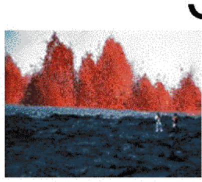
Heat budget & Thermal Evolution

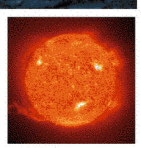
Louise Kellogg, University of California, Davis

- Sources of heat in the Earth
- Transport of heat
- Aspects of thermal evolution
- Can a planetary perspective help?

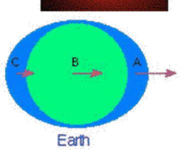
Some powerful numbers...



Total global heat flow: 44 TW



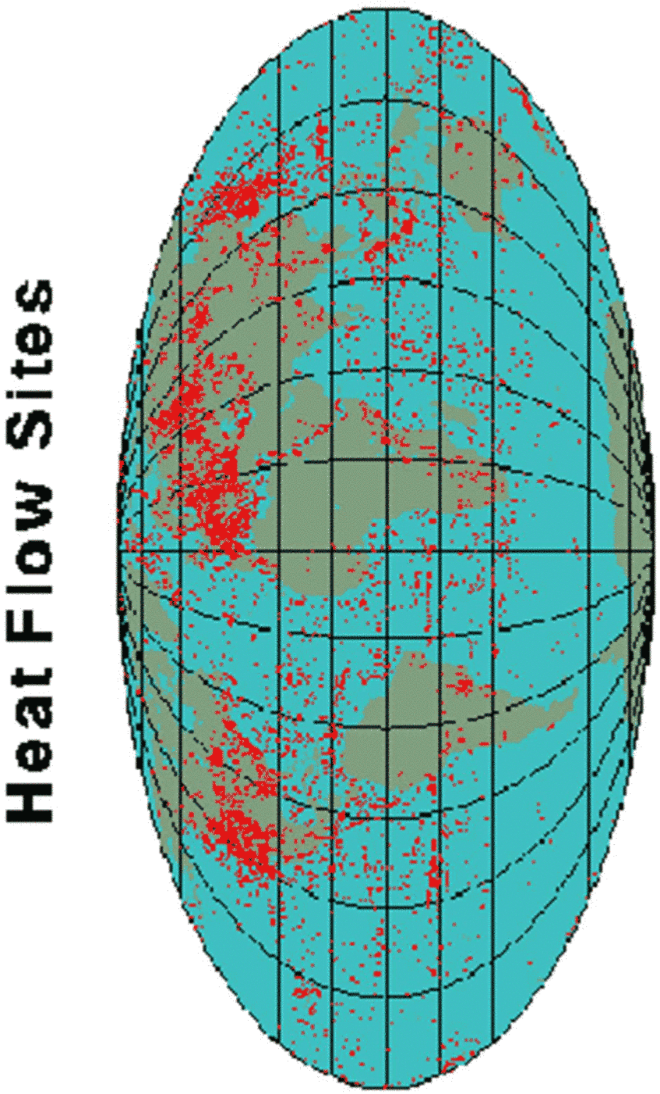
Sun: Solar energy received (and re-radiated) 180,000 TW

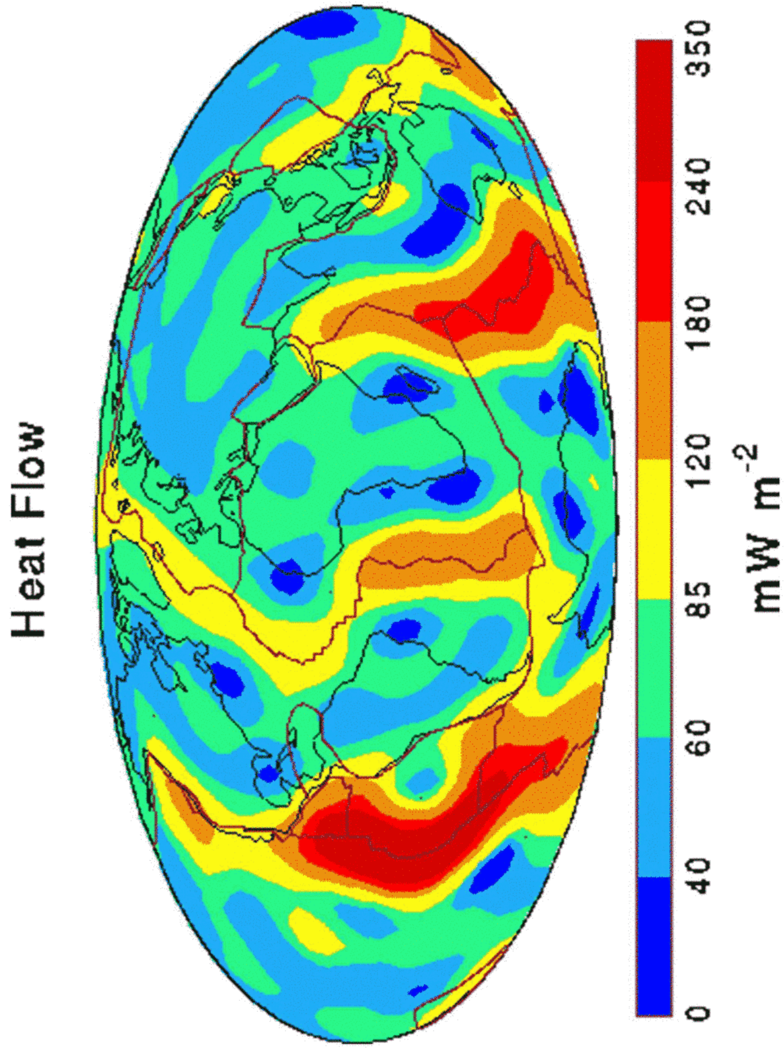


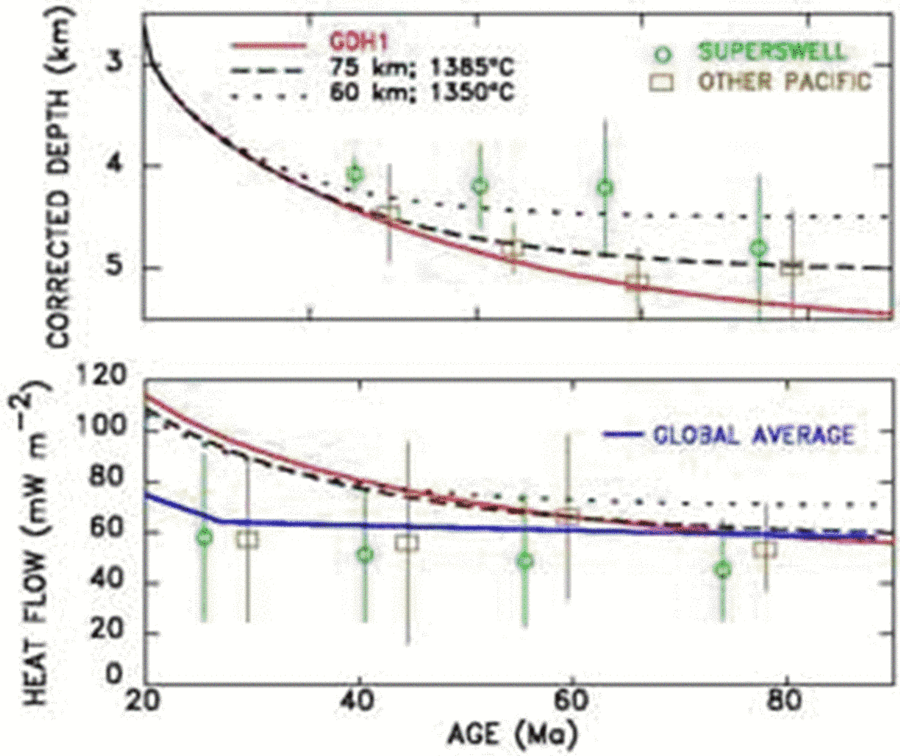
Tides: Dissipation of Earth's rotational energy 3 TW

Earthquakes: Elastic wave energy released 0.03 TW

Humans: World power production 12 TW


Heat Flow Sites


Heat Flow
0
40
60
85
120
180
240
350
mW m^-2



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Mantle Convection: Basic Version of Equations

Conservation of mass: $\nabla \cdot \vec{u} = 0$

Momentum: $-\nabla P + \nabla \cdot \tau + RaT\hat{k} = 0$

Heat flow: $\frac{\partial T}{\partial t} + \vec{u} \cdot \nabla T = \nabla^2 T + Ra_H / Ra$

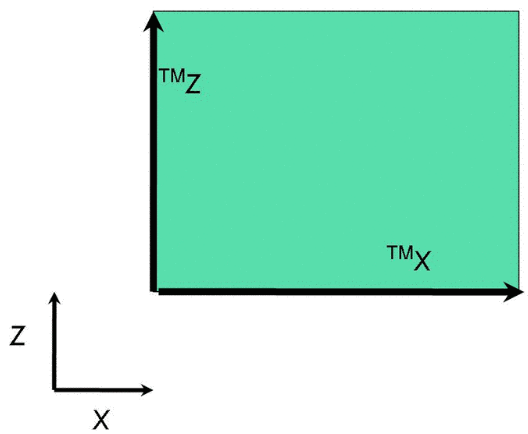
Viscosity varies with temperature and depth: $\mu = \mu_0 e^{-\frac{E^* + Vz}{RT}}$

In this version, two parameters are of primary importance:

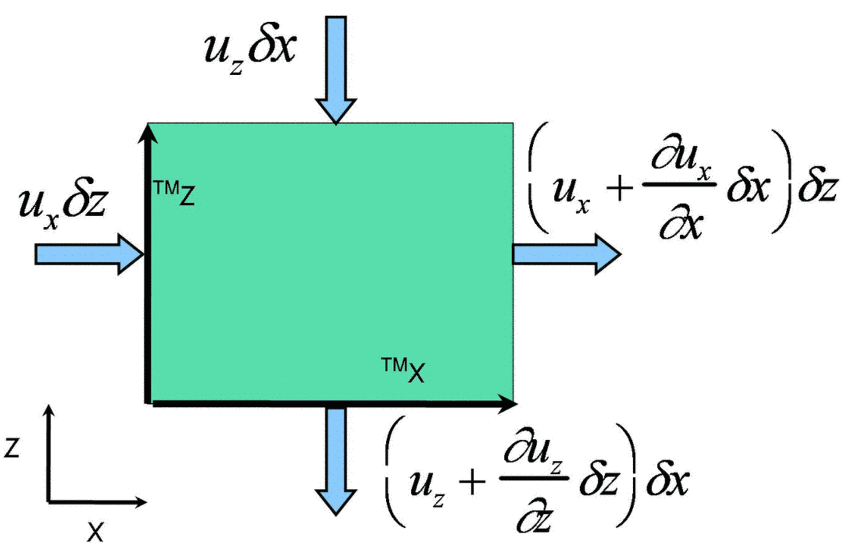
$$Ra = \frac{\rho_m g \alpha \Delta T d^3}{\kappa \mu_0} \qquad Ra_H = \frac{\rho_m^2 g \alpha H d^5}{k \kappa \mu_0}$$

