Thermochemical evolution of Mercury, the Moon and Mars: Constraints from space missions and planetary samples data

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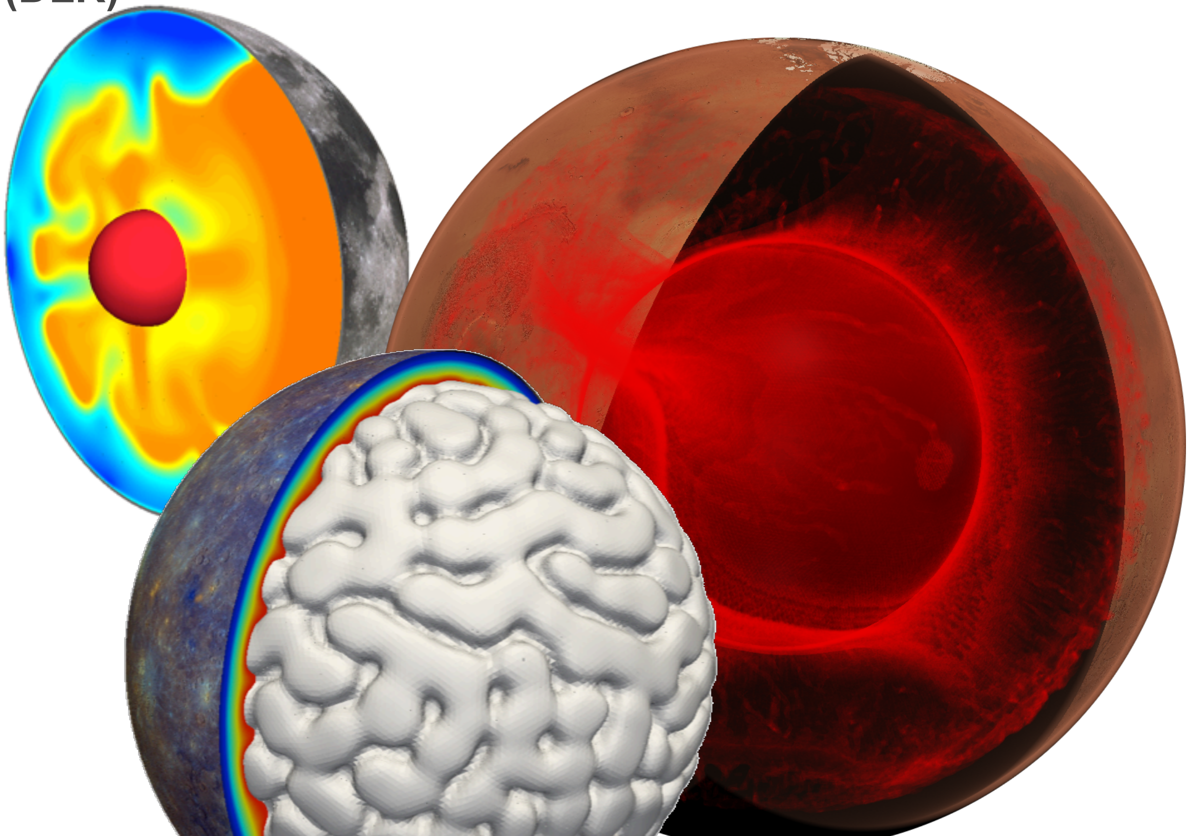
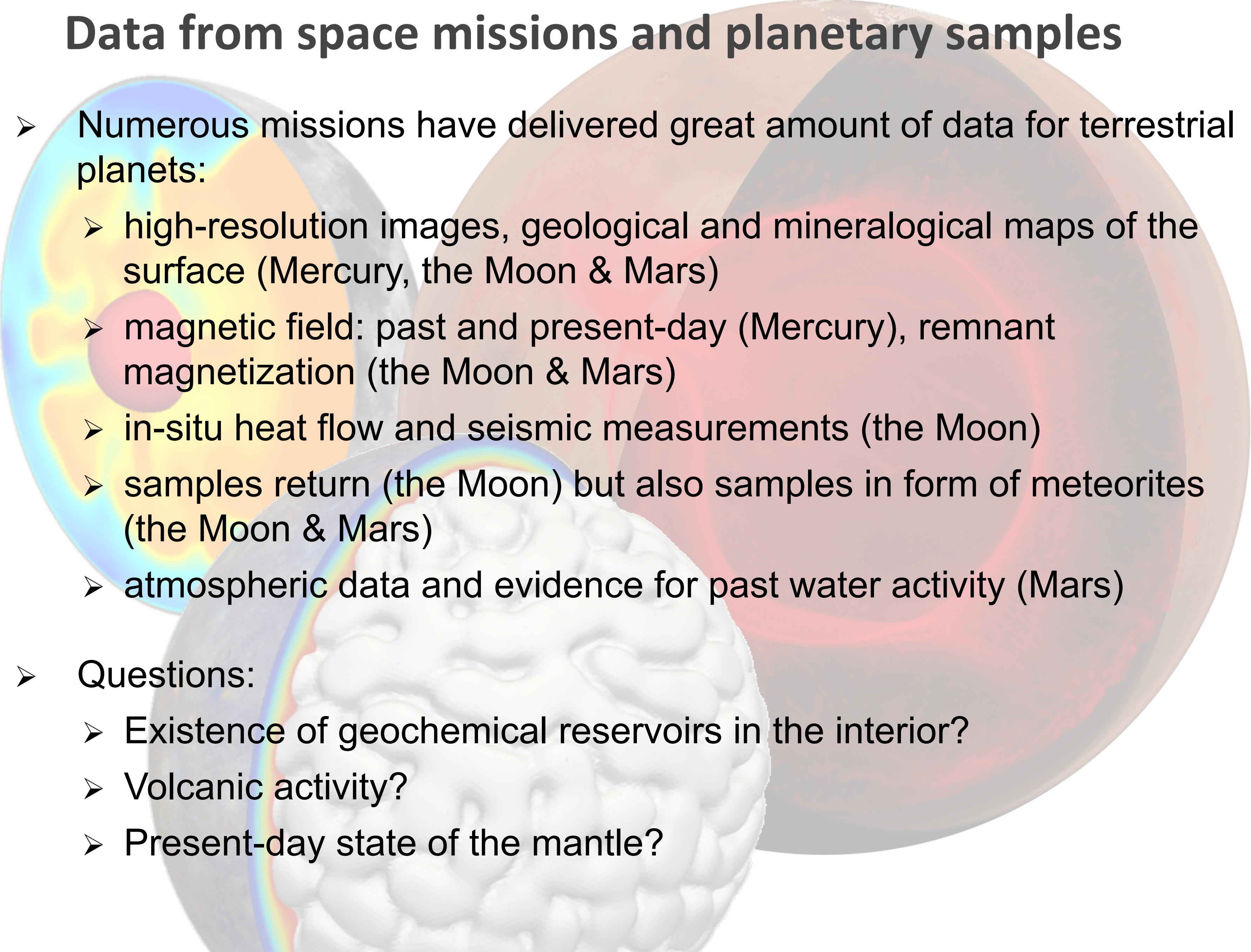


Data from space missions and planetary samples



- Numerous missions have delivered great amount of data for terrestrial planets:
  - high-resolution images, geological and mineralogical maps of the surface (Mercury, the Moon & Mars)
  - magnetic field: past and present-day (Mercury), remnant magnetization (the Moon & Mars)
  - in-situ heat flow and seismic measurements (the Moon)
  - samples return (the Moon) but also samples in form of meteorites (the Moon & Mars)
  - atmospheric data and evidence for past water activity (Mars)

- Questions:
  - Existence of geochemical reservoirs in the interior?
  - Volcanic activity?
  - Present-day state of the mantle?

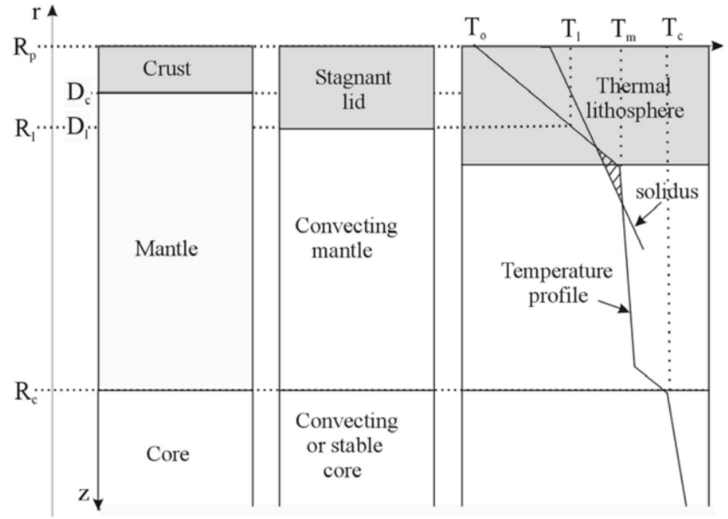
Numerical codes

- Parameterized convection $Nu = \alpha Ra^{\beta}$

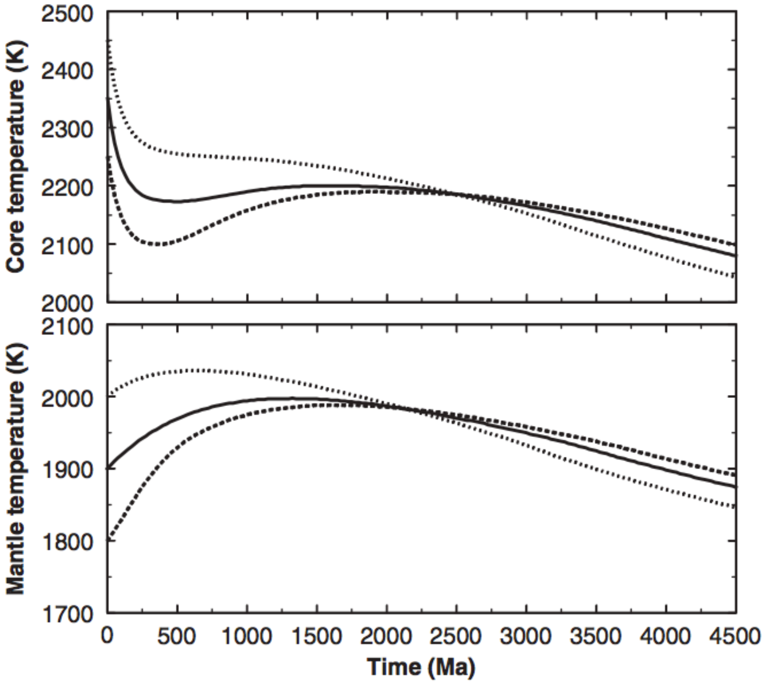
$$\rho_m cp_m V_m \frac{dT_m}{dt} = -q_m A_m + q_c A_c + Q_m V_m$$

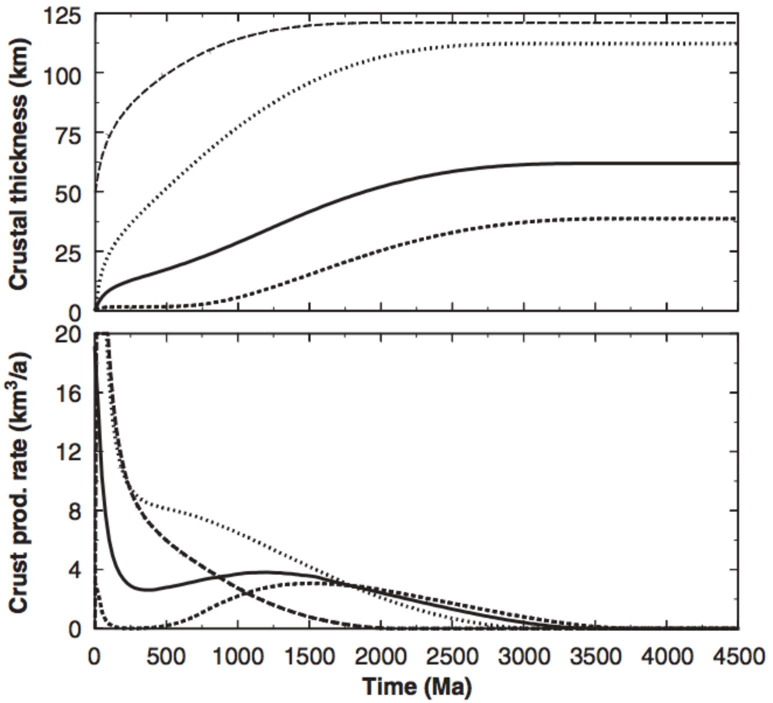
$$\rho_c cp_c V_c \frac{dT_c}{dt} = -q_c A_c$$

$$Q_m = Q_0 \exp(-\lambda t)$$



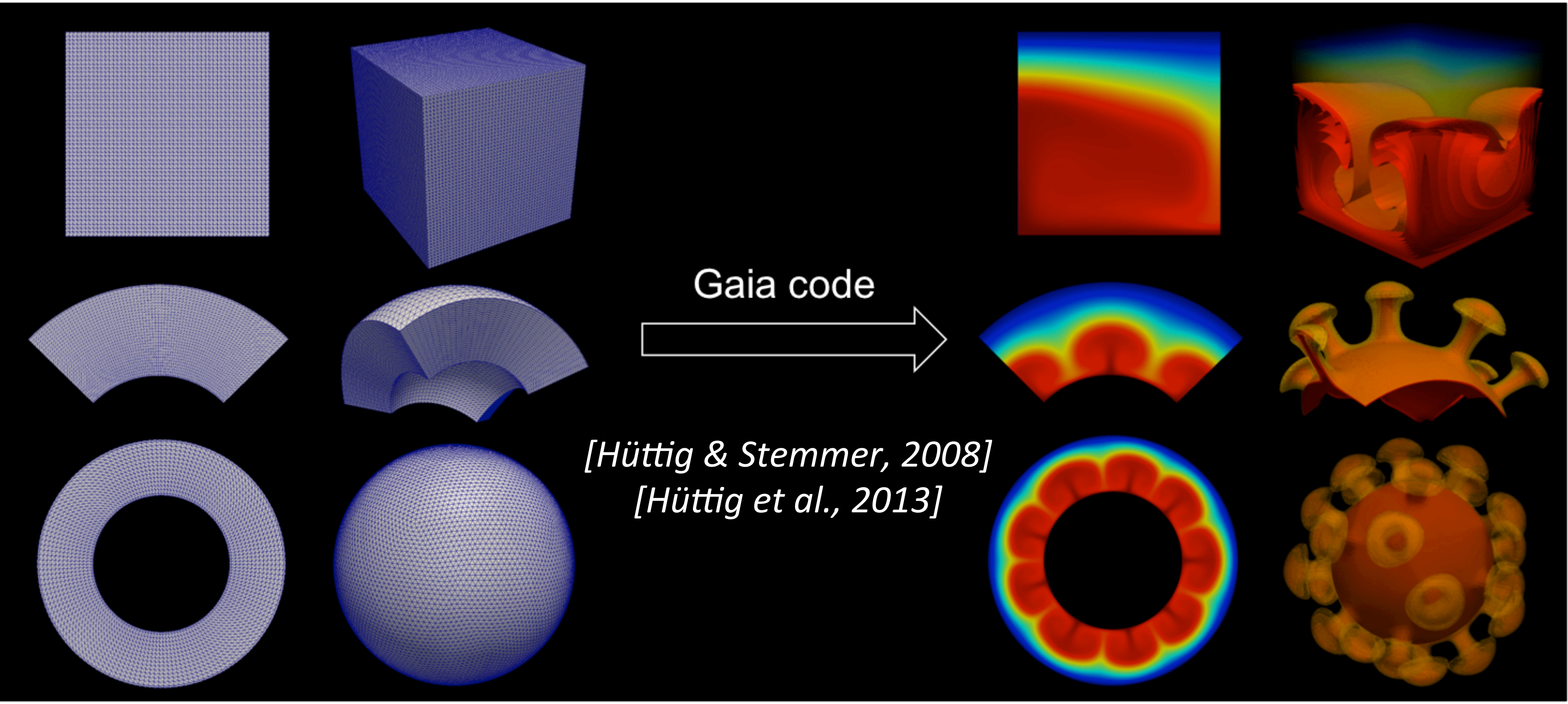
[Breuer & Spohn, PSS, 2006]
[Grott et al., EPSL, 2011]
[Morschhauser et al., Icarus, 2011]





Numerical codes

- Fully dynamical:
  - 2D box, regional / full cylinder
  - 3D box, regional / full spherical shell



[Hüttig & Stemmer, 2008]
[Hüttig et al., 2013]

Numerical codes

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  - 2D box, regional / full cylinder
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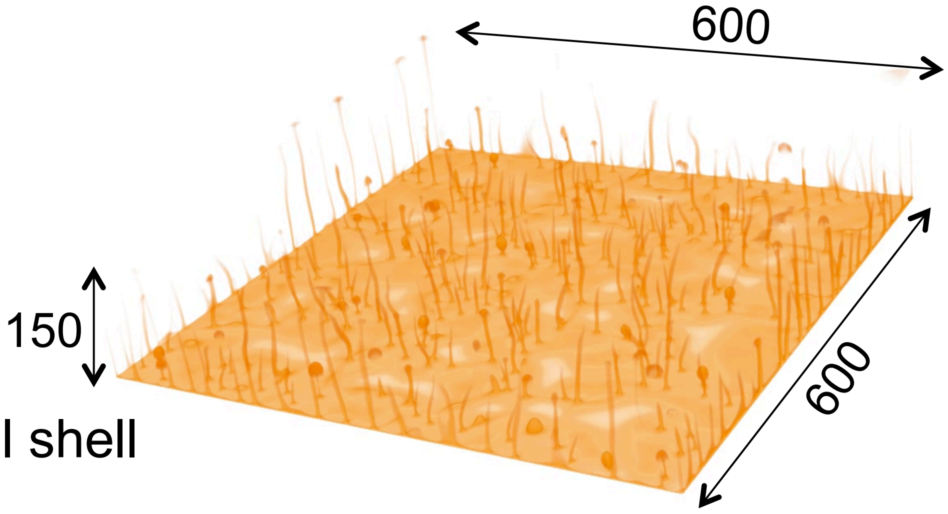
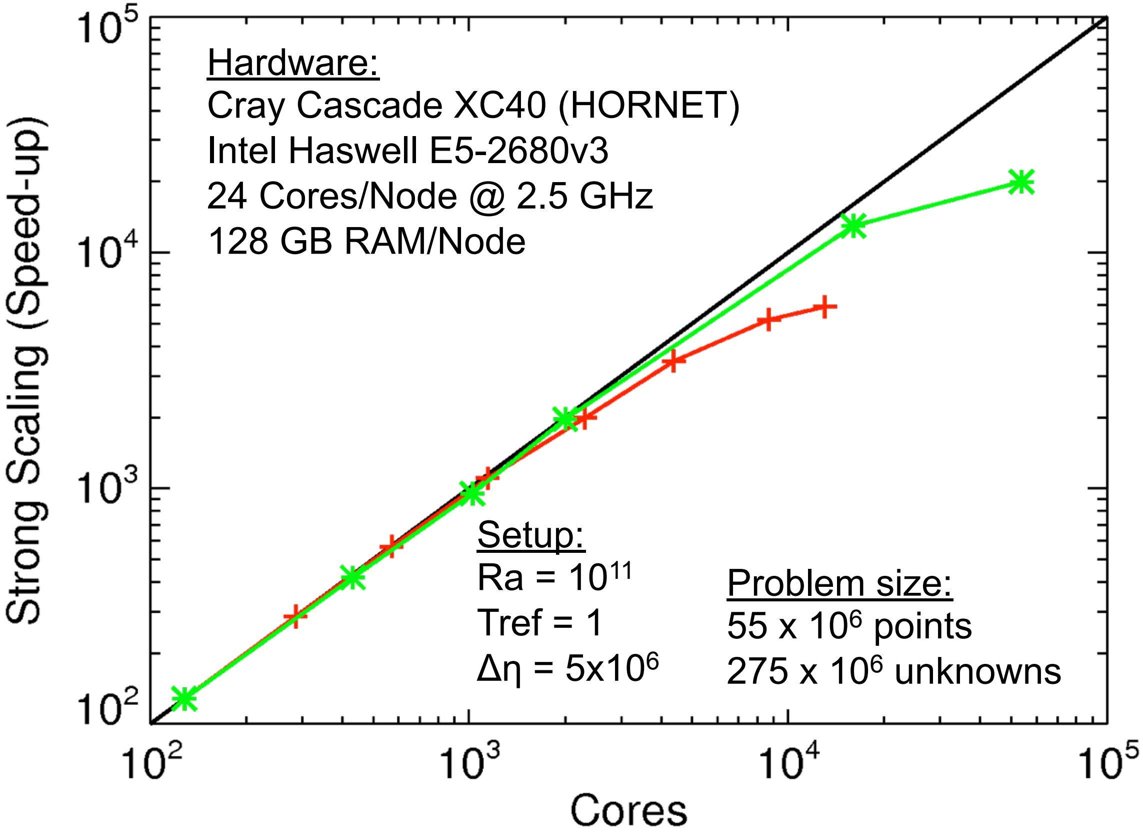


