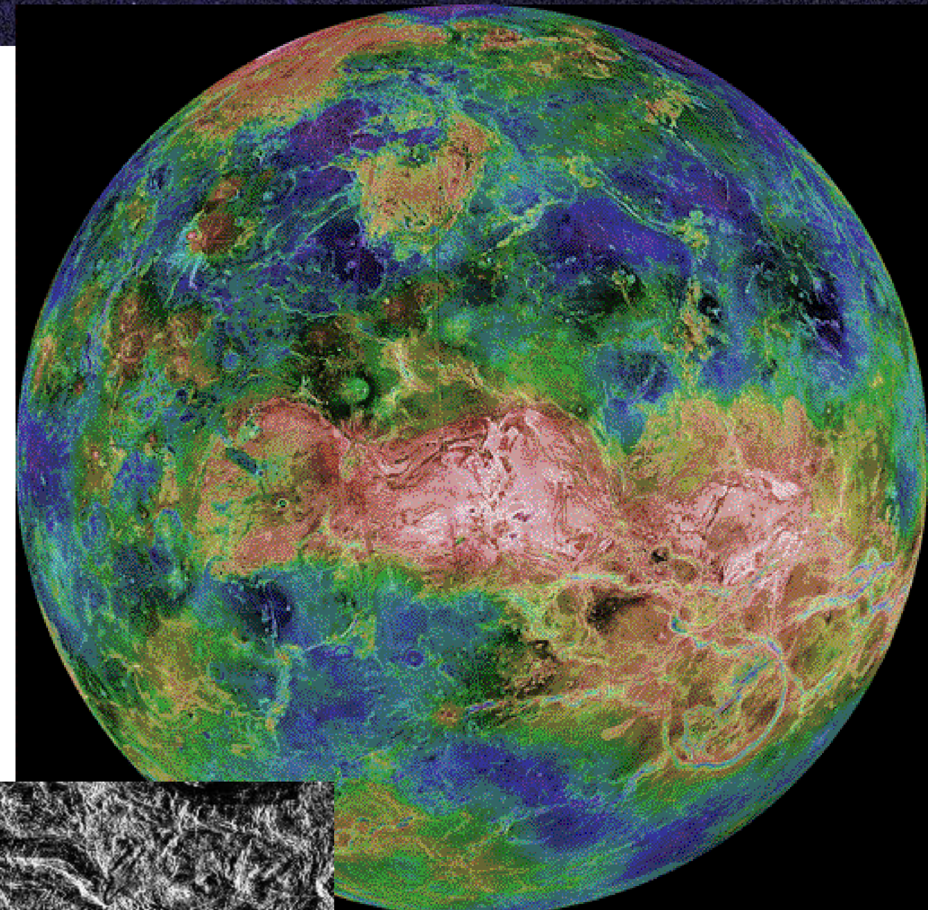
continental-like structures

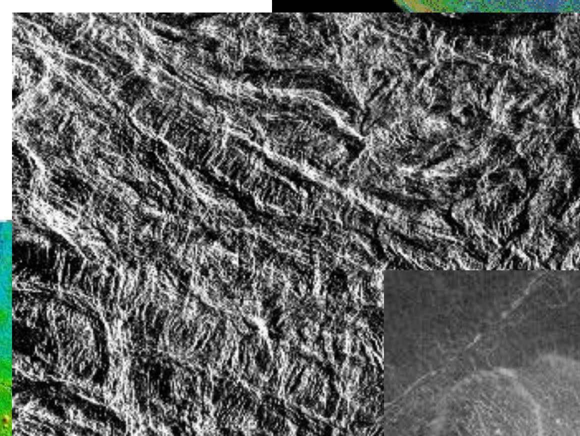
Tessera:

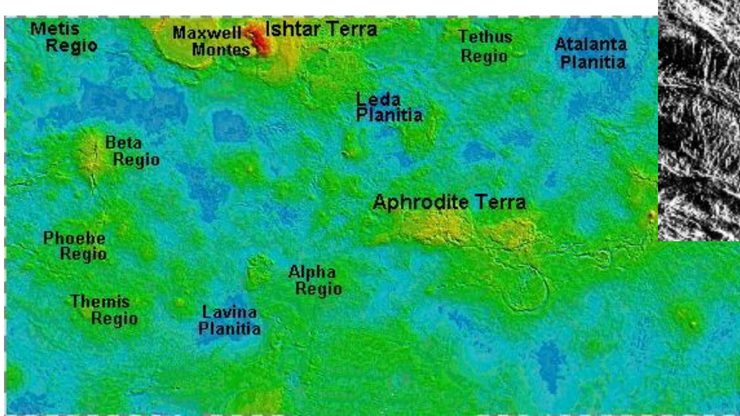
highly deformed crust

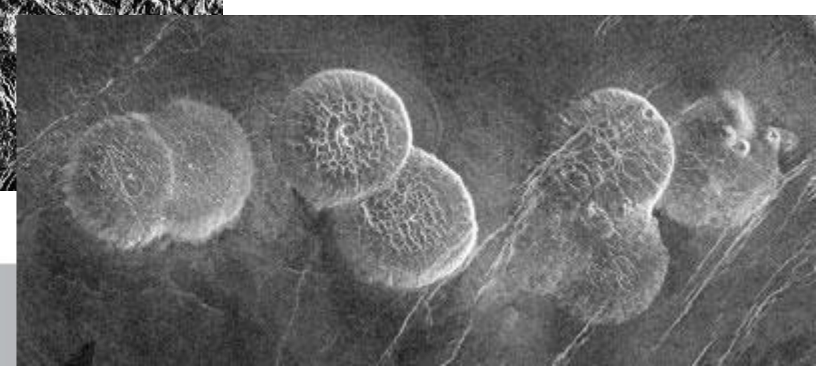
Coronae:

circular domes



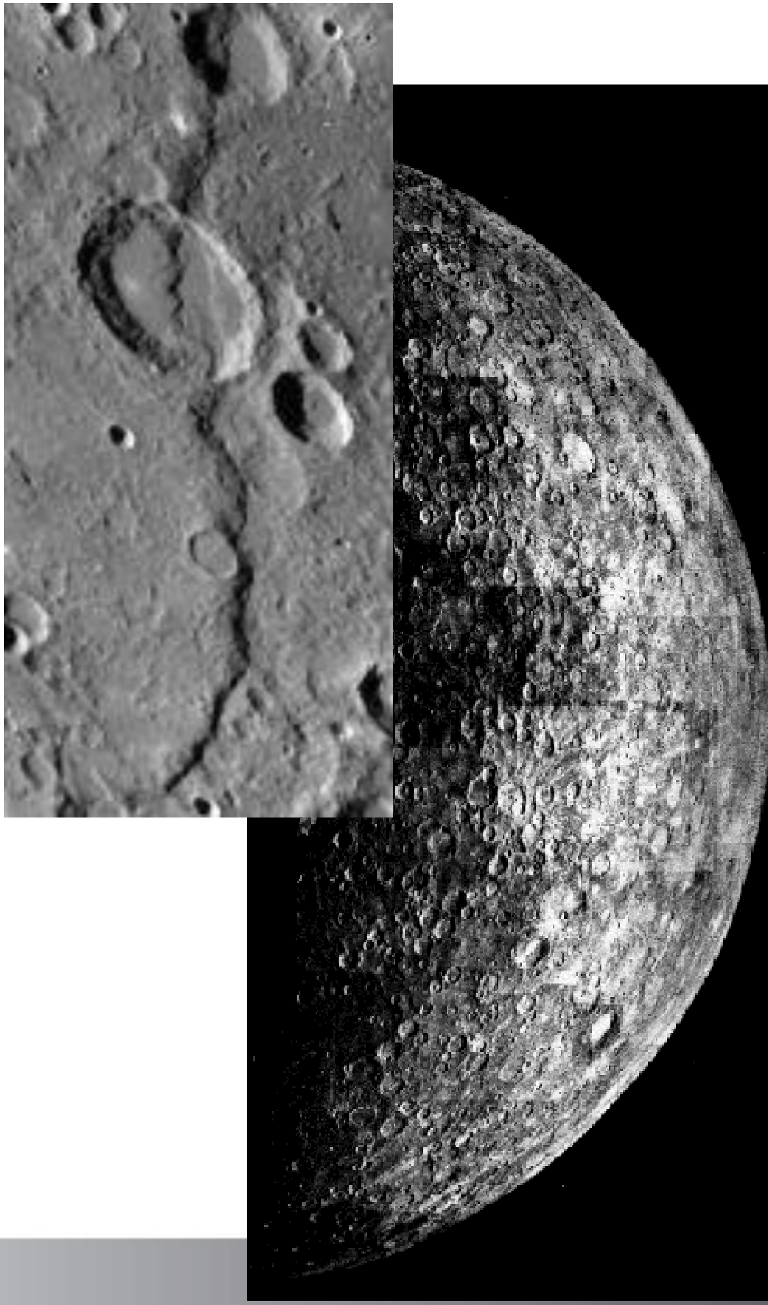






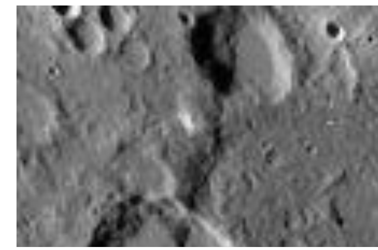
Mercury

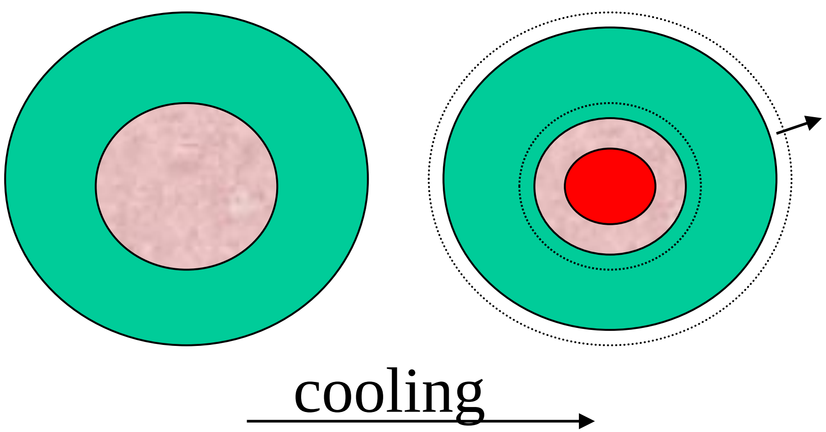
- Crater and scarps cover the surface
  - old surface
  - ~ 7 km decrease of radius caused by thermal contraction since ~4 Ga (Byrne et al. 2014)

