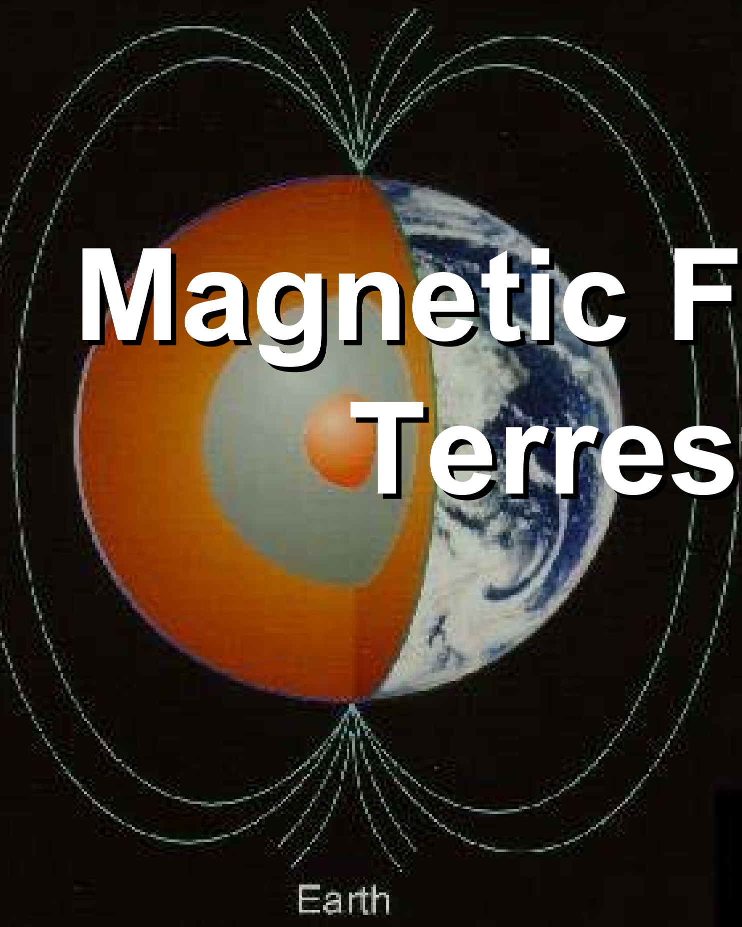
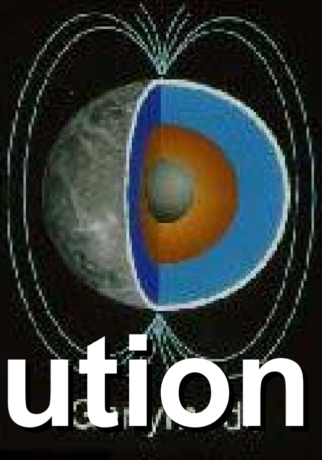
Magnetic Field Evolution of Terrestrial Planets





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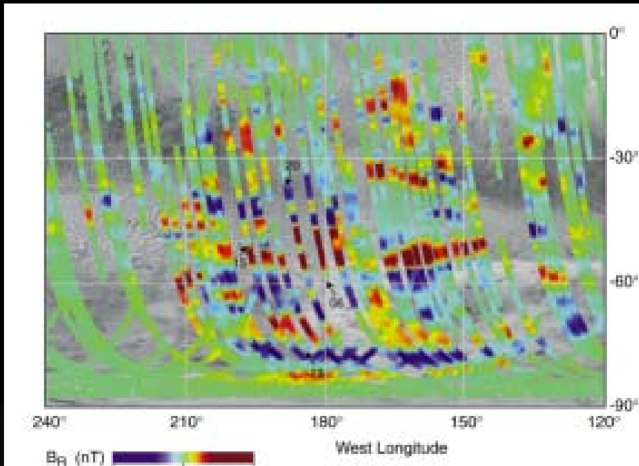


Content Part I

- Introduction
  - Motivation
  - Observations
- Mechanisms for magnetic field generation in the core
  - Thermal convection
  - Compositional convection

Content Part II

- Mantle ‘dictates‘ dynamo action in the core
  - Interior structure and composition
  - Heat sources
  - Heat transport mechansim
- Examples
  - Mars
  - Earth
  - Mercury
  - Galilean Satellites
- Conclusions

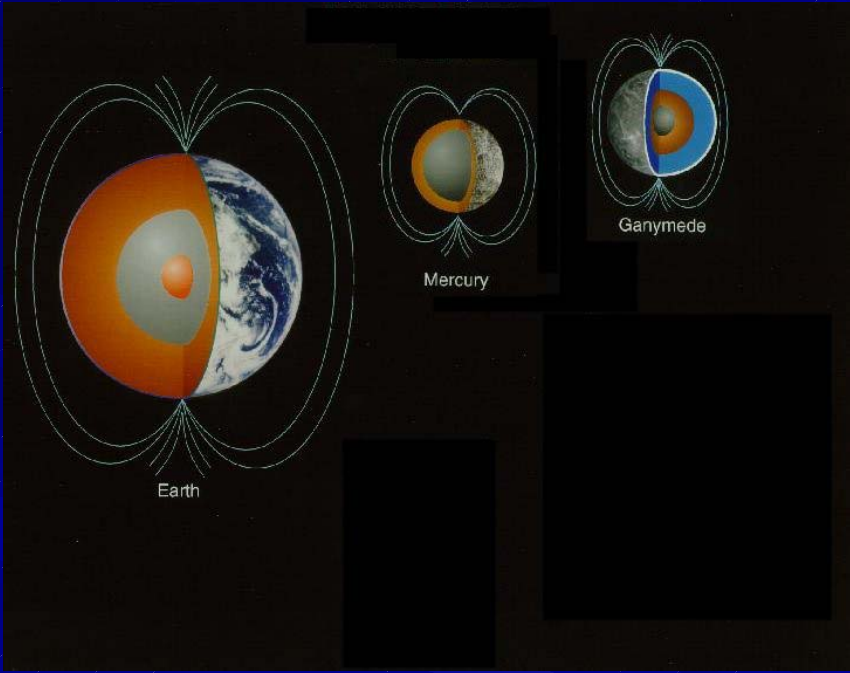
Why it is Interesting to Study the Magnetic Field

- General understanding of dynamo action
- Magnetic field evolution and magnetized crust helps to constrain
    - Thermal evolution
    - Interior structure
    - Geological and tectonic processes
    - Evolution of the atmosphere

Self-Generated Magnetic Field

- Of the terrestrial planets and major satellites, Earth, Mercury, and Ganymede are known to have self-generated magnetic fields

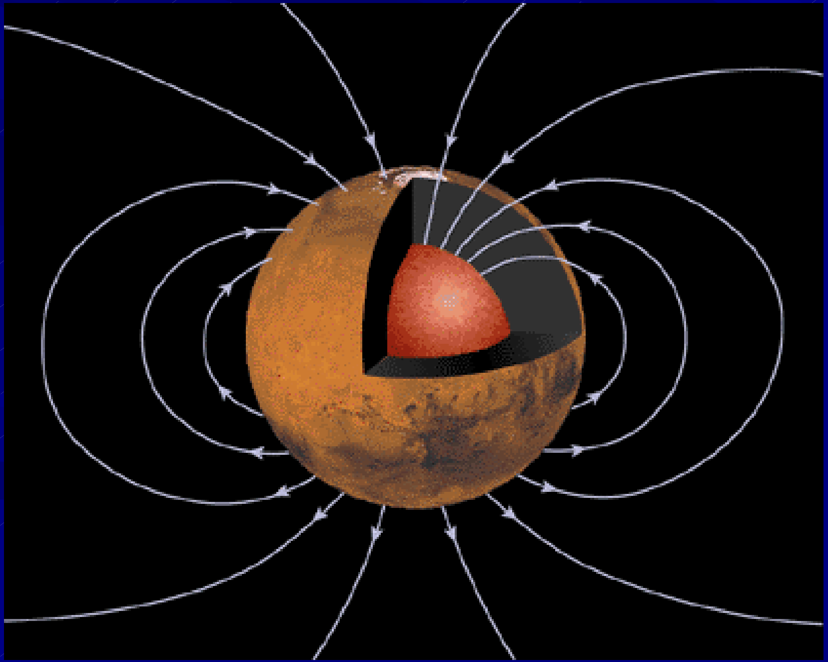
- Mars, Venus, Moon, Io, Europa, and Callisto lack self-generated magnetic fields



Planetary Data

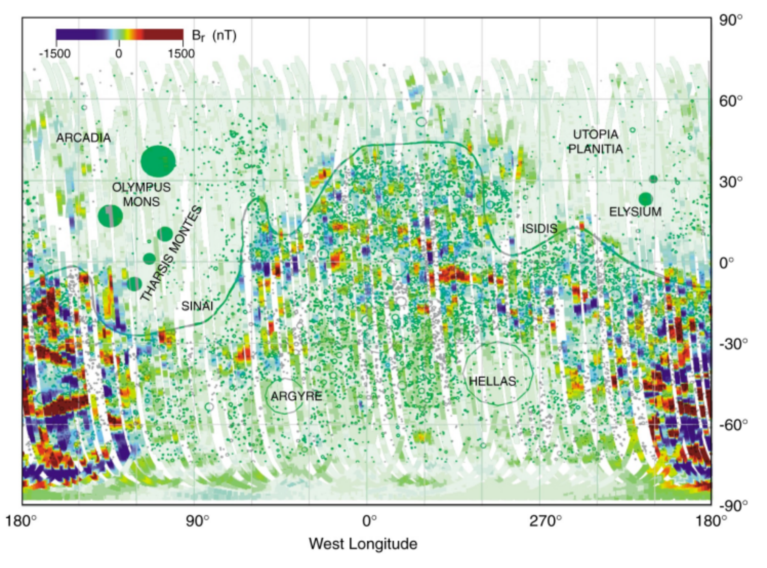
	Mercury	*Venus*	*Earth*	*Mars*	*Ganymede*
Radius	0.38	0.95	1.0	0.54	0.41
Mass	0.055	0.815	1.0	0.107	0.018
Density [kg/m^3]	5430.	5250.	5515.	3940.	1940.
ρ_0 ***[kg/m^3]***	5300.	4000.	4100.	3800.	1800.
Moment of Inertia factor	0.34	?	0.3355	0.3662	0.3105
R_c/R_p	0.8	0.55	0.546	0.5	0.3
Dipole Moment [10^{13} T m^3]	0.43	-	1577.	<0.08	1.4

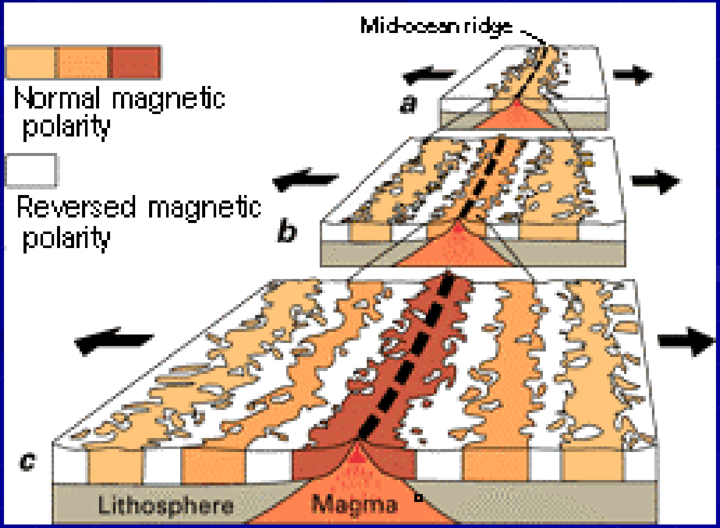
What About
Magnetic Field Generation in the Past?



Remanent Crust Magnetization

- Magnetized crust provides information about
    - History of the magnetic field
    - Geological and tectonic processes





Origin of Remanent Magnetization

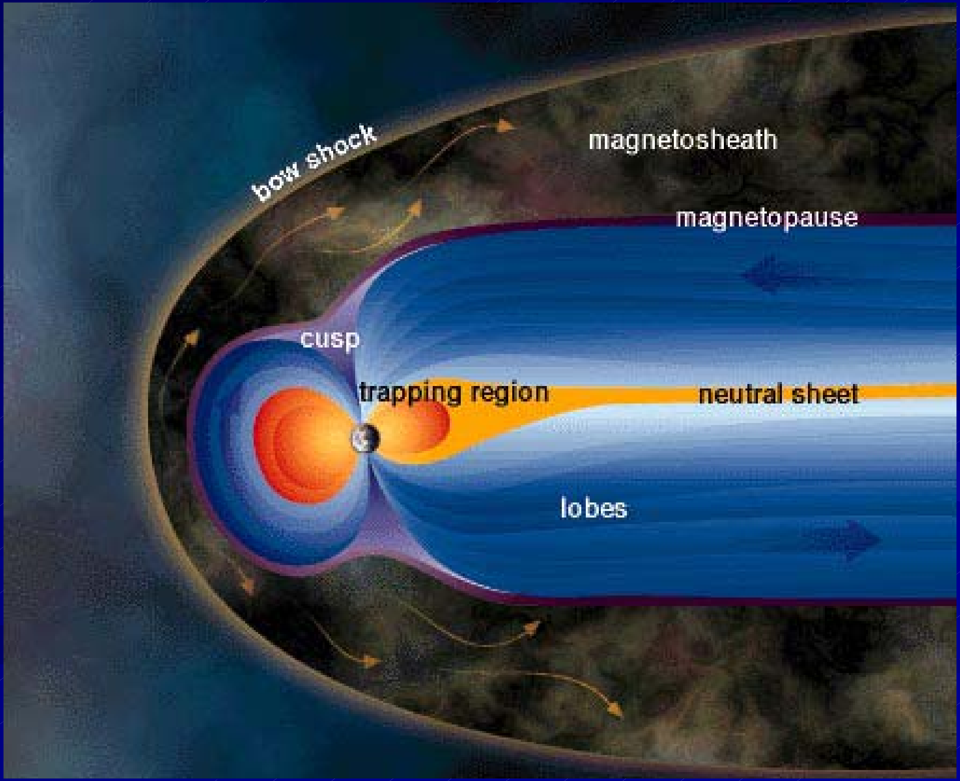
- Thermal remanent magnetization (TRM)
  - If a magnetic mineral is cooled in an ambient magnetic field through a temperature characteristic of the material, **Curie temperature**, it will begin to acquire a very large remanent magnetization. As the material cools through the Curie temperature, domains begin to form, in alignment with the ambient field. The magnetic field is then frozen into the rock and is extremely stable.

Magnetite	~ 853 K
Hematite	~ 953 K
Iron	~ 1043 K

Remanent Crust Magnetization Constraint on Early Dynamo Action

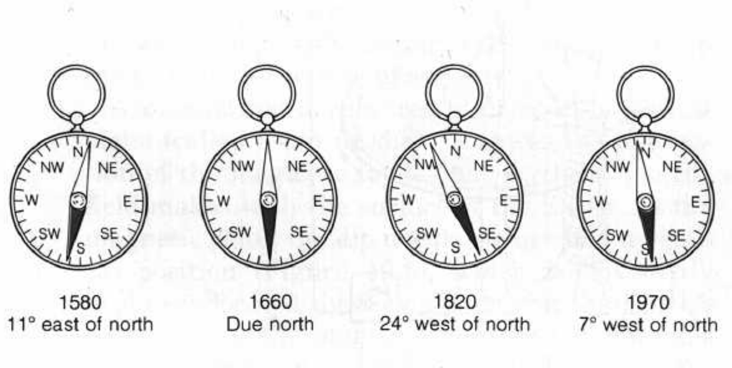
- Earth
  - Age of magnetized crust between 3.5 Ga and today
- Moon
  - Age of magnetized crust between 3.9 and 3 Ga
- Mars
  - Age of magnetized crust between 4.5 and 4 Ga?
- Venus, Mercury
  - No data available

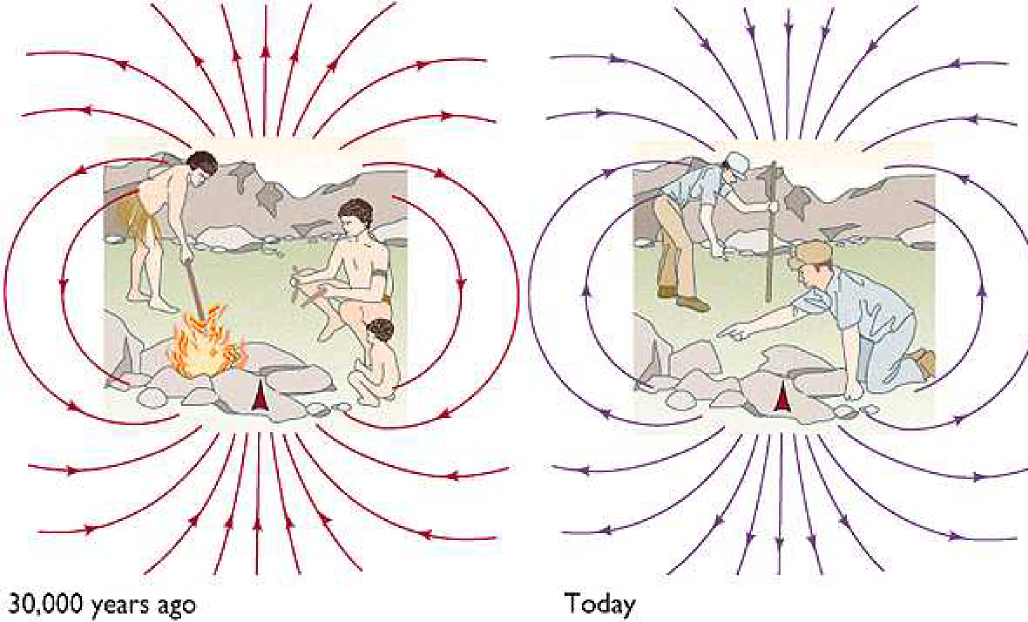
What do we Know About the Origin of the Magnetic Field?