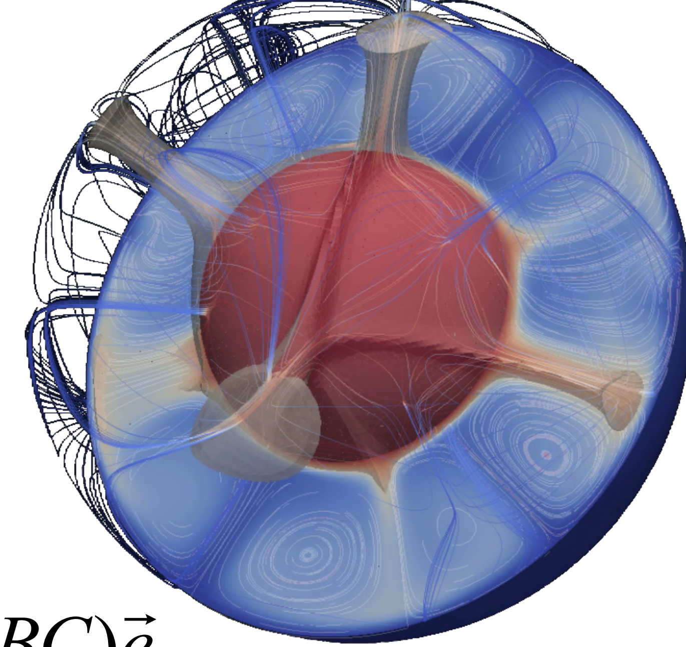


Hardware:
Cray Cascade XC40 (HORNET)
Intel Haswell E5-2680v3
24 Cores/Node @ 2.5 GHz
128 GB RAM/Node

Setup:
Ra = 10^{11}
Tref = 1
$\Delta\eta = 5\mathrm{x}10^6$

Problem size:
55 x 10^6 points
275 x 10^6 unknowns

Thermo-chemical Convection



Conservation equations of

- mass

$$\nabla \cdot \vec{u} = 0$$

- linear momentum

$$\nabla \cdot [\eta(\nabla \vec{u} + (\nabla \vec{u})^T)] - \nabla p = Ra(T - BC)\vec{e}_r ,$$

- thermal energy

$$\frac{\partial T}{\partial t} + \vec{u} \cdot \nabla T - \nabla^2 T = \frac{Ra_Q}{Ra}$$

- material transport

$$\frac{\partial C}{\partial t} + \vec{u} \cdot \nabla C = 0$$

Buoyancy number:

$$\boxed{B = \frac{Ra_C}{Ra} = \frac{\Delta\rho}{\rho\alpha\Delta T}}$$

$$\frac{\partial T_c}{\partial t} = -\frac{\rho_m c p_m}{\rho_c c p_c} \frac{A_c}{V_c} \frac{\partial T_c}{\partial z}$$

$$Ra_Q = Ra_{Q0} \exp(-\lambda t)$$

Tracing of material properties (e.g., density)

Particle in cell method

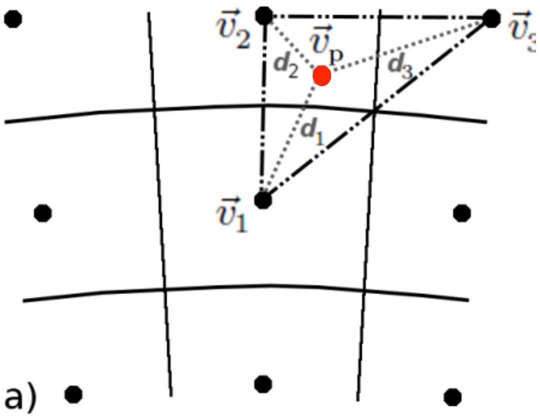
$$\frac{\partial C}{\partial t} + \vec{u} \cdot \nabla C = \frac{DC}{Dt} = 0$$

- solution of a trajectory equation for each particle:

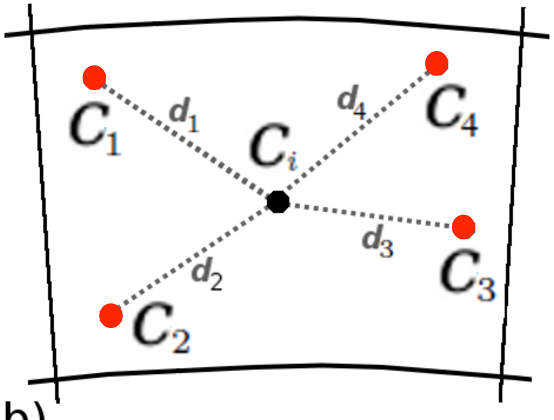
$$\frac{d\vec{x}_p}{dt} = \vec{u}_p$$

- interpolation of material properties from particles to grid

$$C_i = \frac{\sum_j^{n_{part}} C_j \frac{1}{d_j}}{\sum_j^{n_{part}} \frac{1}{d_j}}$$



a)



b)

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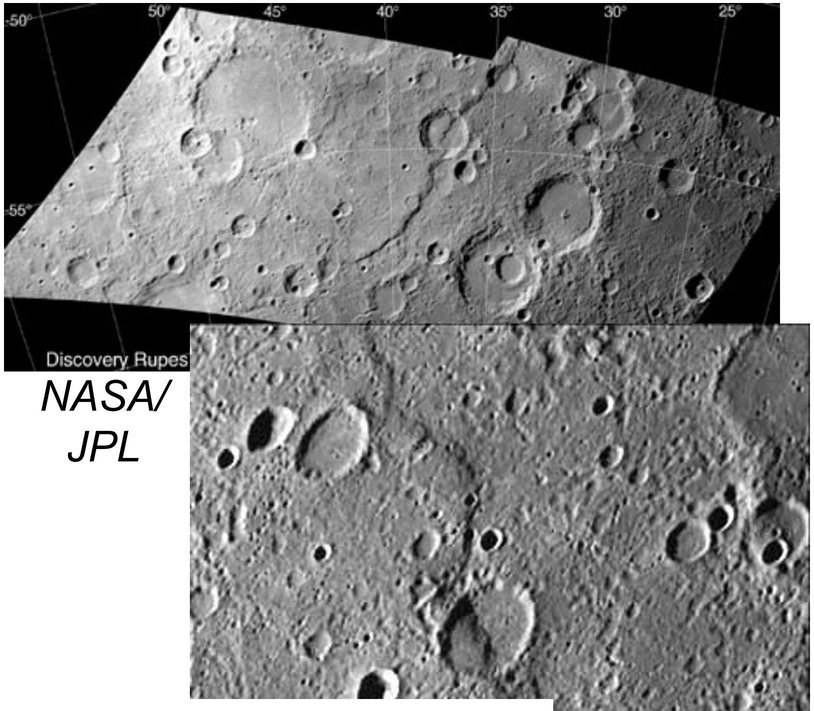
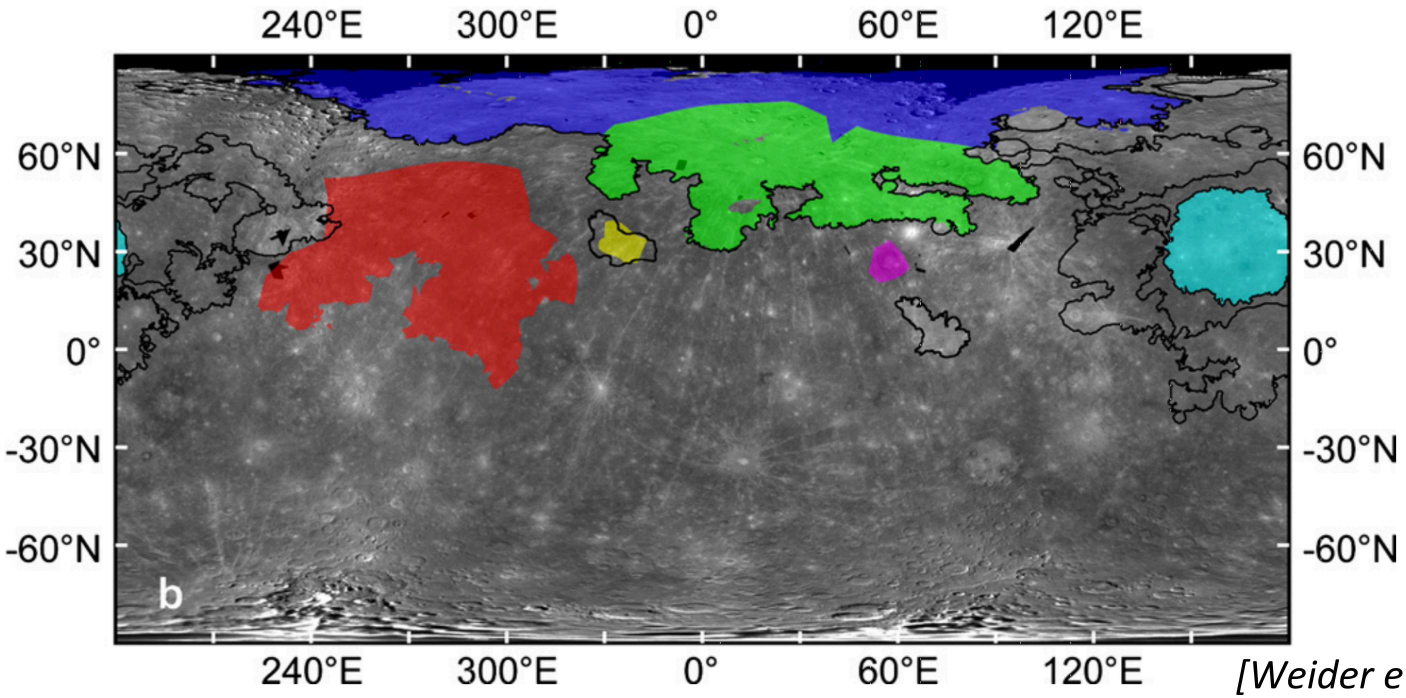
Thermochemical evolution

Mercury:

- ➢ Lobate scarps show that the radius of the planet shrinked by about 3 – 5 km.
- ➢ At least two distinct mantle reservoirs are needed to explain the observed surface lava composition.

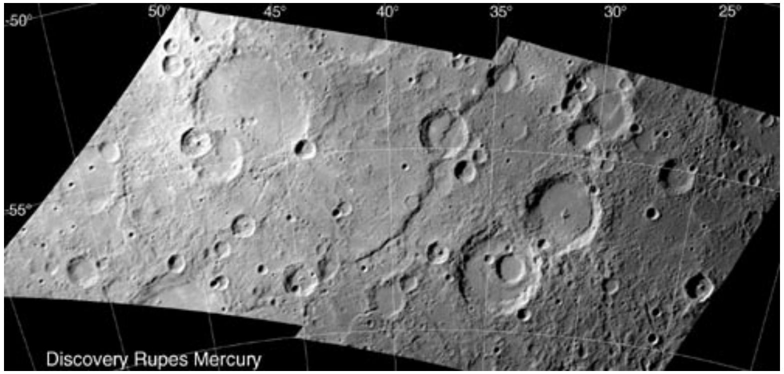


NASA/ JPL

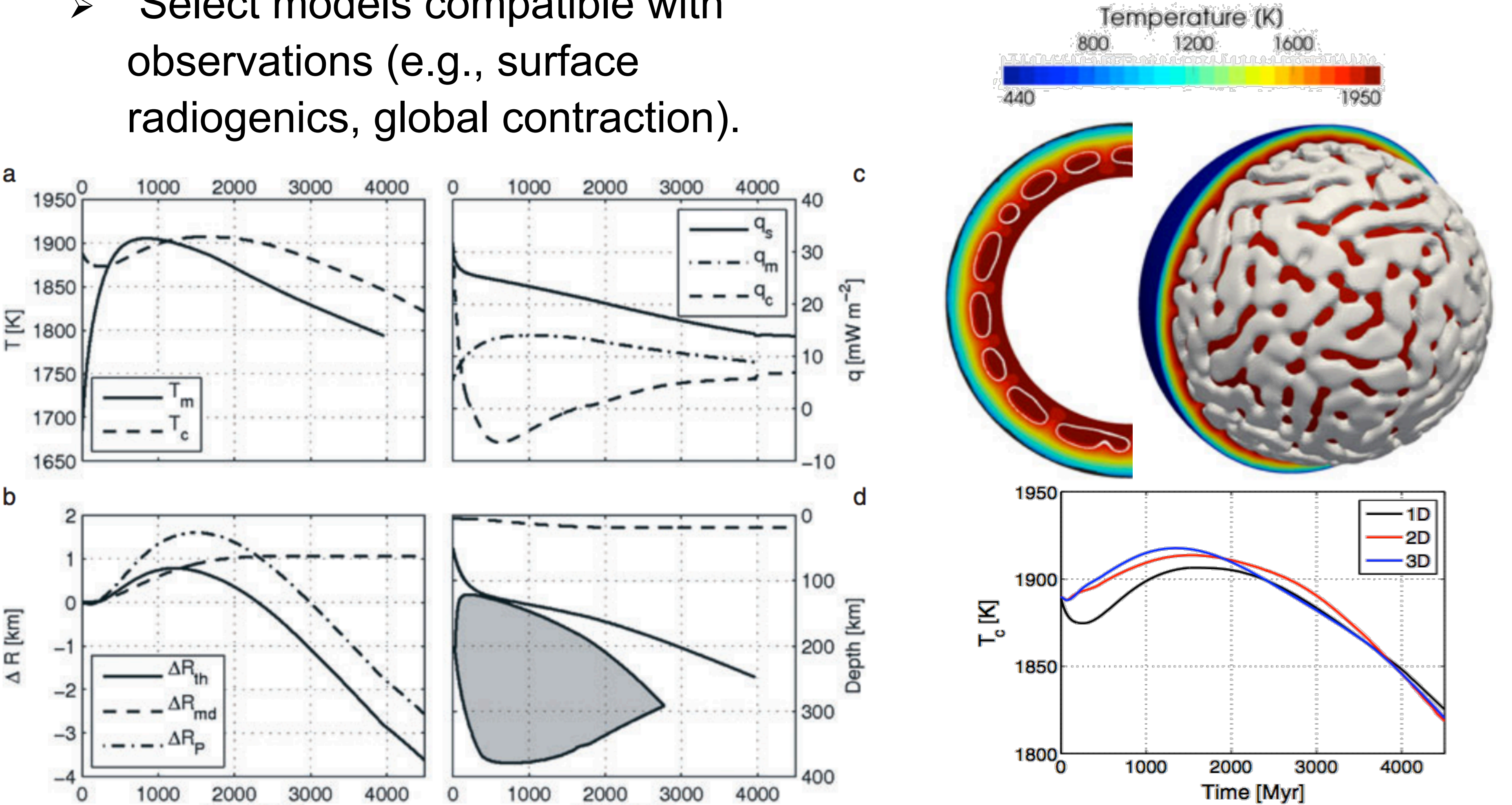


[Weider et al., EPSL, 2015]

Mercury's global contraction

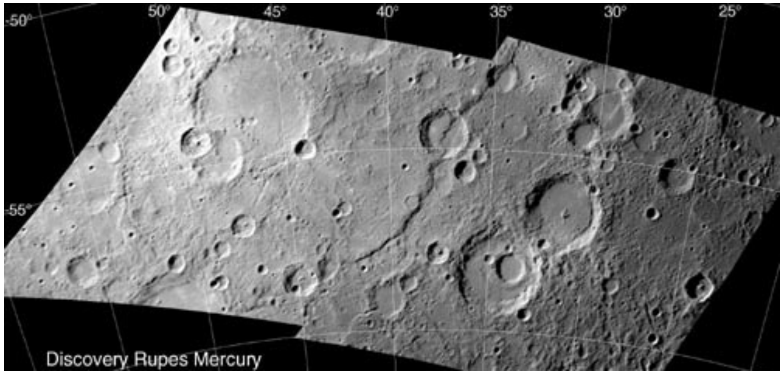


- 1D/2D/3D thermochemical evolution, mantle melting and crust production.
- Select models compatible with observations (e.g., surface radiogenics, global contraction).