Conservation of Mass Equation

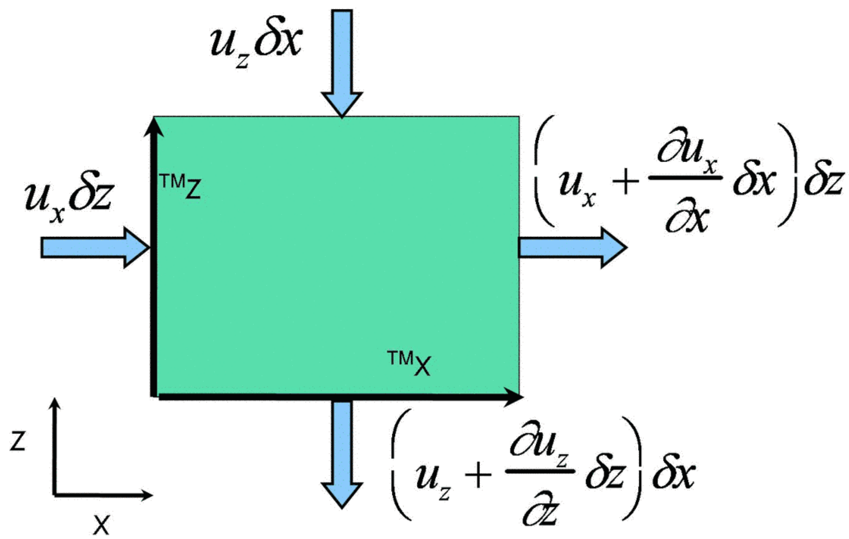
Consider a particle of fluid



For an incompressible fluid,
flow into the box balances the flow out

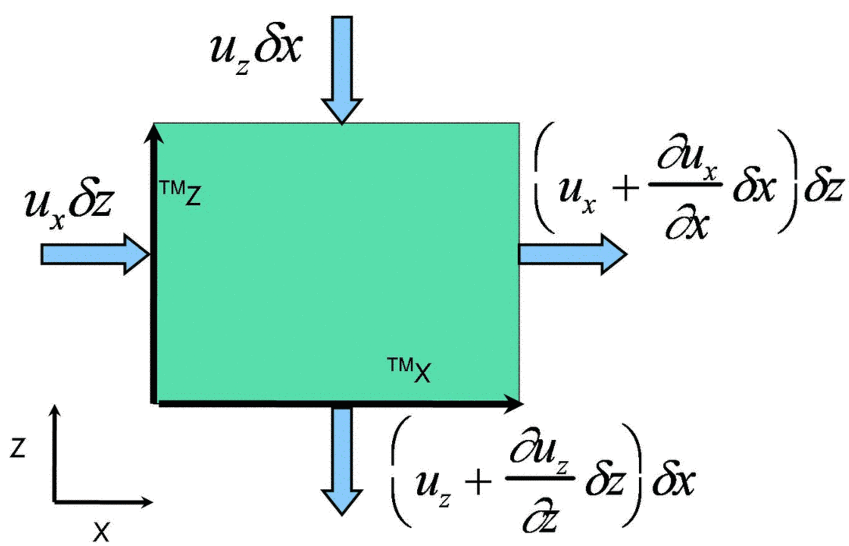


$$\left(u_x + \frac{\partial u_x}{\partial x}\delta x\right)\delta z - u_x\delta z + \left(u_z + \frac{\partial u_z}{\partial z}\delta z\right)\delta x - u_z\delta x = 0$$



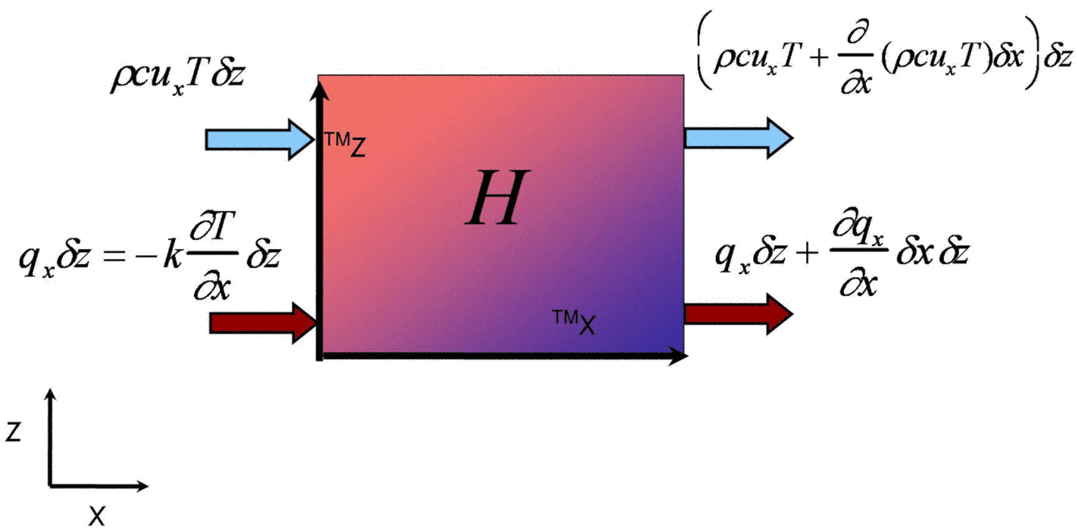
Conservation of Mass Equation

$$\frac{\partial u_x}{\partial x} + \frac{\partial u_z}{\partial z} = 0 \quad \text{that is,} \quad \nabla \cdot \vec{u} = 0$$



Conservation of Energy (heat flow)

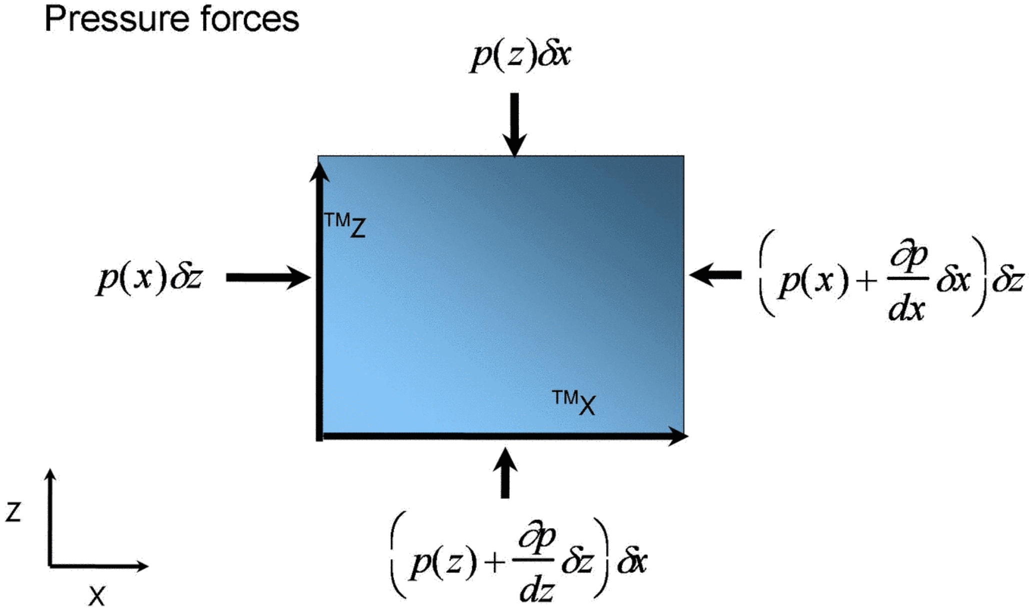
$$\frac{\partial T}{\partial t} + \vec{u} \cdot \nabla T = \nabla^2 T + Ra_H / Ra$$



Momentum (force balance)

$$-\nabla P + \nabla \cdot \tau + RaT\hat{k} = 0 \qquad Ra = \frac{\rho_m g \alpha \Delta T d^3}{\kappa \mu_0}$$

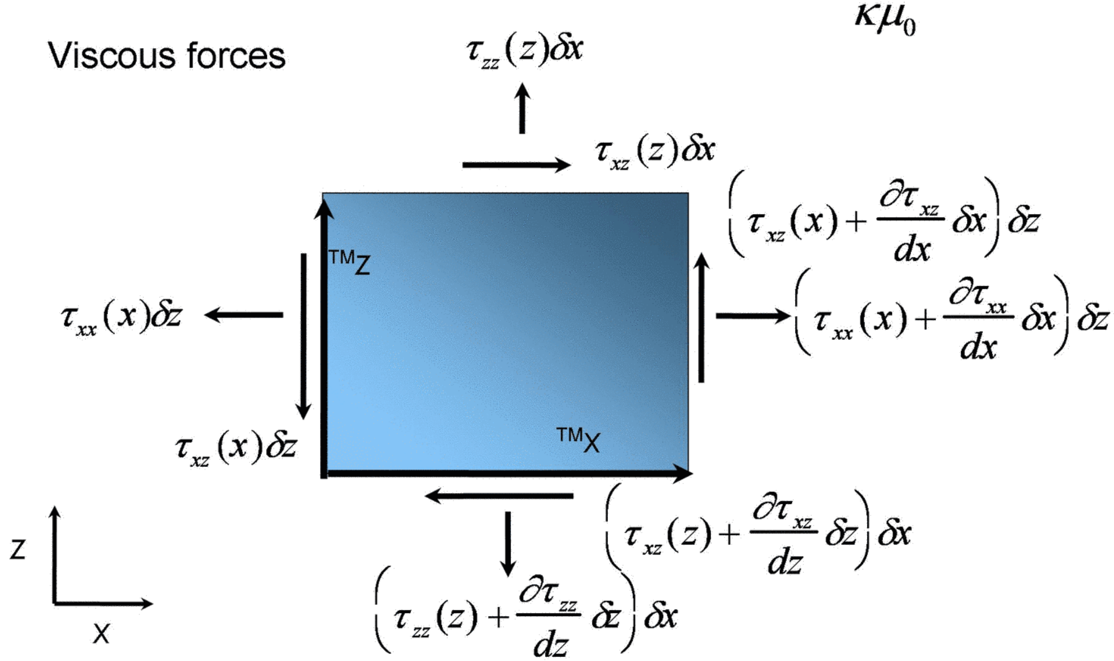
Pressure forces



Momentum (force balance)

$$-\nabla P + \nabla \cdot \tau + RaT\hat{k} = 0 \qquad Ra = \frac{\rho_m g \alpha \Delta T d^3}{\kappa \mu_0}$$