‘Bar Magnet‘ in the Planets?

- No! Variations with time (e.g. polar wander, reversals) can be observed
- No! Magnetic material in the interior is well above the Curie temperature





Decaying Old Magnetic Field?

- Equation of magnetic induction

$$\frac{\partial \mathbf{B}}{\partial \mathrm{t}} = \nabla \times \left(\mathbf{u} \times \mathbf{B}\right) + D_{ma} \nabla^2 \mathbf{B}$$

u velocity field

B magnetic field

t time

Magnetic diffusion coefficient

$$D_{ma} = \frac{1}{\mu \sigma_c}$$

σ_c electrical conductivity

μ magnetic permeabilität

Decaying Old Magnetic Field?

In the case of **u** = 0

Characteristic diffusion time:

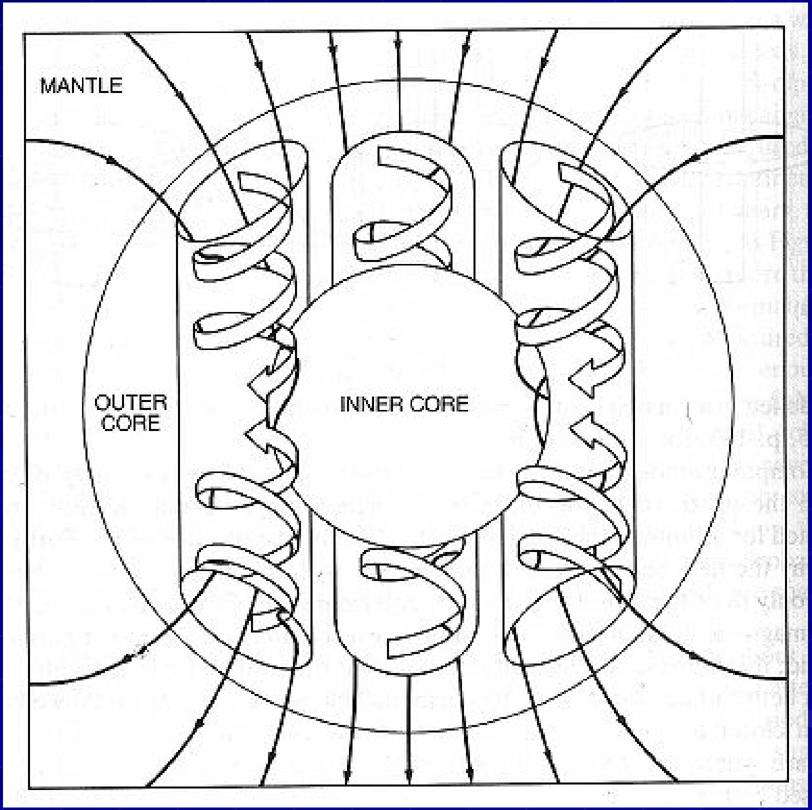
$$\tau_{diff} = \frac{L^2}{D_{ma}} \approx 10^4 \text{ a}$$

L characteristic length scale (planetary radius)

Fast decay of magnetic field is inconsistent with observations

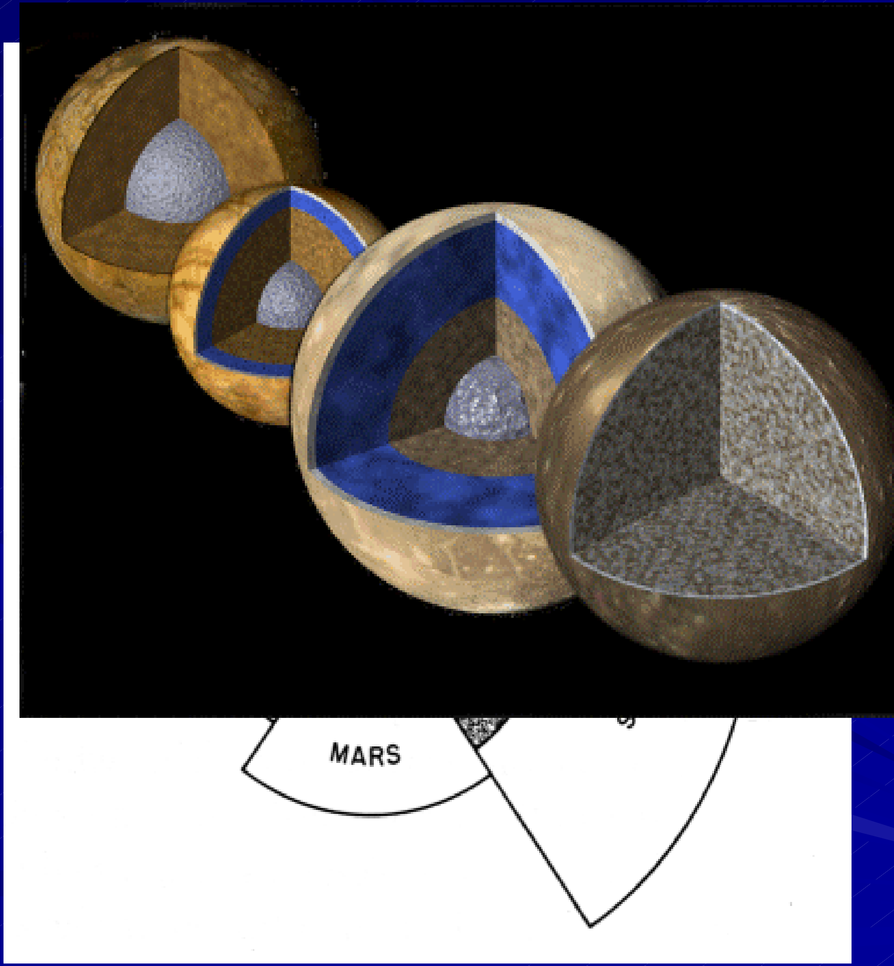
Magnetic Field Generation

- Necessary condictions for existence
  - A conducting fluid
  - Motion in that fluid
  - Cowling's Theorem requires some helicity in the fluid motion



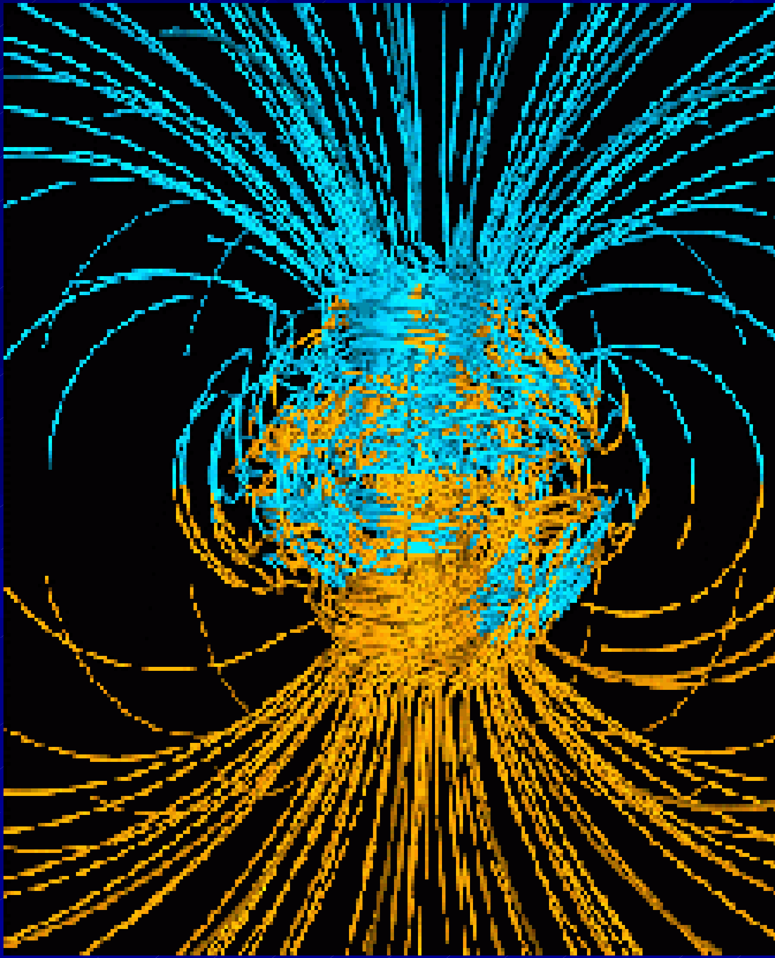
Cores

- The magnetic fields of terrestrial planets and satellites are produced in their cores
- There is little doubt that the planets and most of the major satellites have iron-rich cores



Dynamos

- Hydromagnetic dynamos
  - Thermal dynamos
  - Chemical dynamos

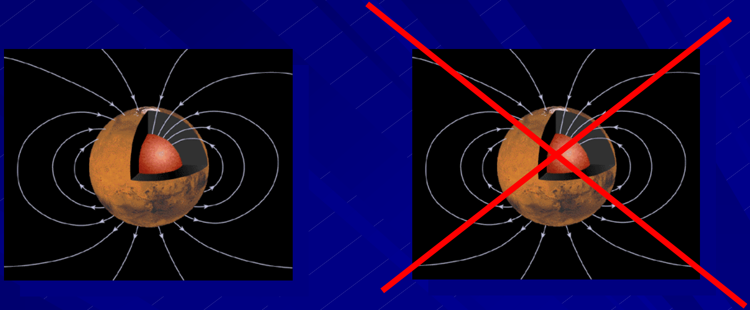


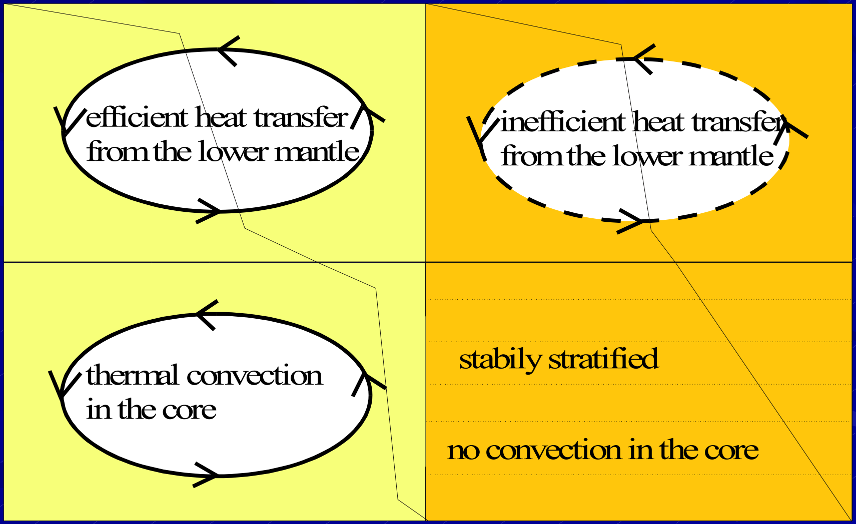
G. Glatzmeier's Dynamo model for Earth

Thermal Dynamo

Fluid motion in the liquid iron core due to thermal buoyancy
(=> cooling from above)

‘Critical‘ heat flow out of the core





efficient heat transfer from the lower mantle

inefficient heat transfer from the lower mantle

thermal convection in the core

stabily stratified

no convection in the core

‘Critical‘ Heat Flow: Heat Flow Along the Core Adiabate

$$q_c = k_c \left(\frac{dT}{dz} \right)_{ad} = k_c \frac{\alpha_c T g}{C_p}$$

k_c	thermal conductivity of the core
T	temperature
z	depth
α_c	thermal expansivity
g	acceleration due to gravity
C_p	specific heat capacity

‘Critical‘ Heat Flow

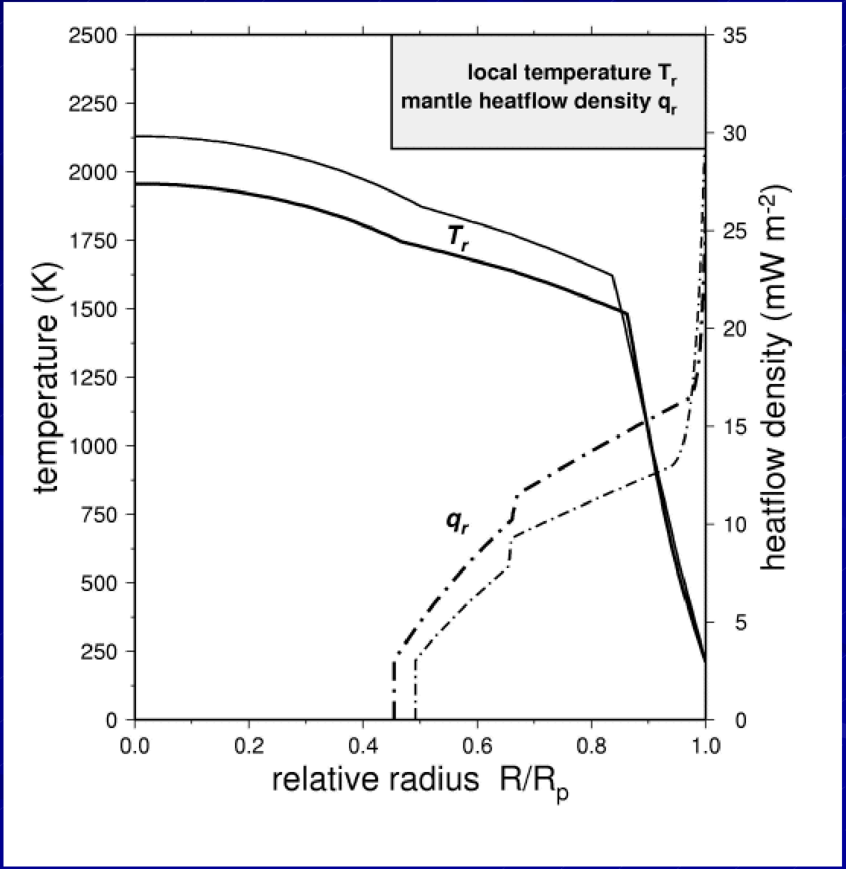
Mars, Mercury	5 - 20 mW/m^2
Earth, Venus	15 – 40 mW/m^2
Galilean Satellites, Moon	< 7 mW/m^2

Large uncertainities due to poorly known parameters

Vigour of Core Convection

- A sufficiently large ΔT between the core and the mantle is required in order to drive thermal convection in the core

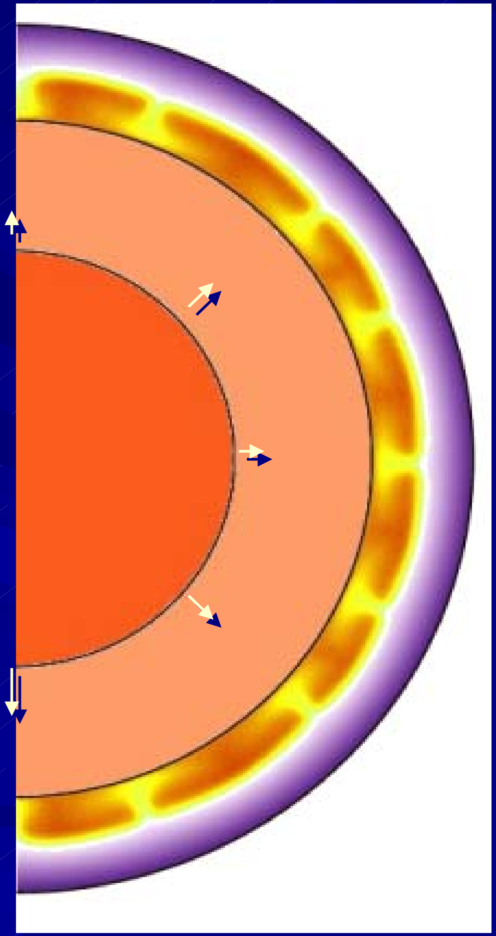
- If ΔT is too small than the core will be cooling by conduction



Sohl and Spohn, 97

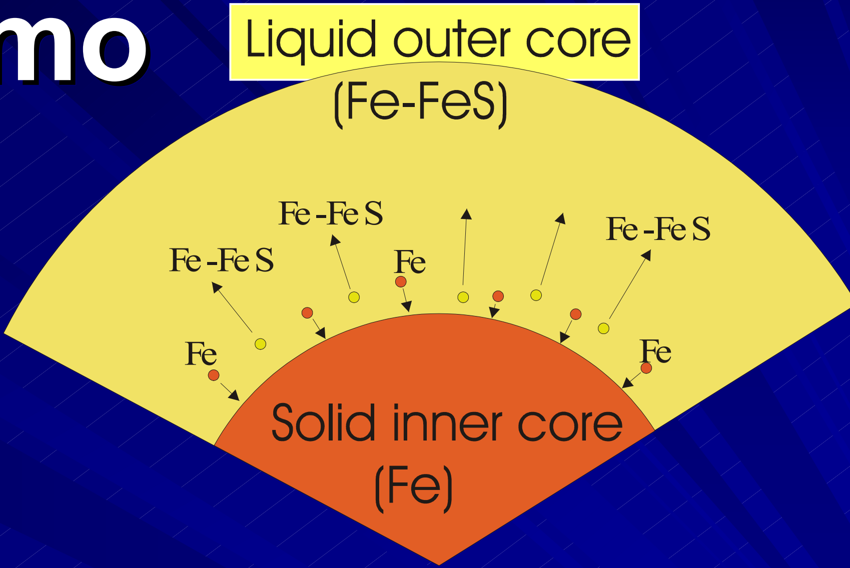
Chemical Dynamo

- Compositional bouyancy released by inner core growth
- Difficult to stop operating