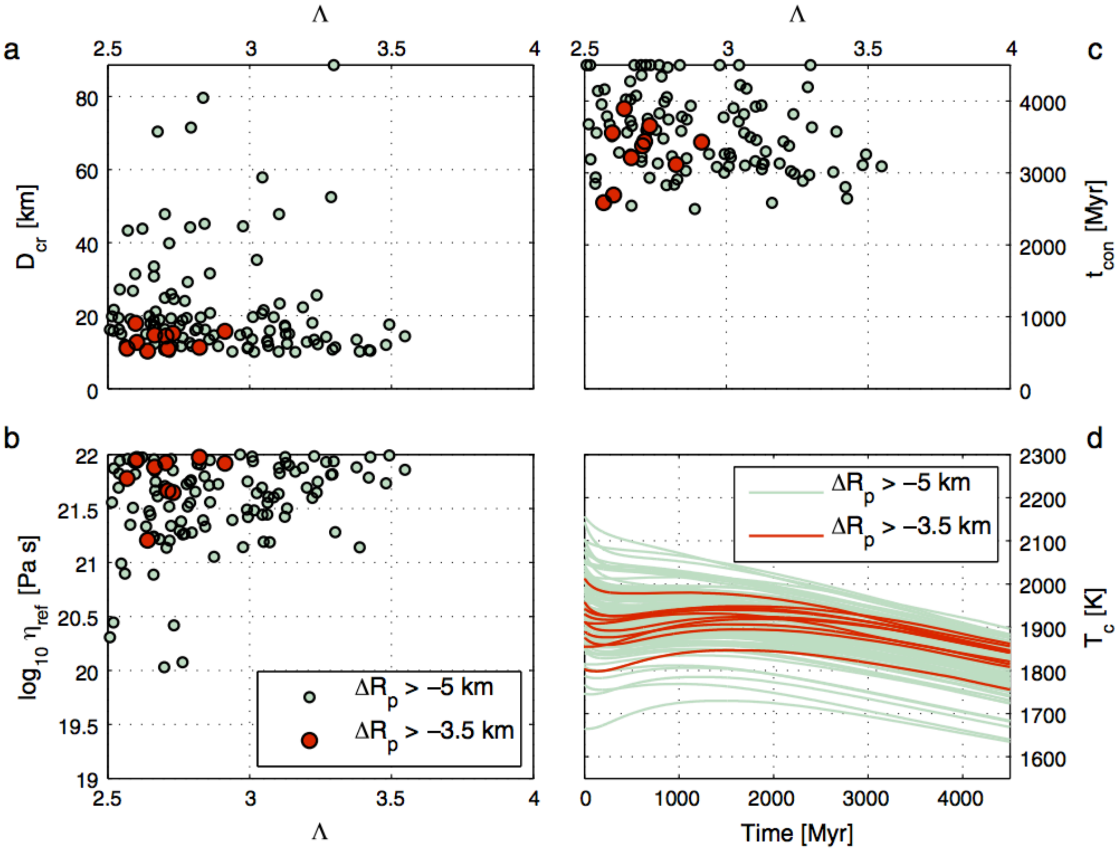
[Tosi et al., JGR, 2013]

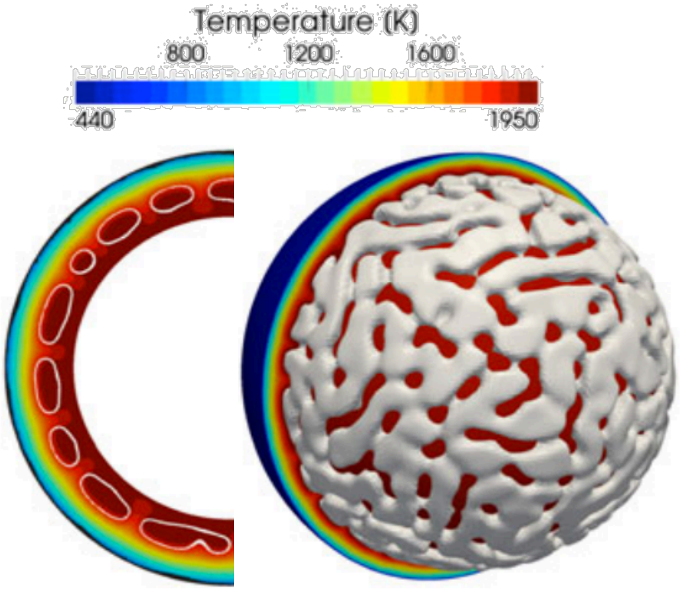
Mercury‘s global contraction

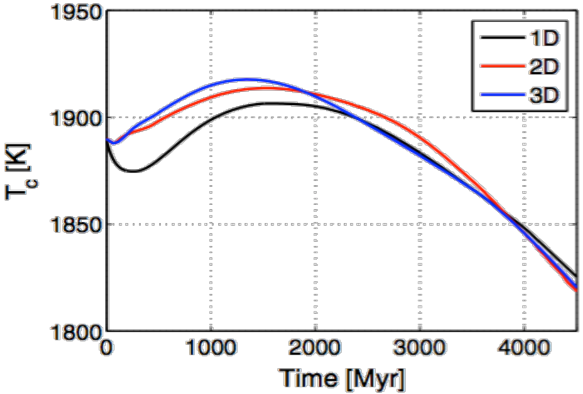


- Most of the admisible models have a conductive present-day mantle.
- Crust enrichment factor: 2.5 – 3.5 and bulk heat source content of:

35 – 62 ppb Th	**20 – 36 ppb U**	**290 – 515 ppm K**



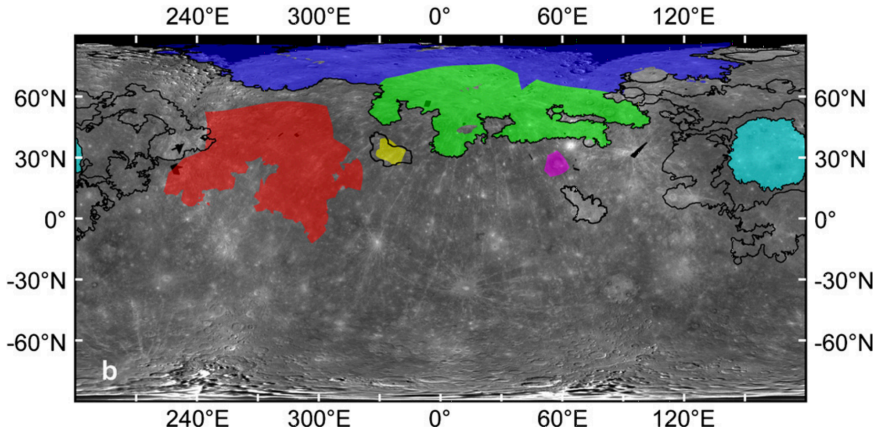


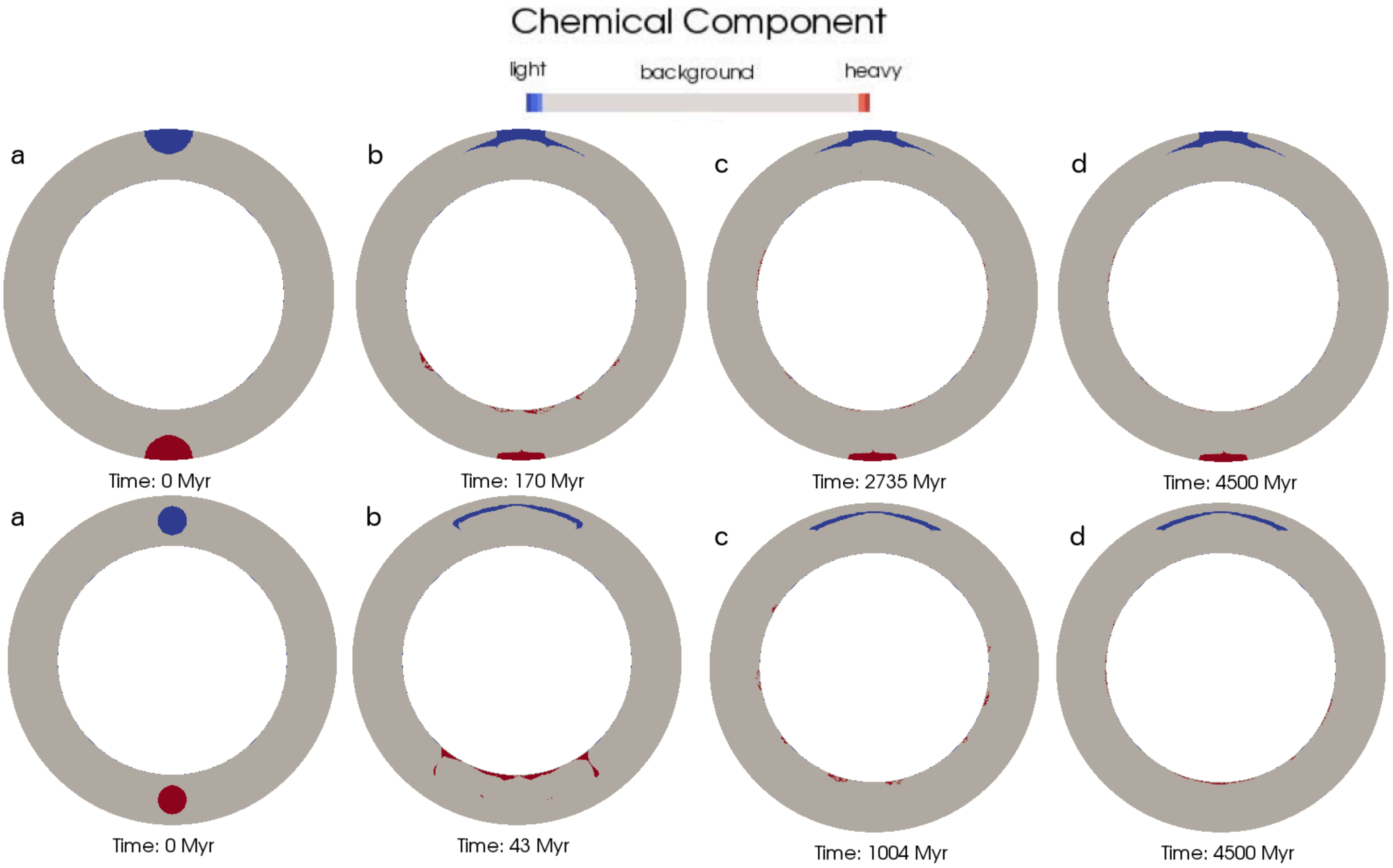


[Tosi et al., JGR, 2013]

Surface lava compositions

- A density contrast of only 20 kg/m^3 can preserve chemical heterogeneous mantle sources.





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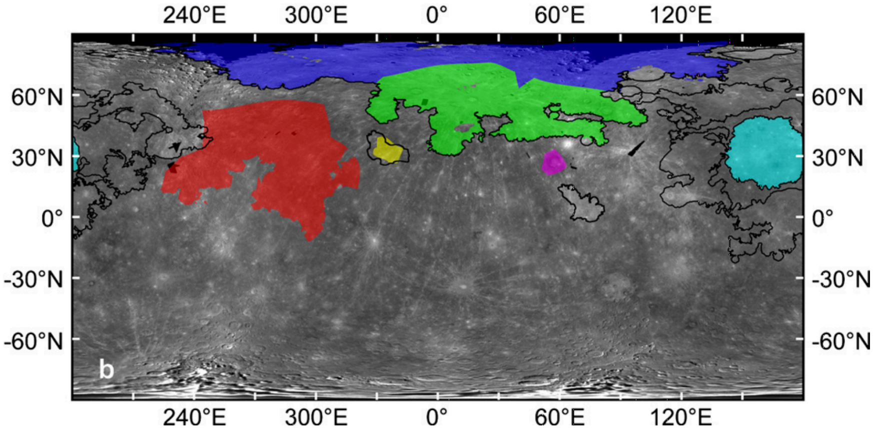
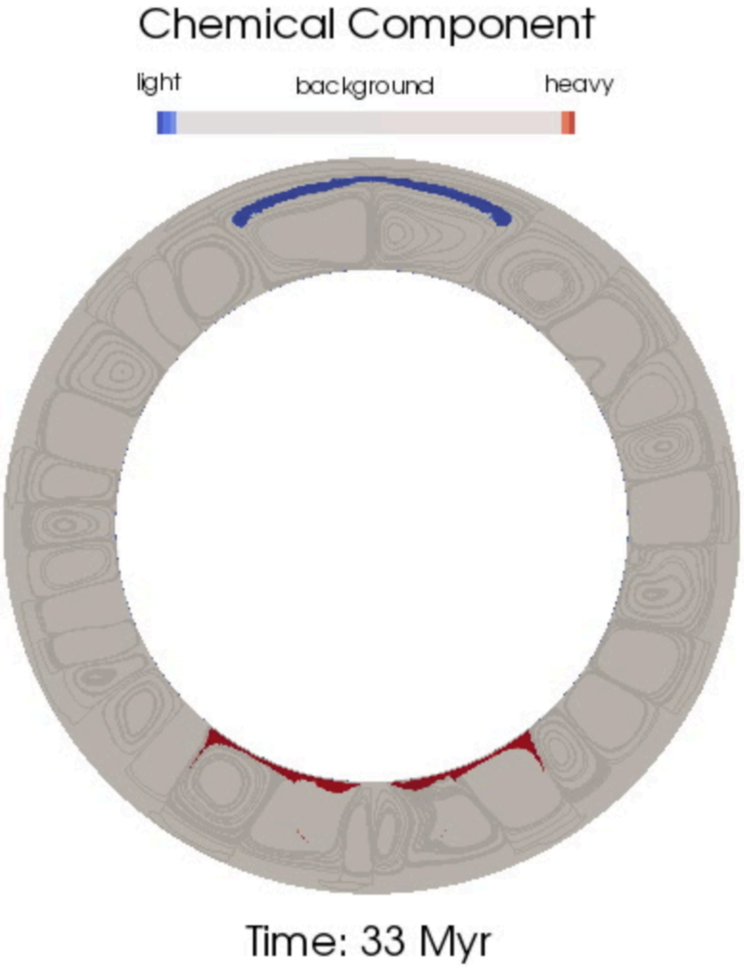
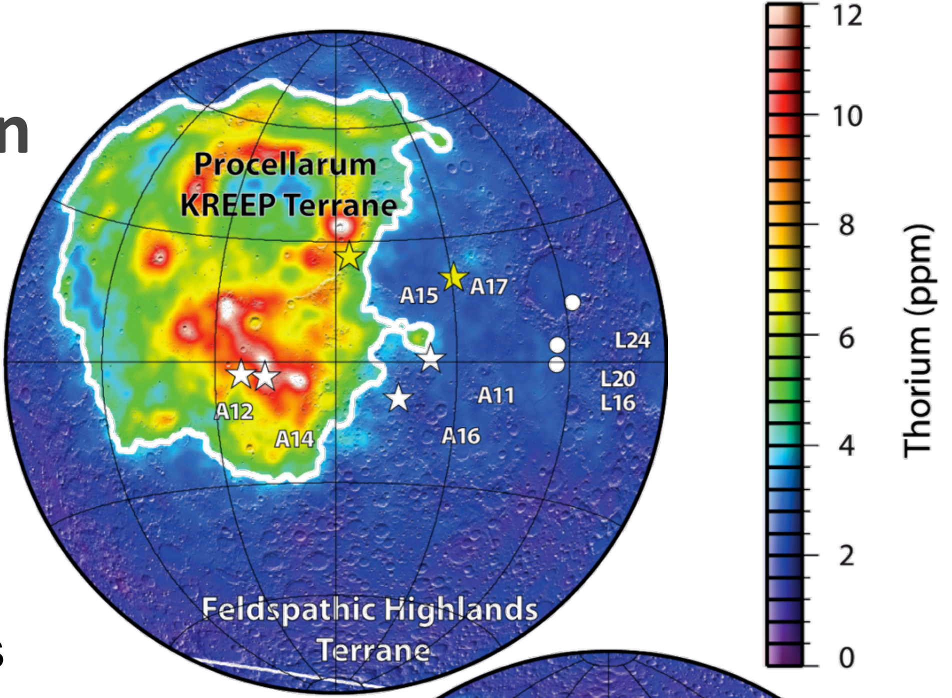
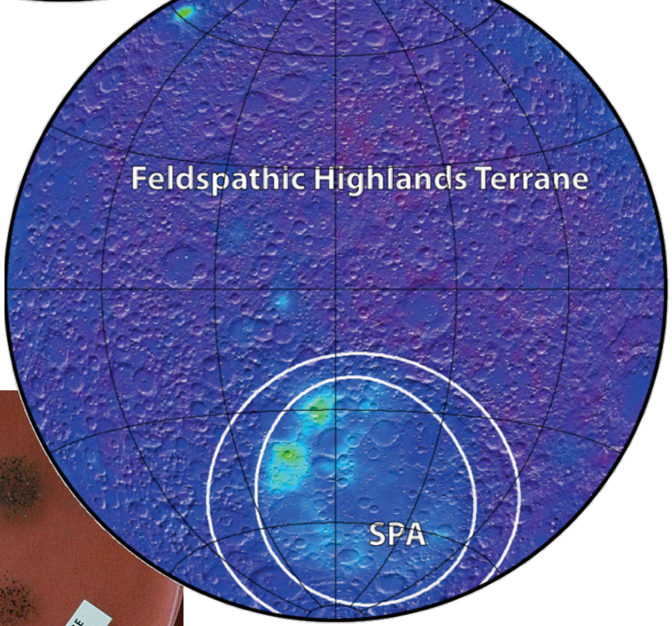




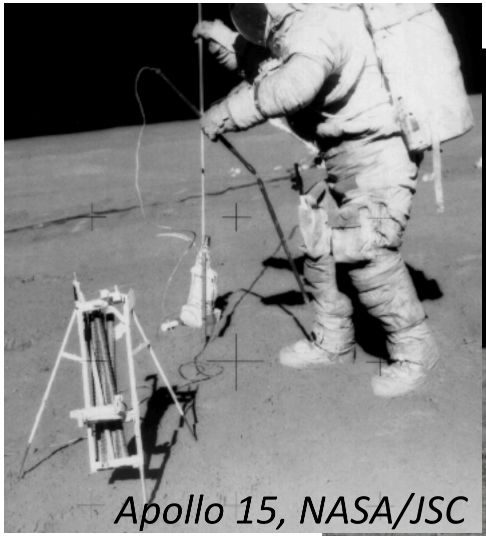
Thermochemical evolution

The Moon:

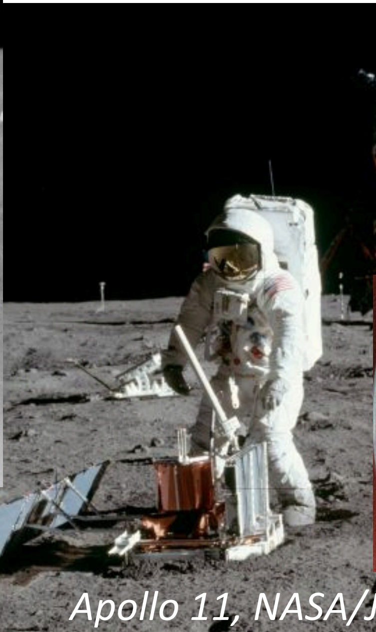
- ➢ In-situ seismic and heat flow measurements to probe its interior.
- ➢ Lunar samples indicate an active dynamo between 4.2 – 3.5 Ga.
- ➢ PKT region could show the traces of a once global lunar magma ocean.