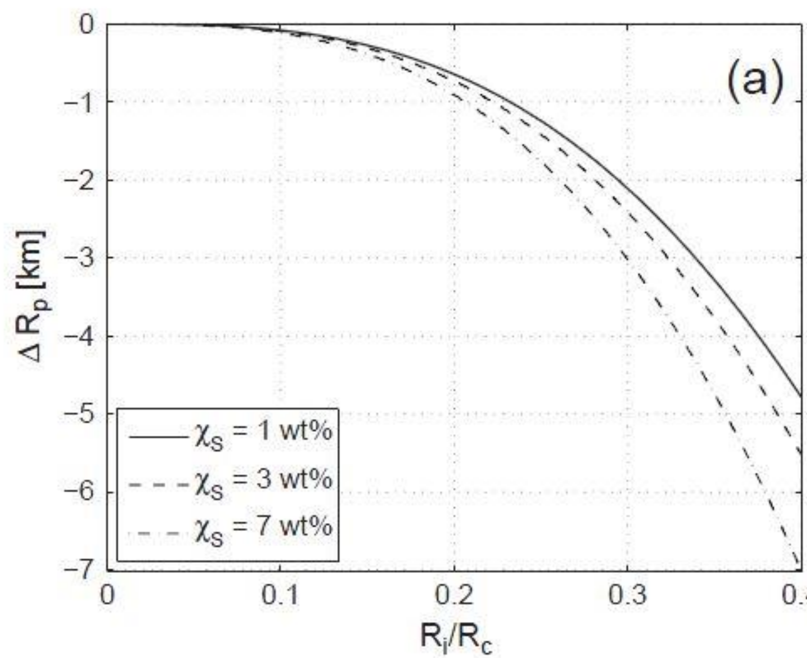


Global contraction of Mercury

- What contributes to the contraction
  - Cooling
  - Phase changes
    (e.g. inner core growth)

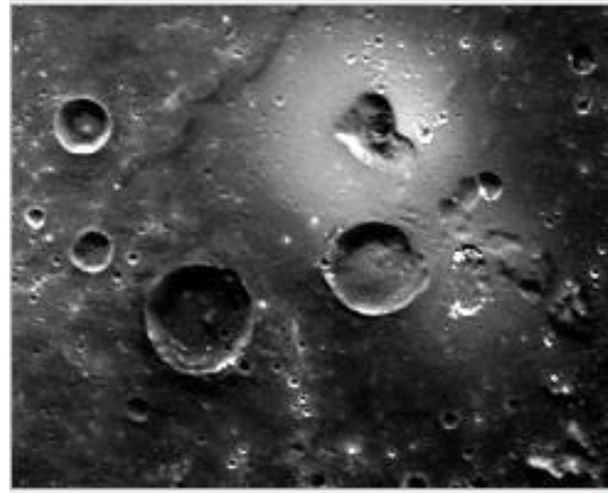


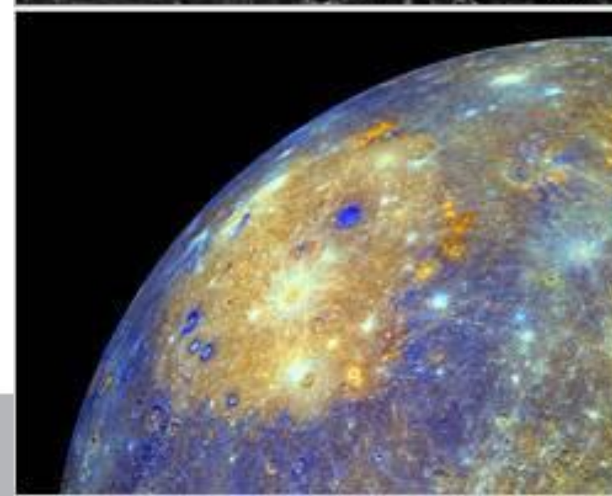




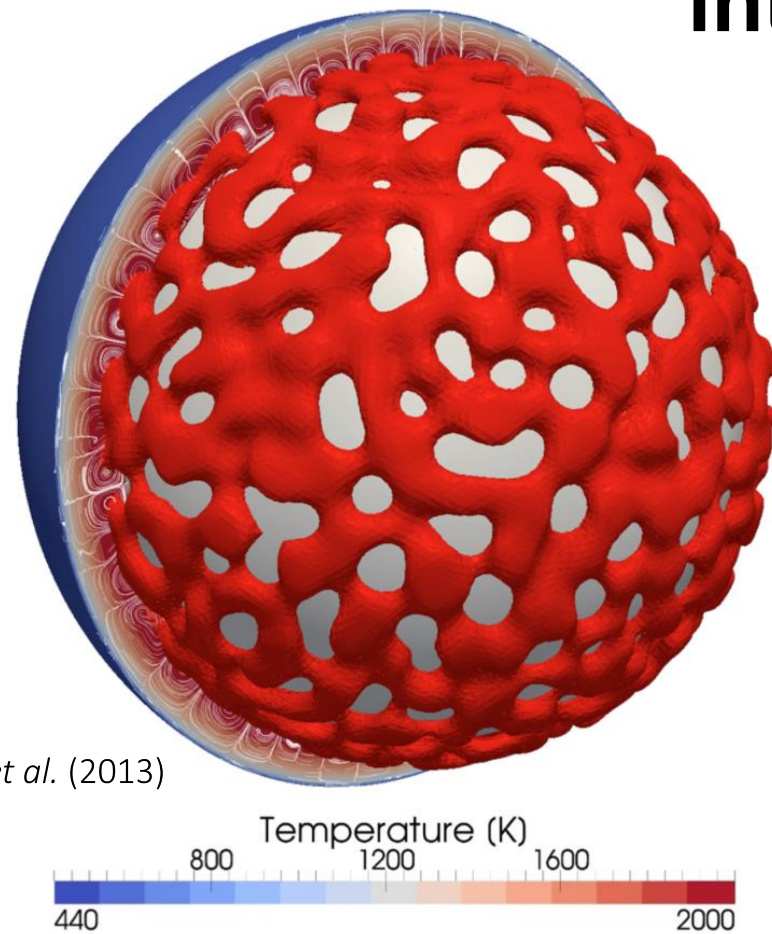
Mercury

- Messenger flybys revealed volcanic resurfacing
- Volcanism more widespread than previously expected

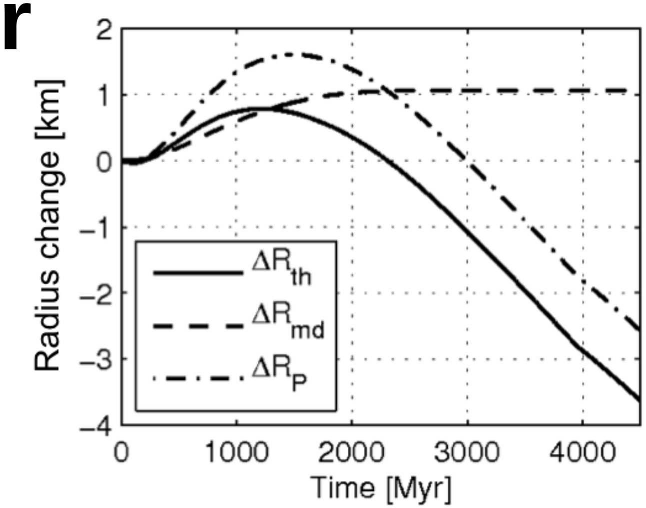


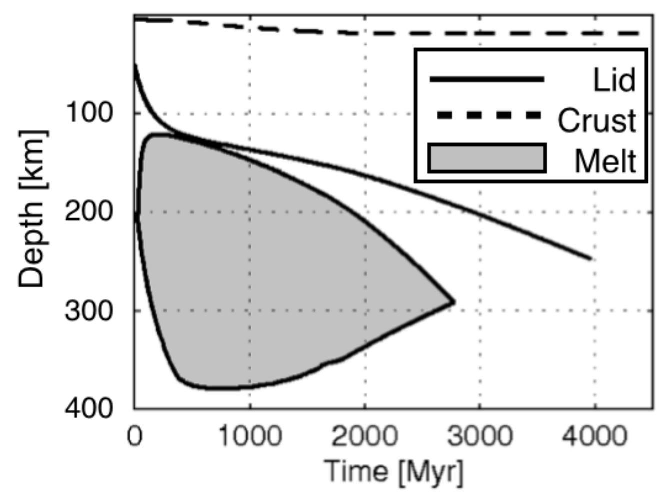


Thermo-chemical evolution of Mercury's interior



Tosi et al. (2013)