Viscous forces

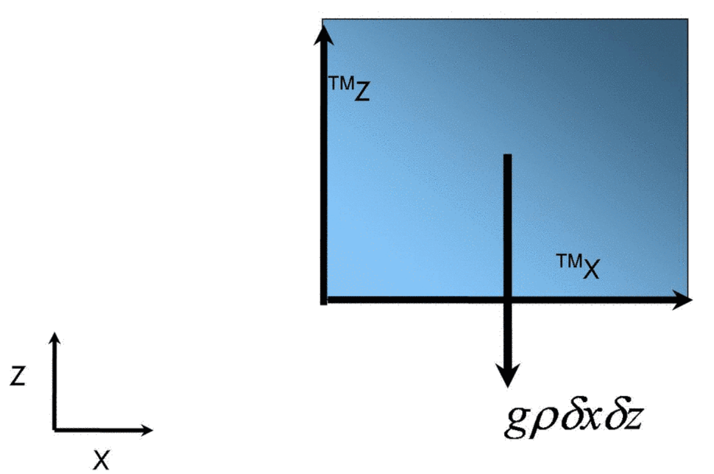


Momentum (force balance)

$$-\nabla P + \nabla \cdot \tau + RaT\hat{k} = 0 \qquad Ra = \frac{\rho_m g \alpha \Delta T d^3}{\kappa \mu_0}$$

Body forces (gravity)

$$\rho(T) = \rho_0(1 - \alpha(T - T_0))$$



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Viscosity varies with temperature and depth: $\mu = \mu_0 e^{-\frac{E^* + Vz}{RT}}$

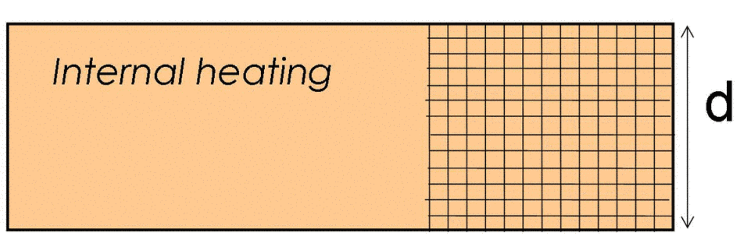
Two parameters are of primary importance:

$$Ra = \frac{\rho_m g \alpha \Delta T d^3}{\kappa \mu_0} \qquad Ra_H = \frac{\rho_m^2 g \alpha H d^5}{k \kappa \mu_0}$$

Numerical methods discretize the region of interest

$$Ra = \frac{\rho_m g \alpha \Delta T d^3}{\kappa \mu_0} \qquad Ra_H = \frac{\rho_m^2 g \alpha H d^5}{k \kappa \mu_0}$$

Top of mantle:
Traction-free surface
Constant temperature



Sides of mantle:
"Wrap around"

Core-mantle boundary:
Traction-free surface
Constant temperature (or constant heat flux)

A simple plume

QuickTime™ and a
Graphics decompressor
are needed to see this picture.

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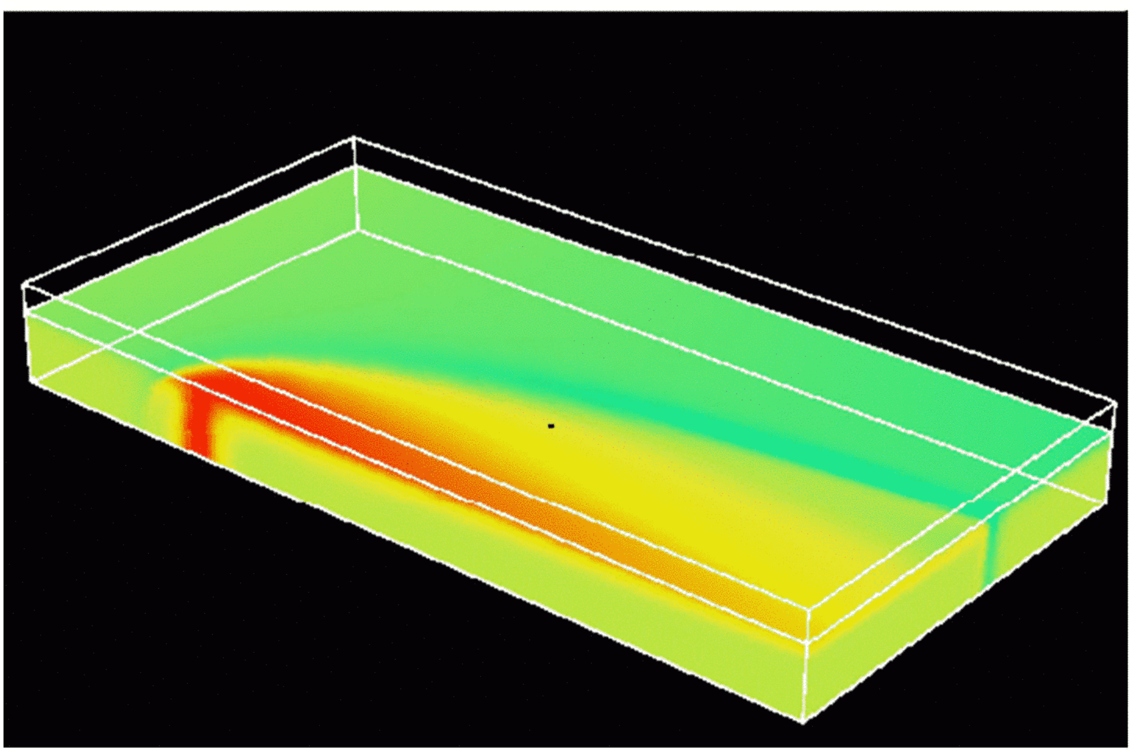


Heat sources from the Earth's interior

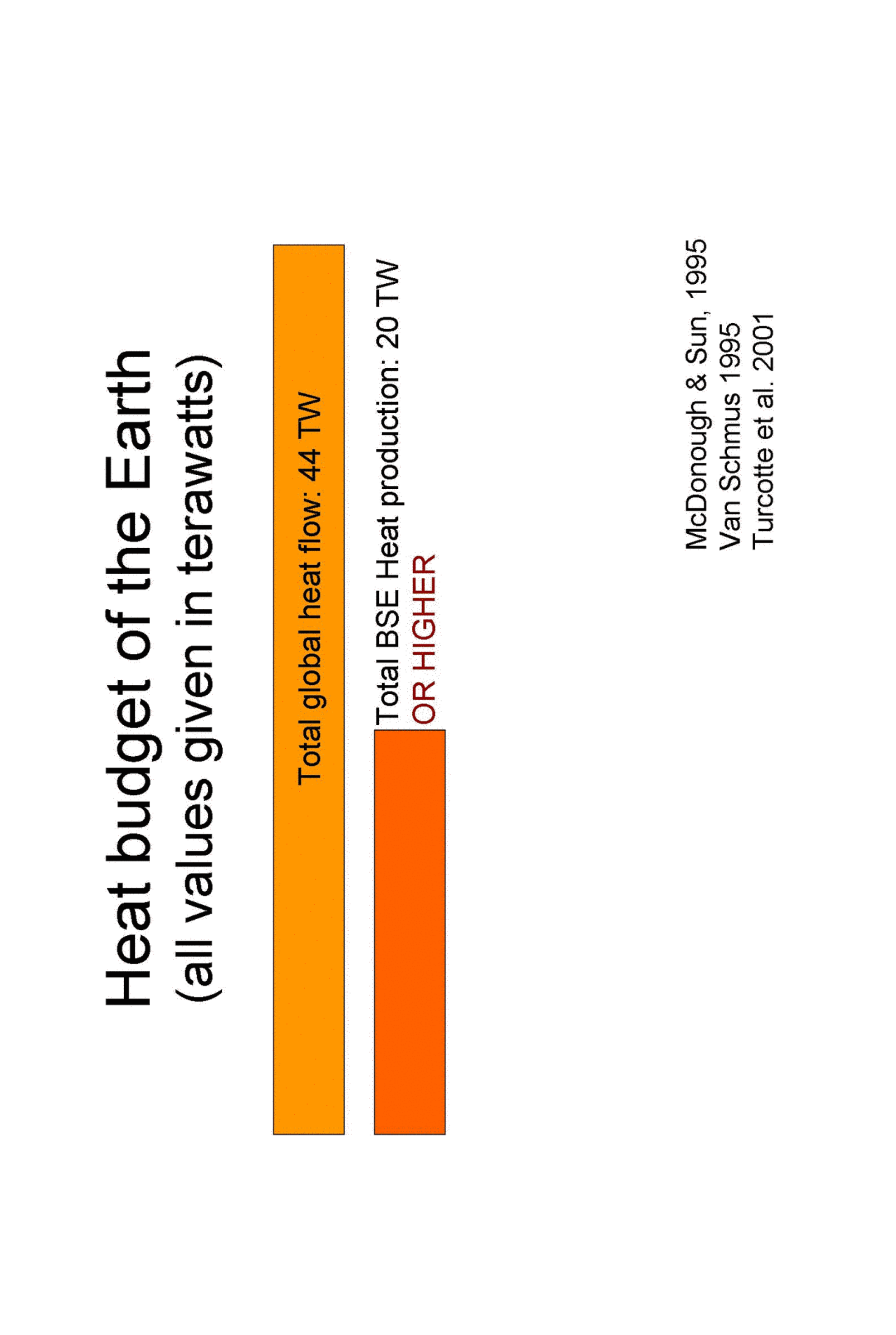
Parent	Daughter	half-life	heat production
238 U	206 Pb	4.49 By	94 ∫ W/kg of U
235 U	207 Pb	0.704 By	570
232 Th	208 Pb	14.0 By	26.6
40 K	40 Ar	1.25 By	27.9