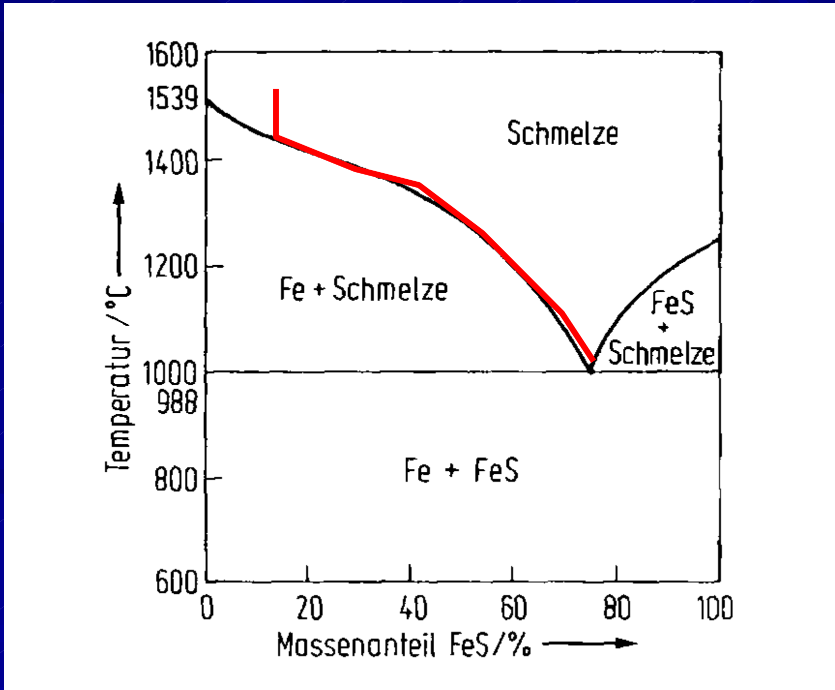


Mercury Model by Conzelmann and Spohn

Chemical Dynamo



- Existence of light alloying elements in the core like S, O, Si
- Core temperature between solidus and liquidus



Core Composition: Earth

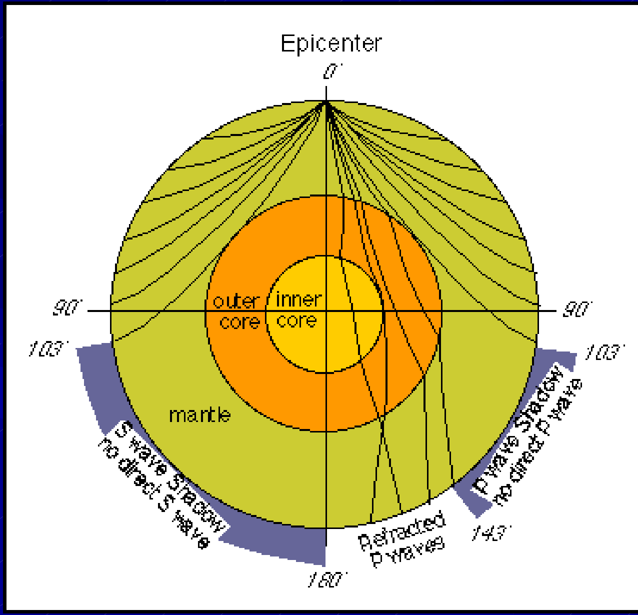
- Indirect information from seismology
  - Seimic velocities of the core constrain density and therefore composition
  - Inner solid core

    2-3% less dense than pure iron

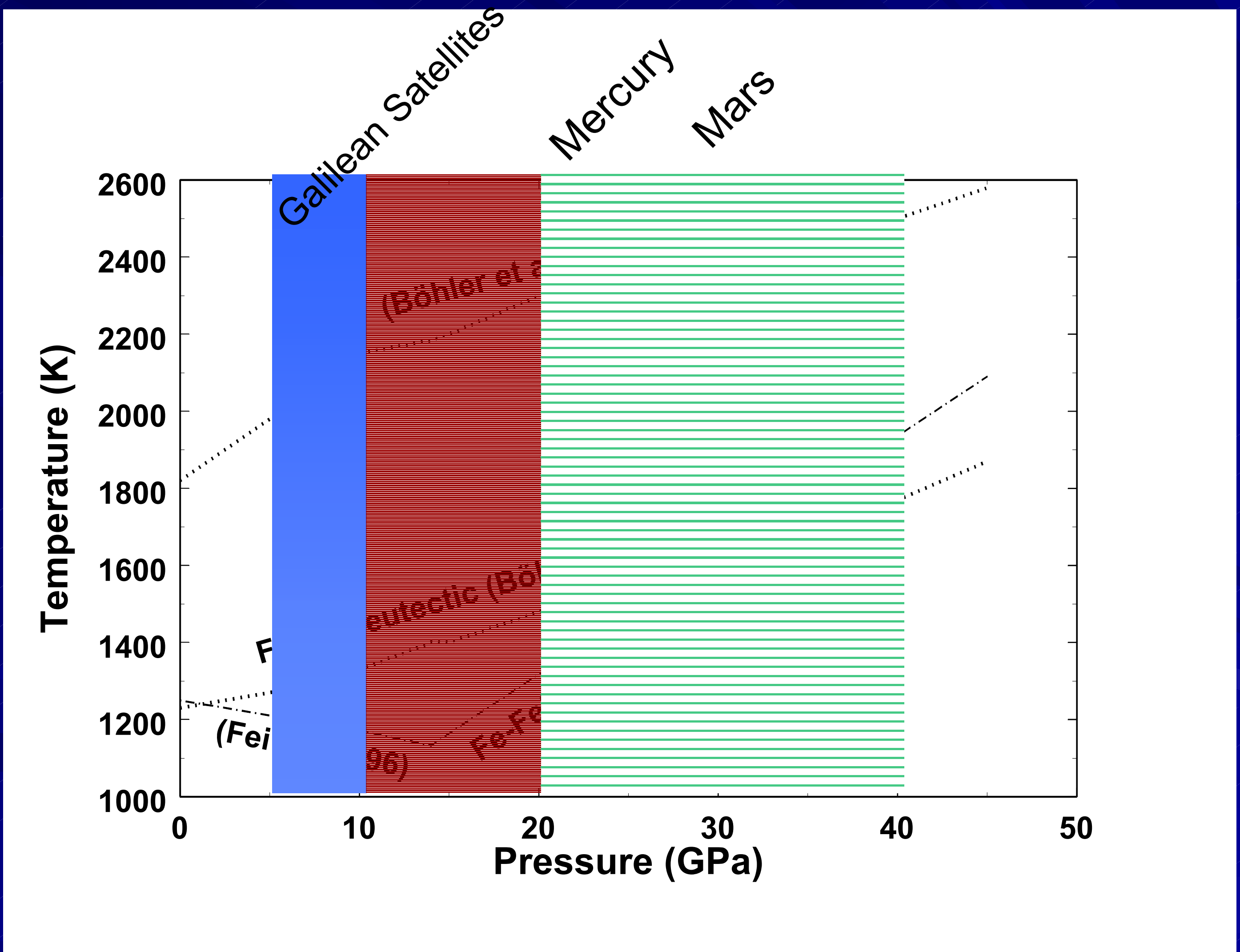
    ~ 8% S / Si ?
  - Outer fluid core

    5-10 % less dense than pure iron

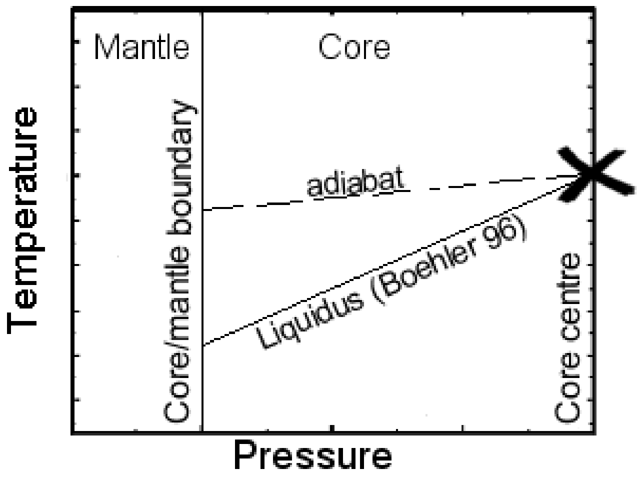
    ~ 8% S / Si and 8% O ?

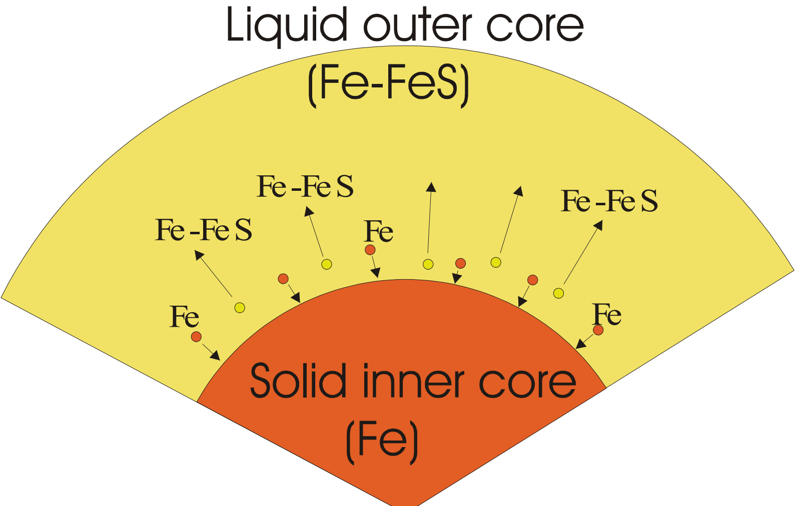


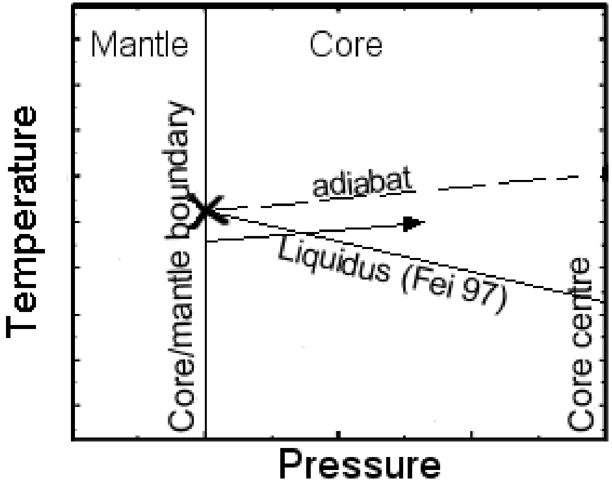
Melting Temperature as Function of Pressure

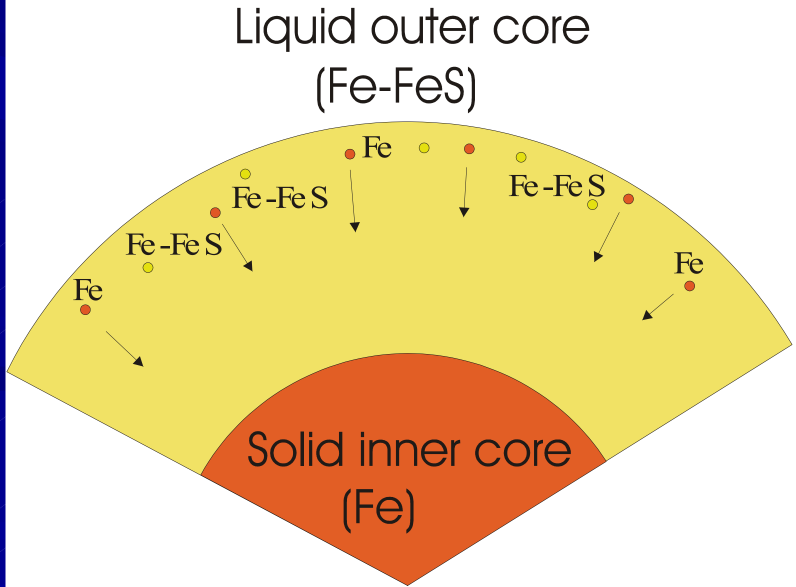


Some Special Cases

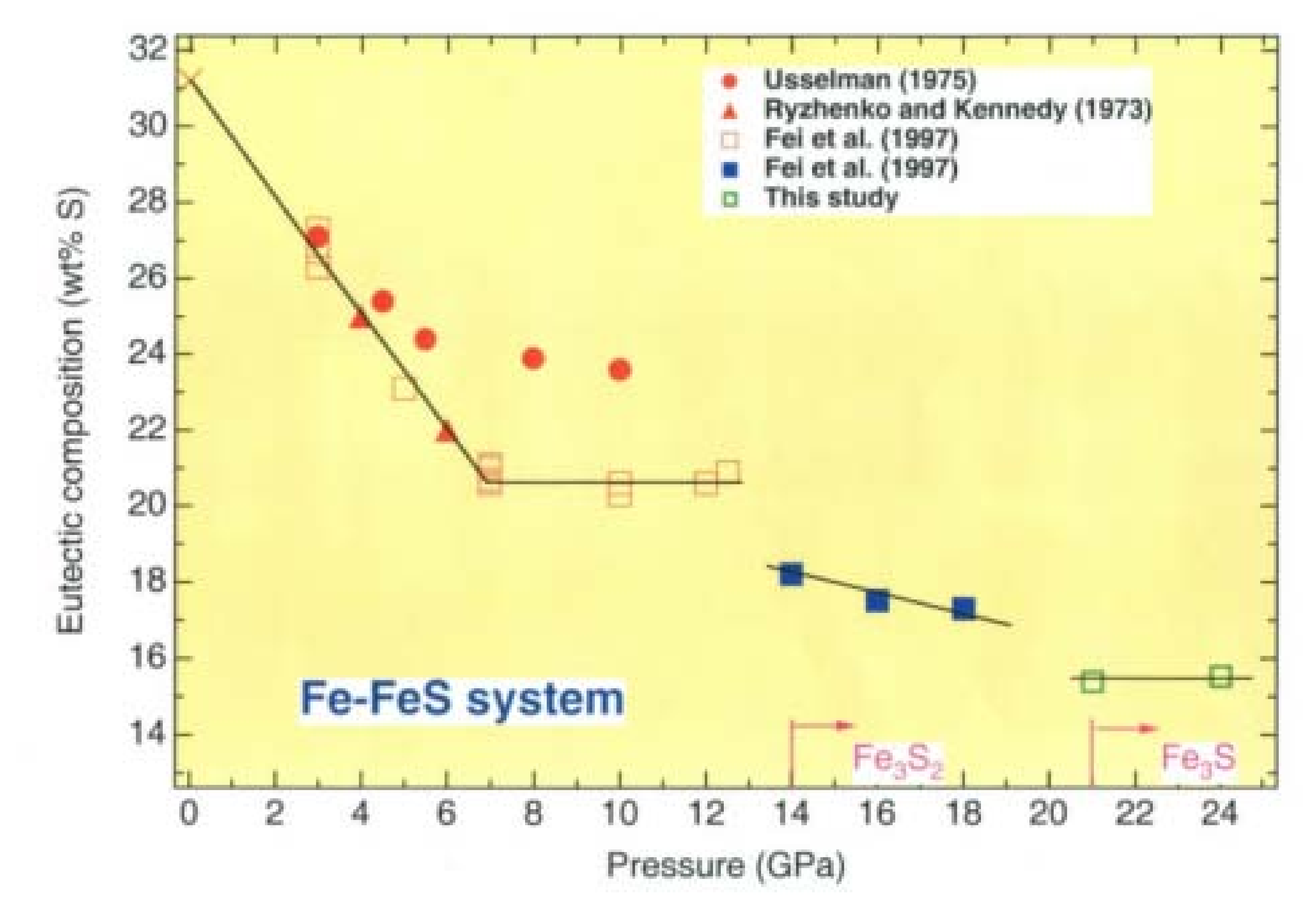








Eutectic Composition



Fei, unpublished

Power Requirements

- Dynamo converts thermal and gravitational energy into magnetic energy
- Power needed to sustain geomagnetic field is set by the ohmic losses (dissipation due to electrical resistance)
- Estimates of ohmic loss for the Earth cover a wide range (0.1 to 3.5 TW)

Ohmic Dissipation

$$\Phi_{diss} = \chi_g E_g \frac{dm_i}{dt} + \chi_t \left(E_L \frac{dm_i}{dt} + \frac{dE_{th}}{dt} - A_c q_c \right)$$

E_g gravitational energy

$\mathrm{E_L}$ latent heat of soldification

$\frac{\mathrm{dm_i}}{\mathrm{dt}}$ rate of inner core growth

$\frac{\mathrm{dE_{th}}}{\mathrm{dt}}$ rate of change of heat content of the core

$\mathrm{A_c} q_c$ heat conducted along adiabat

χ Carnot efficiency factor

Dynamo Efficiency

- Thermal dynamo efficiency is restricted by Carnot efficiency χ_t to be a few percent
- Chemical dynamo is not restricted; $\chi_g = 1$

Some First Conclusions

- Thermal convection (fluid core without inner core growth): Inefficient of dynamo generation