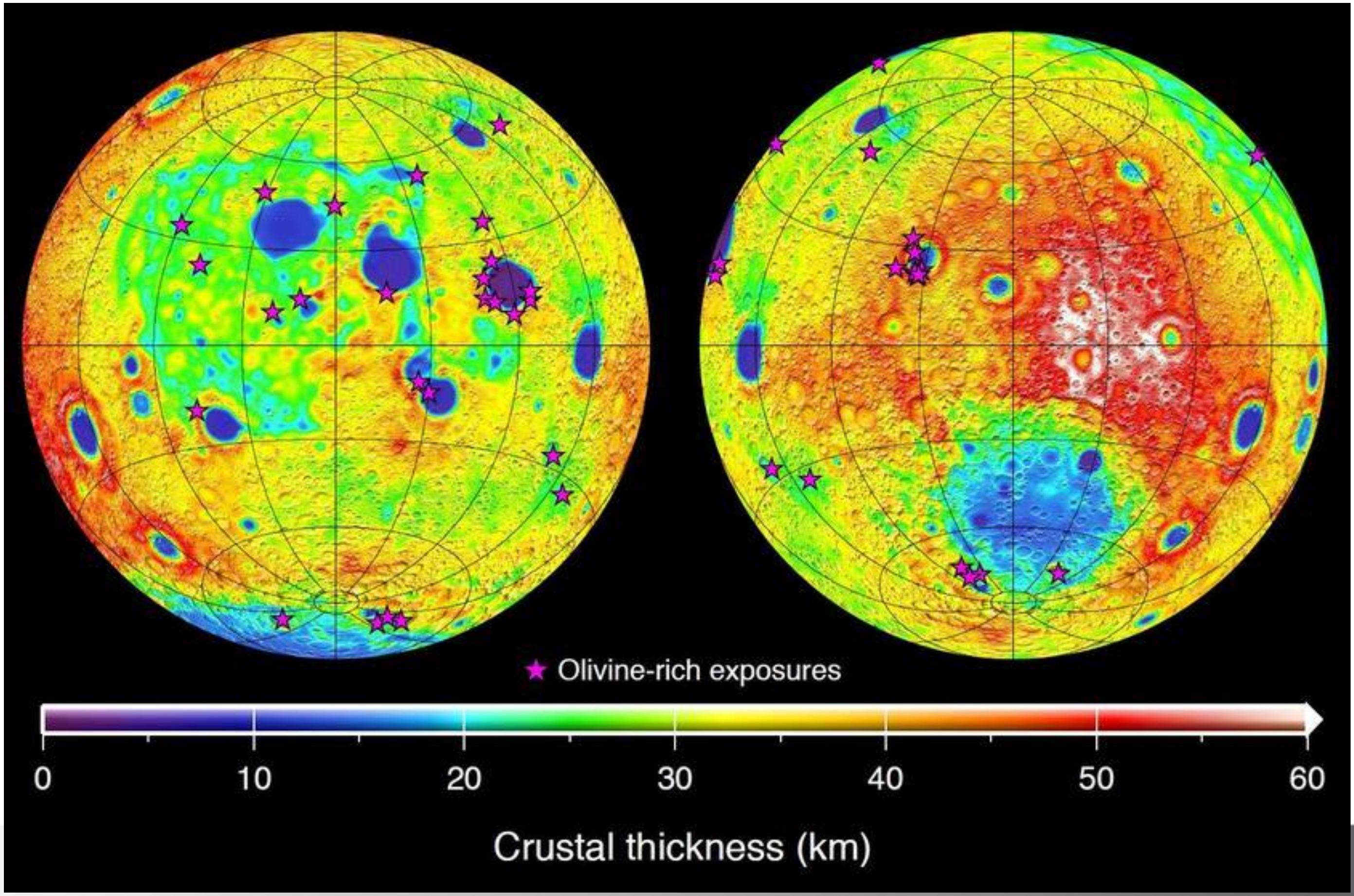




Planetary contraction inferred from surface tectonics and volcanic history from imaging constrain the thermal evolution of the interior and the duration of mantle convection

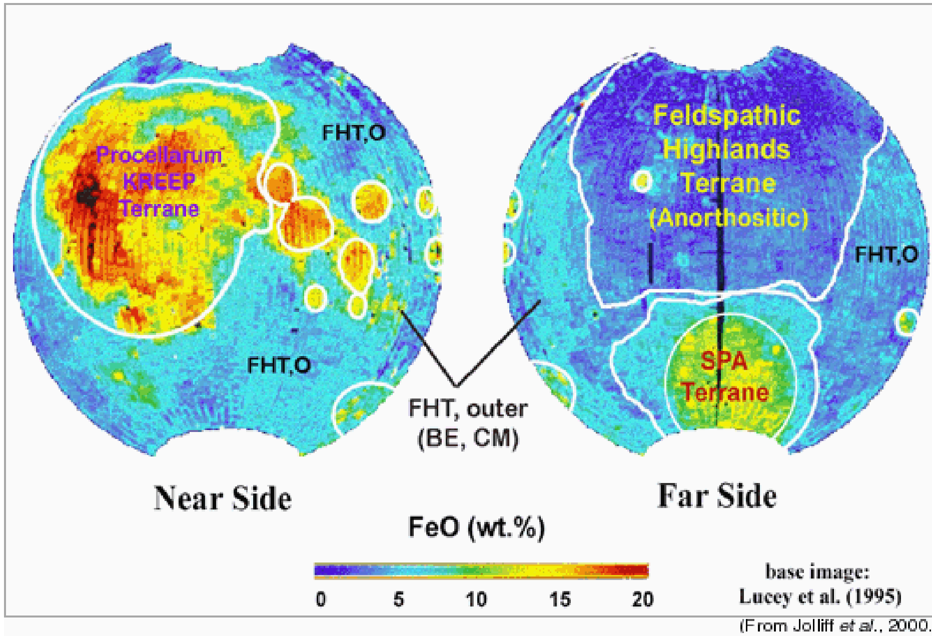
Moon

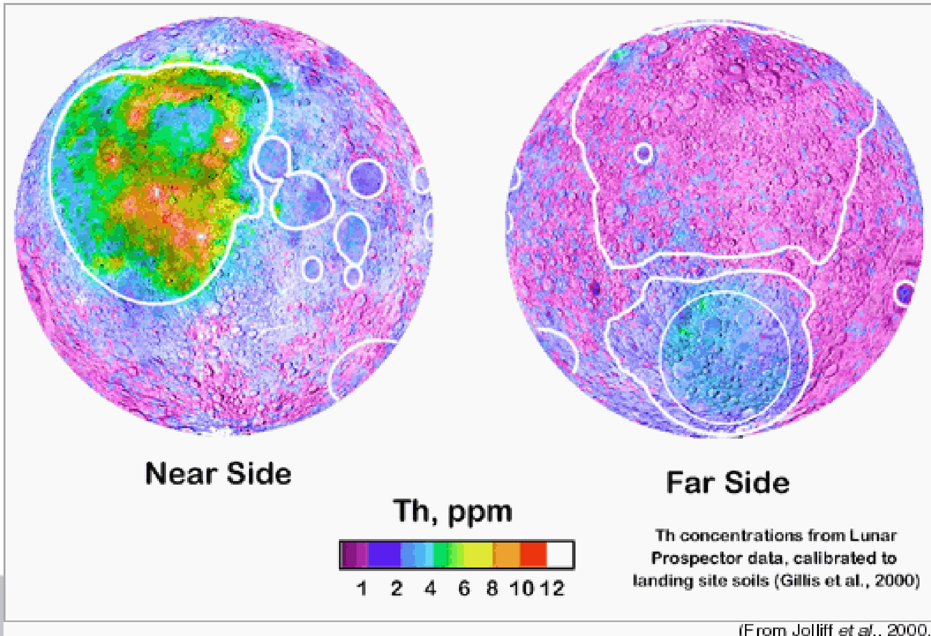
- Crustal dichotomy: lunar mare are primarily found on near side



Moon

- Major geological units:
    - Porcellarum KREEP Terrane (PKT) (enriched in radiogenic heat sources)
    - South Pole Aitken Terrane (SPAT)
    - Feldspathic High-lands Terrane (FHT)