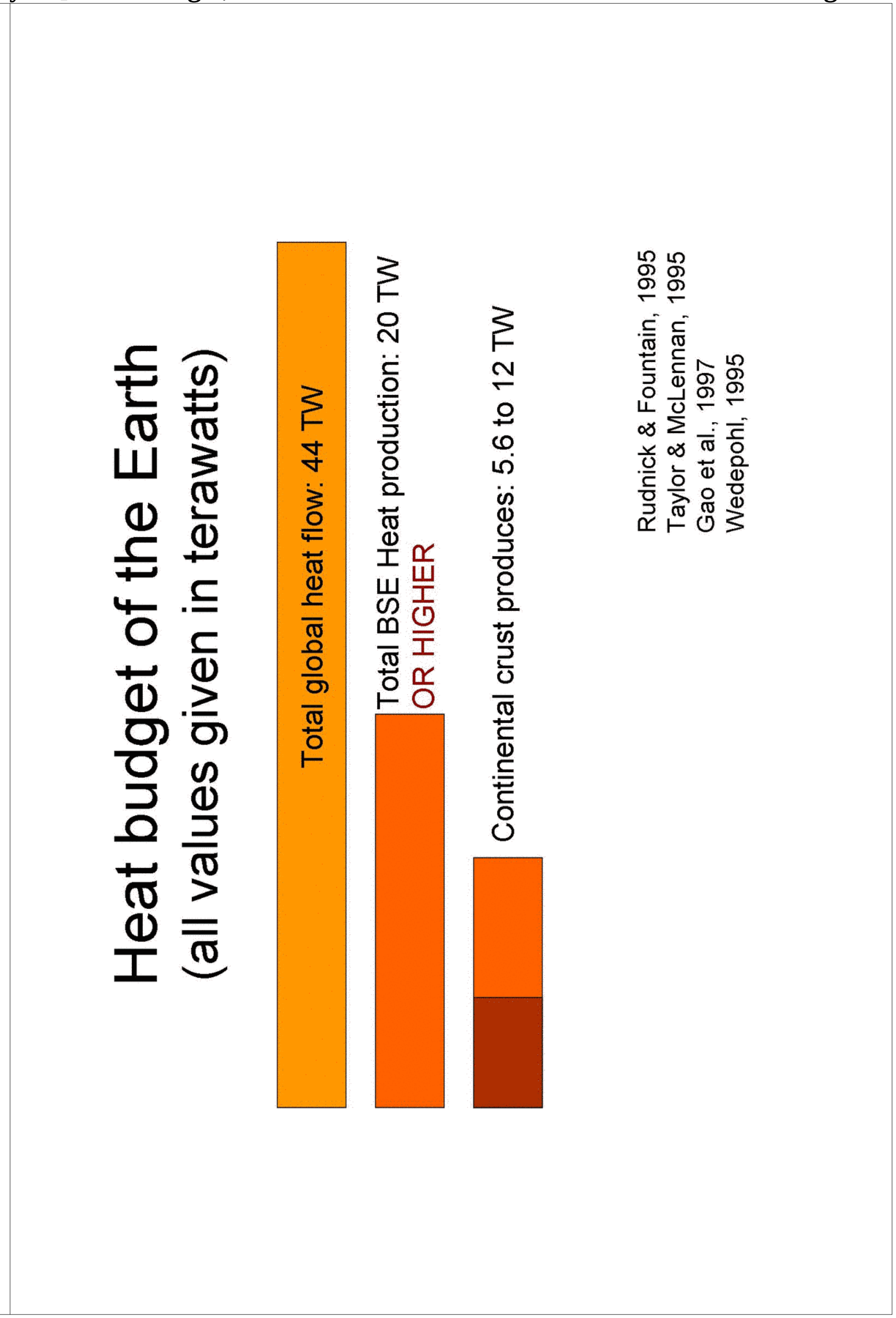
Heat budget of the Earth

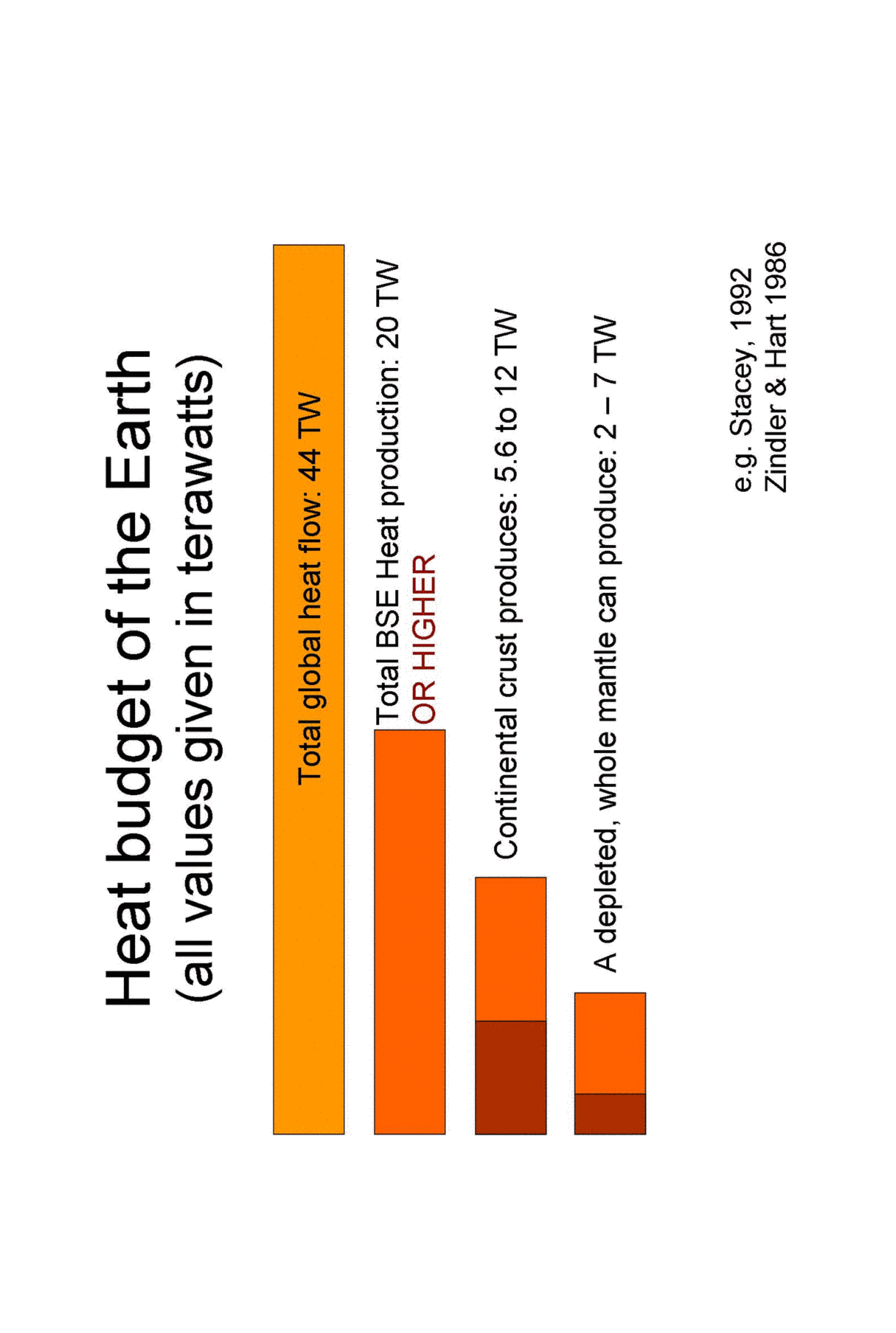
(all values given in terawatts)

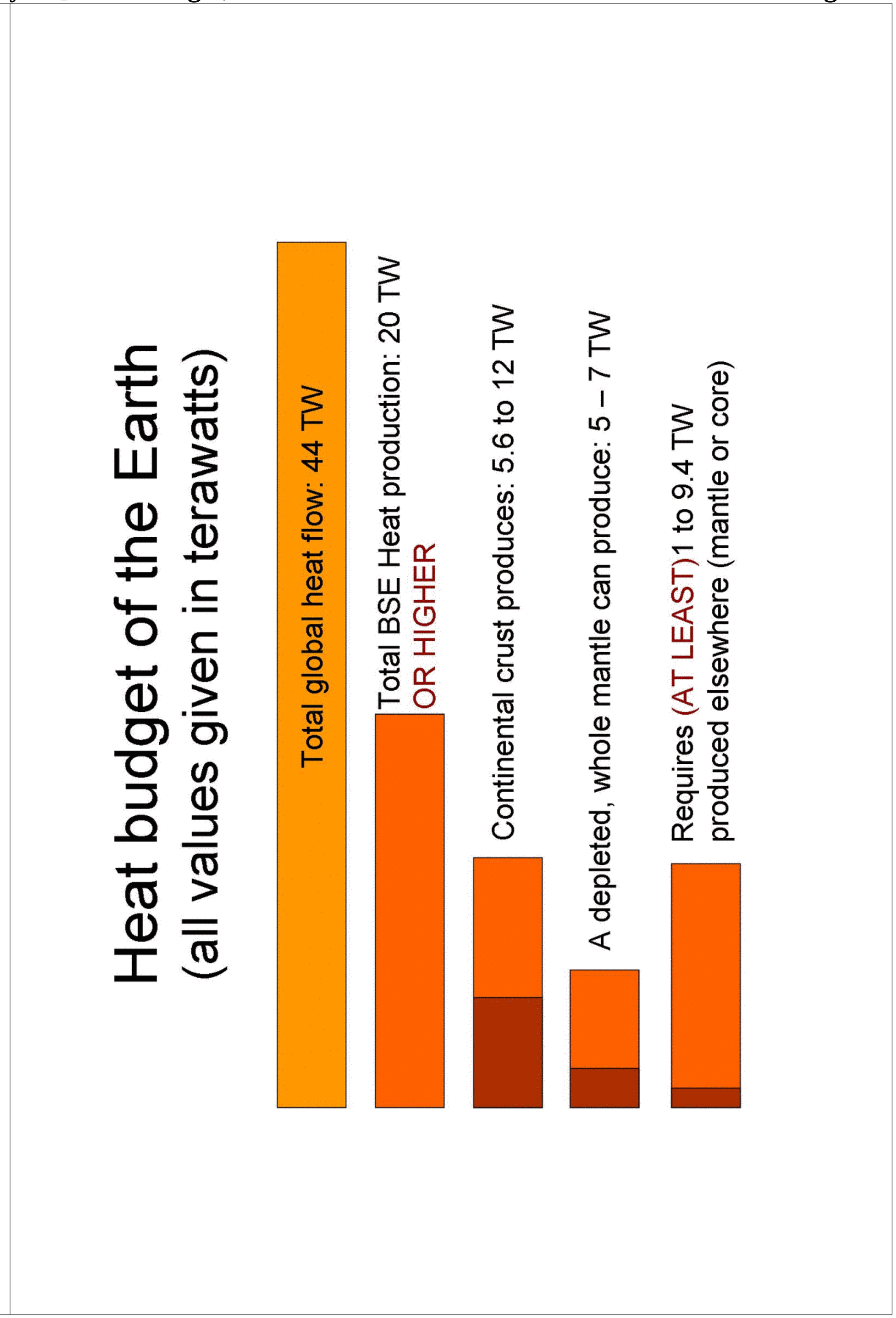
Total global heat flow: 44 TW

Stein, 1995


Heat budget of the Earth
(all values given in terawatts)
Total global heat flow: 44 TW
Total BSE Heat production: 20 TW
OR HIGHER
McDonough & Sun, 1995
Van Schmus 1995
Turcotte et al. 2001