- Compositional convection (inner core growth): Efficient of dynamo generation; difficult to stop

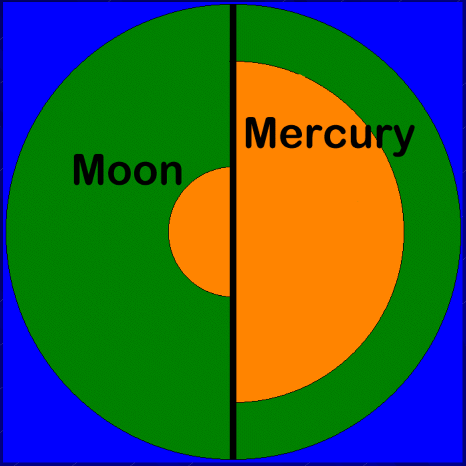
- The mantle determines whether a terrestrial planet has core convection and whether it can have a dynamo

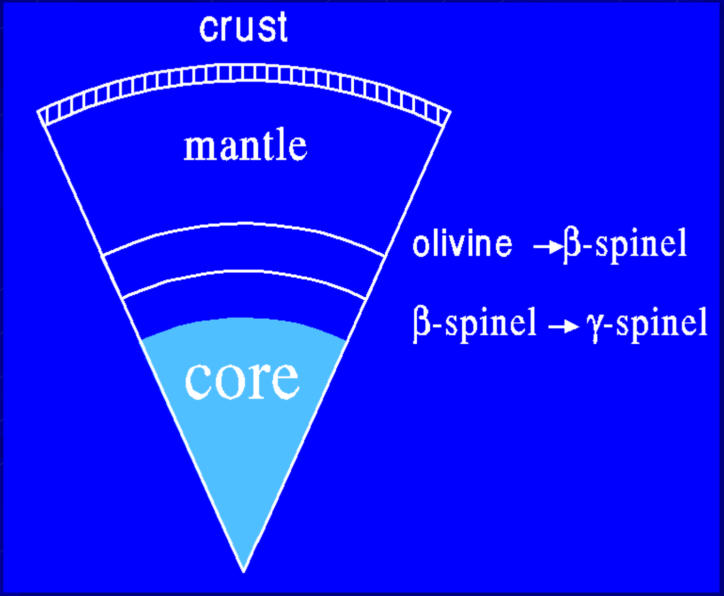
Influence on Thermal Evolution (and Magnetic Field Evolution)

- Interior structure and composition
- Heat sources
- Heat transport mechanisms

Interior Structure and Composition

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





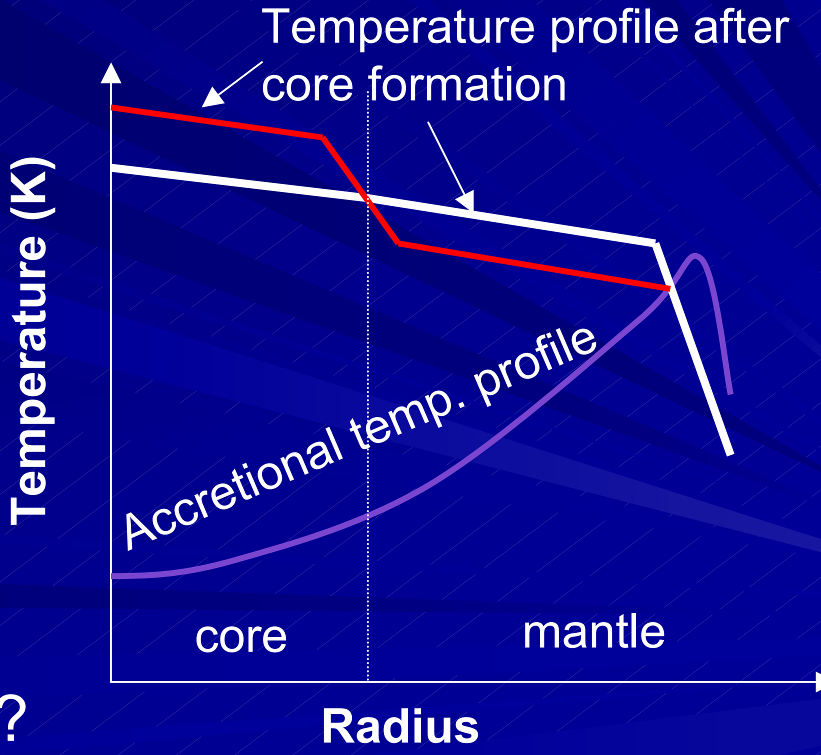
Heat Sources

- Primary energy
  - Accretion
  - Gravitational energy due to core formation

- Decay of radioactive elements
  - Uranium
  - Thorium
  - Potassium

Accretion and Core Formation

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

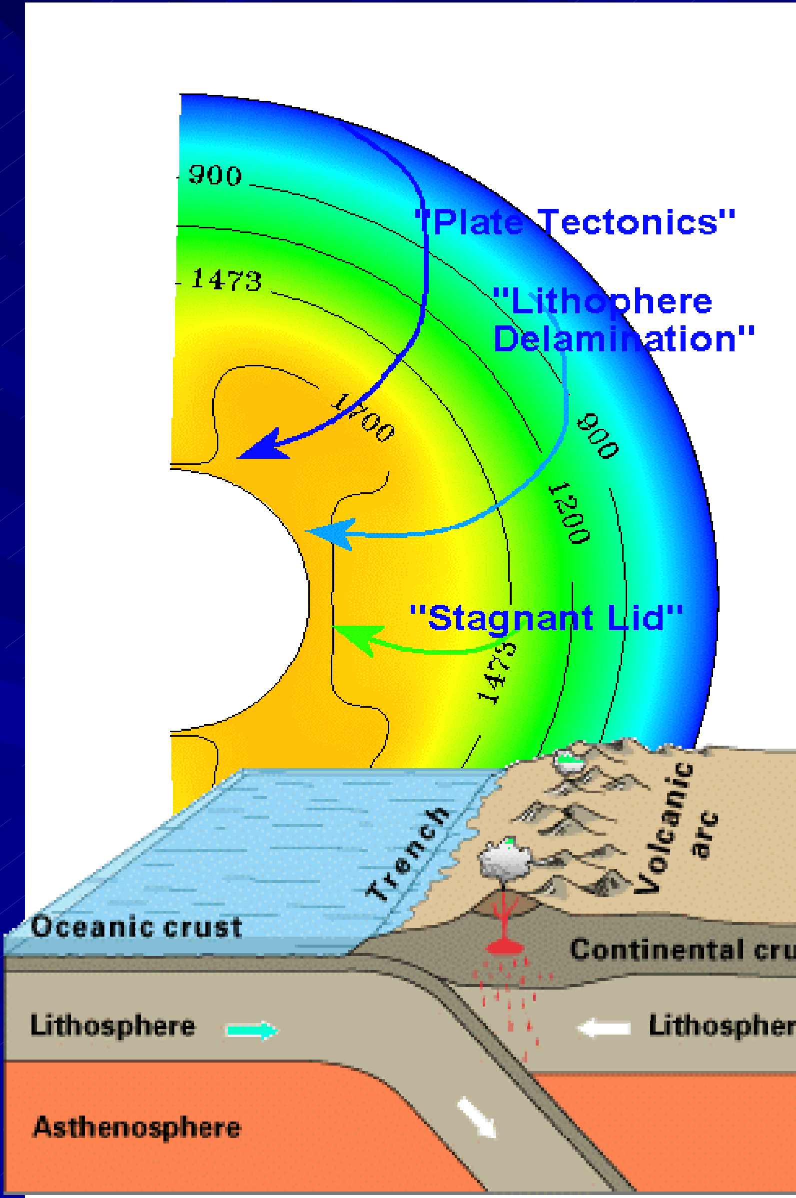


What are the initial thermal conditions after core formation?

Heat Transport Mechanisms

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

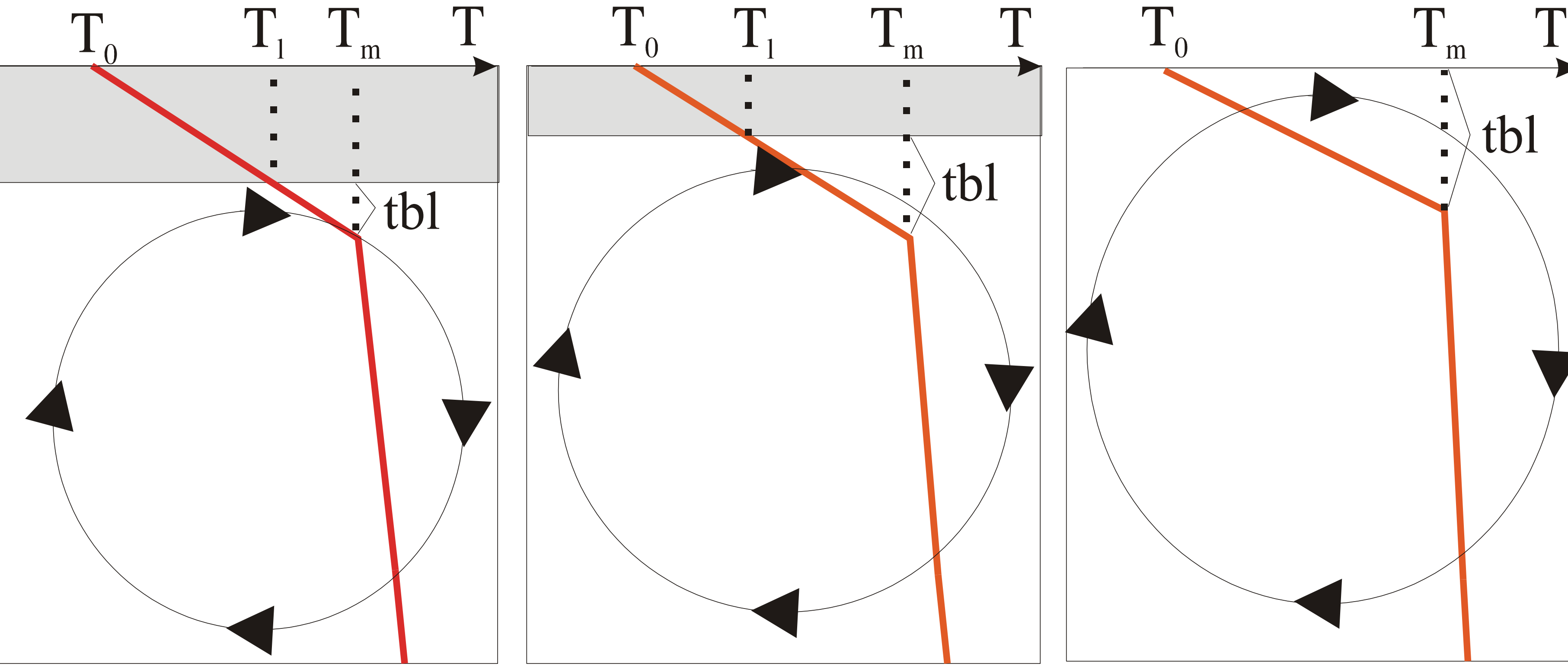
Magma transport (volcanism)



Stagnant Lid

Lithosphere Delamination

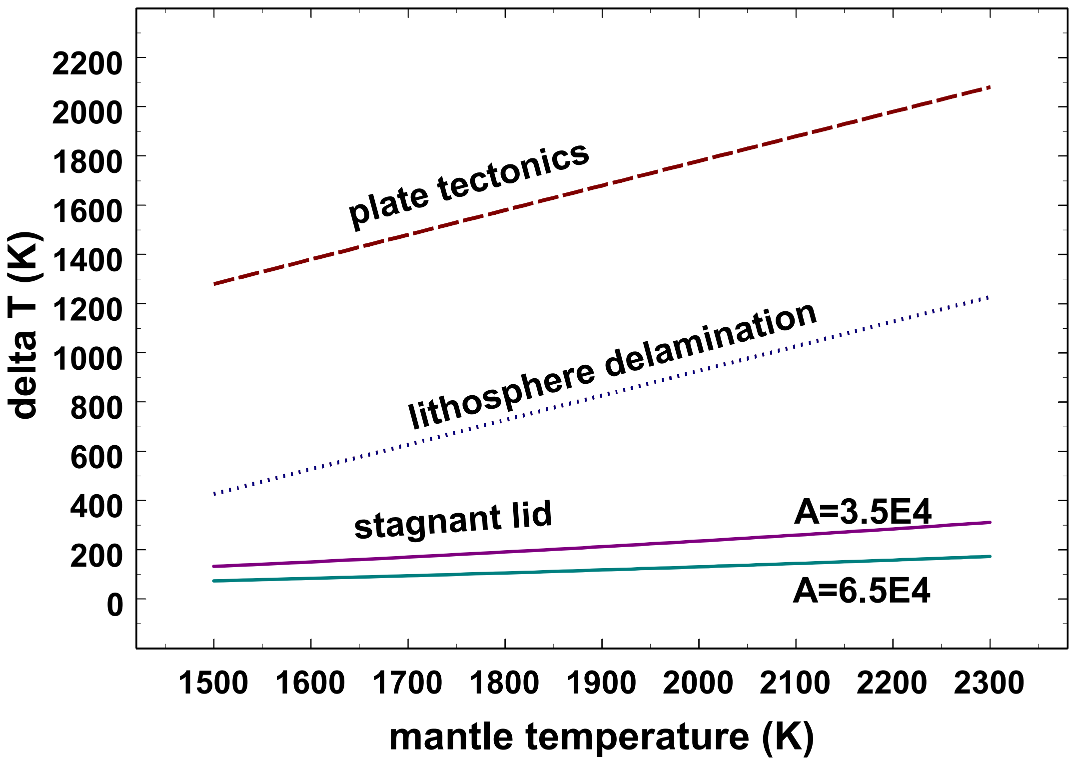
Plate Tectonics



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

$T_1 = 1073$ K

‘Driving‘ Temperature Contrast



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

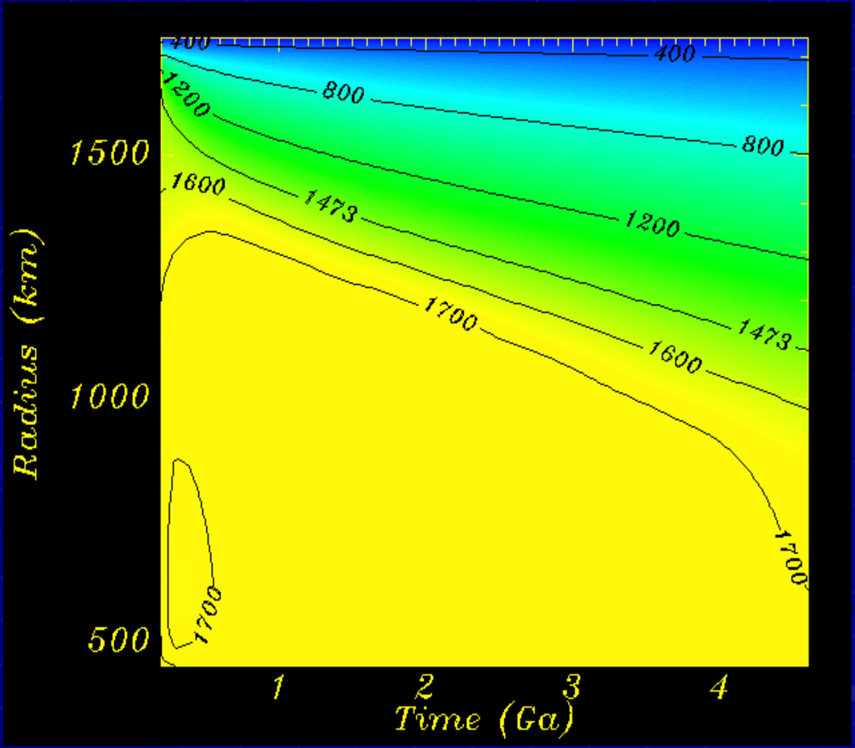
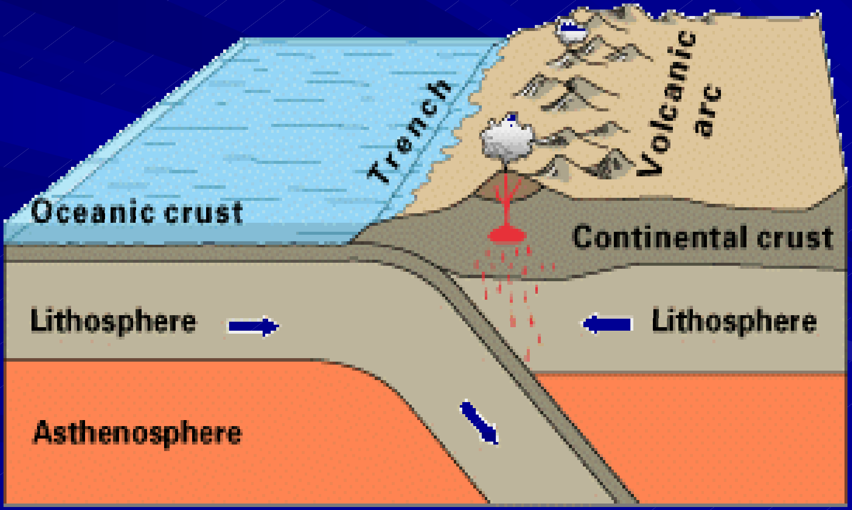