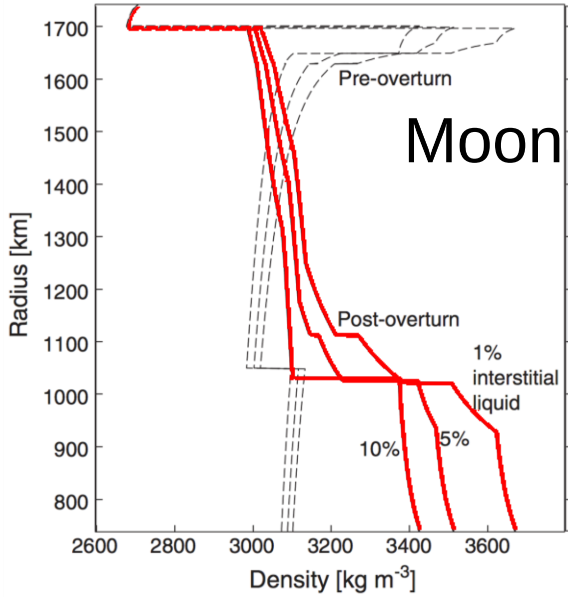


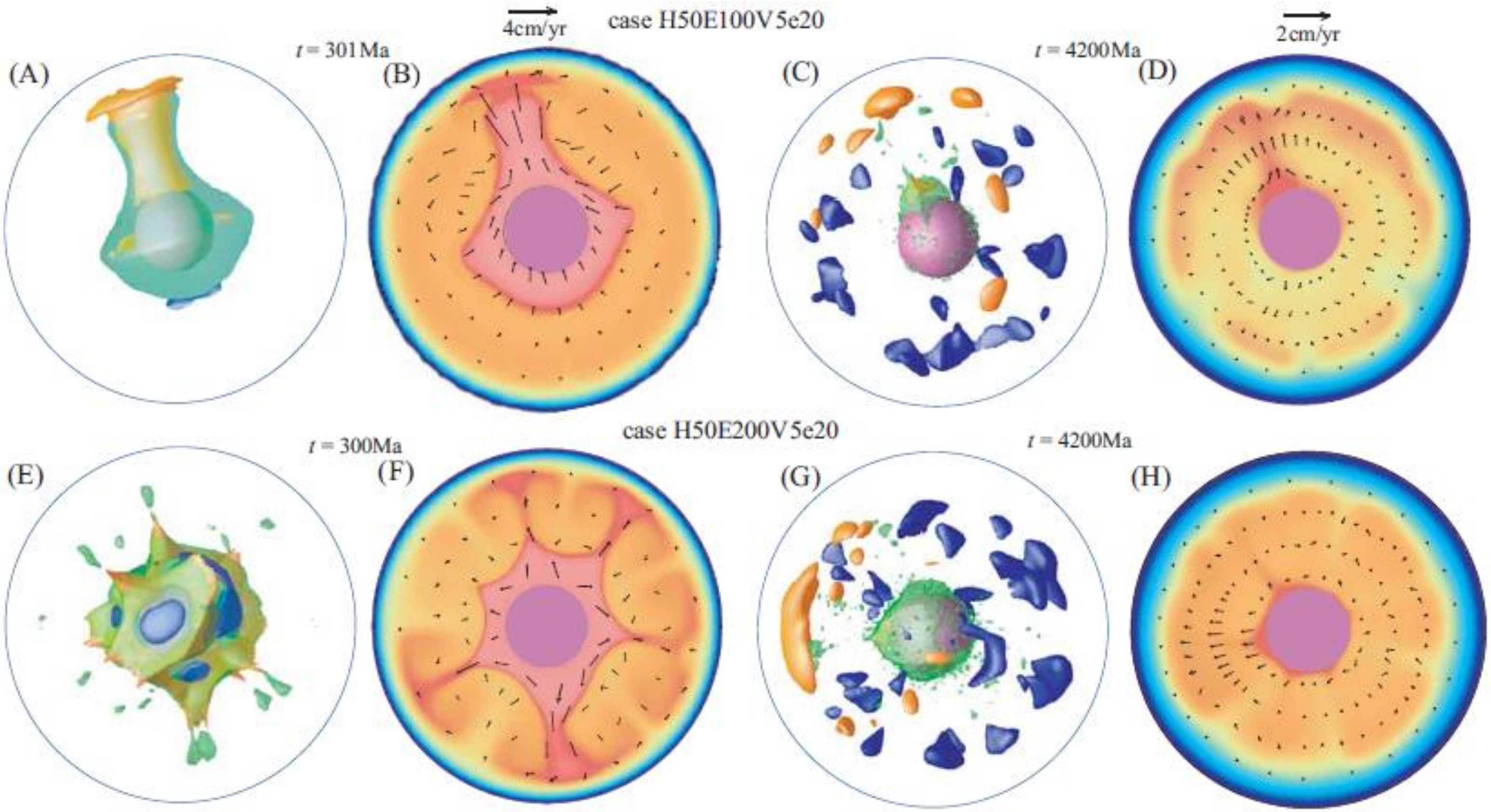


Origin of PKT region

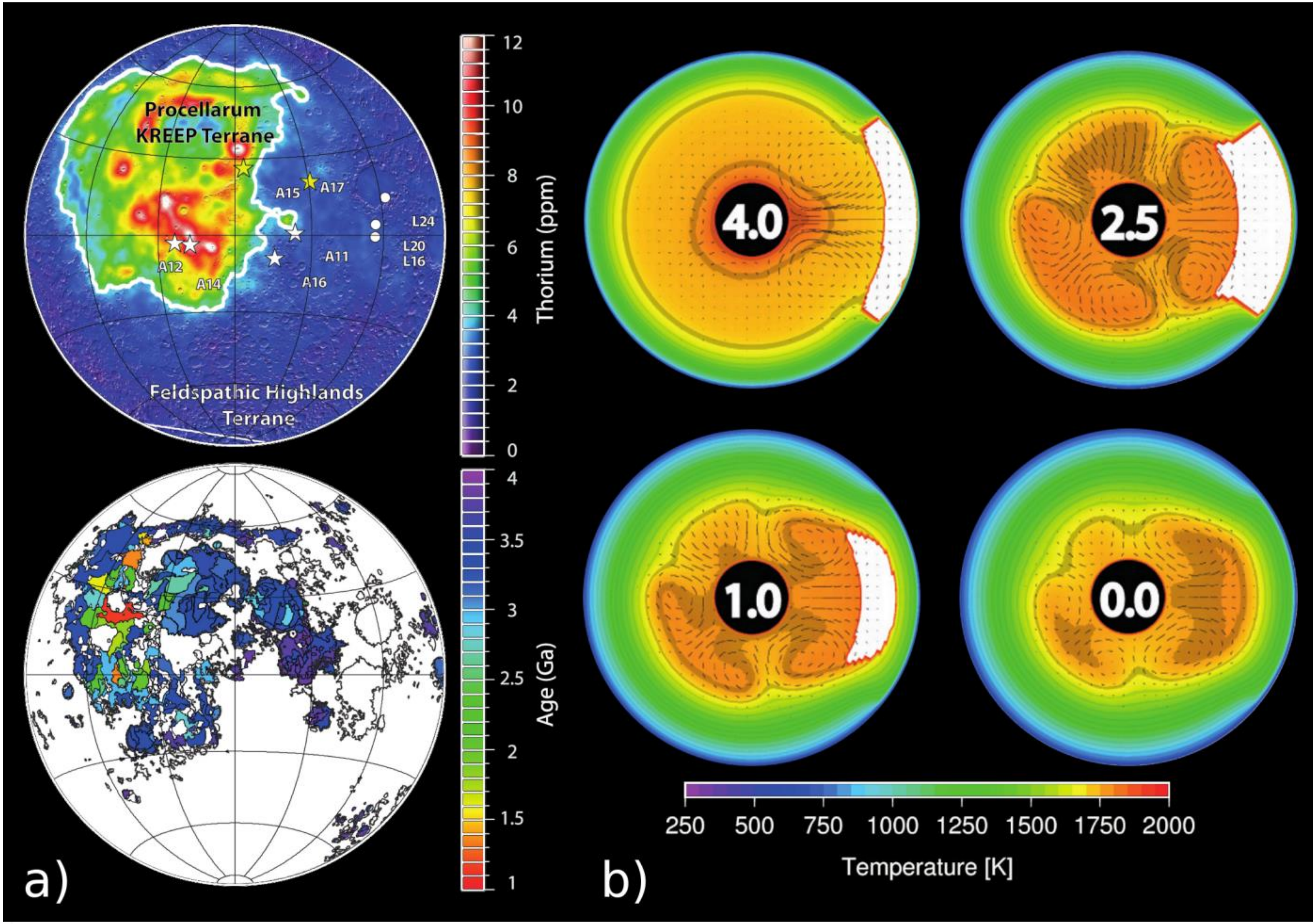
- Magma ocean overturn of dense ilmenite layer below crust
- Low degree instability of dense ilmenite bearing cumulate enriched in KREEP

[Elkins-Tanton et al., 2011]





Zhang et al. 2013



Laneuville et al. 2013

Similarity and Diversity of Terrestrial Planets and the Moon

- Interior structure and composition
- Surface structures (volcanic and tectonic)
- Magnetic field
- Atmosphere