Heat budget of the Earth
(all values given in terawatts)
Total global heat flow: 44 TW
Total BSE Heat production: 20 TW
OR HIGHER
Continental crust produces: 5.6 to 12 TW
Rudnick & Fountain, 1995
Taylor & McLennan, 1995
Gao et al., 1997
Wedepohl, 1995


Heat budget of the Earth
(all values given in terawatts)
Total global heat flow: 44 TW
Total BSE Heat production: 20 TW
OR HIGHER
Continental crust produces: 5.6 to 12 TW
A depleted, whole mantle can produce: 2 – 7 TW
e.g. Stacey, 1992
Zindler & Hart 1986


Heat budget of the Earth
(all values given in terawatts)
Total global heat flow: 44 TW
Total BSE Heat production: 20 TW
OR HIGHER
Continental crust produces: 5.6 to 12 TW
A depleted, whole mantle can produce: 5 – 7 TW
Requires (AT LEAST)1 to 9.4 TW
produced elsewhere (mantle or core)

Mantle viscosity is self-regulating

$$Mc\, dT/dt = MH - Aq$$

$$H = H_0 e^{-\lambda\tau}$$

$$\mu = \mu_0 e^{-\frac{E^* + Vz}{RT}}$$

- As the mantle heats up, the viscosity drops
- More rapid convection cools the mantle
- Viscosity increases in a cooler mantle

$$dT/dt = - Aq/Mc(1 - Ur)$$

Where Ur, the Urey ratio, is MH/Aq

See, for example, Tozer, Geophys. *J. R. Astron. Soc.*, 9, 95-112 (1965) and P*hil Trans. Roy. Soc. London A,* 2588, 252-271 (1965), Schubert et al. *JGR* 85 2511-2518 (1980), Stevenson, *Comptes Rendues Geosciences* **335** 99-111(2003)

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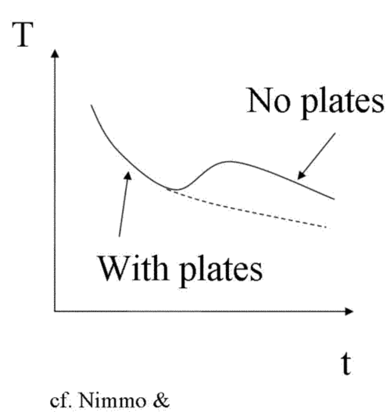
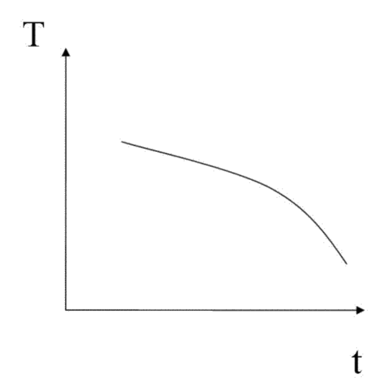
Changes in Convective Style (existence or extent of plate tectonics) can have big effects on Thermal History

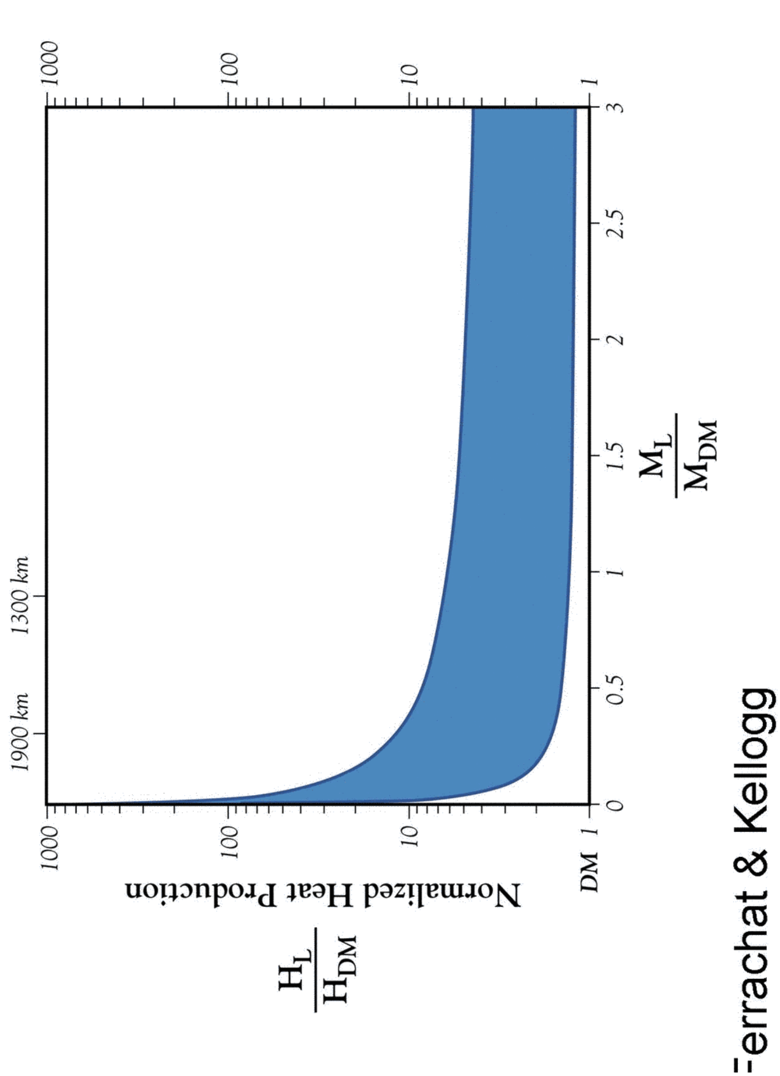


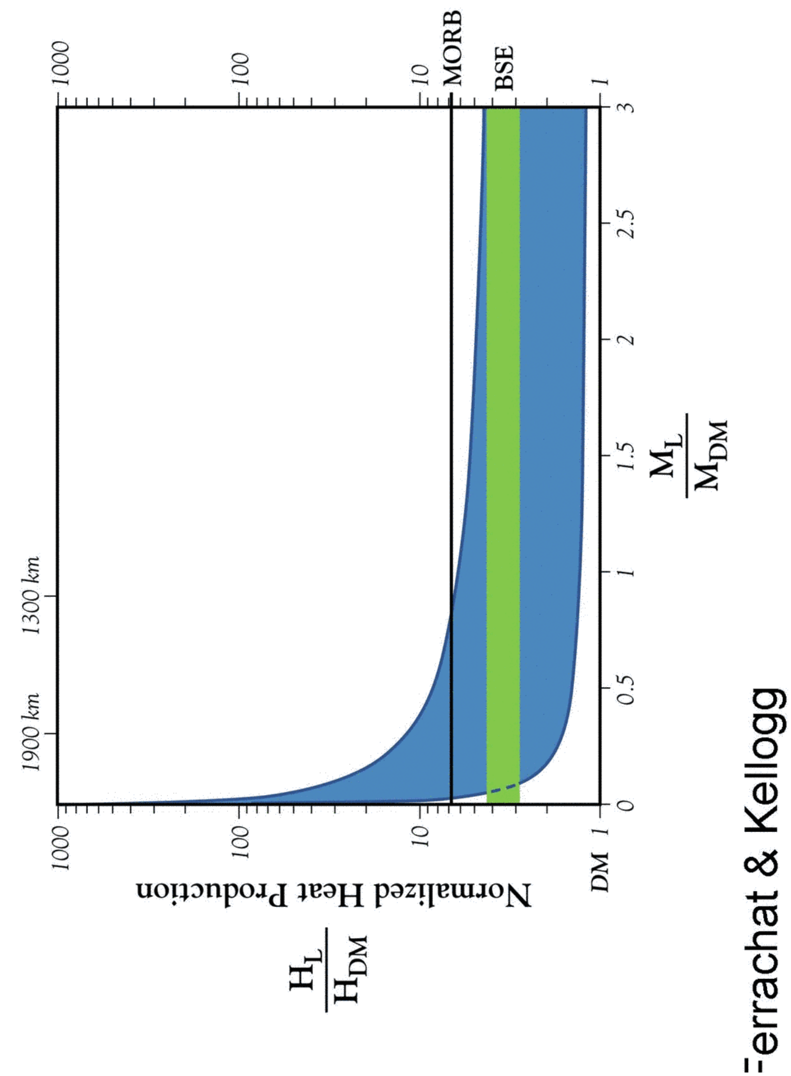
cf. Nimmo & Stevenson(2000) applied to Mars

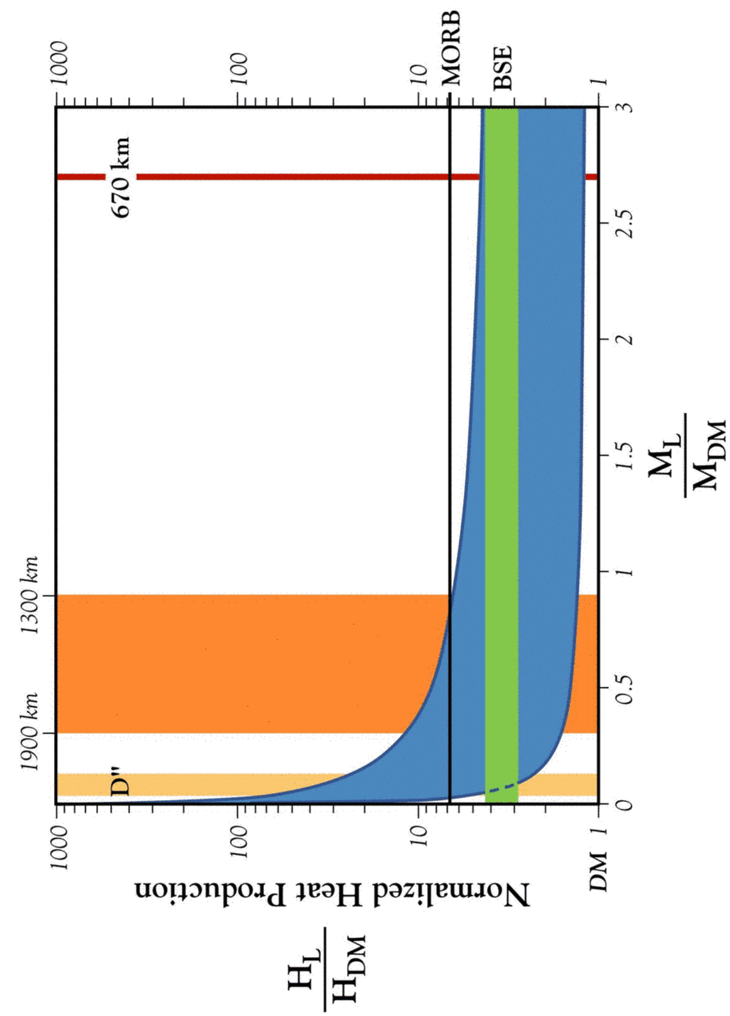


Notice the different <u>curvature.</u> cf. Korenaga, 2003. Also, Conrad & Hager, 1999, Solomatov (1995 onwards).