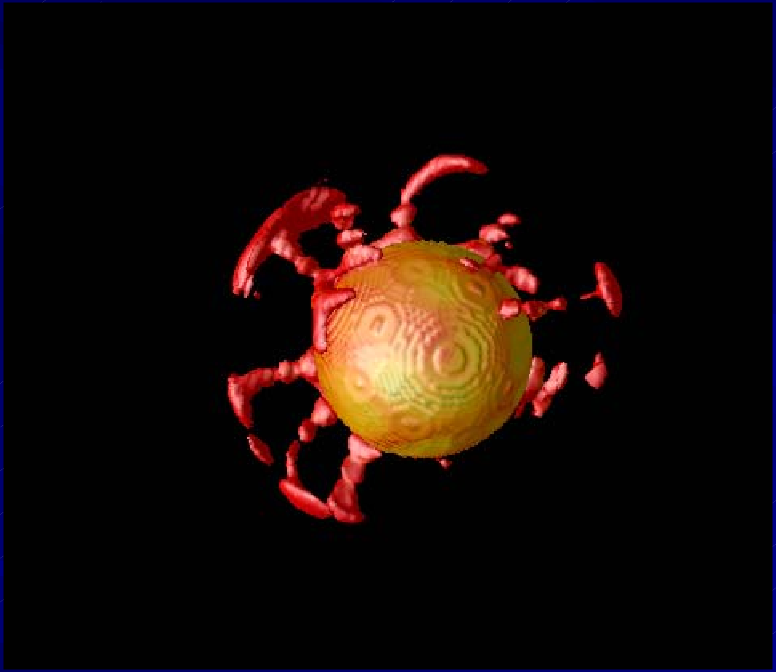
Plate Tectonics

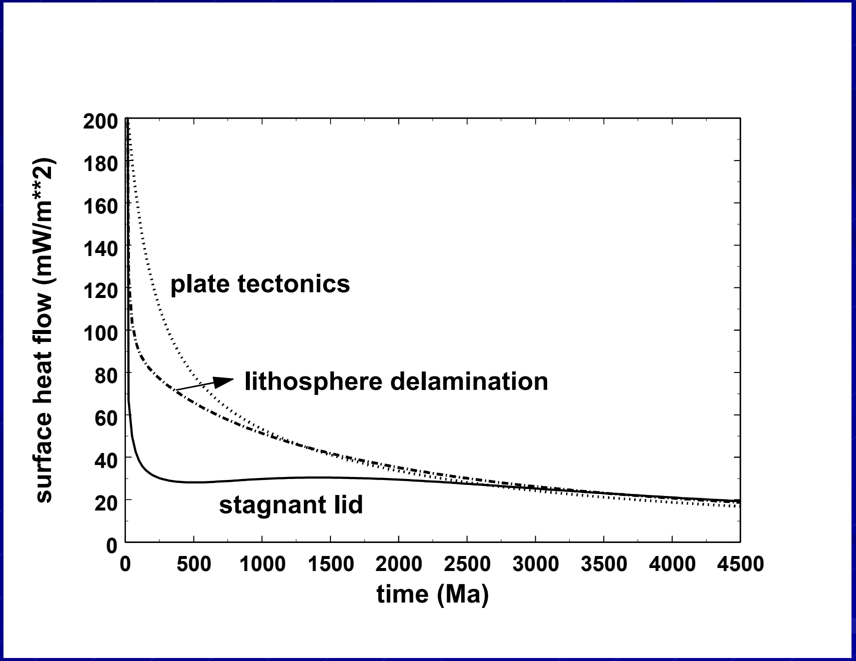
- Plate tectonics is efficiently cools the deep interior of a planet
- Vigorous core convection and inner core growth is more likely



Oceanic-continental convergence

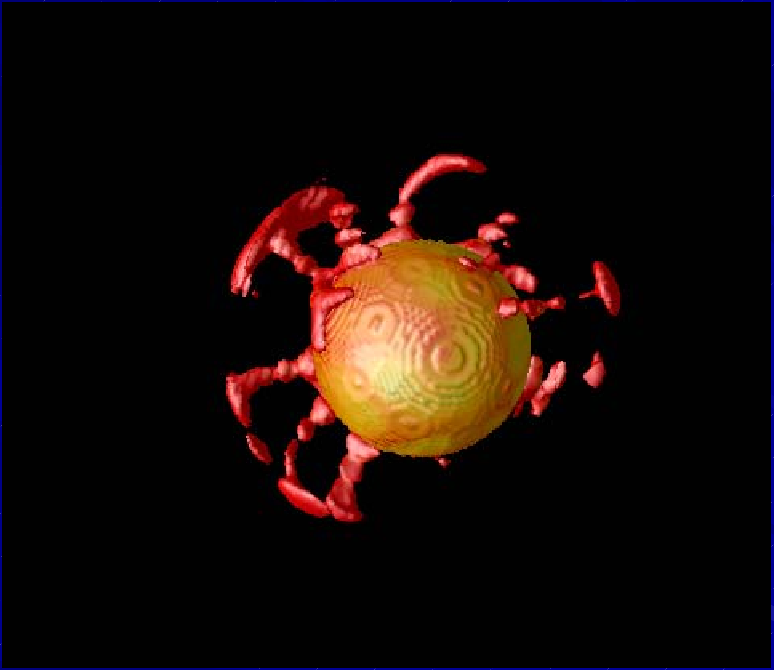
Models to Calculate Thermal Evolution and Mantle dynamics





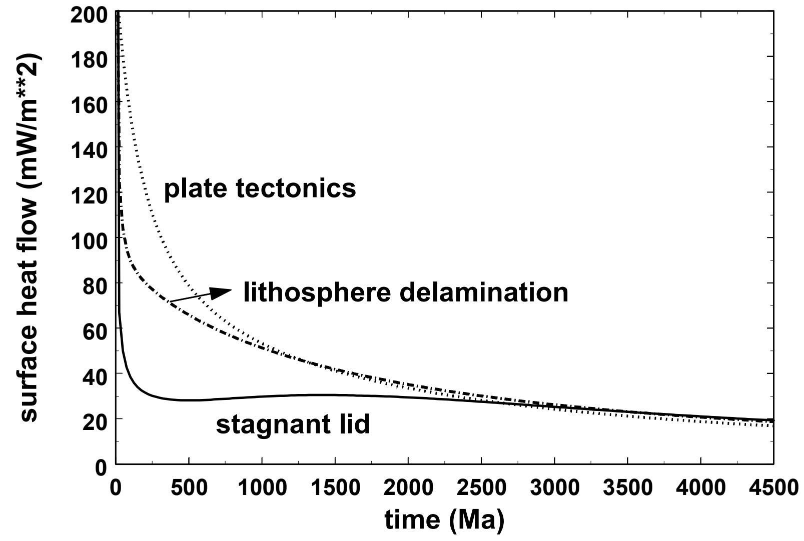
2D and 3D Convection Models

- Full set of hydrodynamic equations
- Local parameters (e.g. temperature, velocity field)
- Mantle flow pattern



Parameterized Models

- Simple scaling laws (e.g. Nu ~ Ra^b)
- Global parameters as function of time (e.g. mean temperature, heat flow)



Parametrized Convection

- Energy equation: Mantle and core

$$\rho_m C_m V_m \frac{dT_m}{dt} = -q_m A_m + q_c A_c + V_m Q_m, \quad q_m = k\frac{\Delta T_s}{\delta_s}, \quad \delta_s \approx const.\ Ra^{-\beta}$$

$$\rho_c C_c V_c \frac{dT_c}{dt} = -q_c A_c + V_c Q_c \qquad q_c = k\frac{\Delta T_c}{\delta_c}, \quad \delta_c = \left(\frac{\chi \mu_c Ra_{cr}}{\alpha\, g\, \Delta T_c}\right)^{1/3}$$

- Lithosphere growth

$$\rho_m C_m \left(T_m - T_l\right)\frac{dl}{dt} = -q_m + k\frac{\partial \mathrm{T}}{\partial \mathrm{z}}\Big|_{z=z_l}$$

- Temperature at the base of lithosphere

Plate tectonics

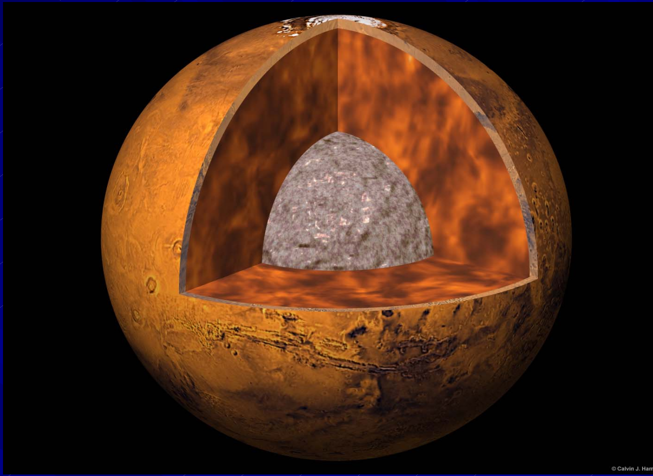
$$T_l = T_s = 220K$$

Stagnant lid convection

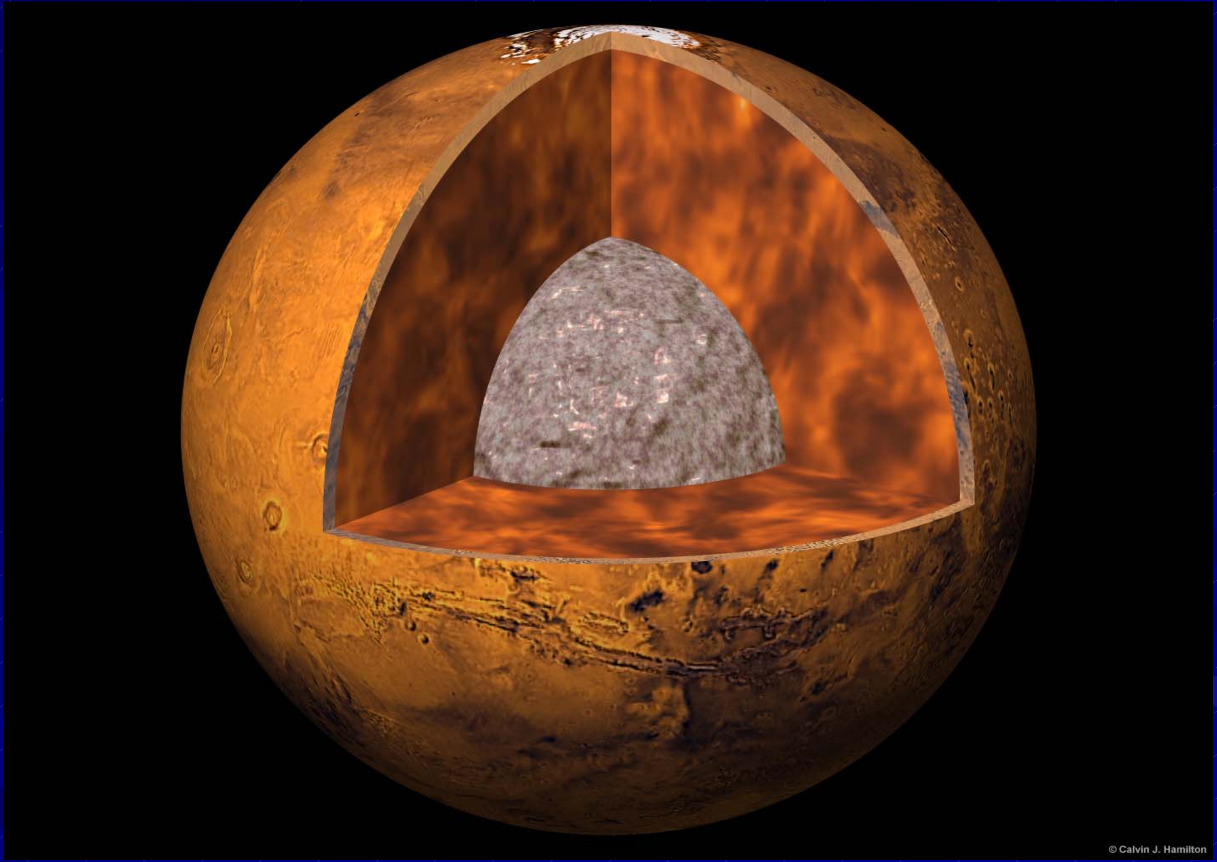
$$T_l = T_m - \Delta T_e$$

SOME EXAMPLES

- Mars
- Earth / Venus
- Mercury
- Ganymede and Europa



MARS



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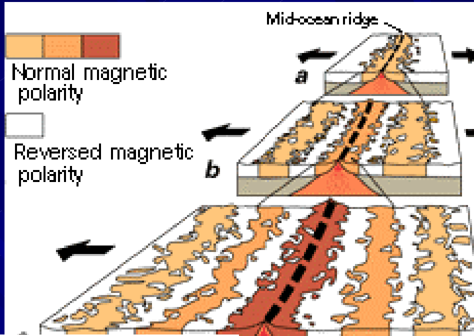
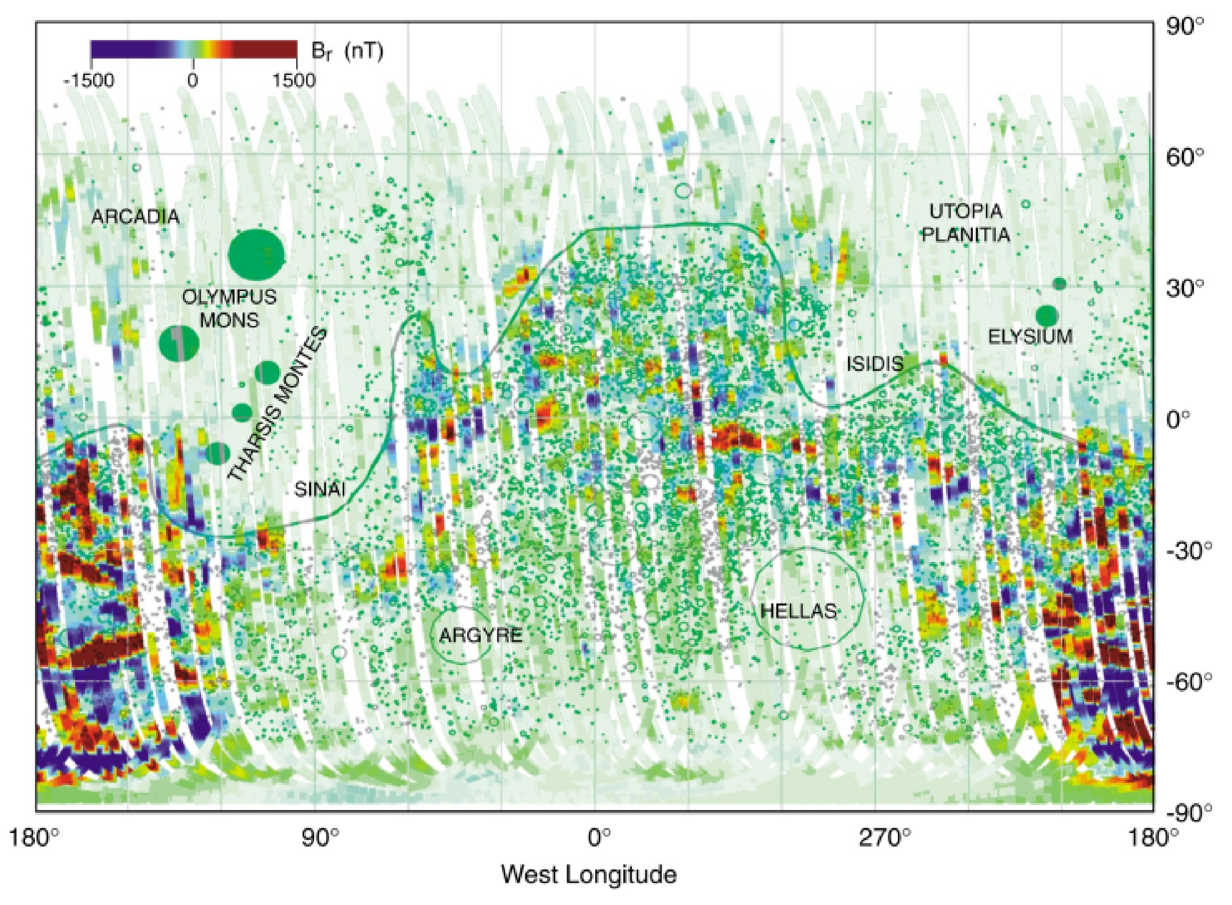
Mars


Mid-ocean ridge
Normal magnetic polarity
Reversed magnetic polarity
a
b
gma


B_r (nT)
-1500
0
1500
ARCADIA
OLYMPUS MONS
THARSIS MONTES
SINAI
ARGYRE
HELLAS
ISIDIS
UTOPIA PLANITIA
ELYSIUM
90°
60°
30°
0°
-30°
-60°
-90°
180°
90°
0°
270°
180°
West Longitude