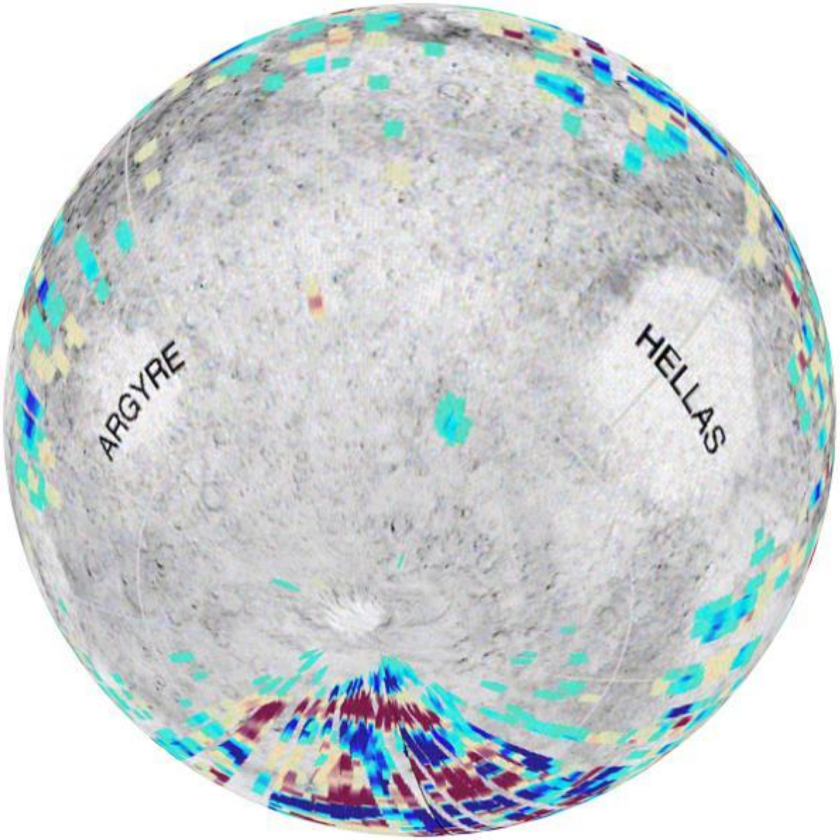
Planetary Data

	Mercury	Venus	Earth	Mars	Moon
Radius	0.38	0.95	1.0	0.54	0.27
Mass	0.055	0.815	1.0	0.107	0.012
Density [kg/m³]	5427.	5243.	5514.	3933.	3344.
MoI	0.346	?	0.3308	0.3662	0.3940
R_c/R_p	0.83	(0.55)	0.546	0.5	0.25
Dipole Moment $[10^{19}\ A\ m^2]$	4.9	-	7980.	-	-

But we find remnant magnetization of old crust on Earth, Mars, the Moon and Mercury (Venus we don‘t know)

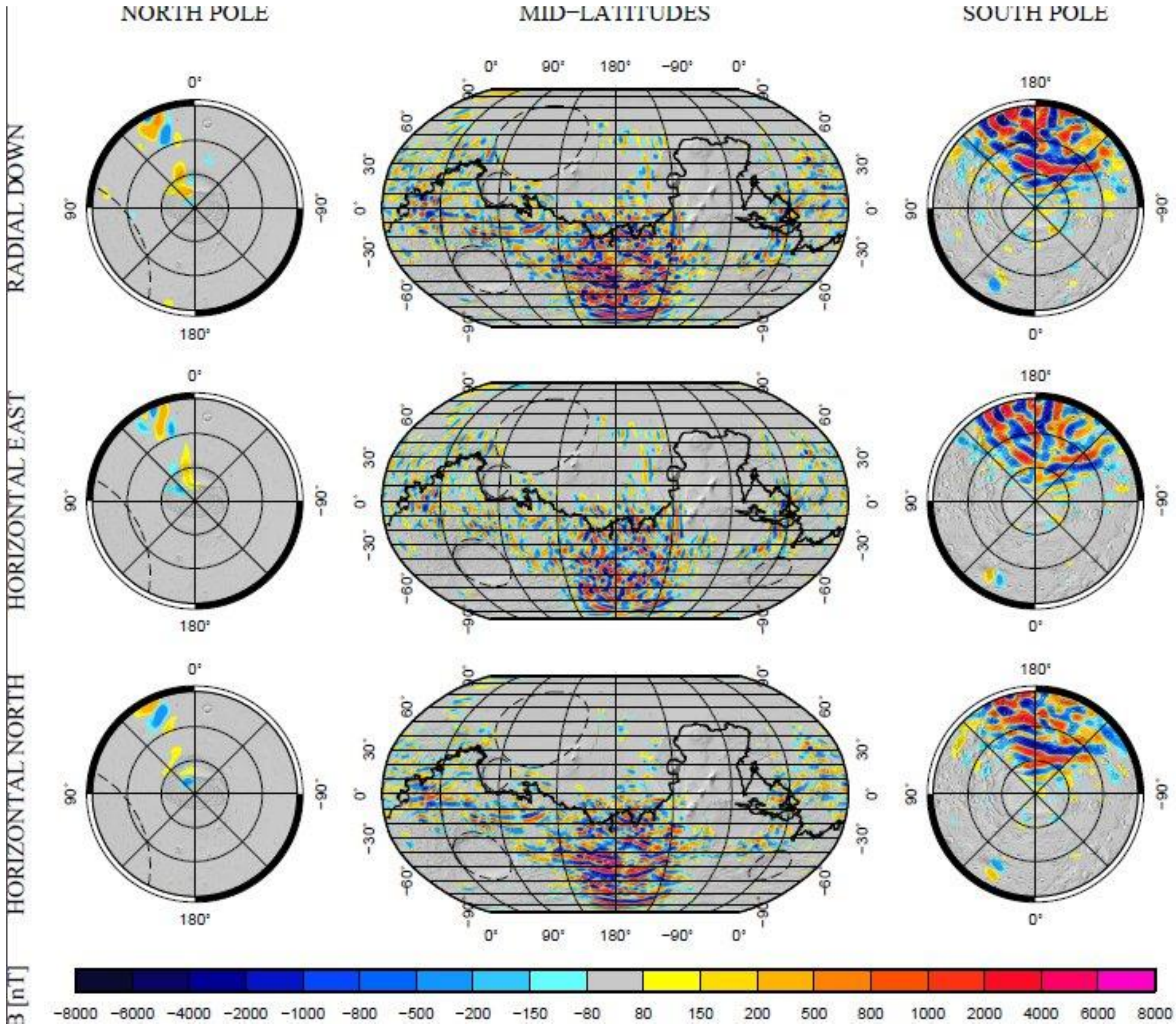
Mars Magnetic field history

- No present-day dynamo
- Strong magnetisation of oldest parts of the Martian crust
- Magnetization of old SNC meteorite
- No magnetisation of large impact basins



⇒ Dynamo action before the large impacts ~4 Ga

DLR



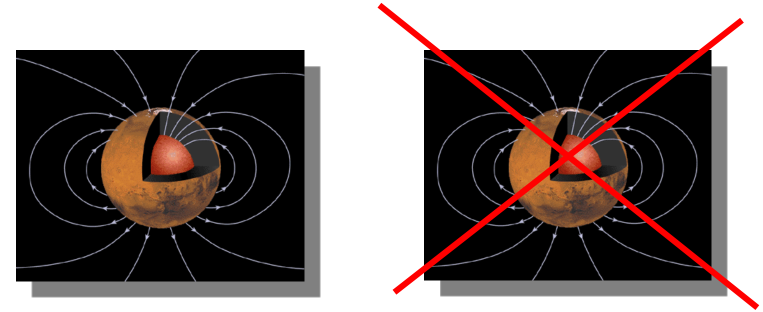
Morschhauser et al. 2014

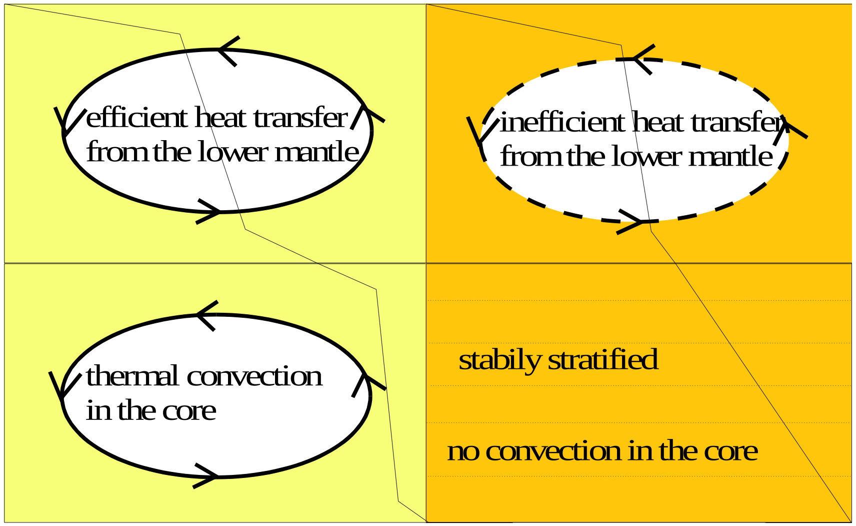
Thermal dynamo

- Fluid motion in the liquid iron core due to thermal buoyancy (=> cooling from above)
- ‘Critical‘ heat flow out of the core

$$q_{crit} = k \frac{\partial T}{\partial r}\bigg|_{ad} = k \frac{\alpha g T}{c_p}$$