Images courtesy of D. Stevenson

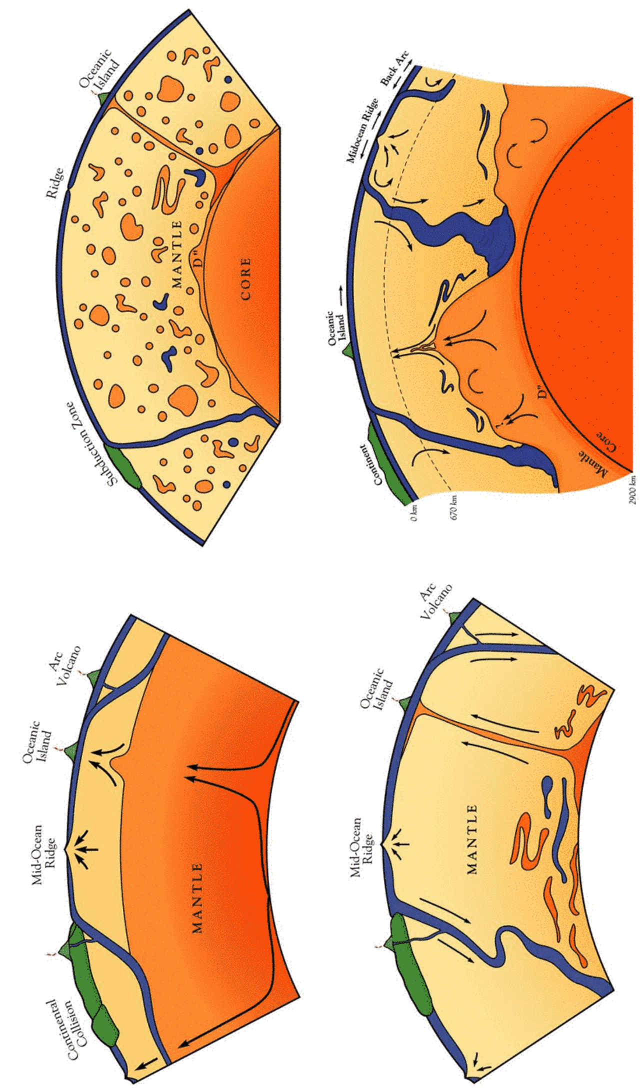

Normalized Heat Production
$\frac{H_L}{H_{DM}}$
$\frac{M_L}{M_{DM}}$
1900 km
1300 km
DM
Ferrachat & Kellogg


Normalized Heat Production
$\frac{H_L}{H_{DM}}$
$\frac{M_L}{M_{DM}}$
1900 km
1300 km
DM
MORB
BSE
Ferrachat & Kellogg



Ferrachat & Kellogg

A menagerie of models



Consider the Earth's mantle as a non-linear, chaotic system

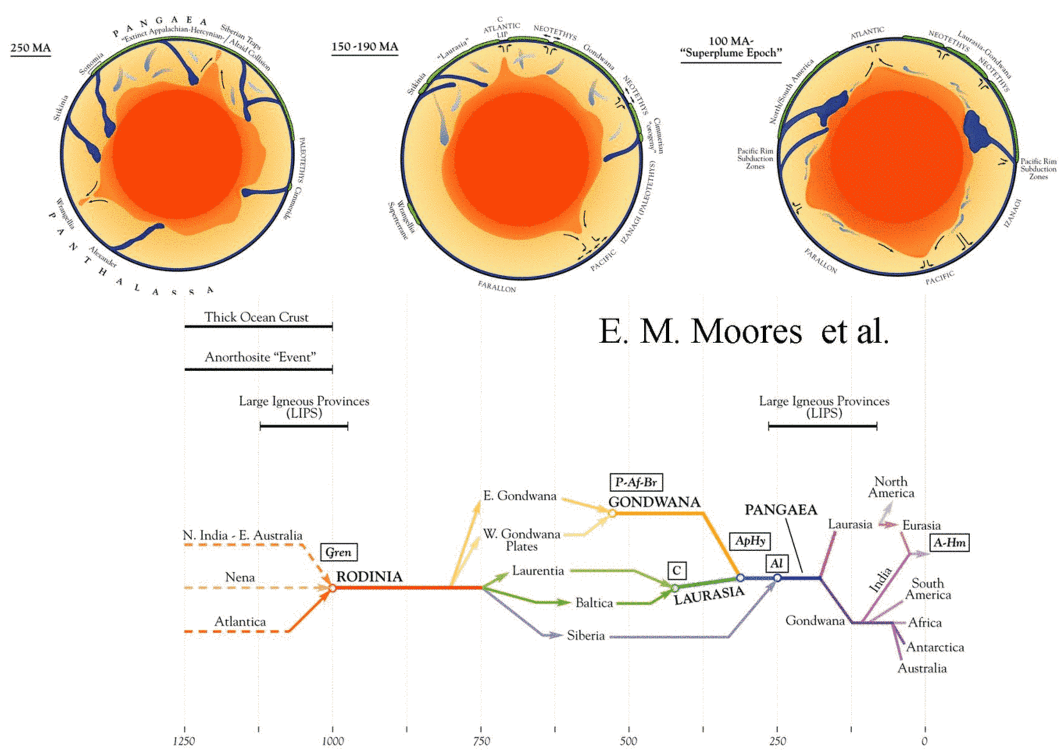
- Conservation of mass: $\nabla \cdot u = 0$
- Momentum: $\nabla^2 u - \nabla \cdot \tau + RaT\hat{k} = 0$ where $\tau = \mu\varepsilon$
- Heat flow: $\frac{\partial T}{\partial t} + u \cdot \nabla T = \nabla(\kappa \nabla T)$

❑ Rayleigh Number - $\mathrm{Ra} = \frac{\rho_m g \alpha \Delta T d^3}{\kappa_0 \mu_0} = 10^7$ - *Referenced to Earth's near surface values*

❑ Flow is driven by basal heating (T= 1) and surface cooling (T=0)

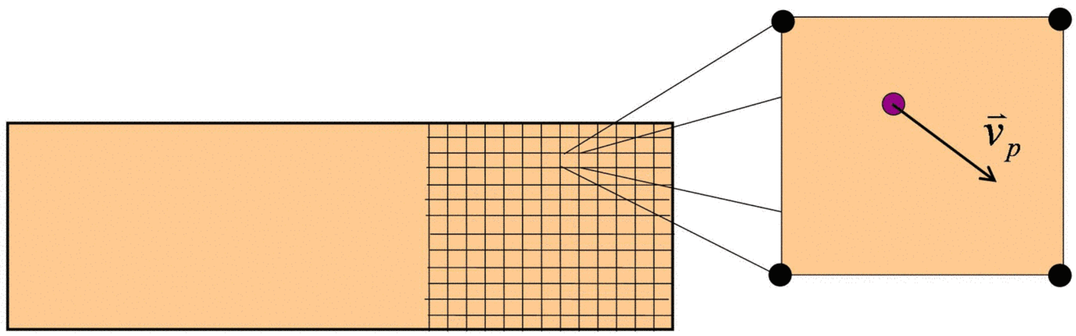
See for example Stewart & Turcotte JGR, 94, 13707-13717 (1989)

Thermal and dynamical evolution of the Earth?



E. M. Moores et al.

We use passive "tracers" to track trace elements (such as strontium isotopes)



Example of mixing in 2-D internally heated convection

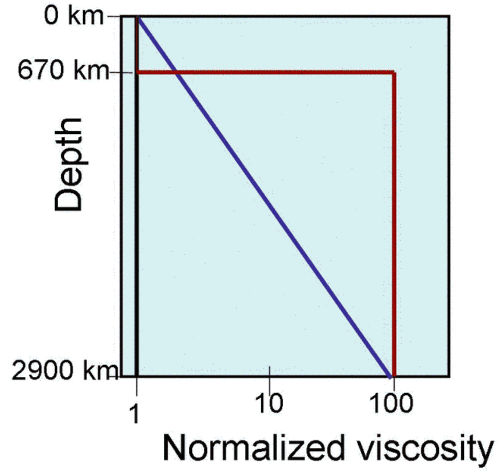
QuickTime™ and a Sorenson Video decompressor are needed to see this picture.

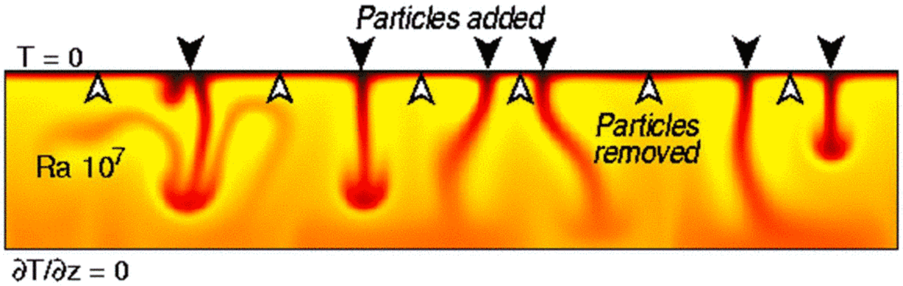
300,000 particles
Starting position

Numerical method: the finite element method

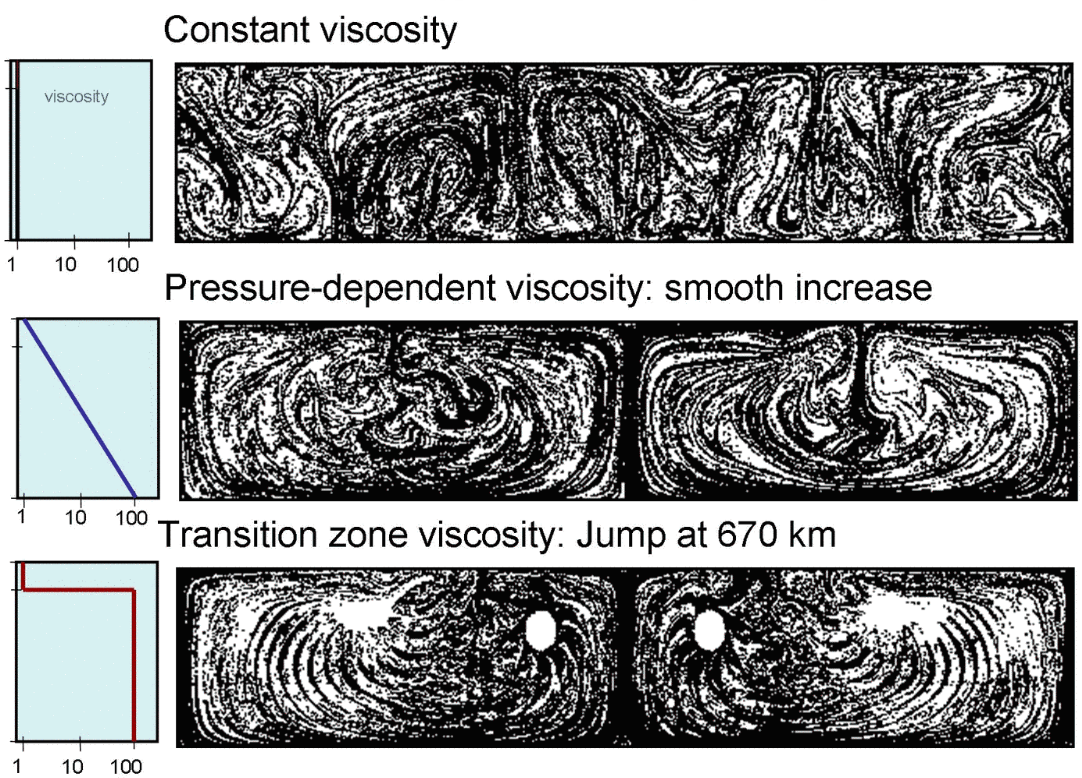
Hunt and Kellogg 2000

Mixing in 2-D with particles
- Added at subduction zones
- Removed at mid-ocean ridges