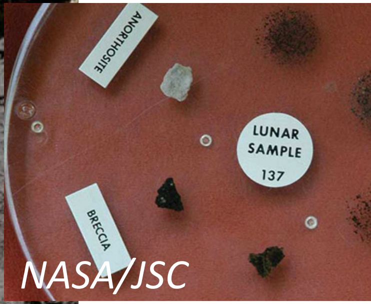
[Lawrence et al., JGR, 2003]



Apollo 15, NASA/JSC



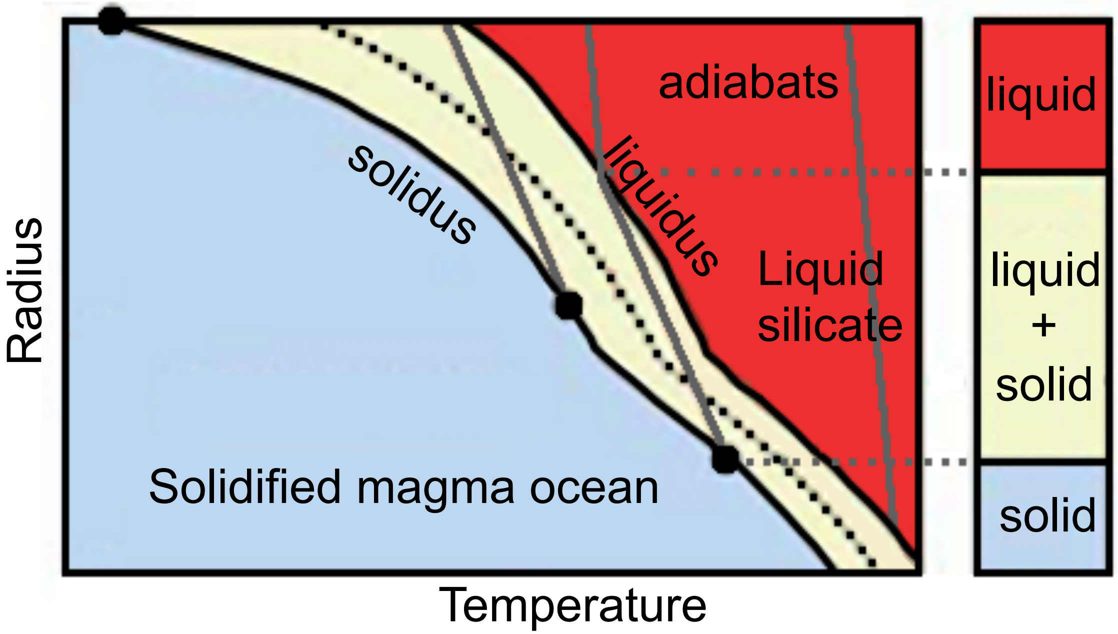
Apollo 11, NASA/JSC

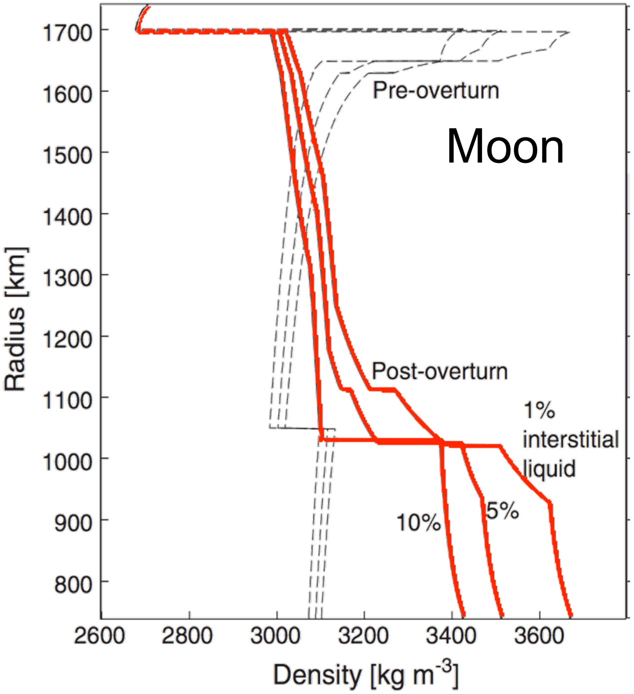


NASA/JSC

Lunar Magma Ocean

- Bottom-up freezing of the magma ocean.
- Formation of the anorthosite crust before complete solidification.
- Materials enriched in KREEP and iron (e.g. ilmenite) close to the surface.
- Possible overturn of the ilmenite-rich layer and accumulation at the CMB.

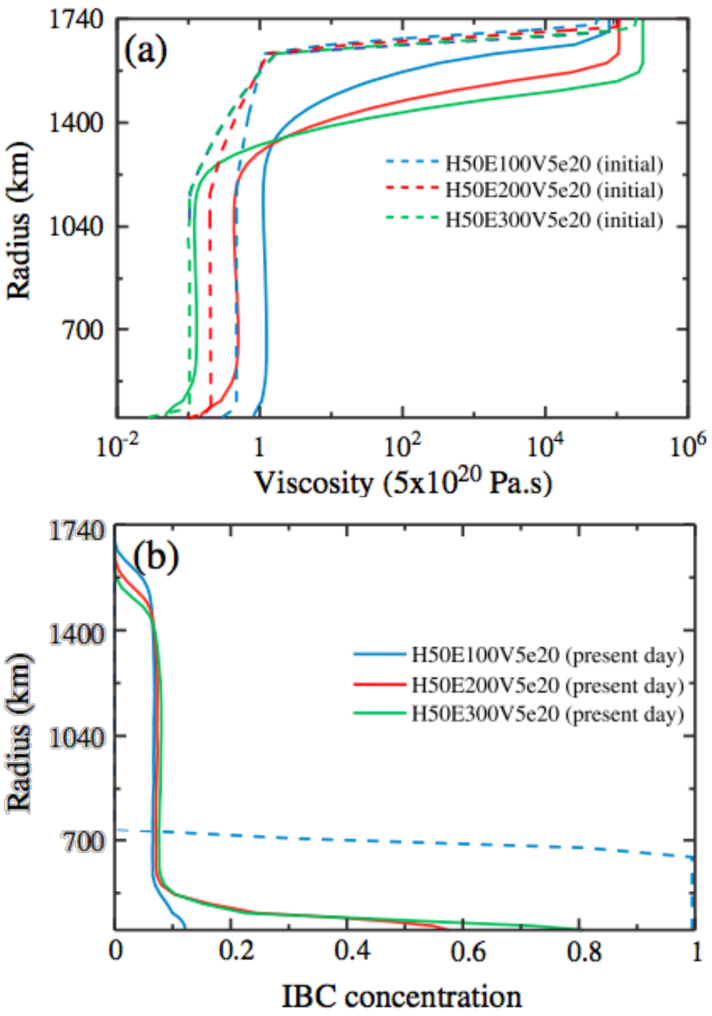


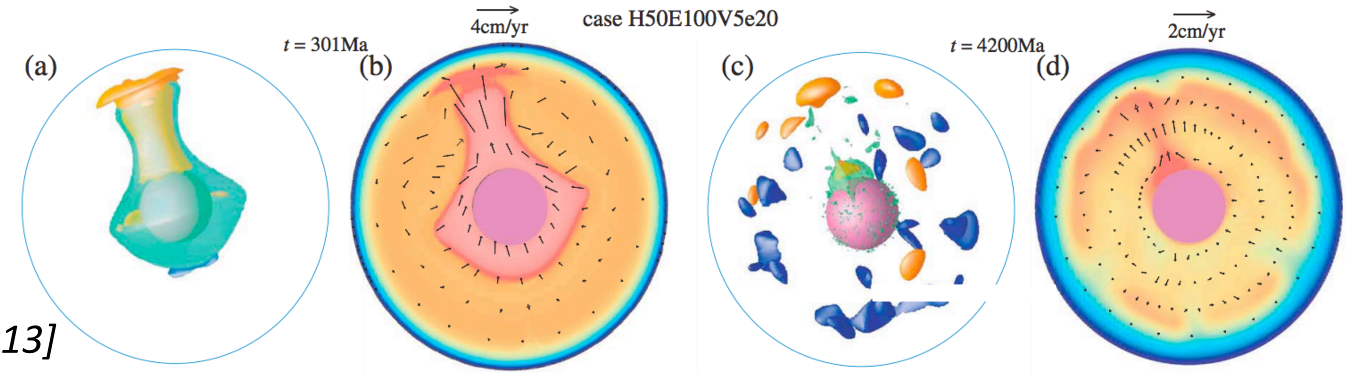


[Elkins-Tanton et al., 2002, 2003, 2005, 2008, 2011]

Subsequent lunar evolution

- Initial IBC rich layer located at the CMB with $\Delta\rho$ = 90kg/m^3 between IBC layer and mantle ($B \approx 0.5$).
- A weak temperature dependence of the viscosity is necessary to produce a single upwelling.
- The CMB heat flux:
  - Negative values from 4.1 to 3.9 Ga (the heating stage of the IBC-rich layer)
  - Positive values from 3.9 to 3.2 Ga (after the instability of the IBC-rich layer)

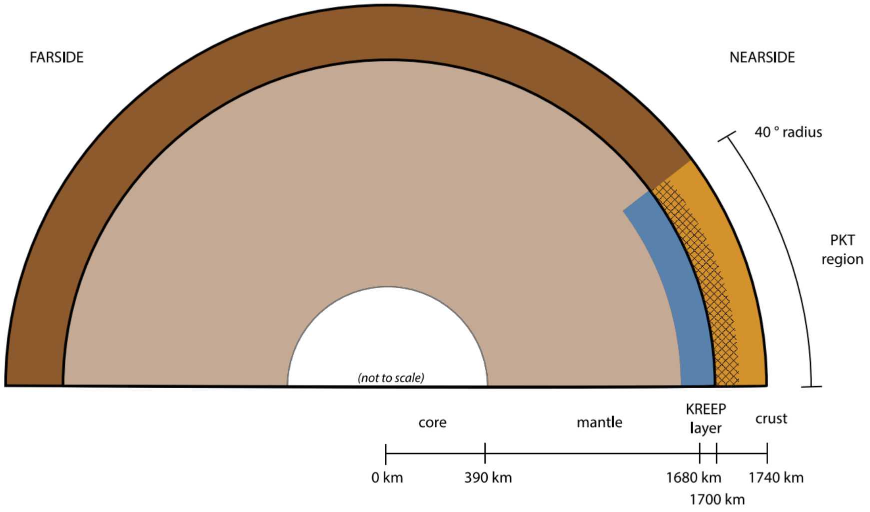


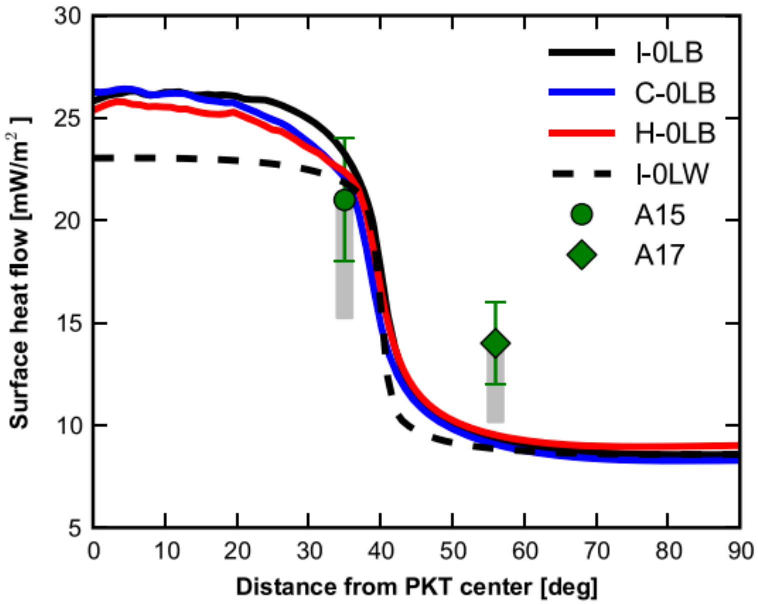


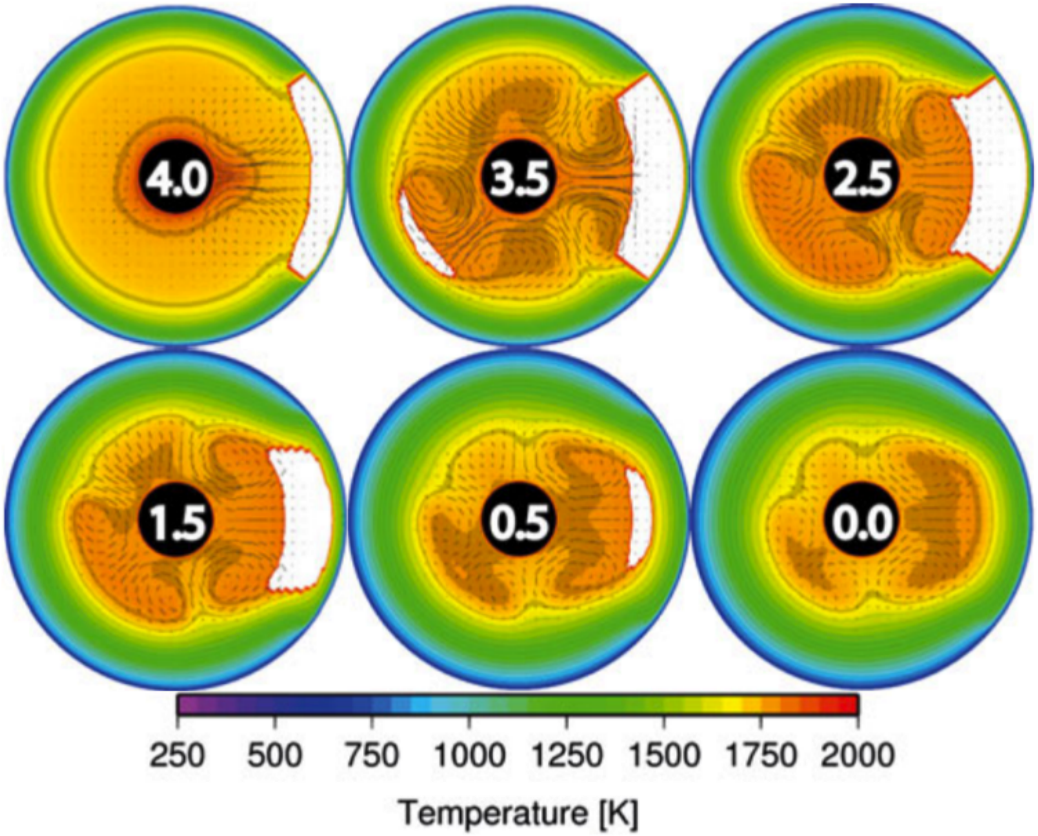
[Zhang et al., JGR, 2013]

Subsequent lunar evolution

- Asymetric thermal evolution due to the concentration of heat sources in the PKT region.
- Partial melt production in the mantle largely consistent with the timing and distribution of mare volcanism.



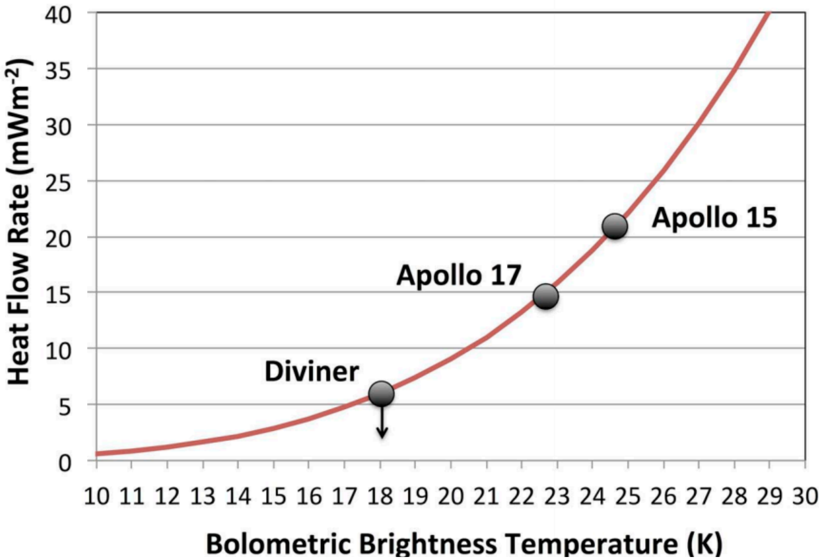




[Laneuville et al., JGR, 2013]

Present-day heat flux

- Heat flow maps using crustal thickness data from *Clementine* and radiogenics distribution from *Lunar Prospector*
- A further heat flow value has been obtained with *Diviner (LRO)*

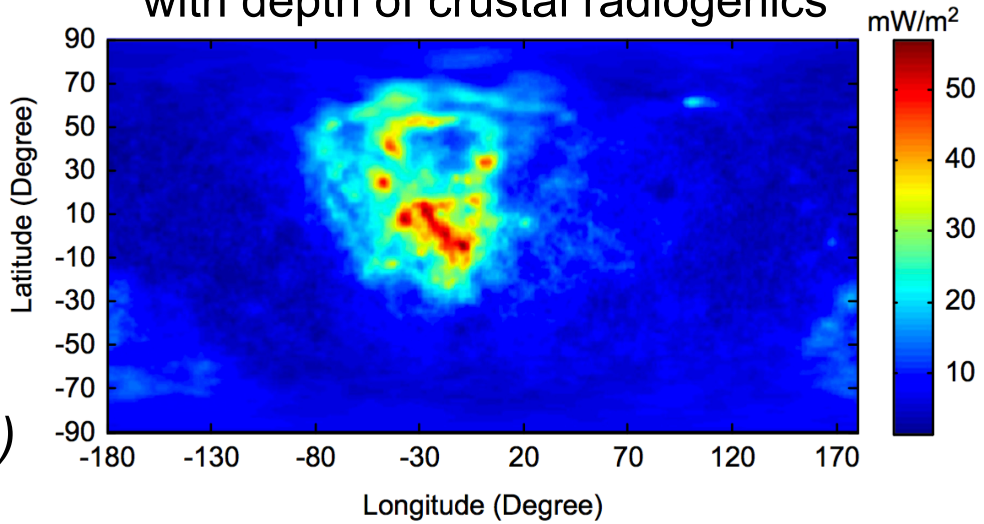


$$H_{solar} + H_{ir} + k\frac{dT}{dz} + F_{horiz} + F_{\uparrow} = F_{ir} = \varepsilon\sigma T^{4}$$