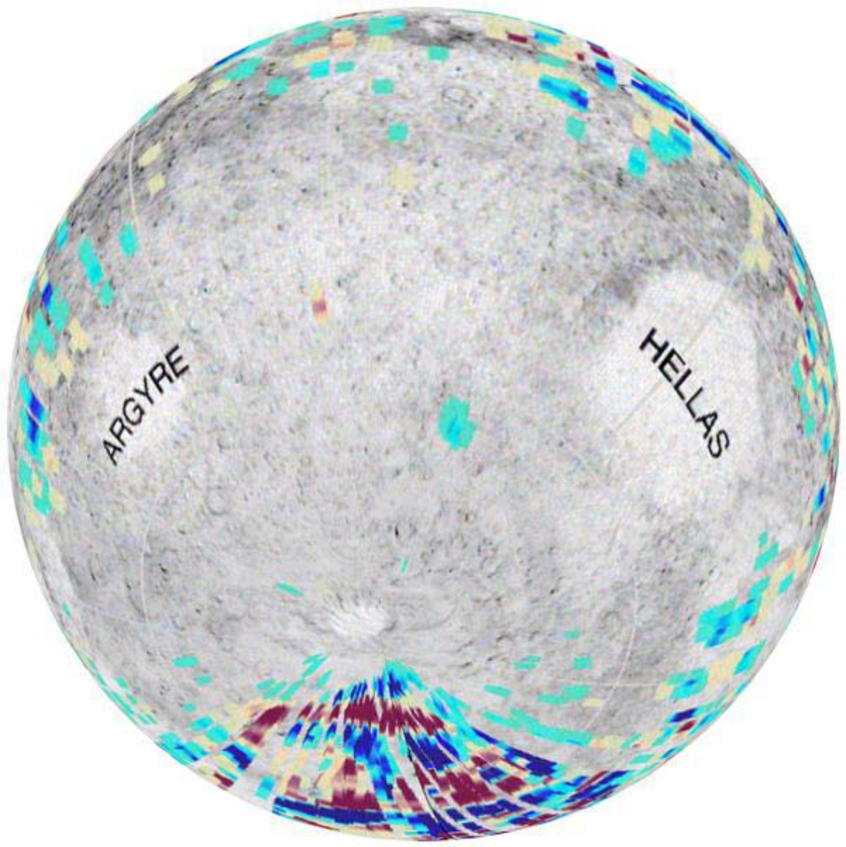
Magnetic Field History

- No present-day dynamo
- Magnetisation of oldest parts of the Martian crust
- No magnetisation of large impact basins

⇒ Dynamo action before the large impacts ~4 Ga

or

⇒ Dynamo action after large impacts



\`The Great Nothing\`

Dynamo Action Before Large Impacts

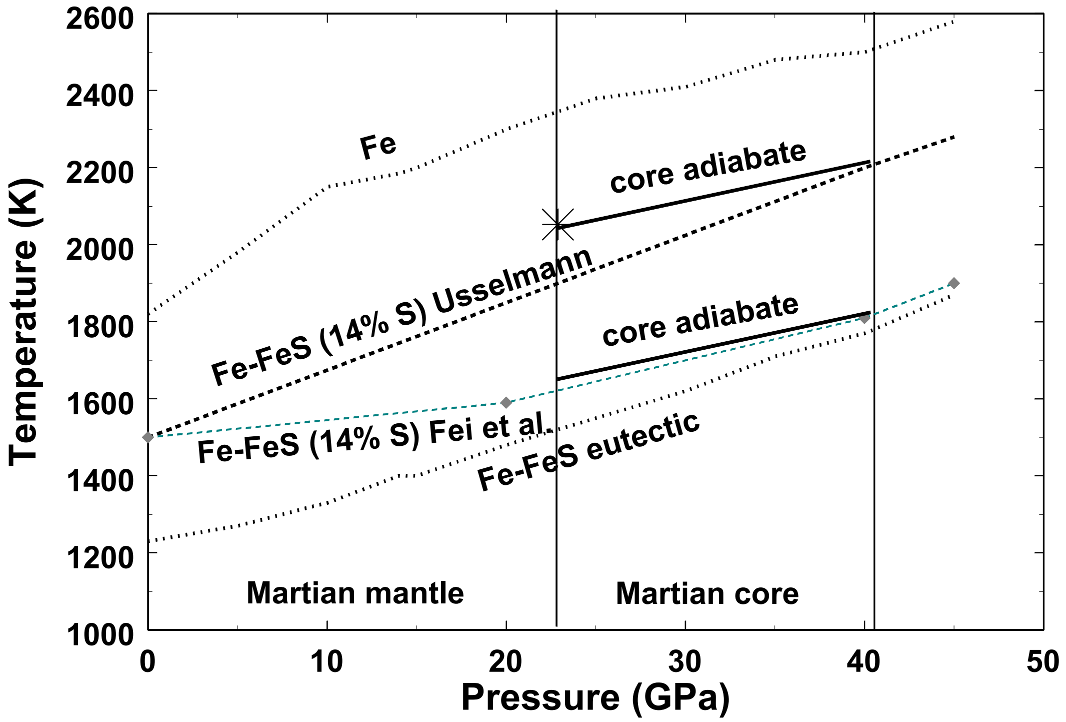
- Pro
  - ‘Easiest‘ explanation: (old surface – magnetized, young surface – non-magnetized)
  - Magnetization of old SNC meteorite (age 4.4 Ga)

- Contra
  - Thermal dynamo not very efficient
  - Difficult to explain the non magnetized areas in the southern hemisphere
  - Northern hemisphere has old crust below young surface but almost no magnetization

Dynamo Action After Large Impacts

- Pro
  - Inner core growth more efficient
  - Non magnetized area in the southern hemisphere
- Contra
  - Chemical Dynamo difficult to stop
  - Late strong crust production (e.g. plutonism) necessary but not observed on the surface
  - Early Hesperian volcanic plains (about 3.7 – 3.2 Ga) show no magnetization

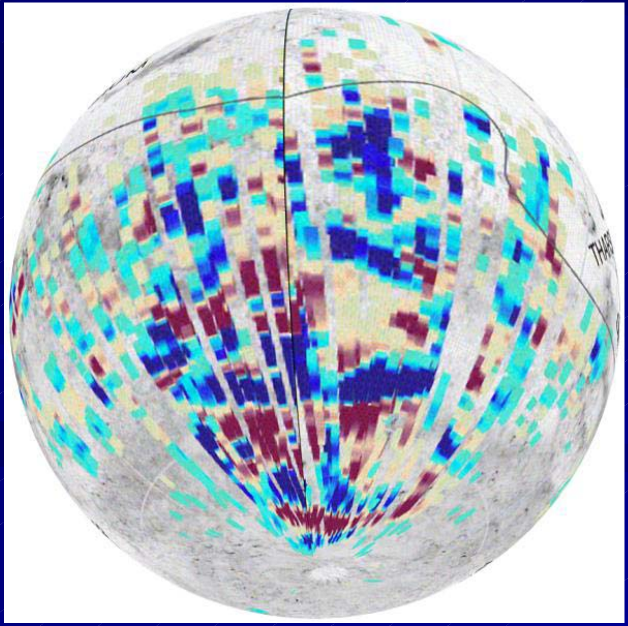
Melting Temperatures in the Martian Core

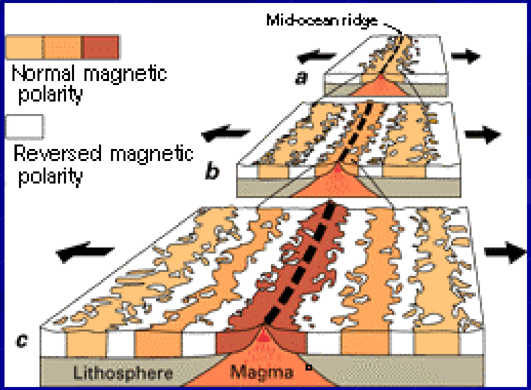


Thermal Evolution Models

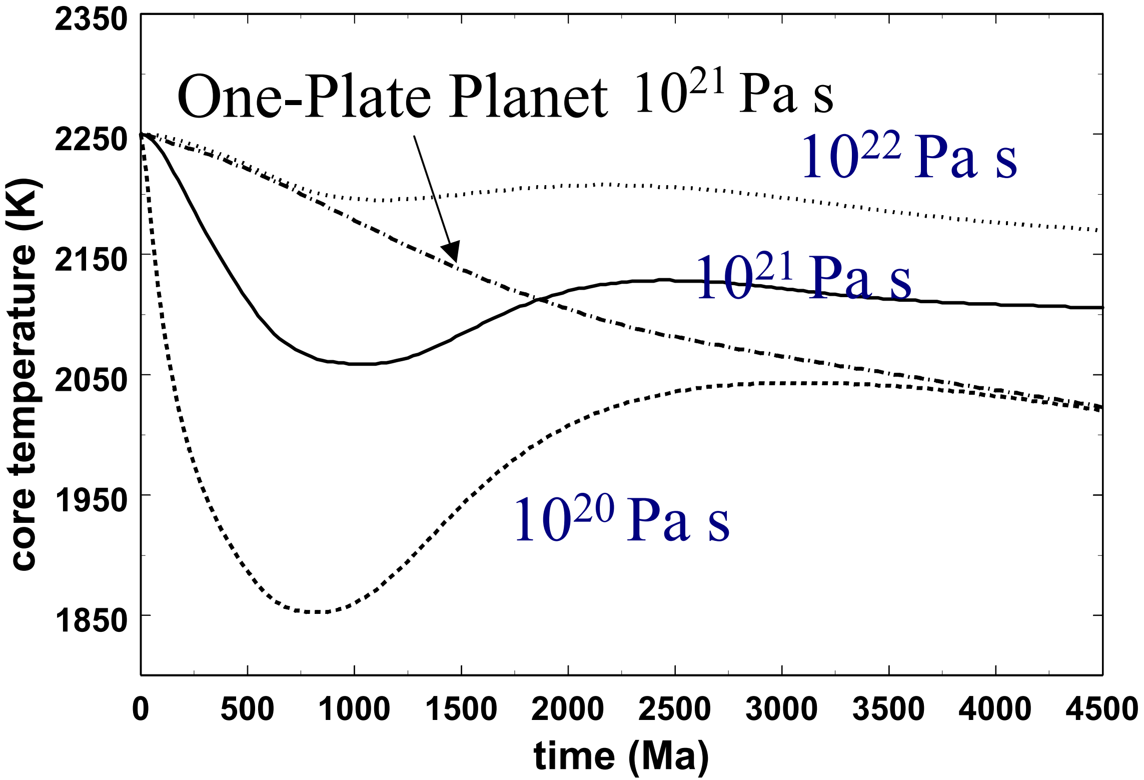
- What can we learn about Mars from the constraints on the magnetic field history?
  - Heat transfer mechanism
    (plate tectonics versus stagnant lid convection)
  - Composition
    (dry versus wet Martian mantle)

Early Plate Tectonic Regime
Versus
Stagnant Lid Convection



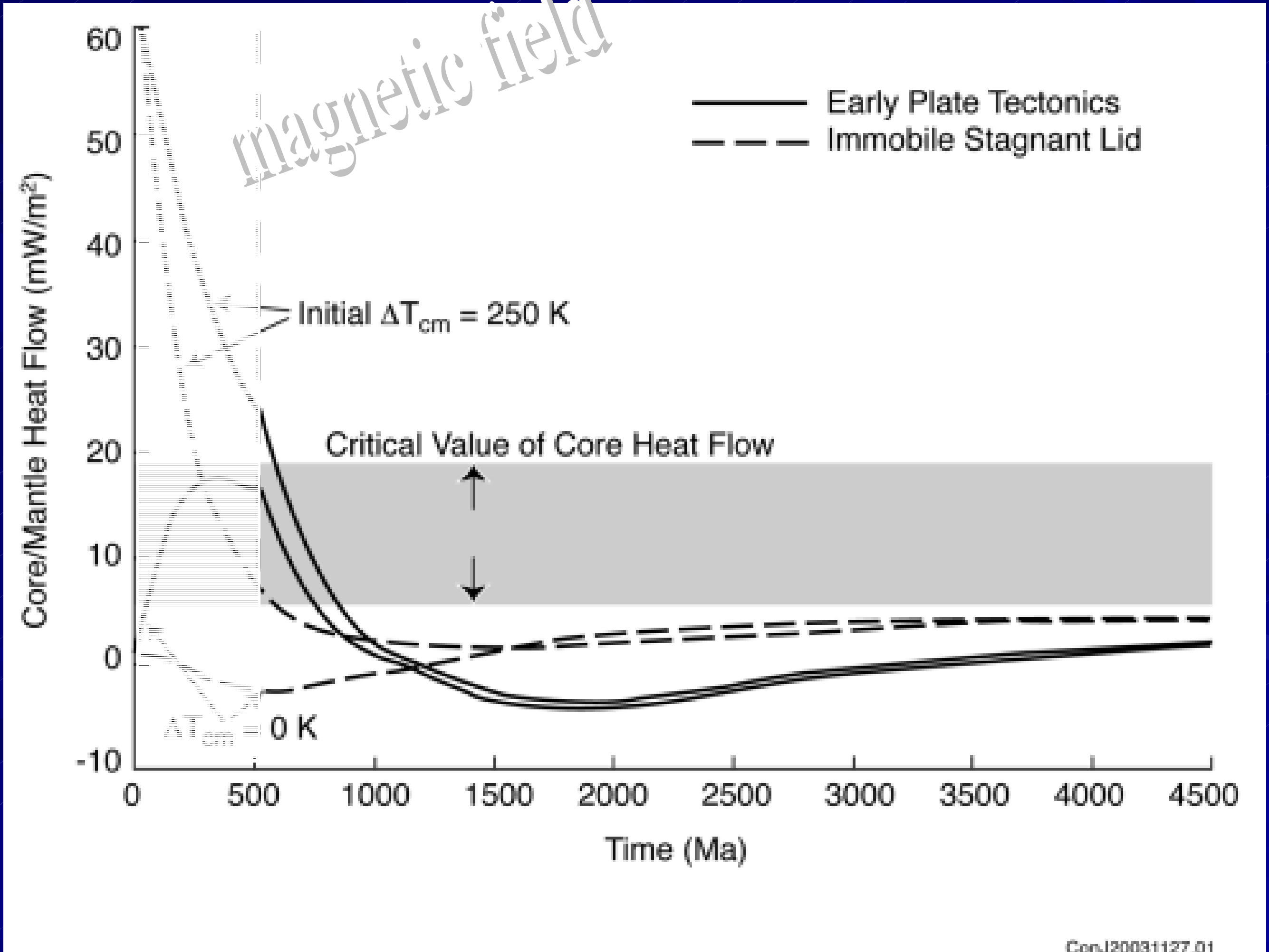


Core Temperatures



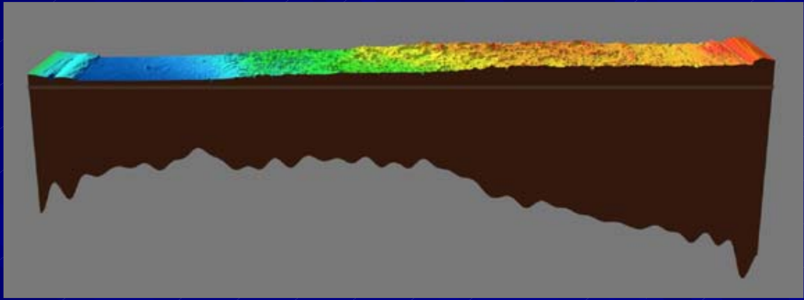
- Early Martian (thermal) dynamo possible with a superheated core

magnetic field

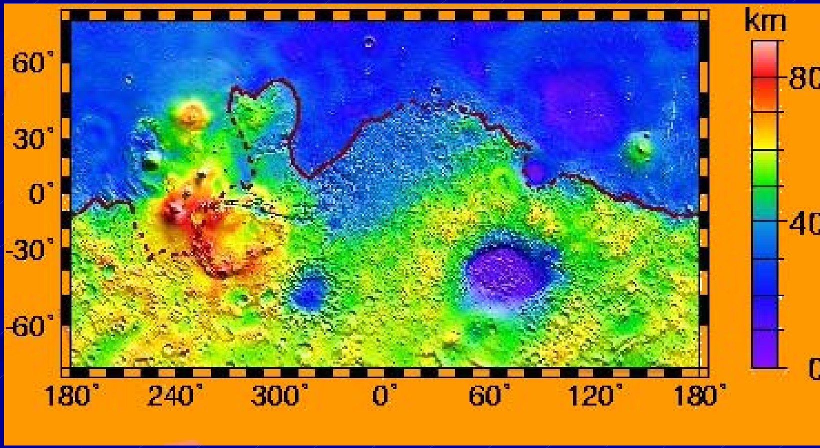


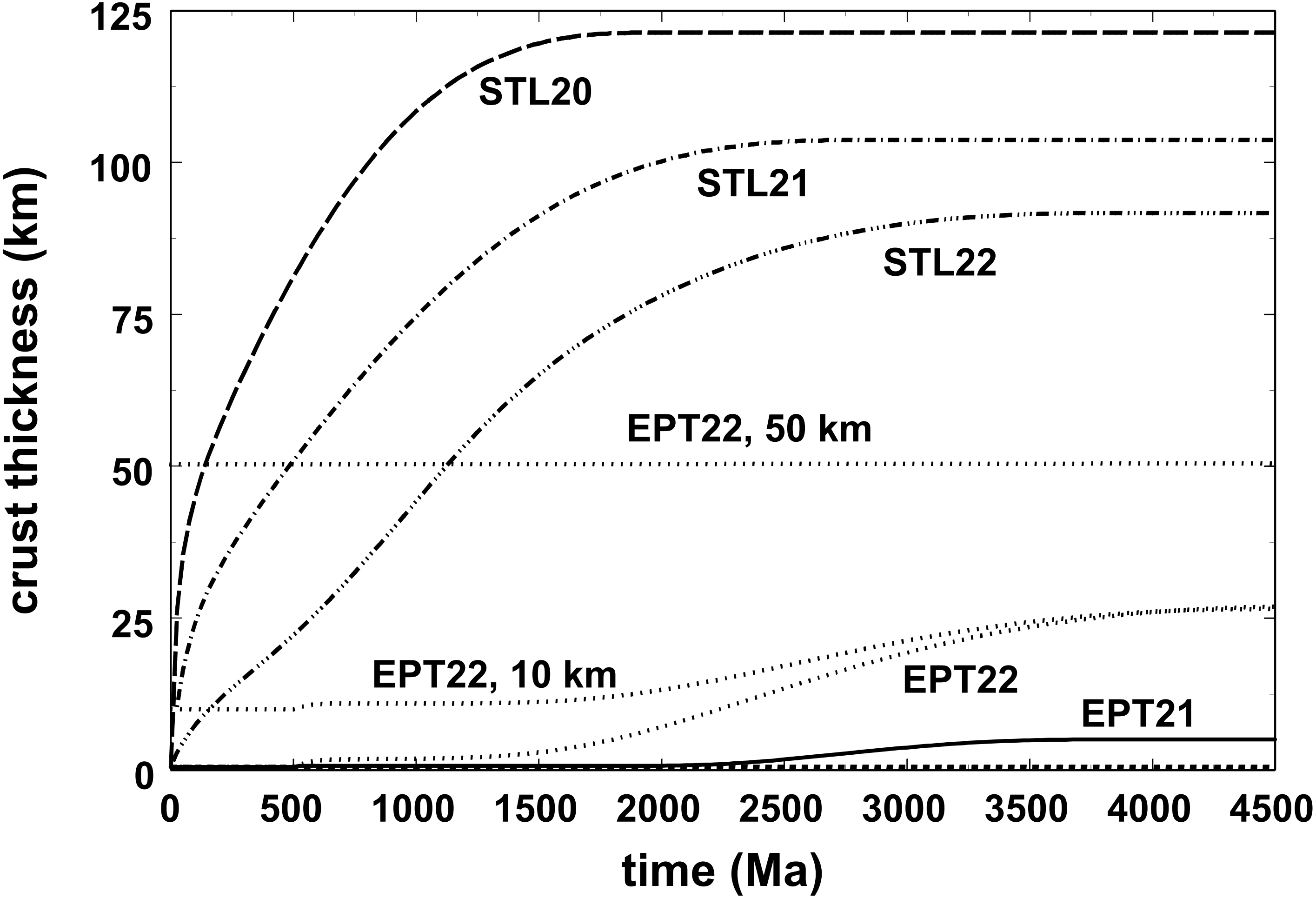
Crustal Evolution: Additional Information