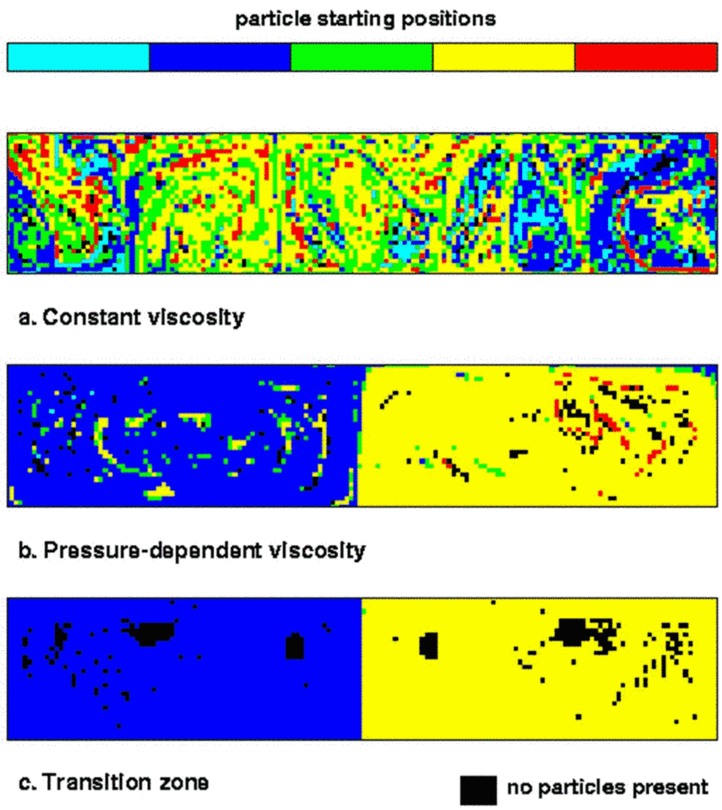




Hunt & Kellogg - effect of viscosity on mixing

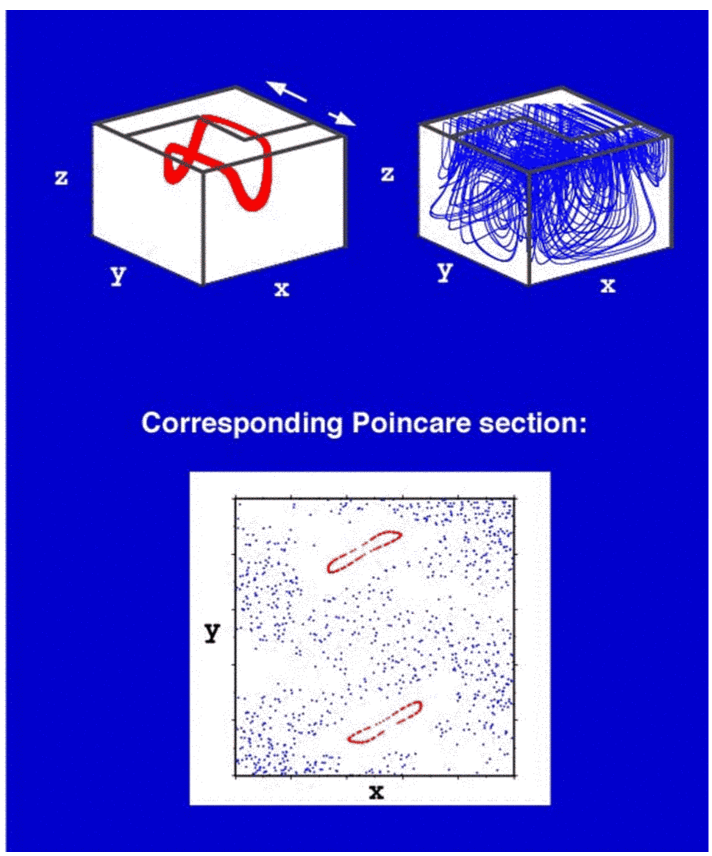


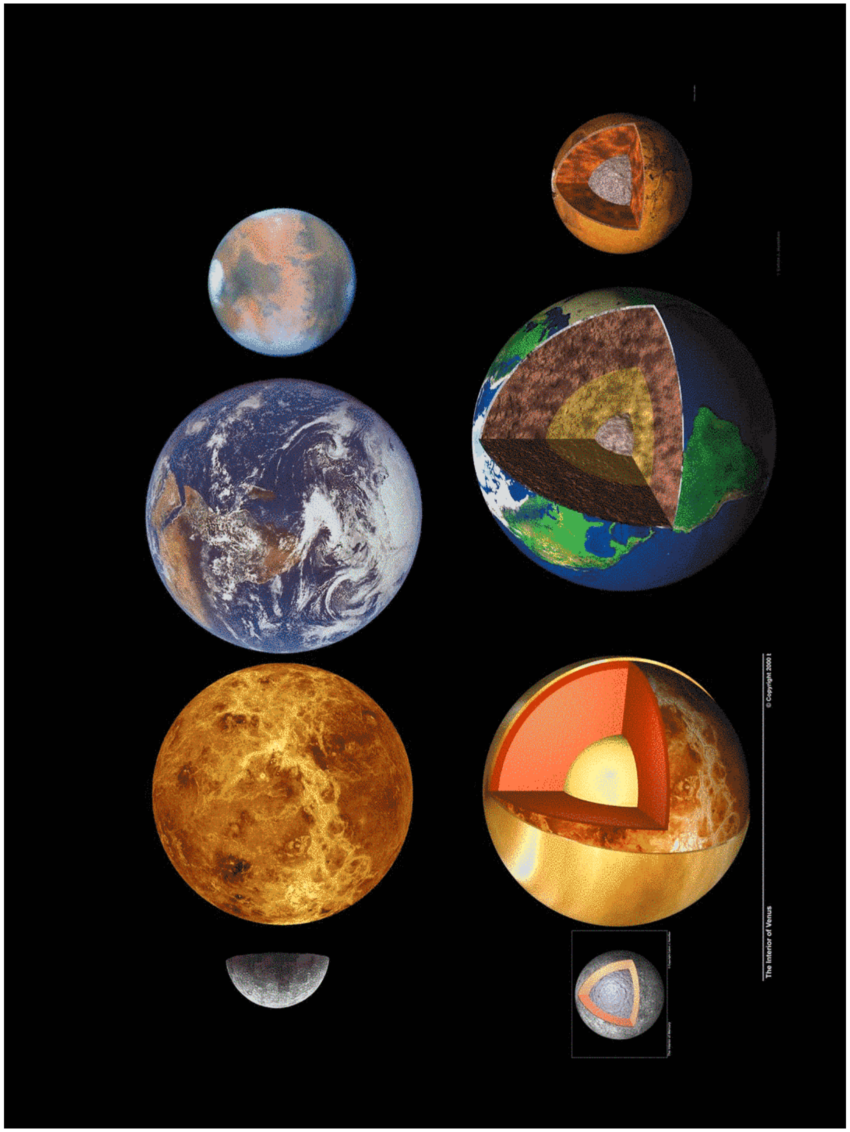
Initial location of particles (Hunt and Kellogg model)



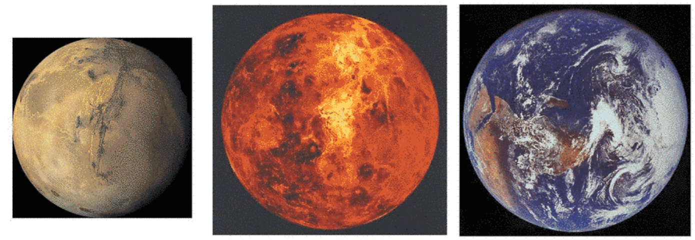
Ferrachat & Ricard, Mixing in 3-D plate driven flows

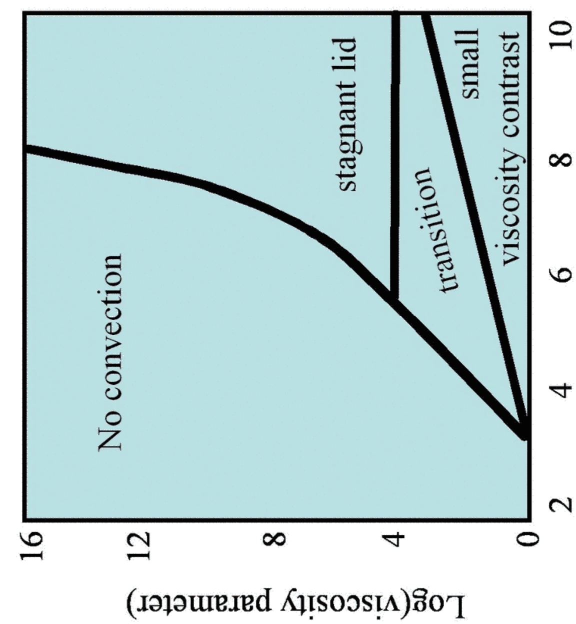
Chaotic trajectories occur even in steady-state flows