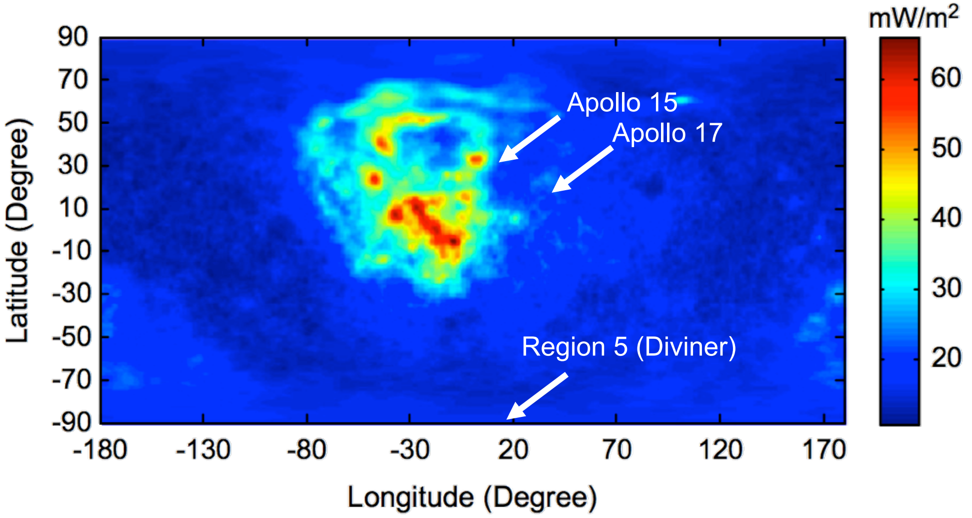
$$F_{\uparrow} < 6 \text{ mWm}^{-2}$$

[Paige et al., LPSC, 2016]

Crustal heat flow
Assumed an exponential decrease with depth of crustal radiogenics



Surface heat flow (Crust + Mantle)
Assumed mantle contribution 9.1mW/m^2

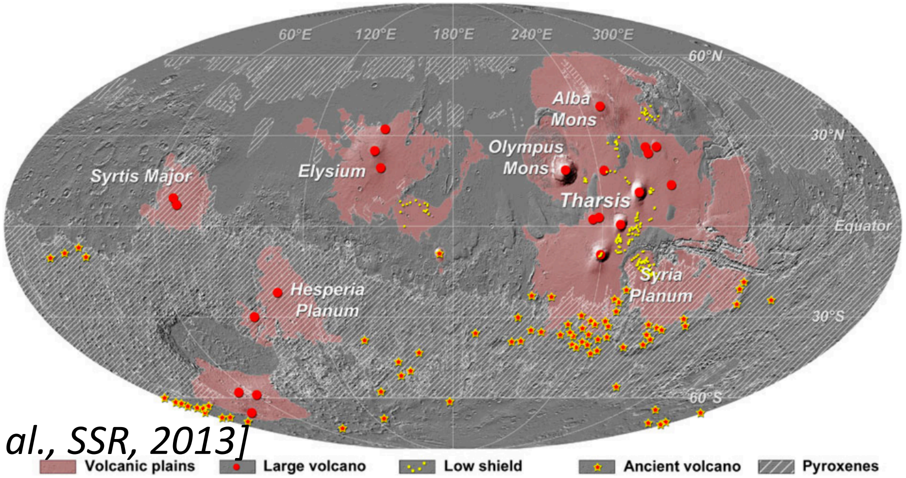


[Zhang et al., Acta Astronautica, 2014]

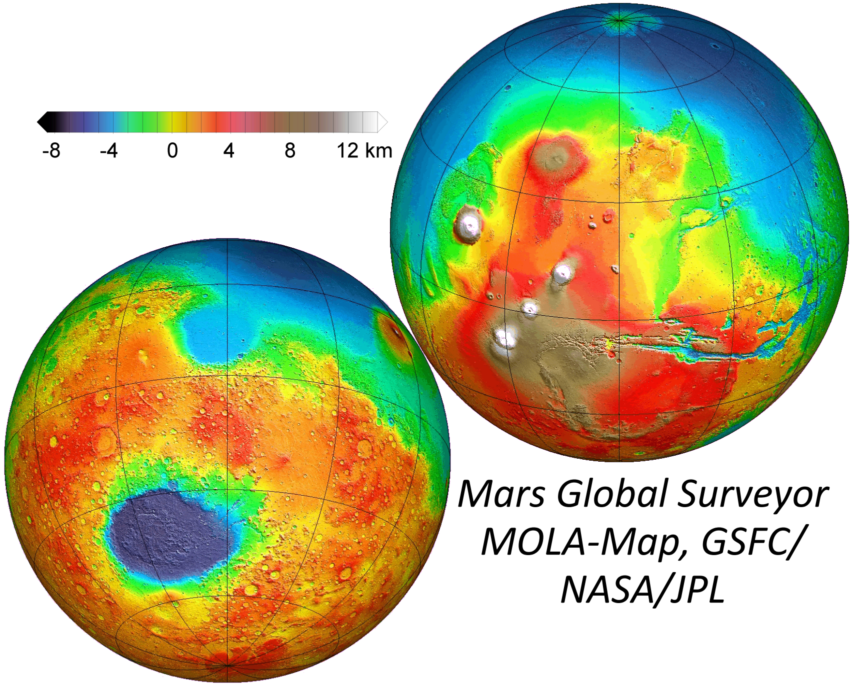
Thermochemical evolution

Mars:

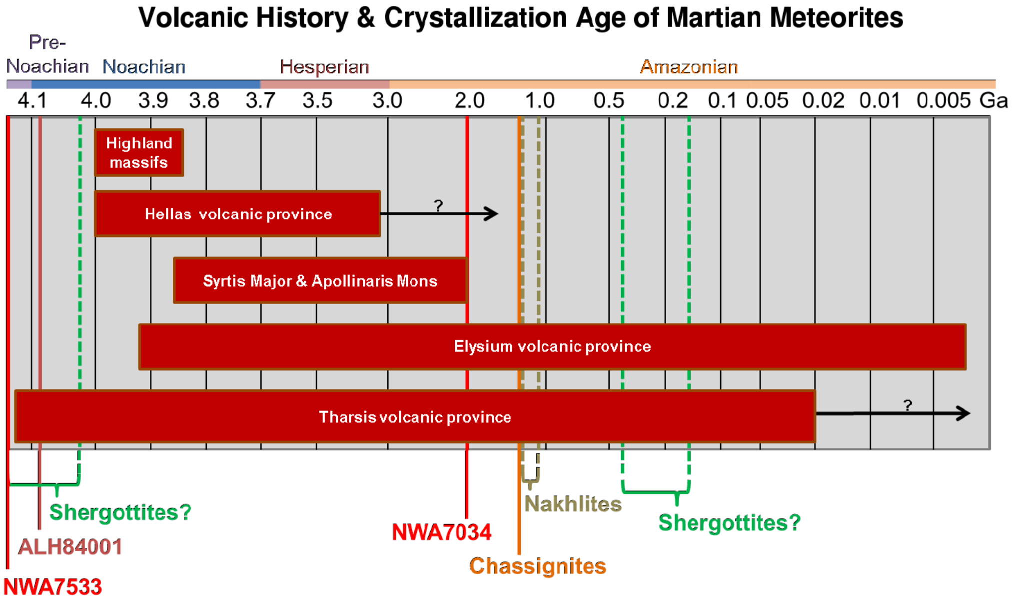
- Crustal dichotomy is one of the oldest features.
- Samples in form of meteorites hint at large degrees of heterogeneity in the interior.
- Long lasting volcanic activity.



[Grott et al., SSR, 2013]

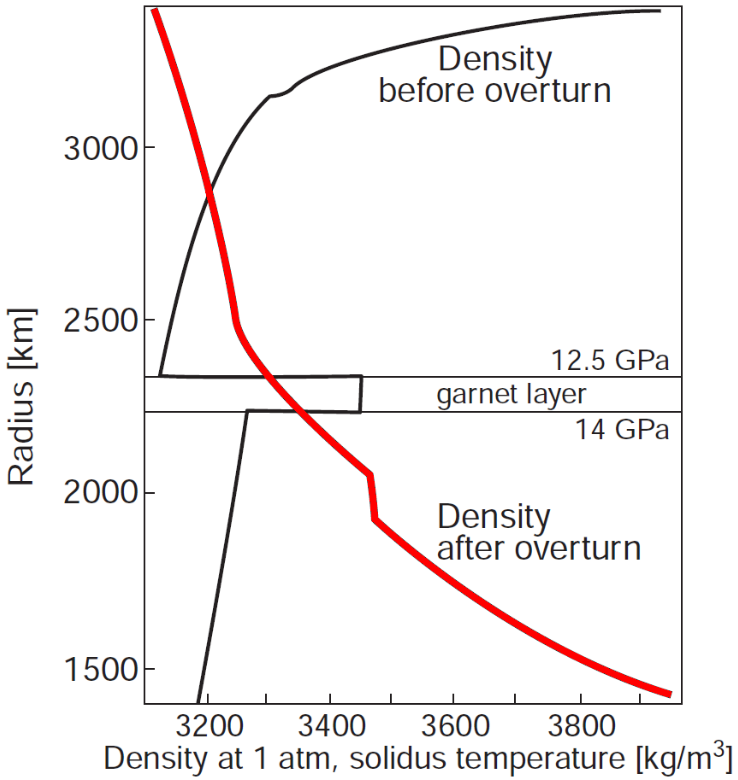


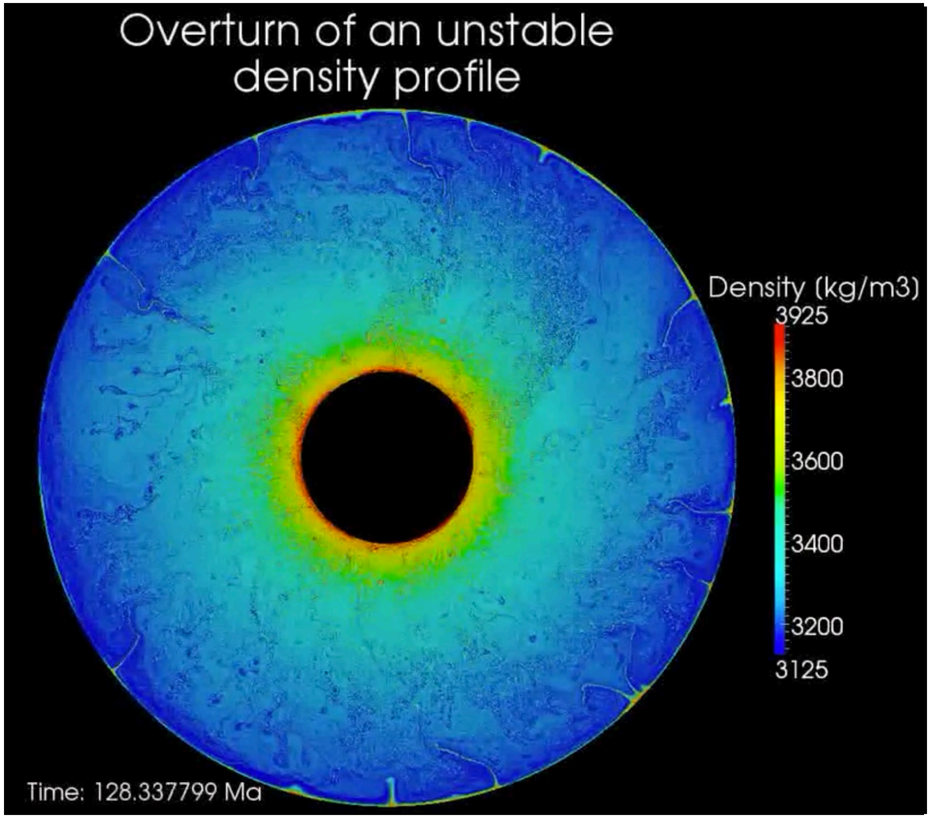
Mars Global Surveyor MOLA-Map, GSFC/ NASA/JPL


Volcanic History & Crystallization Age of Martian Meteorites

Magma Ocean Cumulate Overturn

- Fractional crystallization ⇒ unstable density gradient ⇒ overturn ⇒ stably stratified mantle
- Late mantle cumulates enriched in incompatible heat producing elements ⇒ upon overturn, heat sources accumulate at the CMB





[Elkins-Tanton et al., EPSL, 2005]

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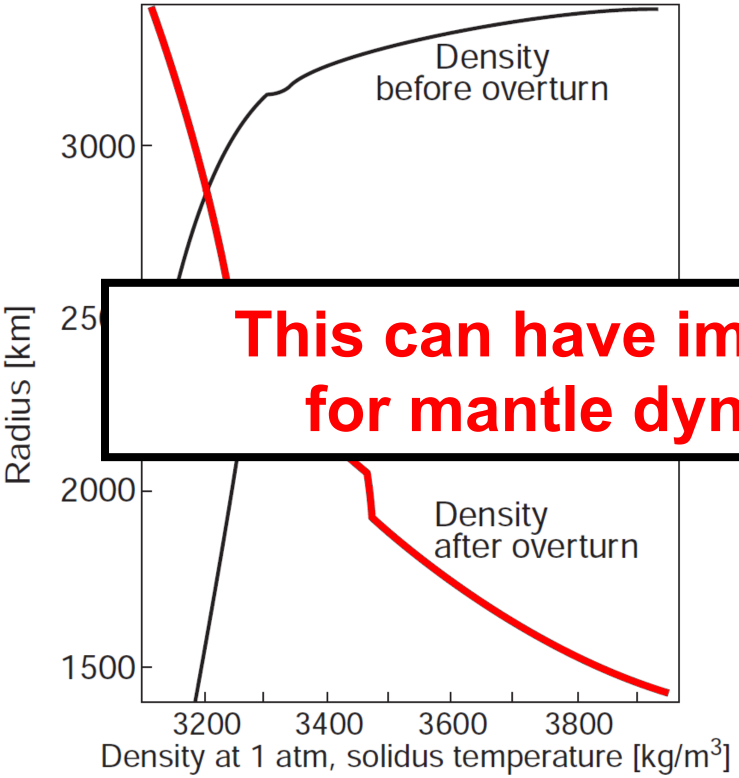
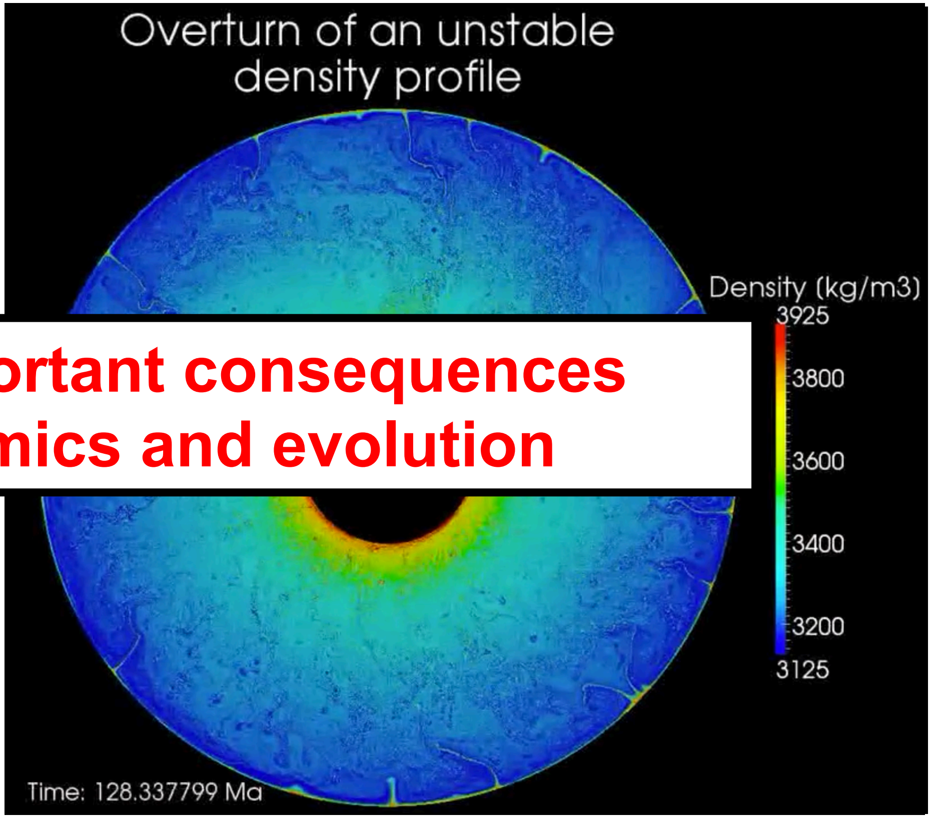
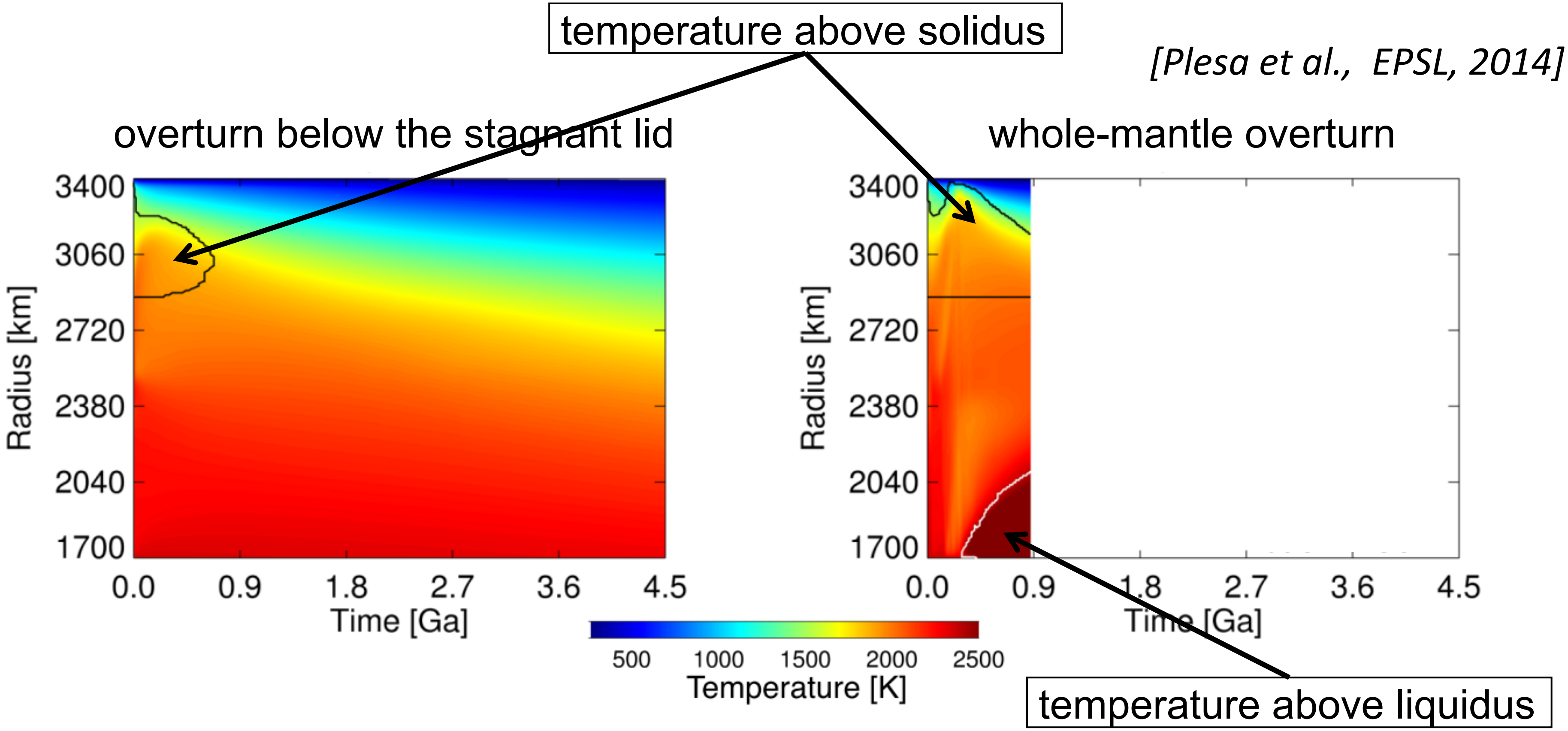