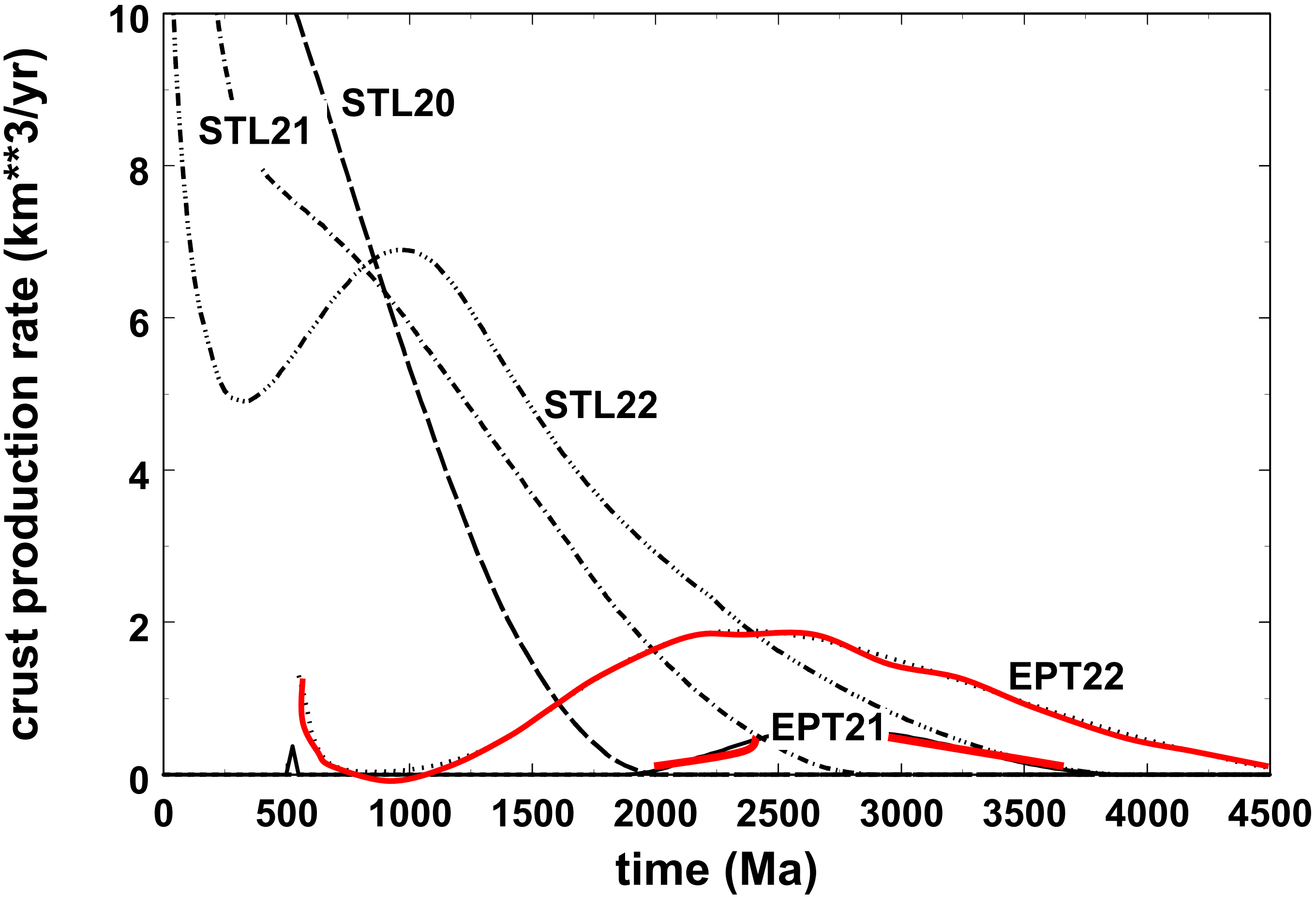
- Average crust thickness:
  50 – 120 km



- Strong decrease of crustal productivity since the Noachian
- Recent volcanism?

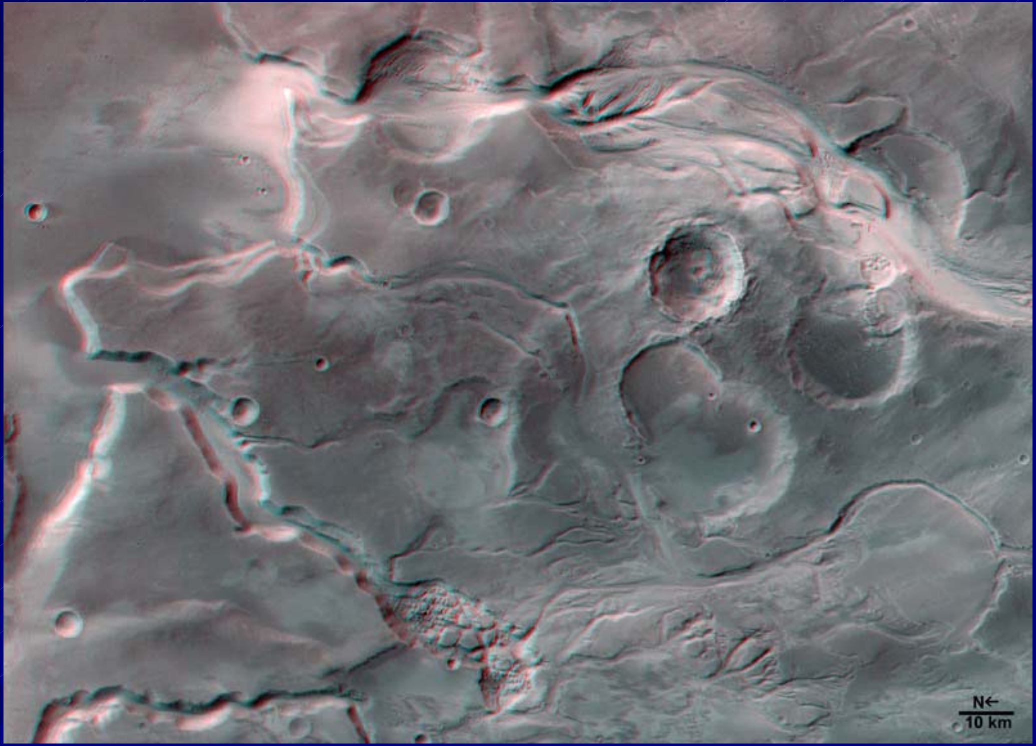



STL20
STL21
STL22
EPT22, 50 km
EPT22, 10 km
EPT22
EPT21
crust thickness (km)
125
100
75
50
25
0
0
500
1000
1500
2000
2500
3000
3500
4000
4500
time (Ma)

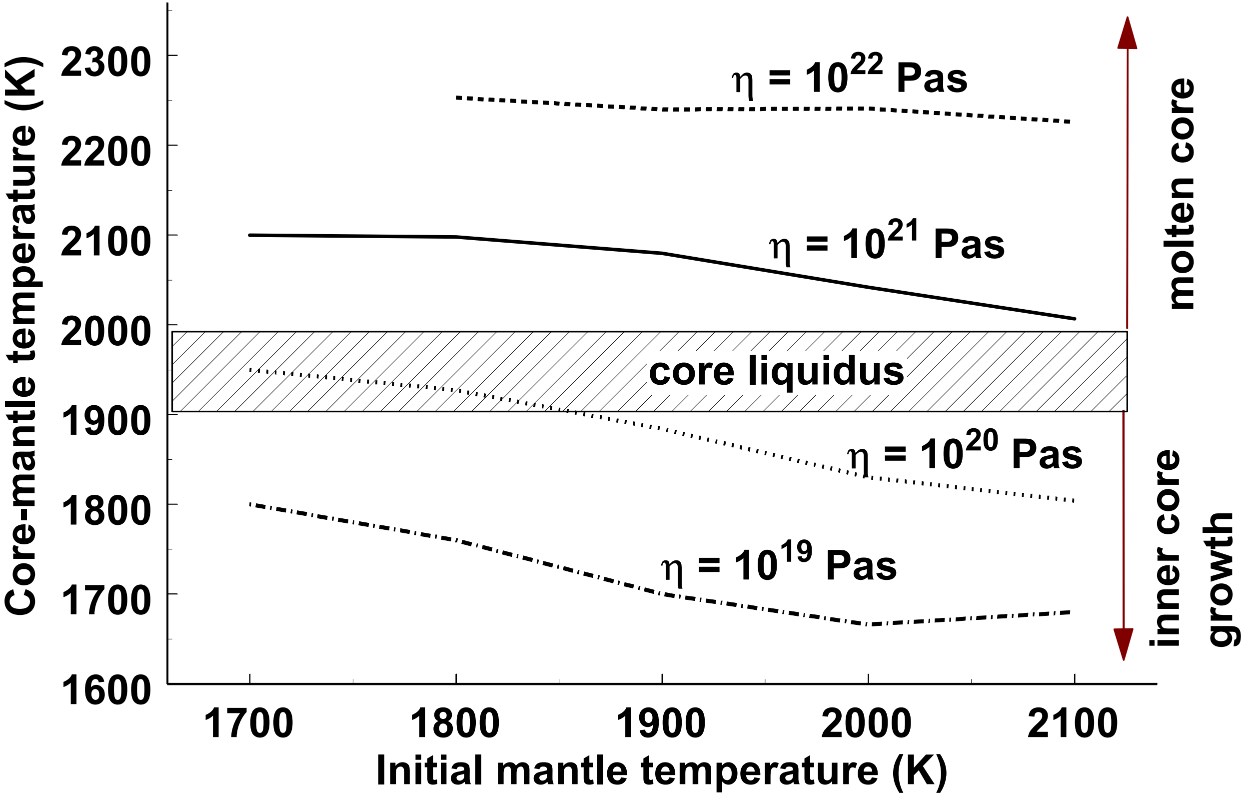


crust production rate (km**3/yr)
10
8
6
4
2
0
0
500
1000
1500
2000
2500
3000
3500
4000
4500
time (Ma)
STL20
STL21
STL22
EPT21
EPT22

Dry Versus Wet Martian Mantle



- Models with a weak/wet viscosity show present-day inner core growth for Usselman data; inconsistent with the lack of a present dynamo.



Conclusions Mars I

- Early plate tectonics consistent with early strong magnetic field
- Crustal evolution shows a peak in crustal activity 2 Ga ago and an average crust thickness smaller than 50 km, inconsistent with observation
- Stagnant lid convection consistent with early magnetic field if the core is superheated by more than 100 K.

Conclusions Mars II

- In case of Usselman data and 14 wt.% S, a dry, stiff mantle is more likely to explain early magnetic field
- In case of Fei data (strong decrease of S with pressure in eutectic composition) and 14 wt.% S, Mars indirectly ‘proves‘ that a thermal dynamo can exist

Earth

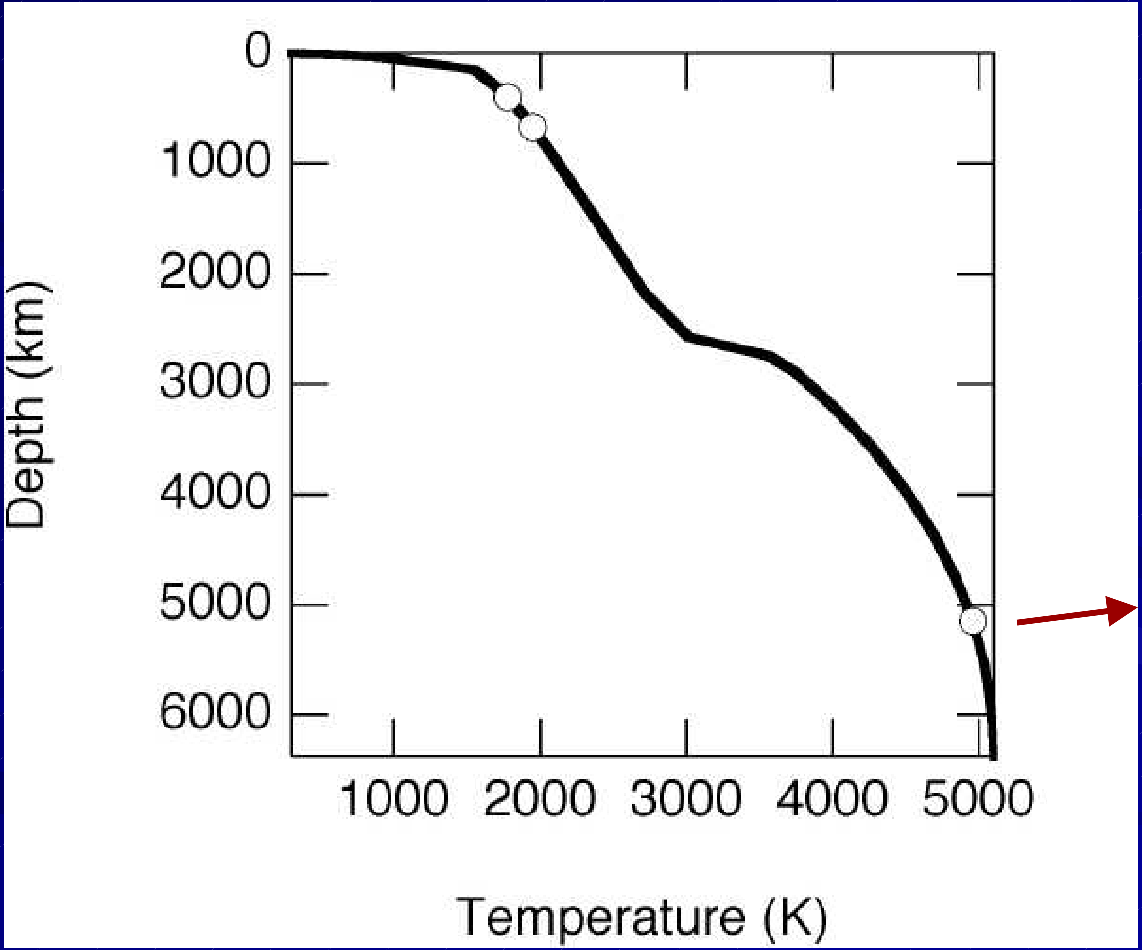
Questions

- When did the inner core growth?
- Themal dynamo active before inner core growth?
- Radioactive elements in the core?