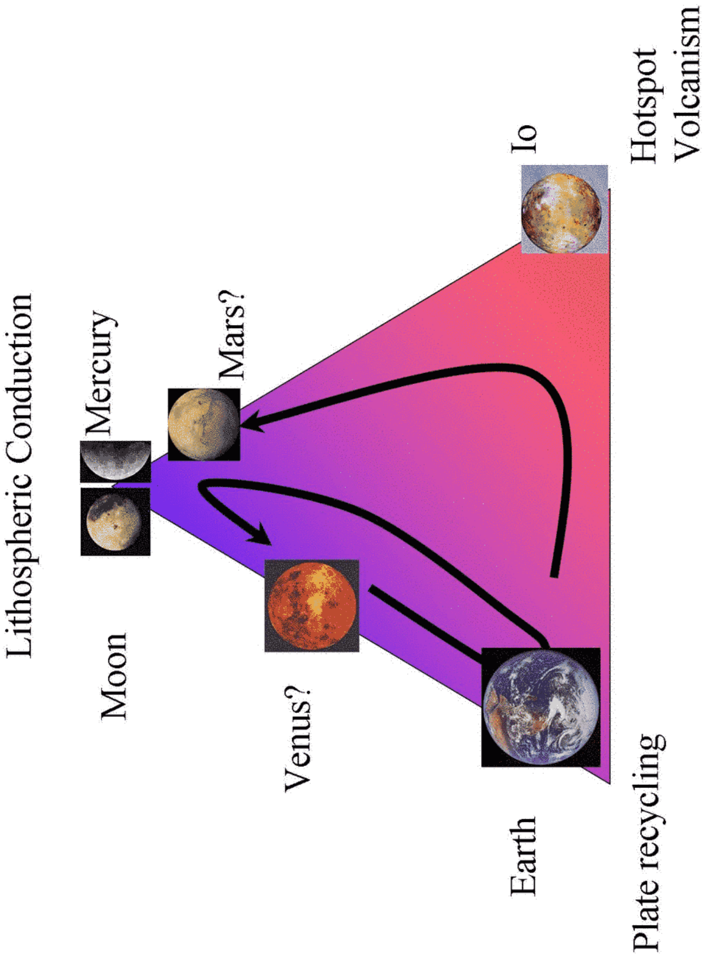


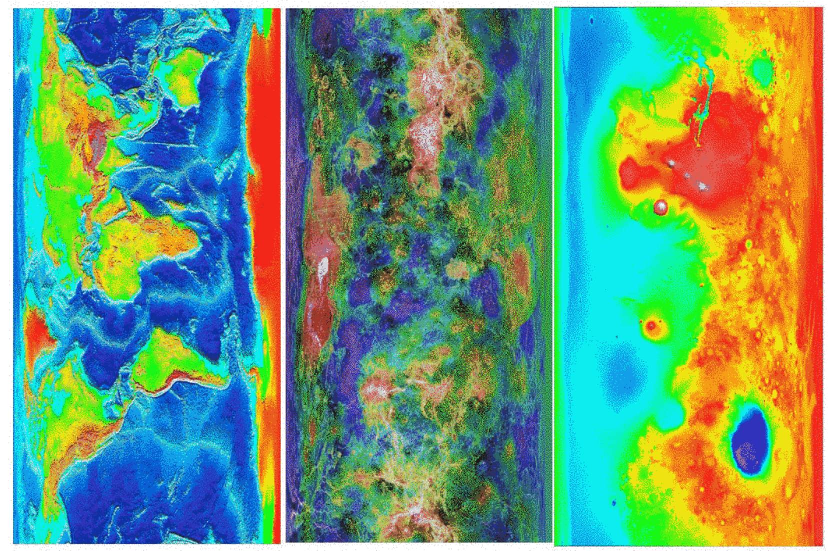
A planetary perspective on mantle flow

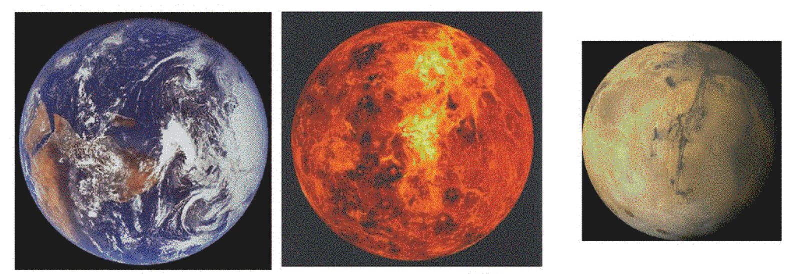


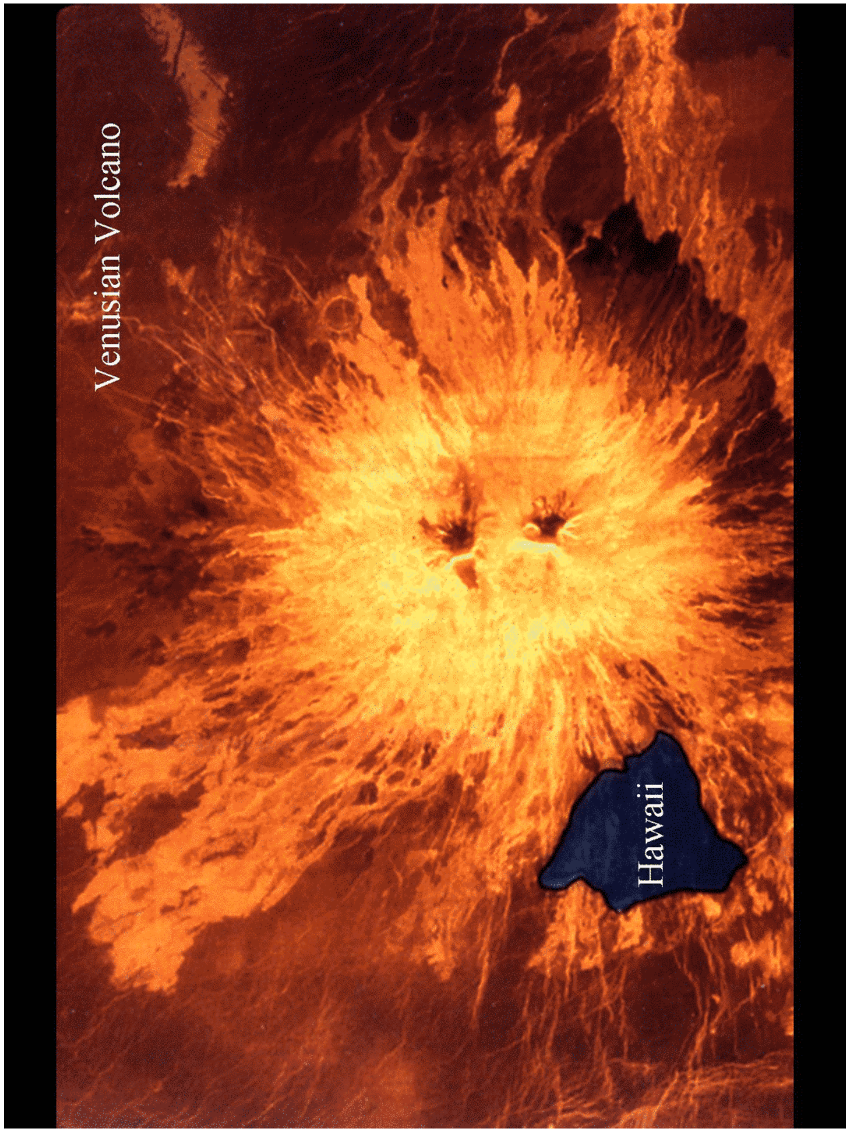


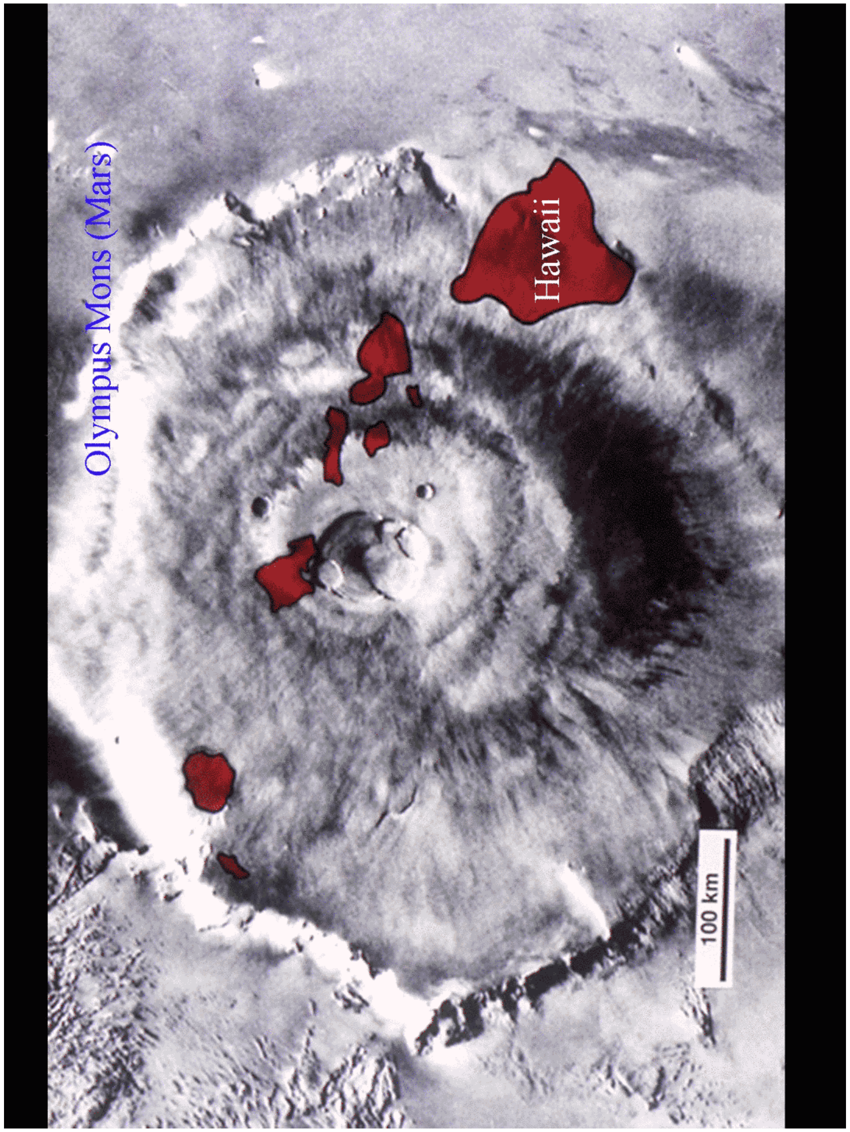
See, for example, work by Solomatov, Moresi, Lenardic, Stevenson, others


Lithospheric Conduction
Moon
Mercury
Mars?
Venus?
Io
Earth
Hotspot
Volcanism
Plate recycling






Venusian Volcano
Hawaii


Olympus Mons (Mars)
Hawaii
100 km