$$q_{cm} > q_{crit}$$

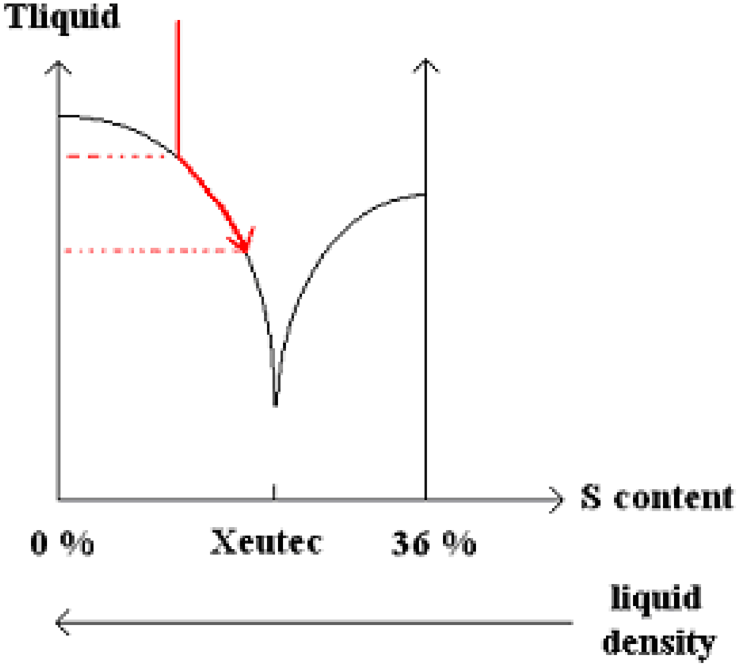




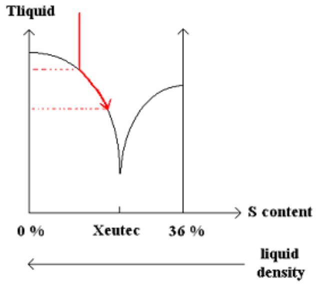
DLR

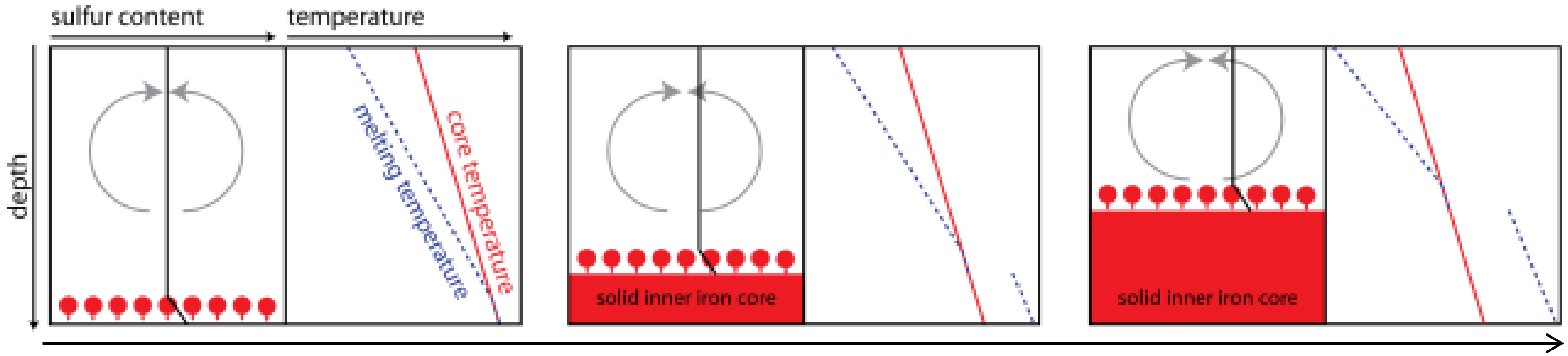
Chemical Dynamo

- Existence of light alloying elements in the core like S, O, Si
- Core temperature between solidus and liquidus
- Compositional bouyancy released by inner core growth



‘Classical‘ Inner Core Growth

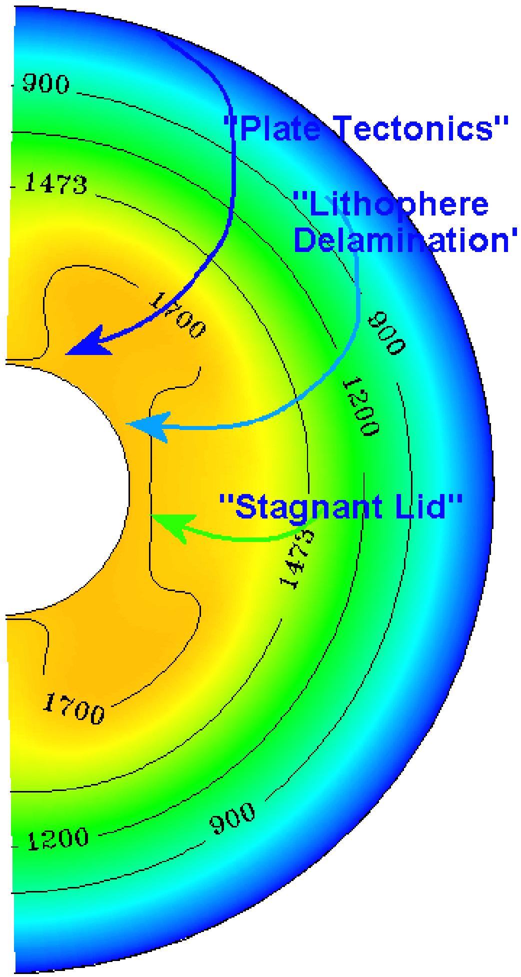




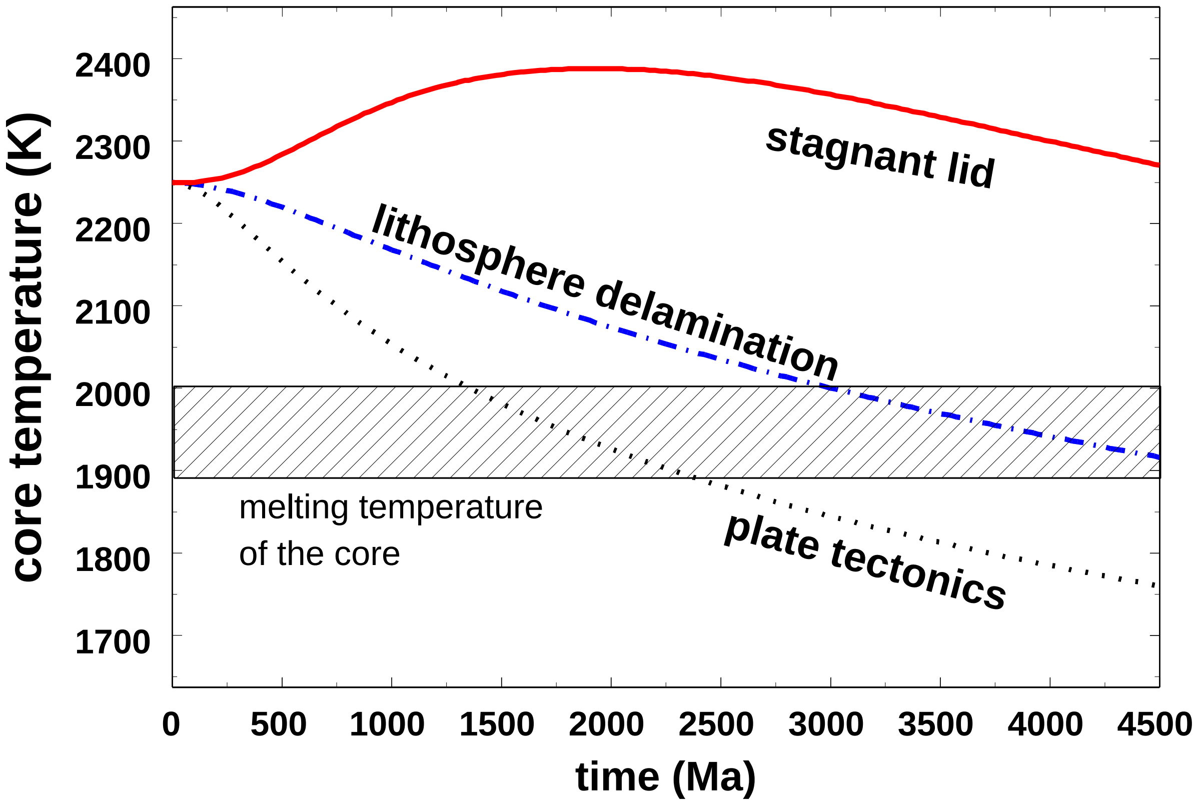
Magnetic field generation in upper fluid core

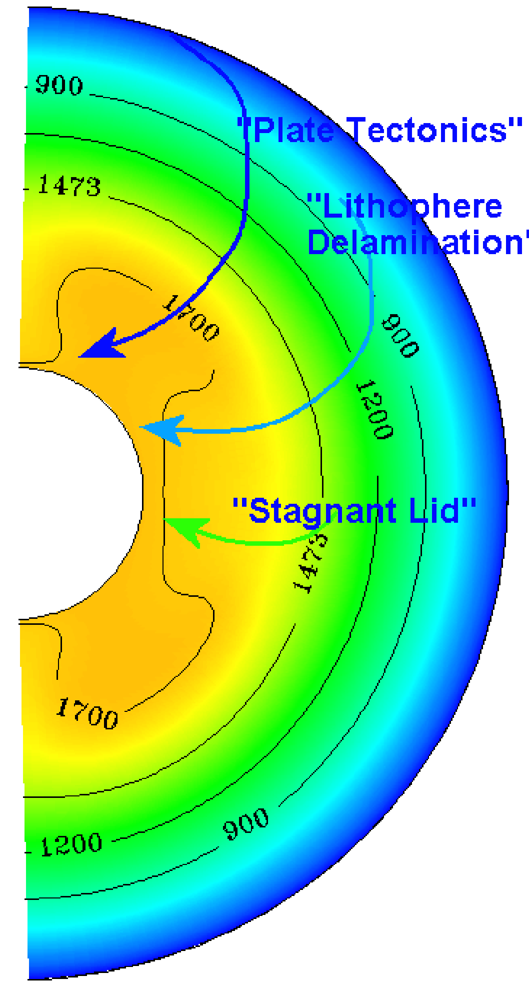
Thermal and chemical dynamo

- Mantle 'dicates' core cooling and dynamo action
- To study the magnetic field evolution of a planet, we need to know the heat transport mechanism in the mantle