[Elkins-Tanton et al., EPSL, 2005]



This can have important consequences for mantle dynamics and evolution

Subsequent evolution after overturn



temperature above solidus

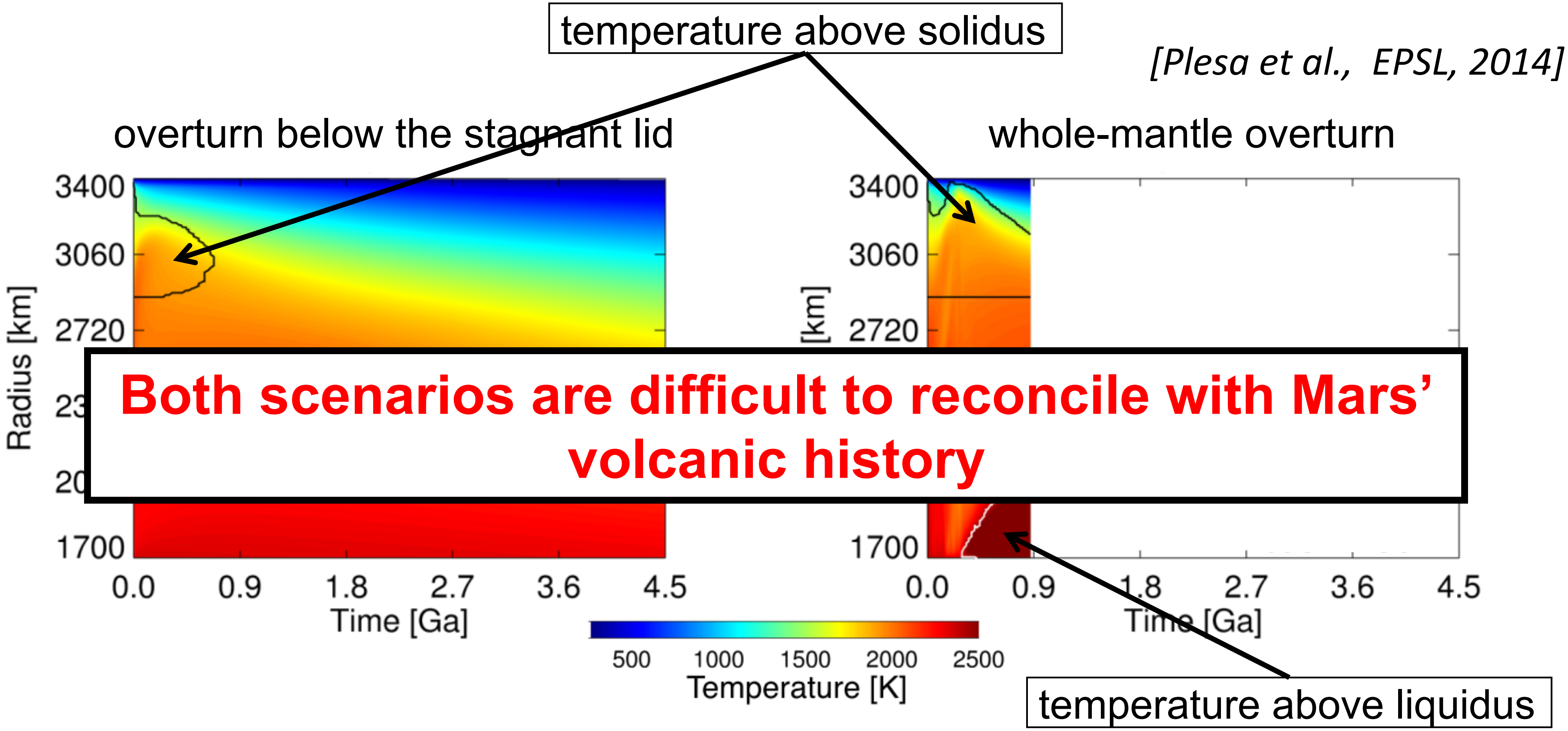
[Plesa et al., EPSL, 2014]

temperature above liquidus

- Overturn below the stagnant lid: mantle cools conductively, short phase of mantle melting (< 1Ga).
- Whole-mantle overturn: mantle overheating above the CMB, temperatures above the liquidus, melt likely negatively buoyant.

Subsequent evolution after overturn

[Plesa et al., EPSL, 2014]

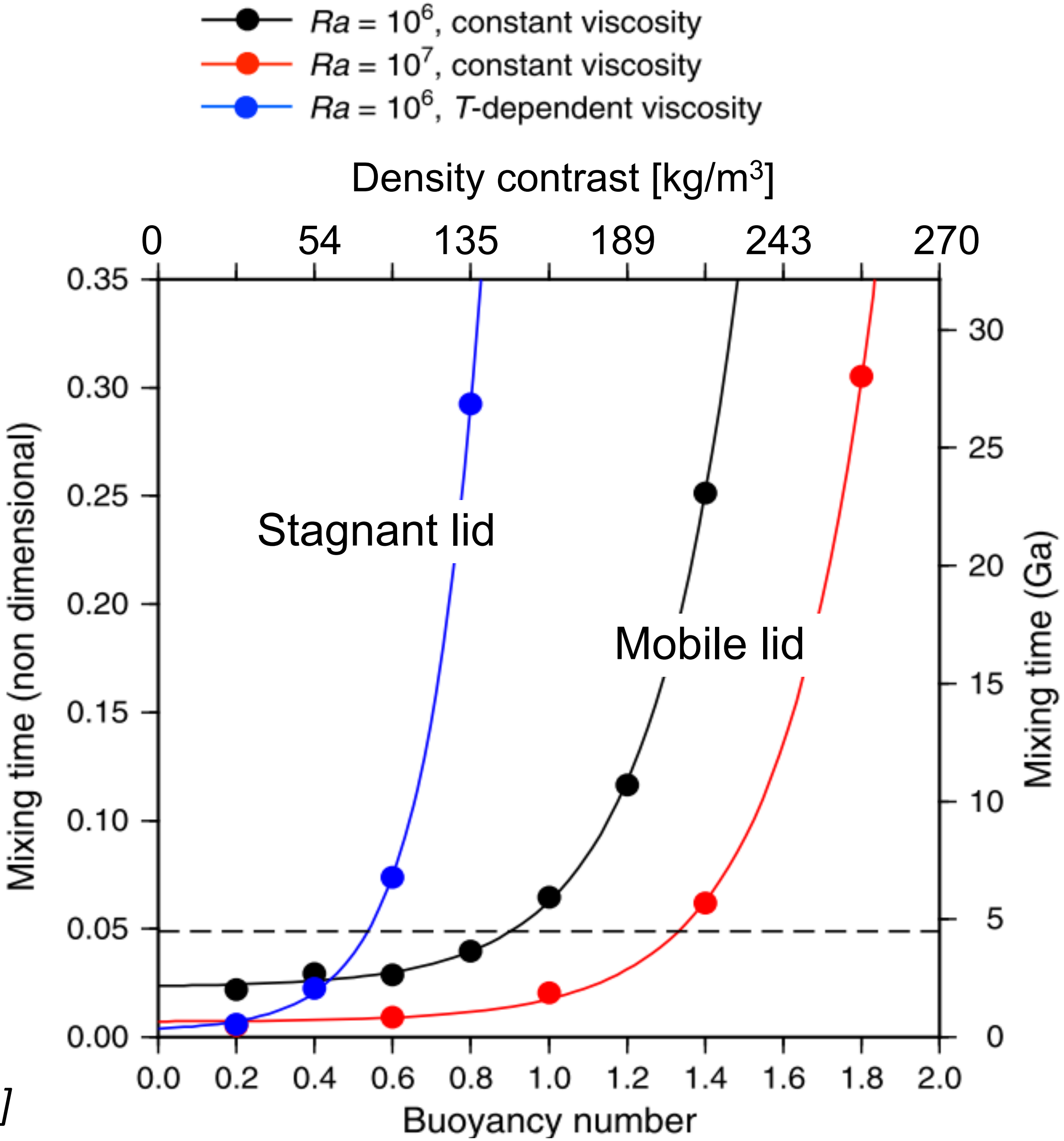


Both scenarios are difficult to reconcile with Mars' volcanic history

- Overturn below the stagnant lid: mantle cools conductively, short phase of mantle melting (< 1Ga).
- Whole-mantle overturn: mantle overheating above the CMB, temperatures above the liquidus, melt likely negatively buoyant.

Reservoir stability: Mixing time scaling

- Mixing time scales exponentially with the density contrast (Δρ).
- Mixing only occurs for the smallest values of Δρ (i.e, Δρ < 60kg/m^3).
- For a one-plate planet it is very difficult to erase chemical heterogeneities via mantle mixing apart from the smallest Δρ.



[Tosi et al., JGR, 2013]

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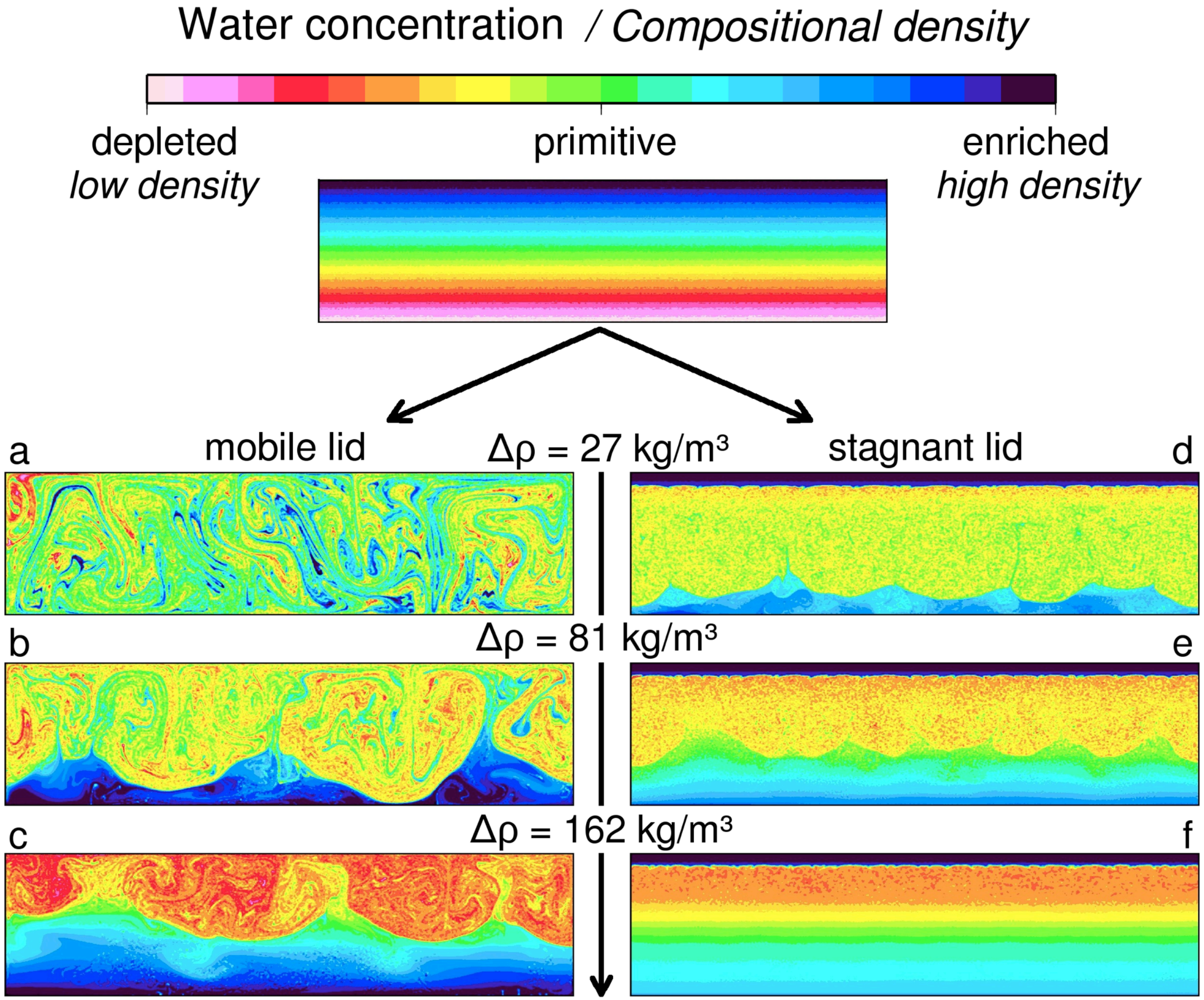


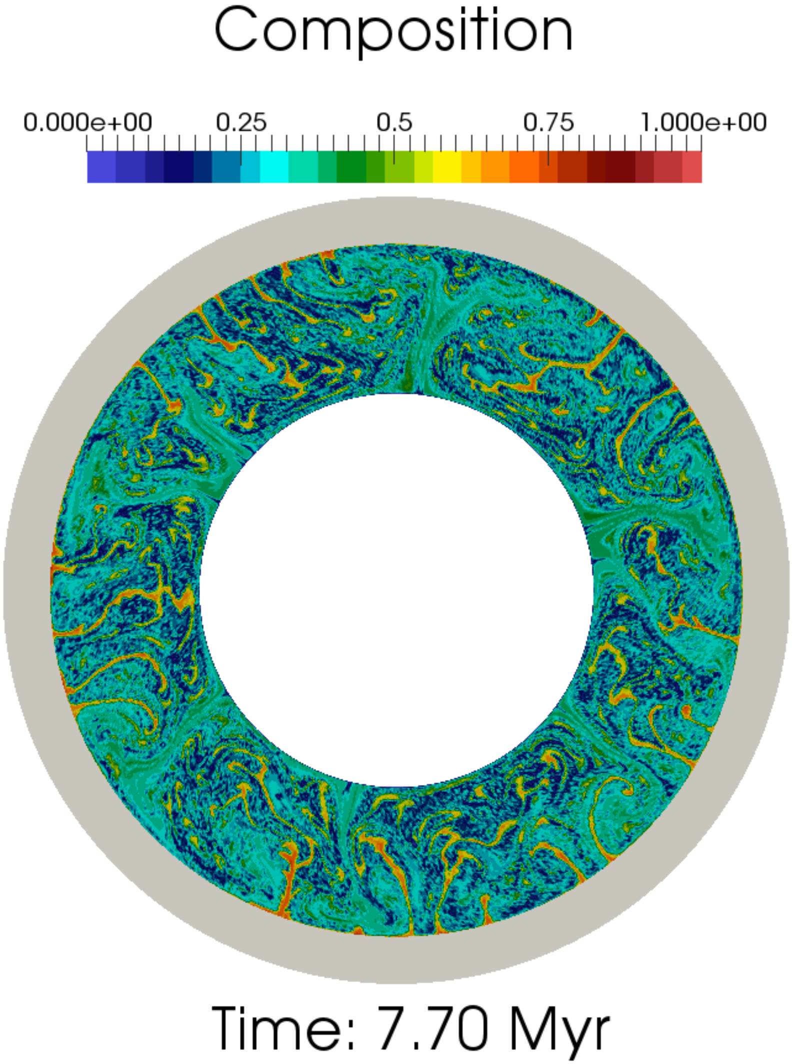
[Tosi et al., JGR, 2013; Breuer et al., MAPS, 2016, in press]

Mixing during the magma ocean crystallization

- Onset of solid state convection may occur provided that the crystallization time is longer than 1 Myr.

- Mixing of chemical heterogenities may take place during the MO crystallization.

- Chemical heterogeneities may be reduced or even erased during the crystallization phase.

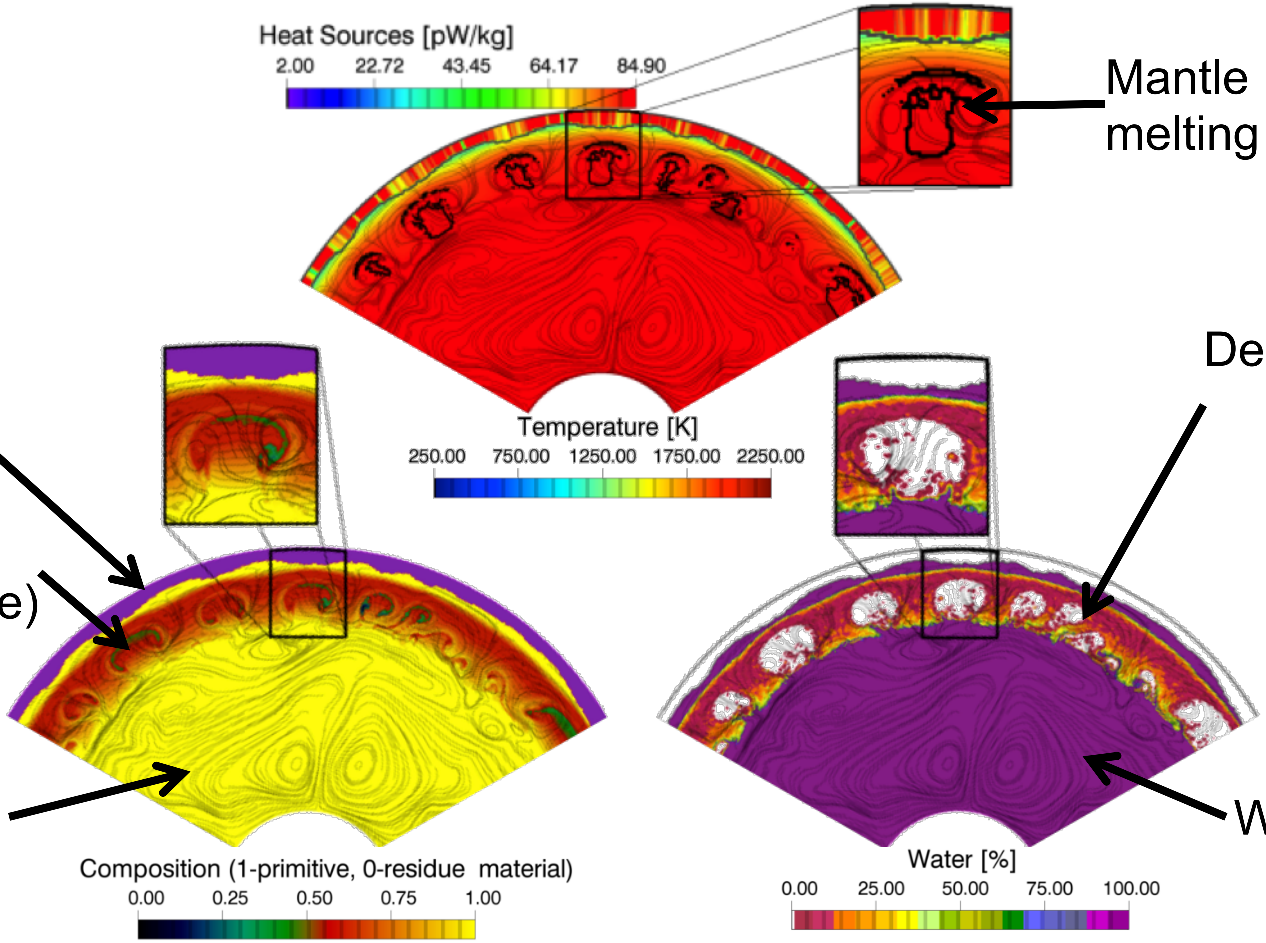


Other sources of chemical heterogeneities?