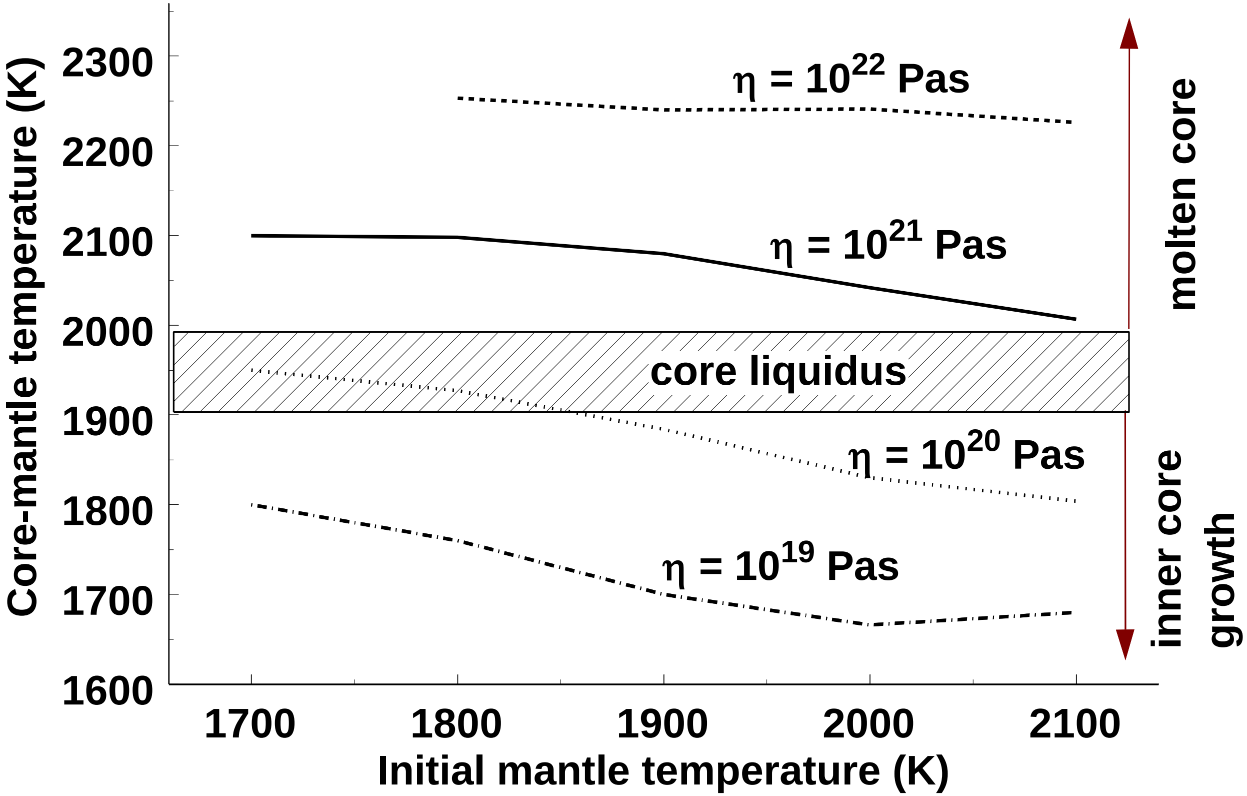


Temperature evolution of the core

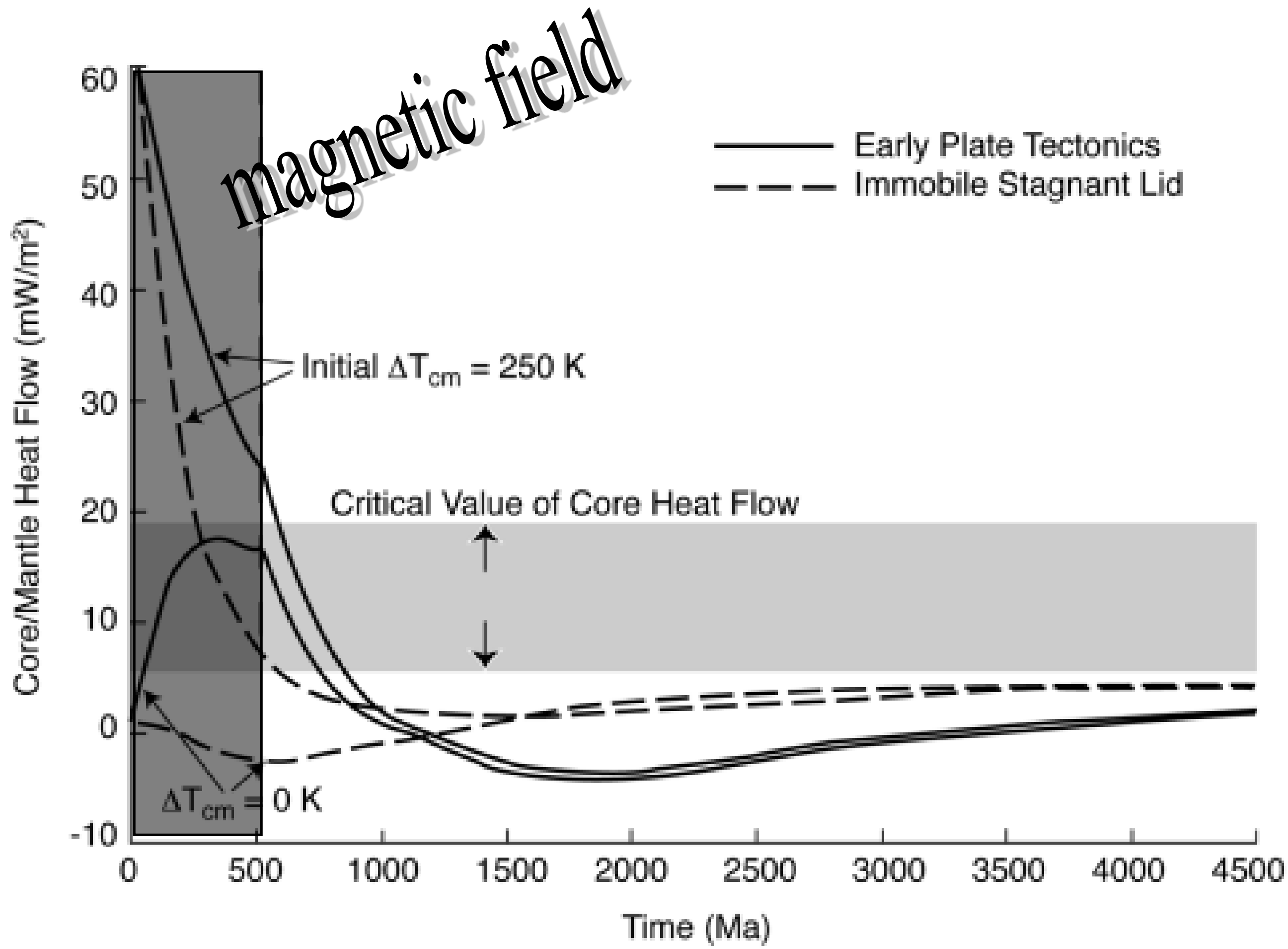




Present core-mantle temperature depending on mantle rheology



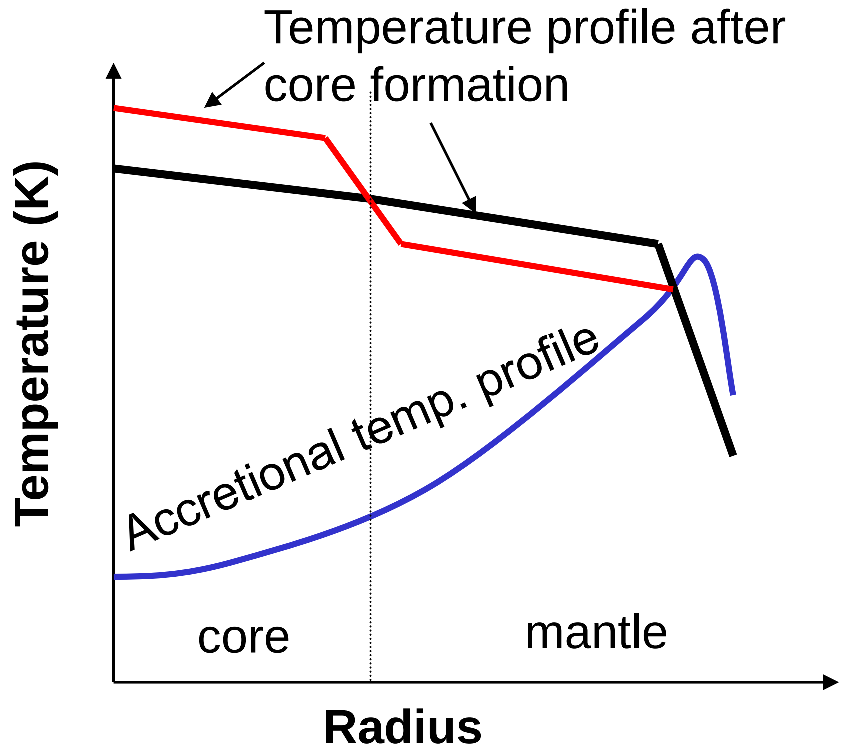
Early Martian (thermal) dynamo possible with a superheated core



Breuer and Spohn, 2003

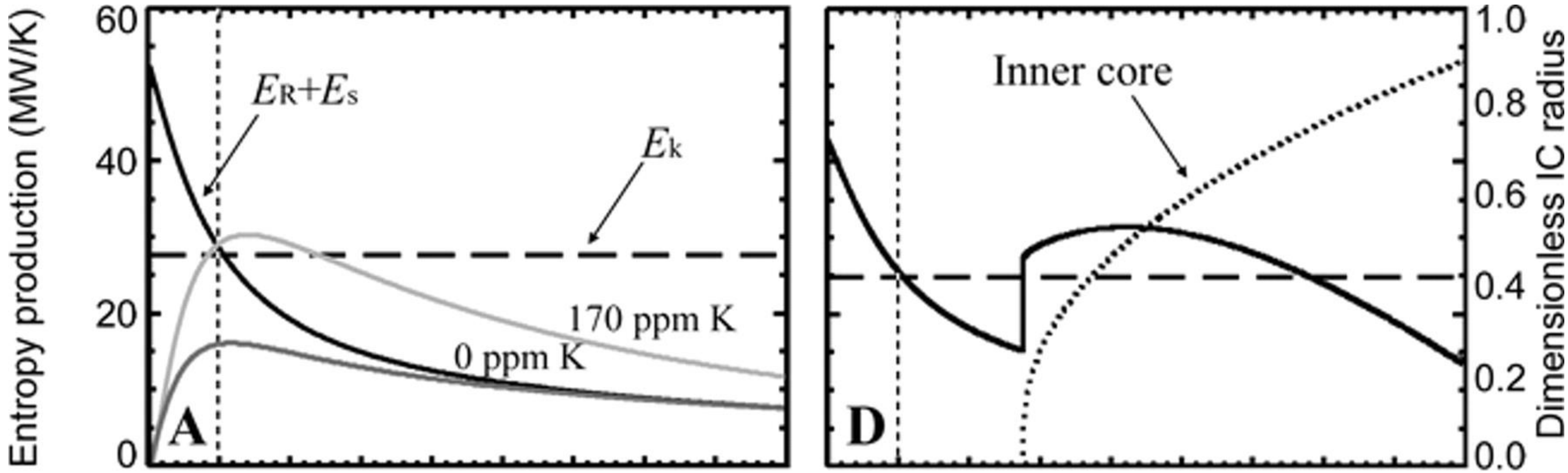
Accretion and Core Formation

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



What are the initial thermal conditions after core formation?

Chemical dynamo lasts ~ 2 Ga inconsistent with observations



Williams and Nimmo, 2004

Present favorite scenario

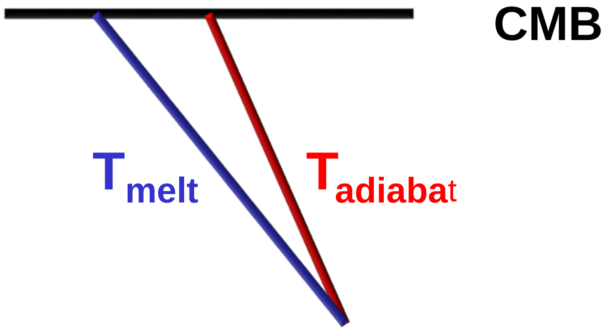
- Early thermally driven dynamo
- Cessation of the core due to mantle cooling → heat flow decreases below heat flow along the core adiabat
- Other models assume that a large impact can stop the dynamo (e.g. Roberts et al. 2009; Arkani-Hamed 2010)
- but all scenarios predict

present-day Martian core is entirely fluid --
Venus similar magnetic evolution?

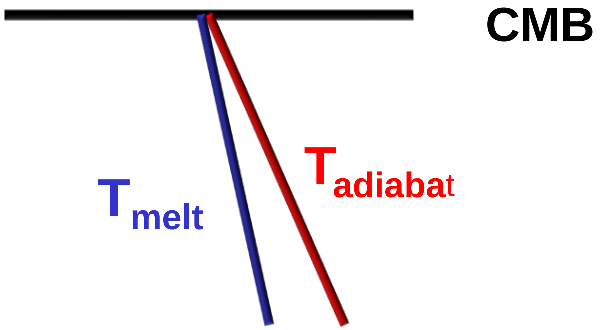
DLR

Where does crystallization in the core occur?

$$\left.\frac{dT}{dP}\right|_{melt} > \left.\frac{dT}{dP}\right|_{adiabat}$$



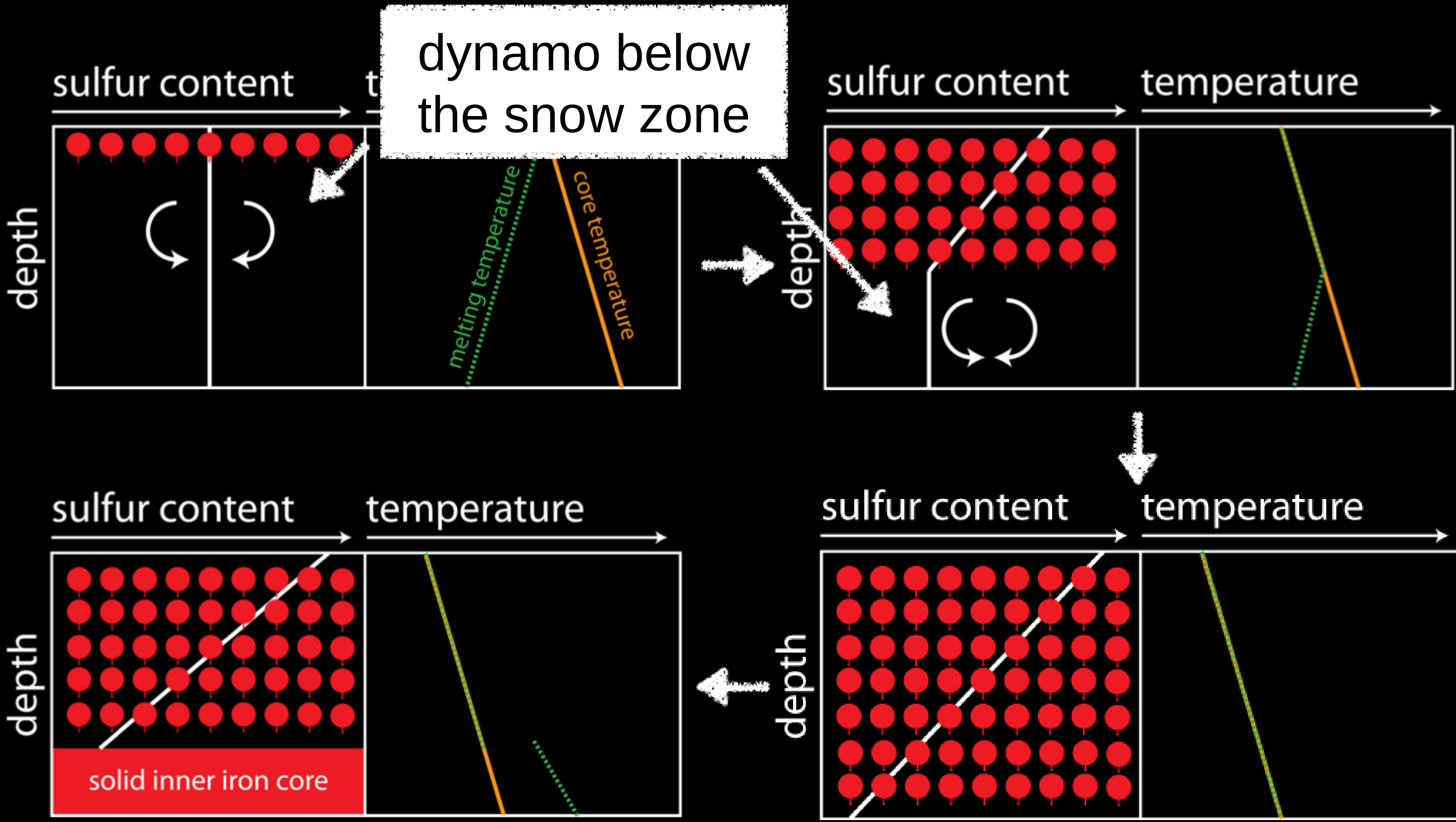
$$\left.\frac{dT}{dP}\right|_{melt} < \left.\frac{dT}{dP}\right|_{adiabat}$$



dT/dP_{melt} can be even negativ

DLR

The Fe-snow regime



Similarity and Diversity of Terrestrial Planets and the Moon

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- Atmosphere

Planet	 Mercury	 Venus	 Earth	 Mars
T_P, K (°C)	437 (163)	232 (-41)	255 (-18)	209 (-64)
T_{obs}, K (°C)	~440 (167)	735 (462)	288 (15)	215 (-58)
Atmosphere: Pressure, kPa composition [trace gases]	none	9300 CO_2(0.965), N_2(0.035), [SO_2, Ar]	101 N_2(0.78), O_2(0.21), Ar(0.009), [CO_2, H_2O]	0.64 CO_2(0.95), N_2(0.03), Ar(0.02), [O_2, CO]

Terrestrial planets and the Moon

- Each planet has its own specific history depending on interior structure and composition, distance to the sun and initial conditions (e.g. magma ocean phase)