[Maurice et al., in prep.]

Chemical heterogeneities from partial melting

- Subsequent partial melting of the mantle



Mantle melting

Crust

Depleted mantle (light residual mantle)

Primitive mantle

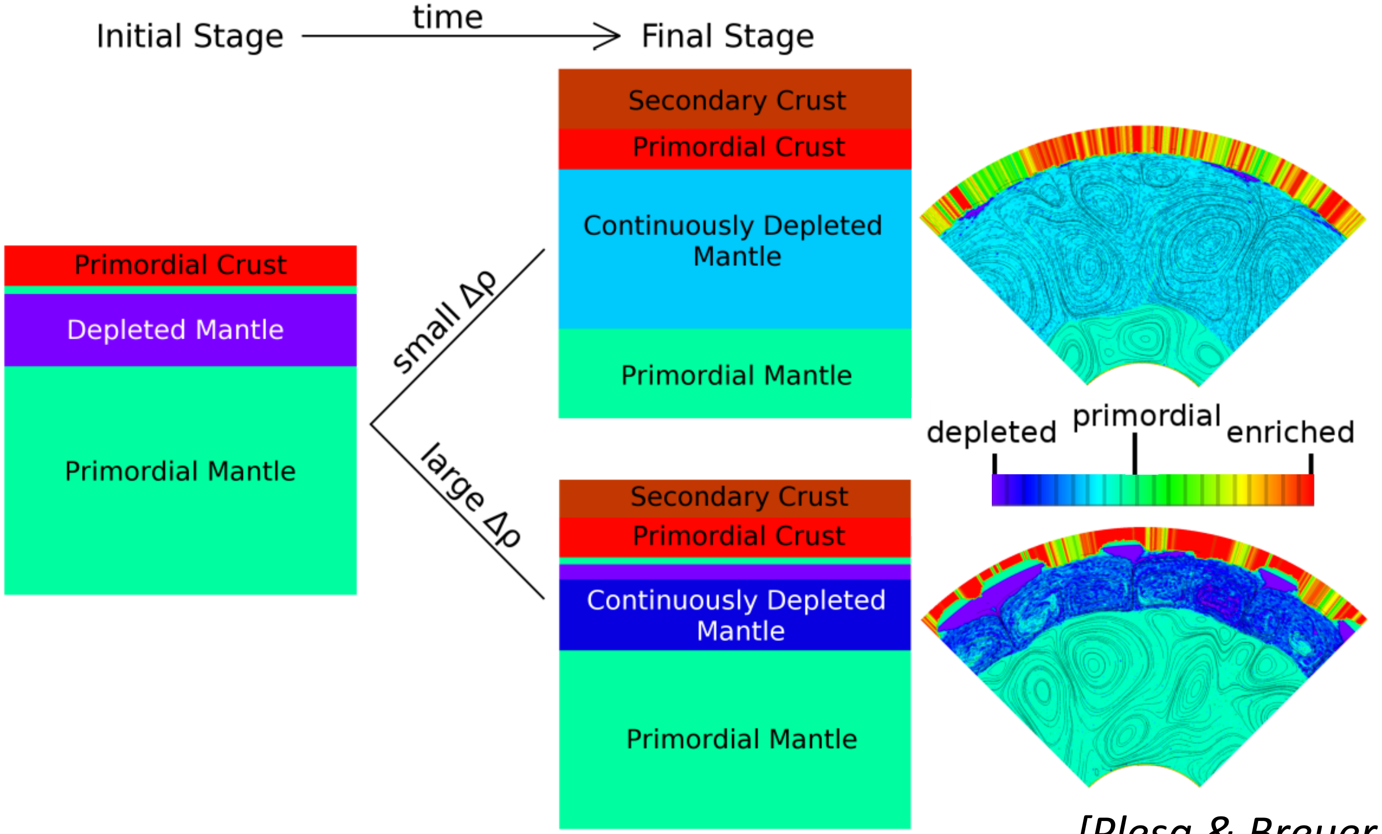
Dehydrated (stiff) mantle

Wet mantle

[Plesa & Breuer, PSS, 2014]

Chemical heterogeneities from partial melting

- Generation of reservoirs by partial melting and secondary differentiation:



[Plesa & Breuer, PSS, 2014]

- Reservoirs may change/new reservoirs can form depending in particular on the density difference between primordial and depleted mantle.

Chemical heterogeneities from partial melting

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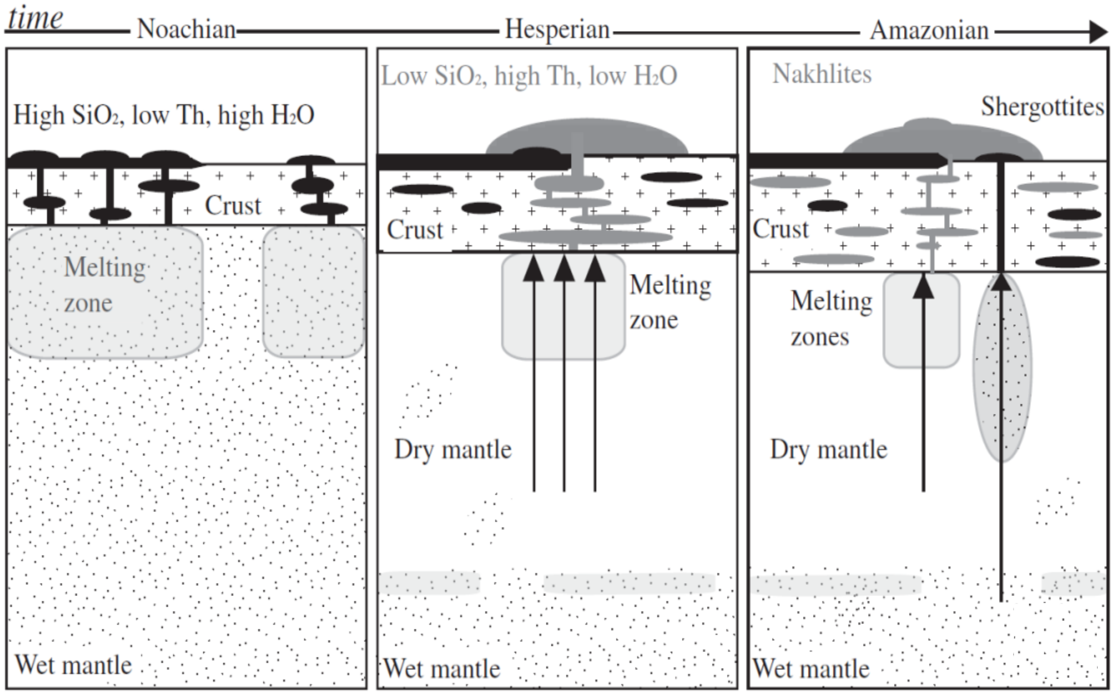
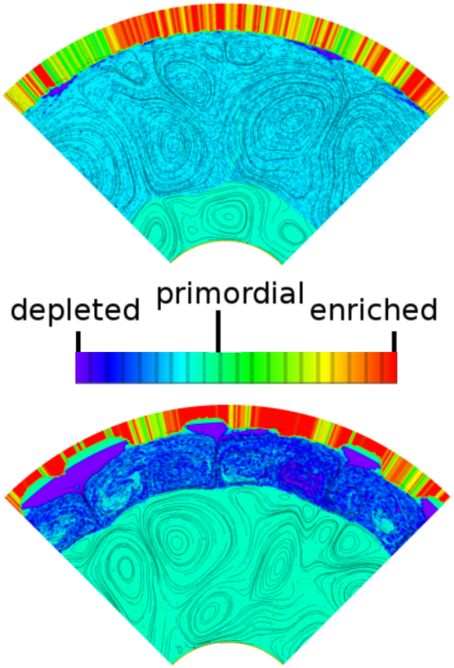


[Balta & McSween, GSA, 2013]



[Plesa & Breuer, PSS, 2014]

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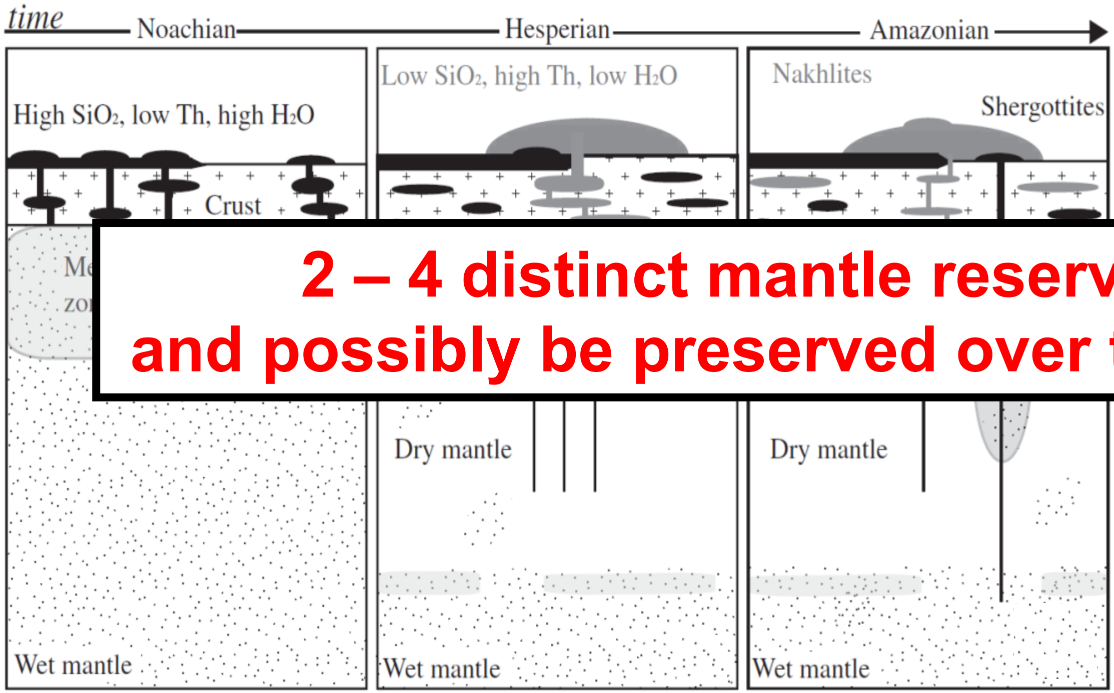

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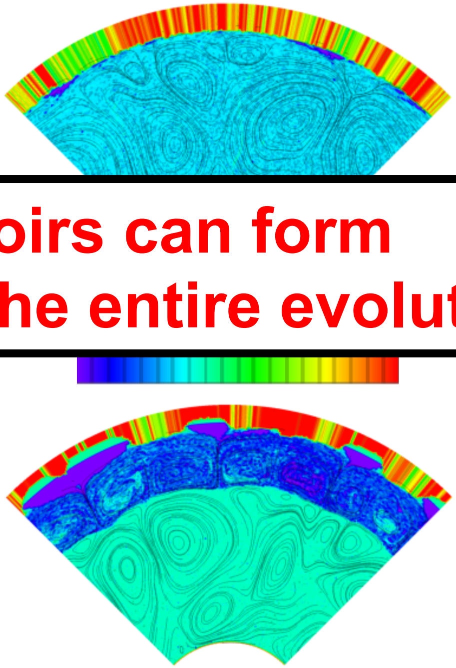
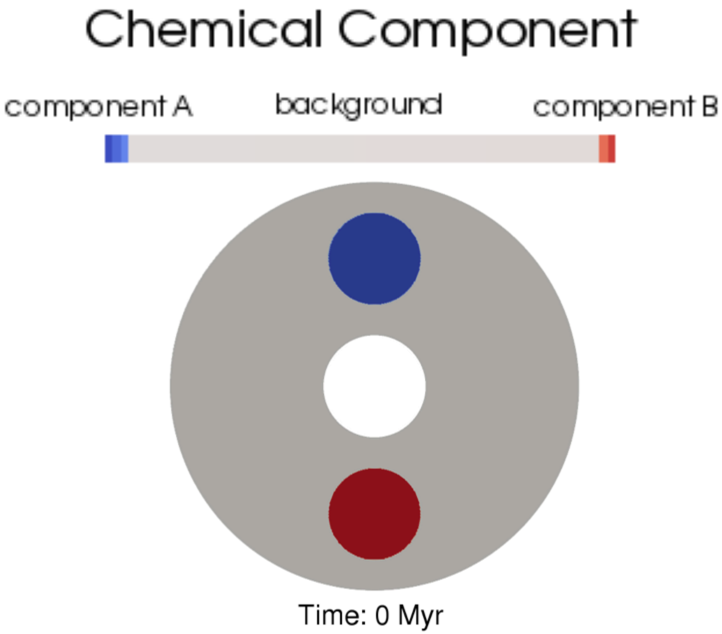


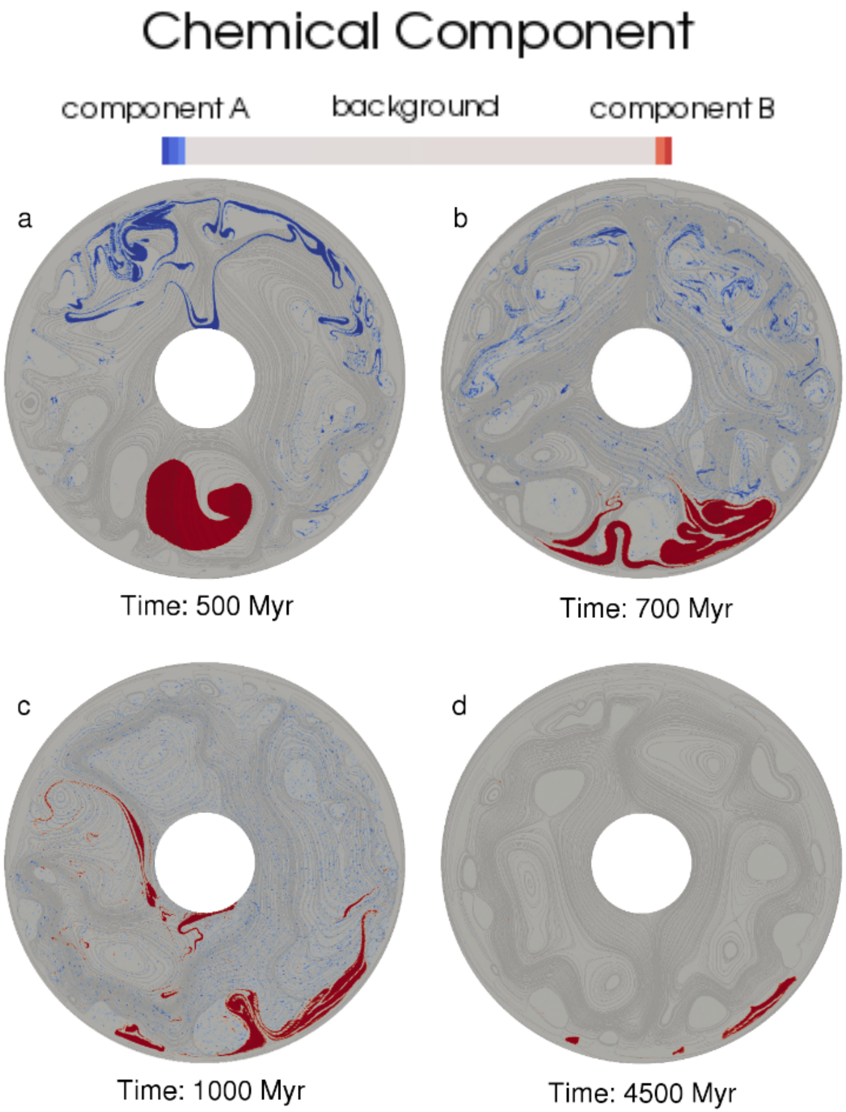
[Plesa & Breuer, PSS, 2014]

2 – 4 distinct mantle reservoirs can form and possibly be preserved over the entire evolution

- Reservoirs may change/new reservoirs can form depending in particular on the density difference between primordial and depleted mantle.

Local sources of chemical heterogeneities





[Breuer et al., MAPS, 2016, in press]

Two components located initially in the mantle with the following properties:

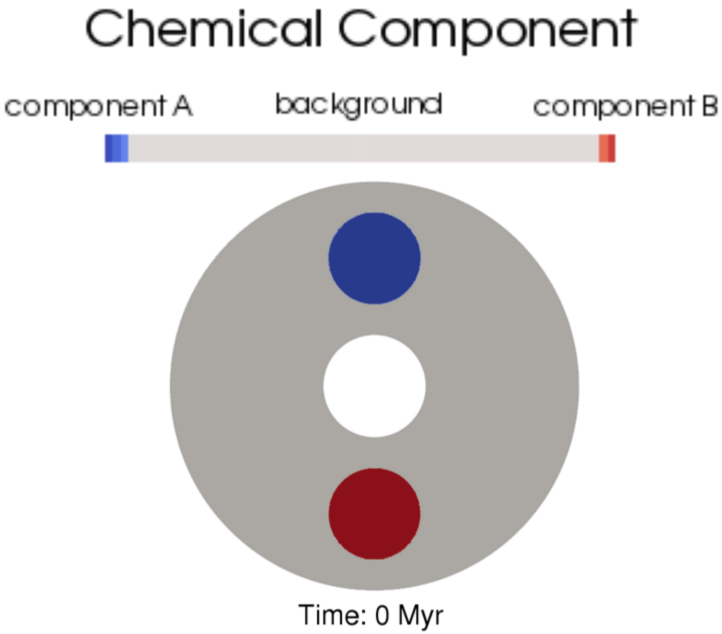
- Density:

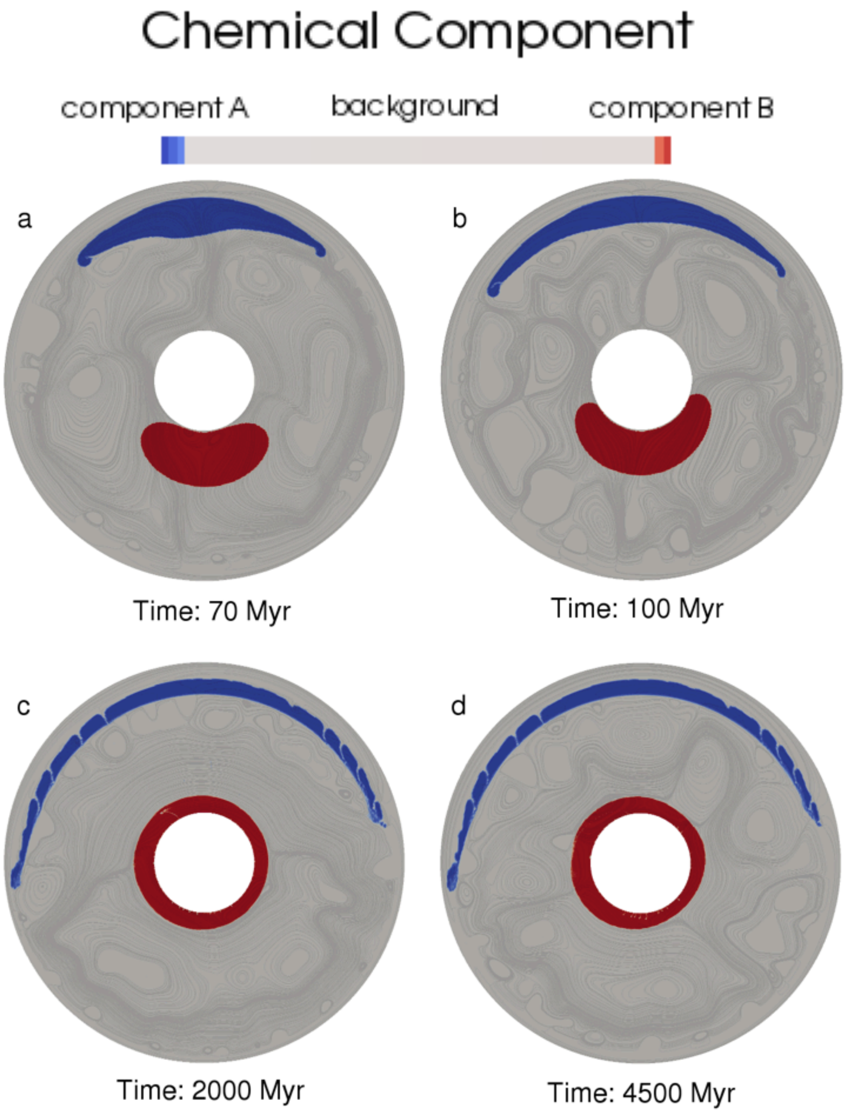
  $\rho_A = \rho_B = \rho_{background}$
- Viscosity:

  $\eta_A = 10 \times \eta_{background}$

  $\eta_B = 100 \times \eta_{background}$

Local sources of chemical heterogeneities





Two components located initially in the mantle with the following properties:

- Density:

  $\rho_A = \rho_{background} - 60\ kg/m^3$

  $\rho_B = \rho_{background} + 60\ kg/m^3$

- Viscosity:

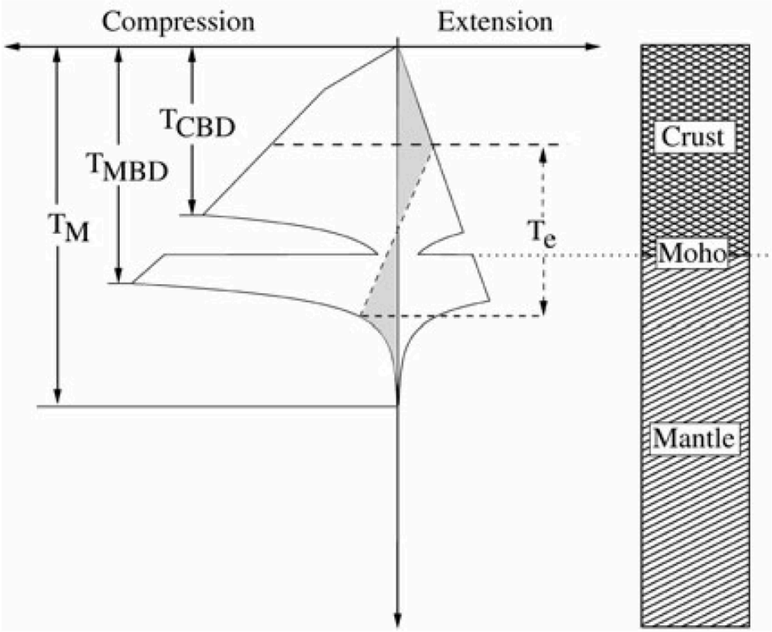
  $\eta_A = \eta_B = 100 \times \eta_{background}$

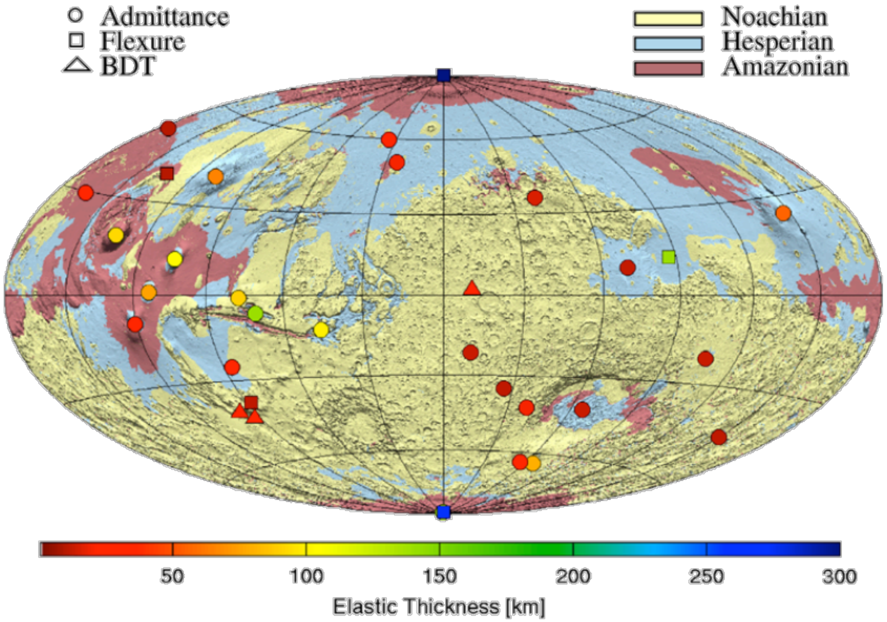
[Breuer et al., MAPS, 2016, in press]

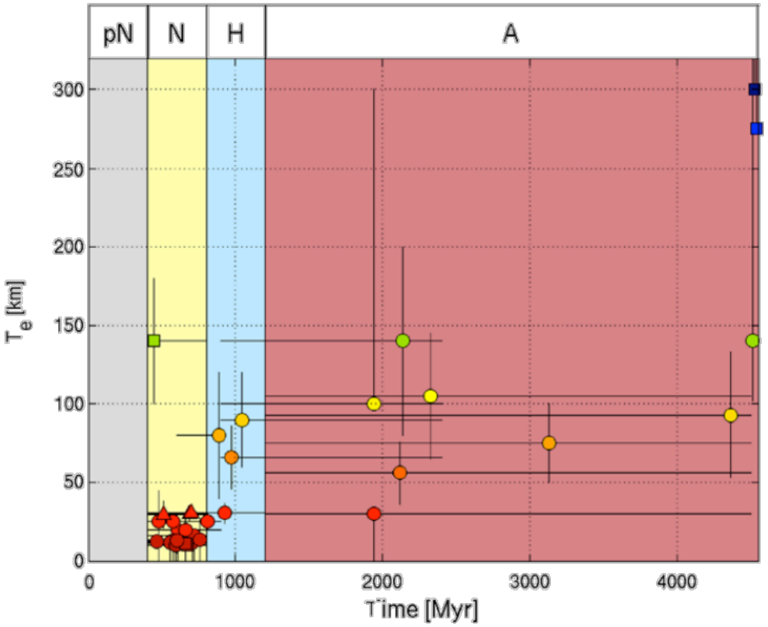
Thermochemical evolution

Mars:

- Elastic lithosphere thickness data available for Mars show planetary cooling over time:
  - Noachian: 20 – 30 km
  - Present-day: > 300 km



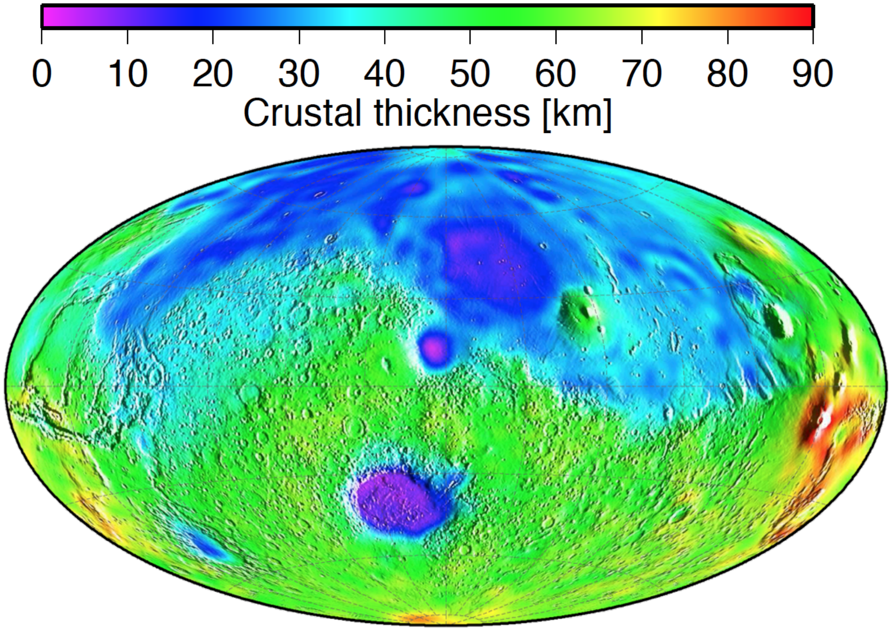


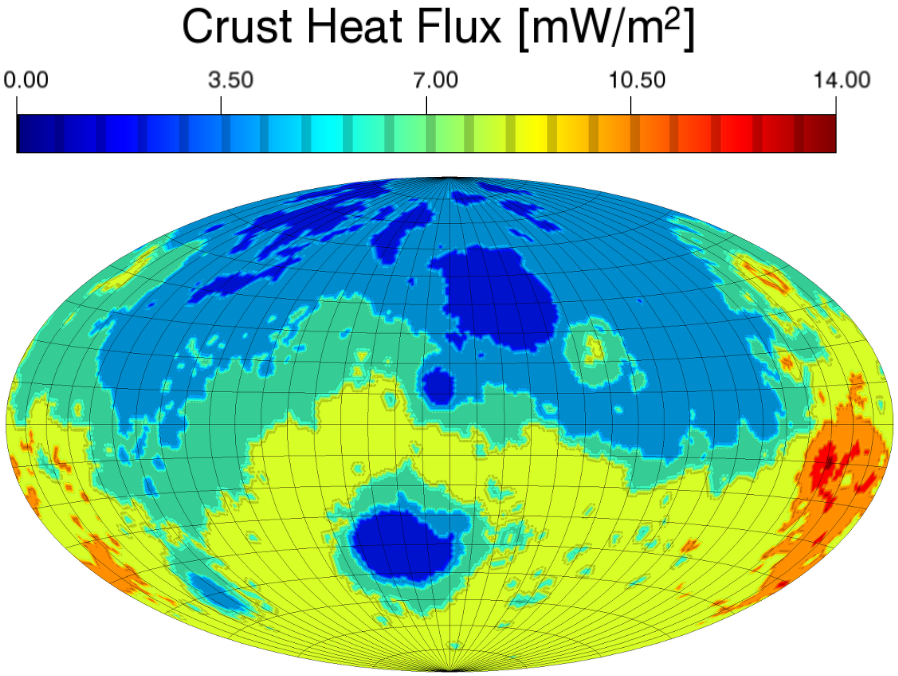


[Grott et al., SSR, 2013]

Evolution of the elastic lithosphere thickness

- 3D fully dynamical simulations of interior thermal evolution.
- Assuming a crust
  - using the model of Neumann et al. [2004]
  - with low conductivity (3 W/mK)
  - enriched in HPE [Hahn et al., 2011]
- Investigate the role of the convection planform on the elastic thickness and heat flow variations.



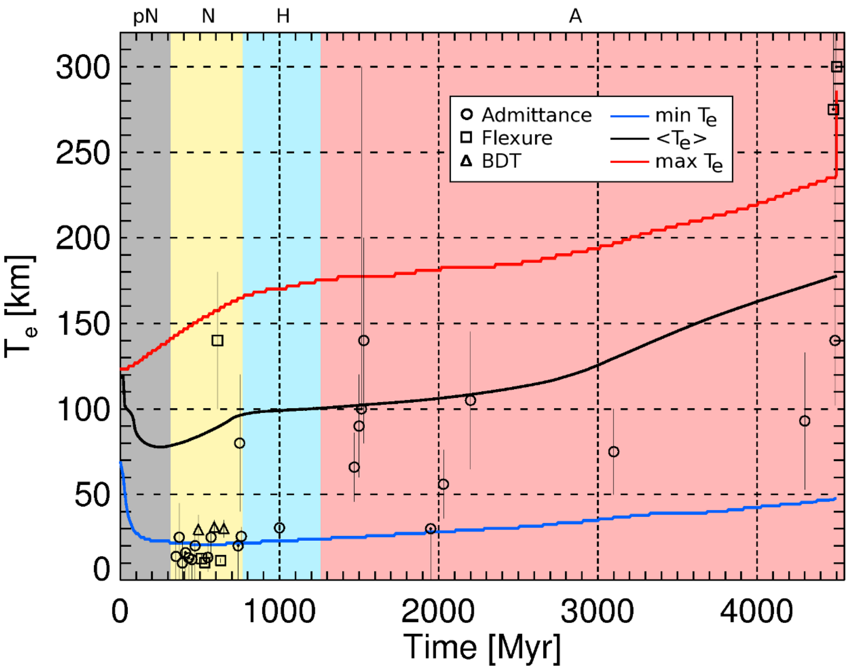


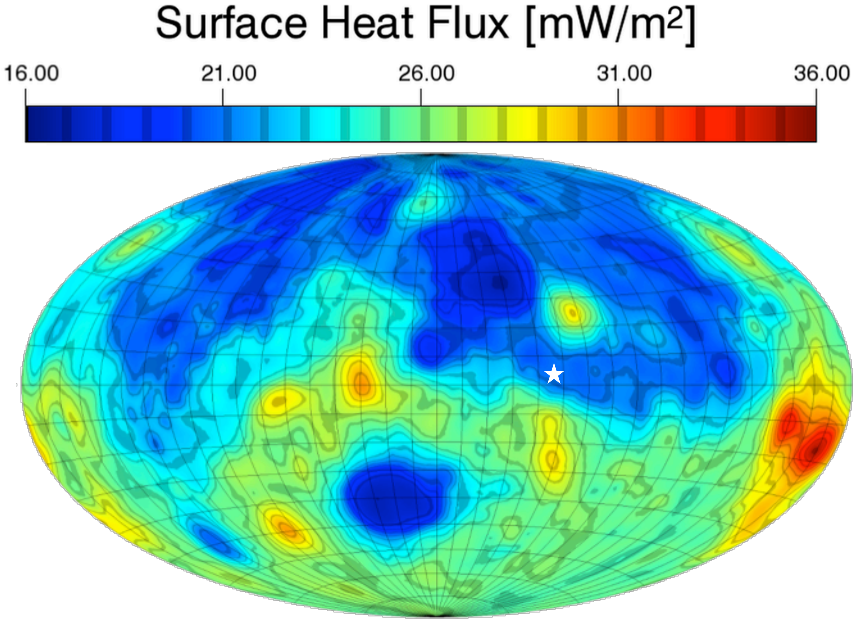
[Plesa et al., JGR, in review]

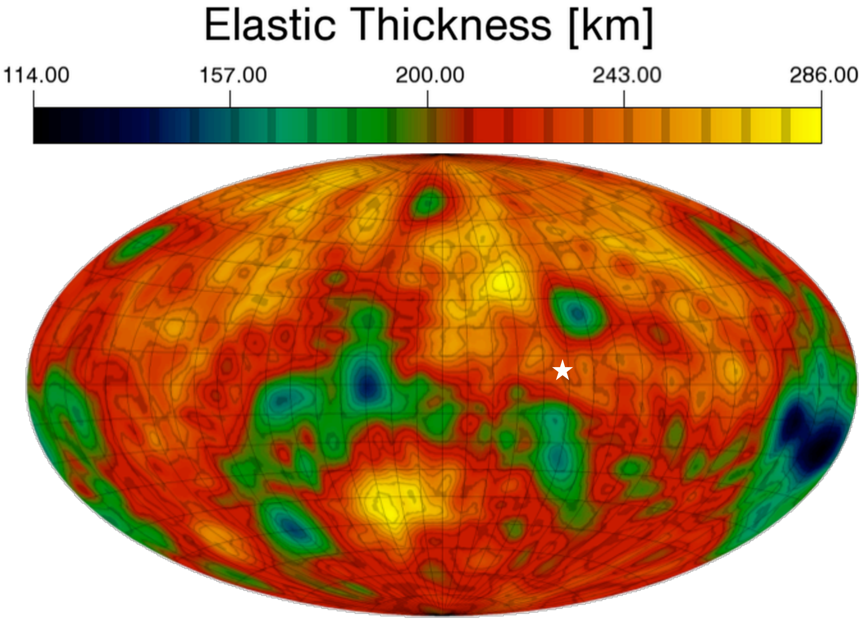
Evolution of the elastic lithosphere thickness

- Mantle plumes can produce large variations of T_e.
- Present-day surface heat flux and T_e distribution is dominated by the crust structure.

[Breuer et al., MAPS, 2016, in press]




Surface Heat Flux [mW/m²]


Elastic Thickness [km]

[Plesa et al., JGR, in review]

Conclusions

- Need of more sophisticated thermochemical models that can be compared with data from missions and samples.
- Still many open questions:
  - Mercury southern hemisphere
    (BepiColombo 2018)
  - Moon far side & sample return
    (Chang‘e missions 2017 & 2018)
  - Mars in-situ seismic & heat flow measurements & sample return
    (InSight 2018 & ExoMars & Mars 2020)