Constraints on Thermal Evolution Models for the Earth

- Observed magnetic field evolution
- Surface heat flow
- Inner core radius

Present-Day Temperature Profile

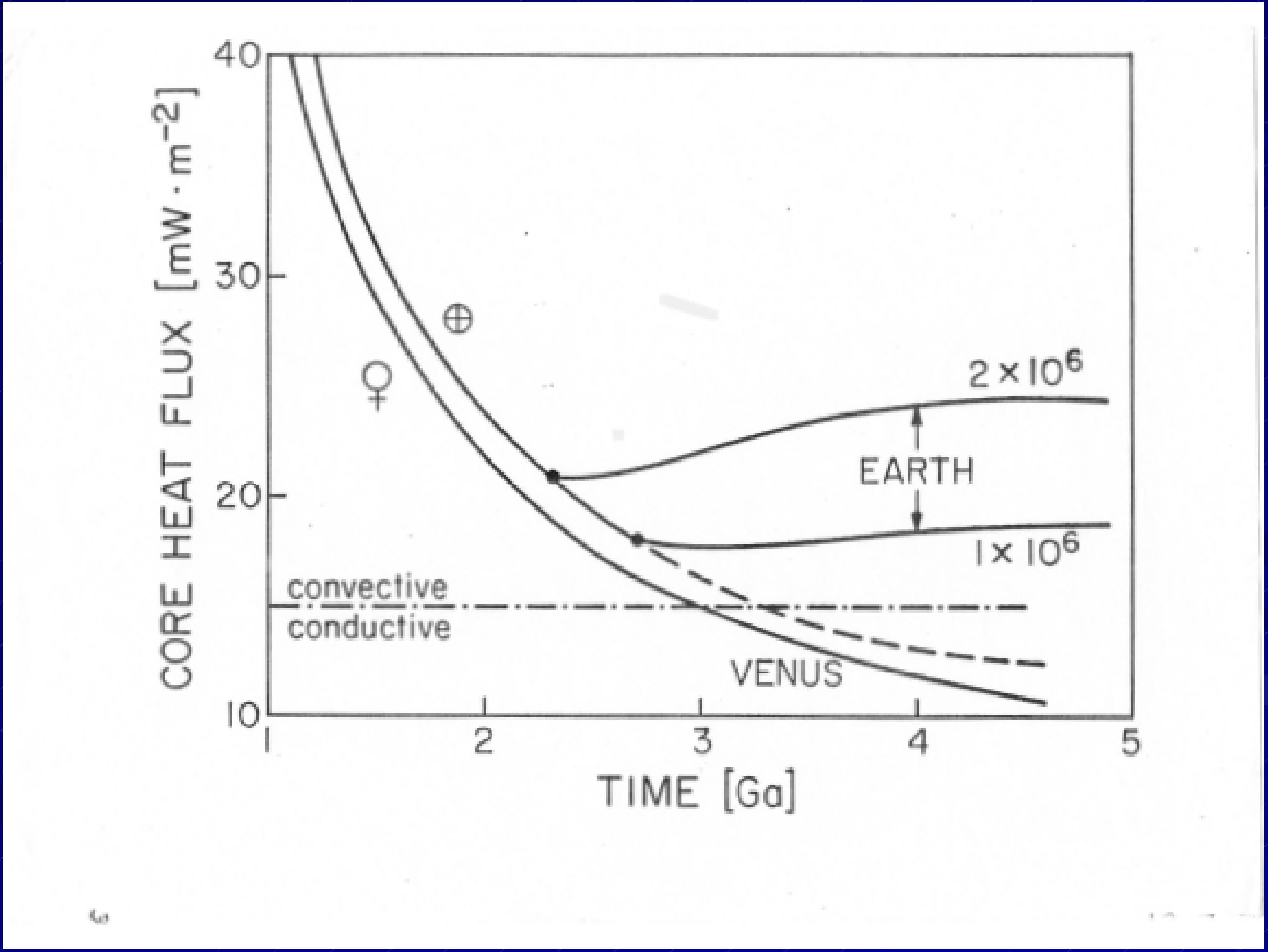


Melting temperatu
at the ICB varies
between
4800 - 5500 K

General Considerations

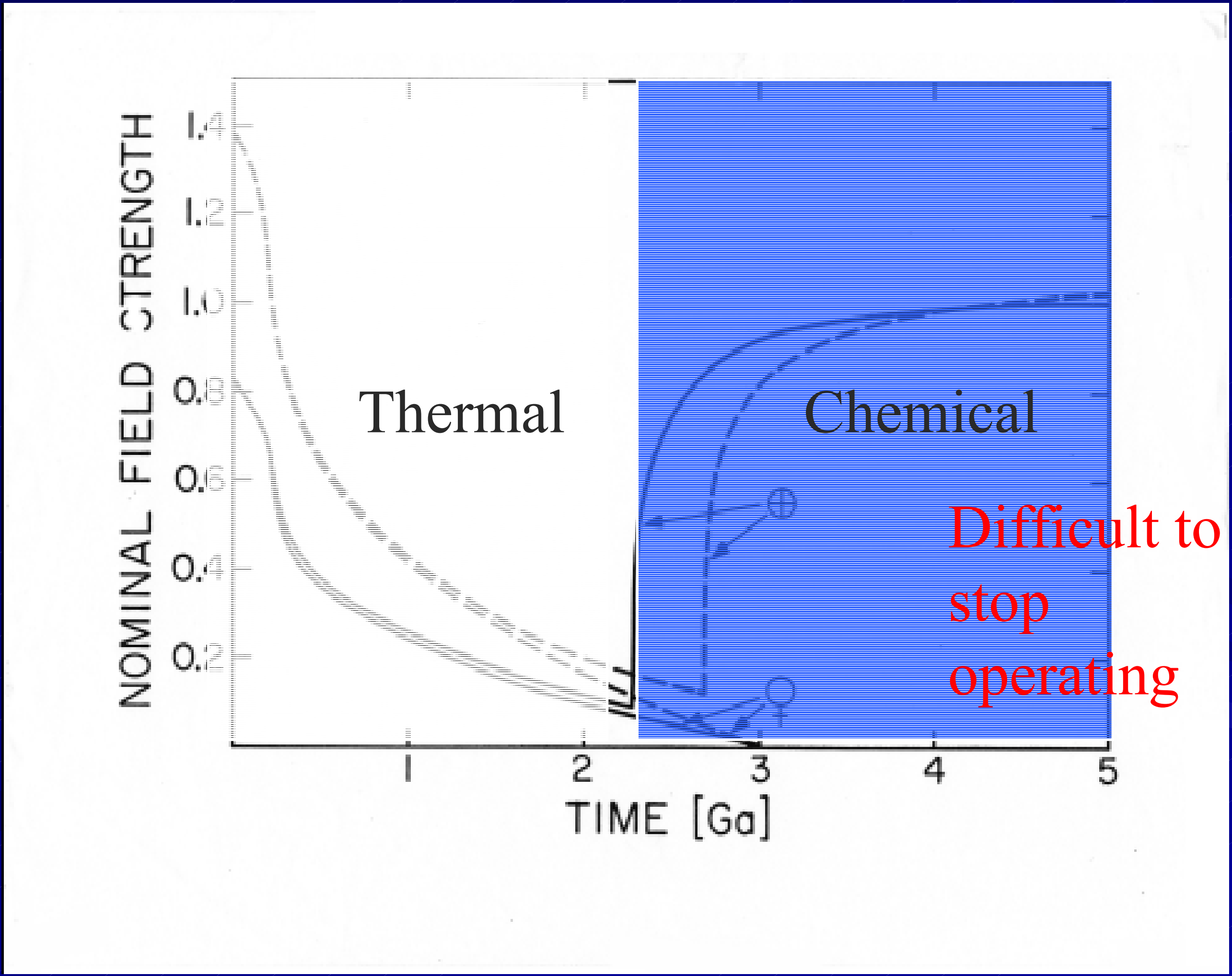
- Core-mantle heat flow too low
  - Thermal convection is shut off
  - Rate of generation of compositional bouyancy by the soldification of the core becomes too low to sustain geodynamo
- Core-mantle heat flow too large
  - Rapid growth of inner core; young age of the core
  - Requirements on primordial heat become more severe as the age of the inner core decreases.

Evolution of the Earth's Core-Mantle Heat Flow



Stevenson et al., 1983

Evolution of the Earth's Magnetic Field



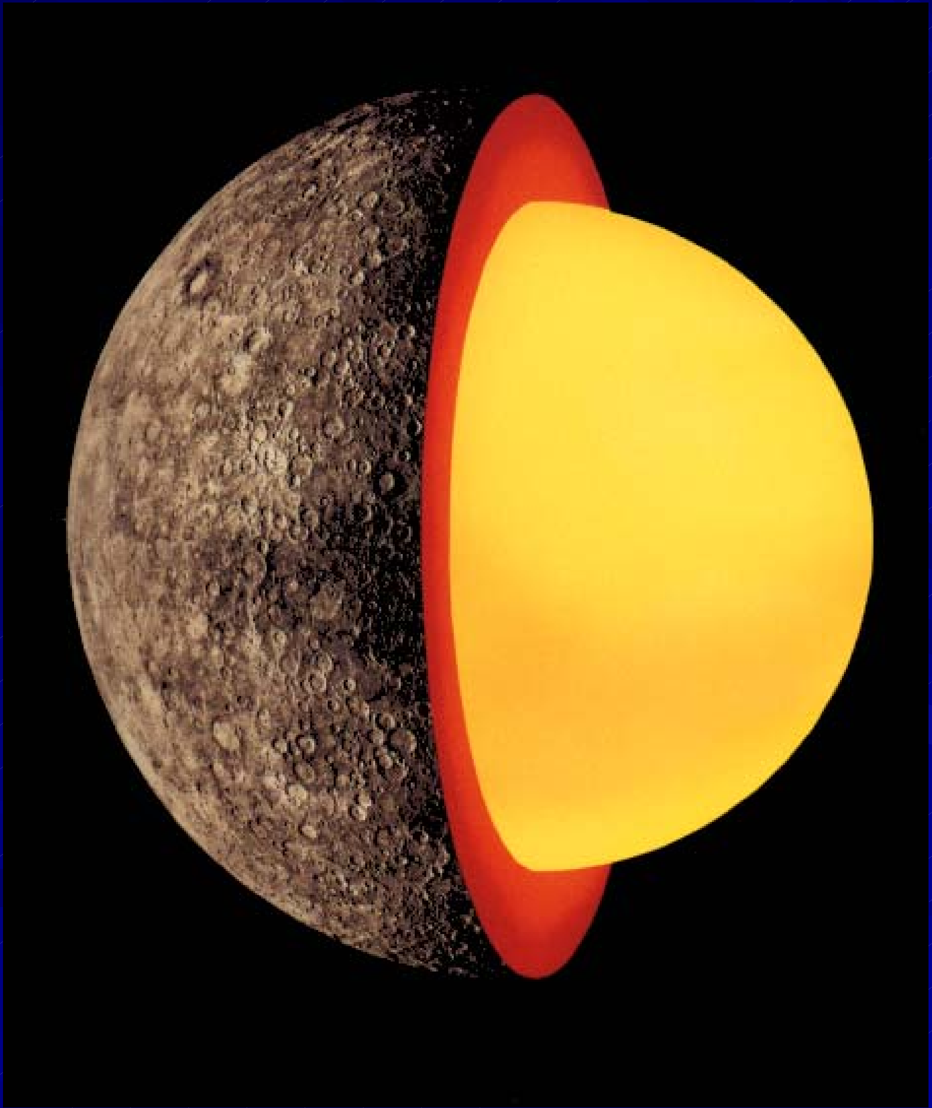
Difficult to stop operating

Stevenson et al., 1983

Conclusions Earth

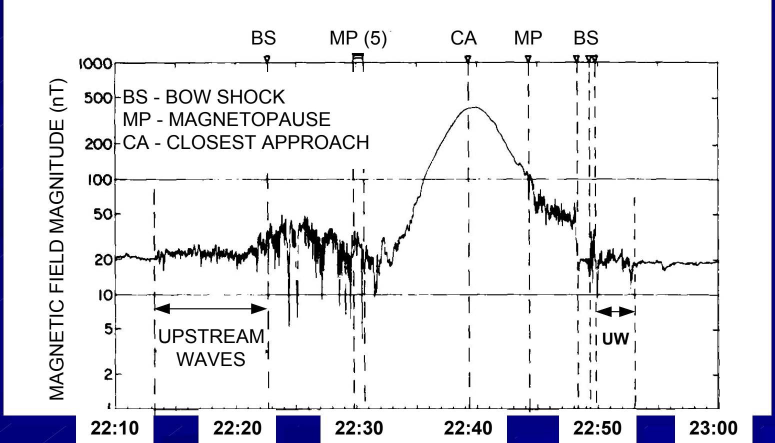
- Current thermal evolution models show growth of inner core between 1.5 and 3 Ga
  - Magnetic field must be powered by thermal dynamo in the early evolution
  - Potassium in the core is required depending mainly on ohmic dissipation and the core adiabate
- Onset of inner core depends on
  - Solidus and adiabate of the core
  - Content of radioactive elements e.g. K (the higher the potassium content the younger the inner core growth)
  - Two-layered versus one-layered convection

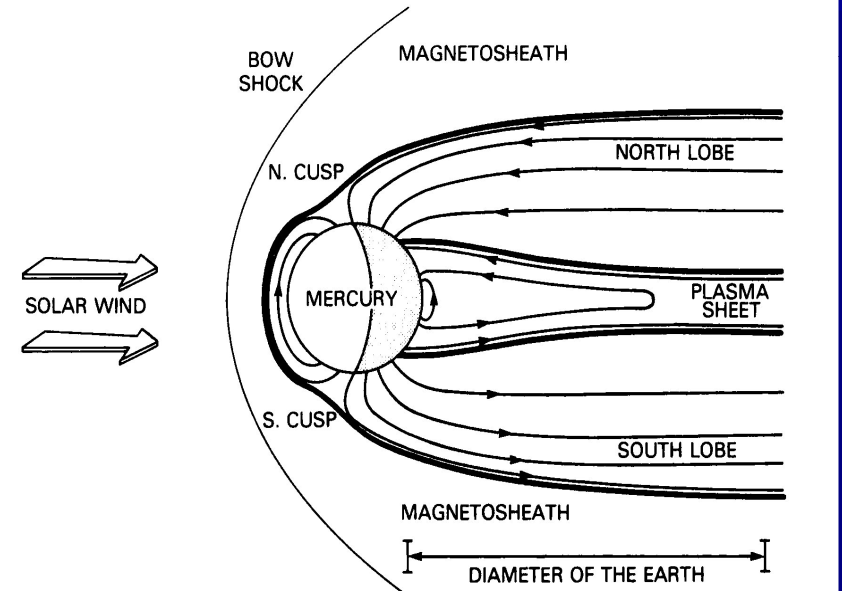
Mercury



Mercury’s Magnetic Field

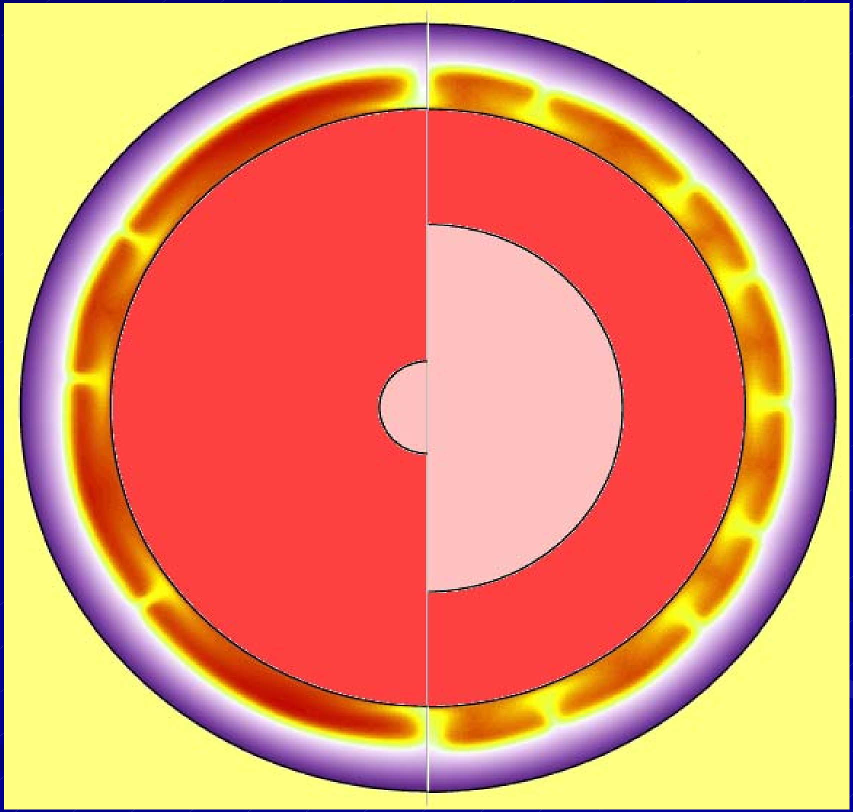
- Mariner 10 discovered Mercury's planetary magnetic field and magnetosphere
- The planetary magnetic field is sufficient to stand off the solar wind (at least most of the time)





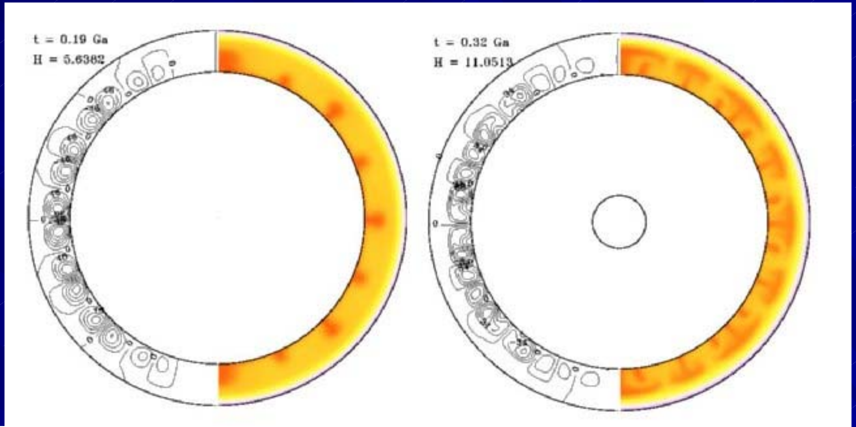
Thermal History

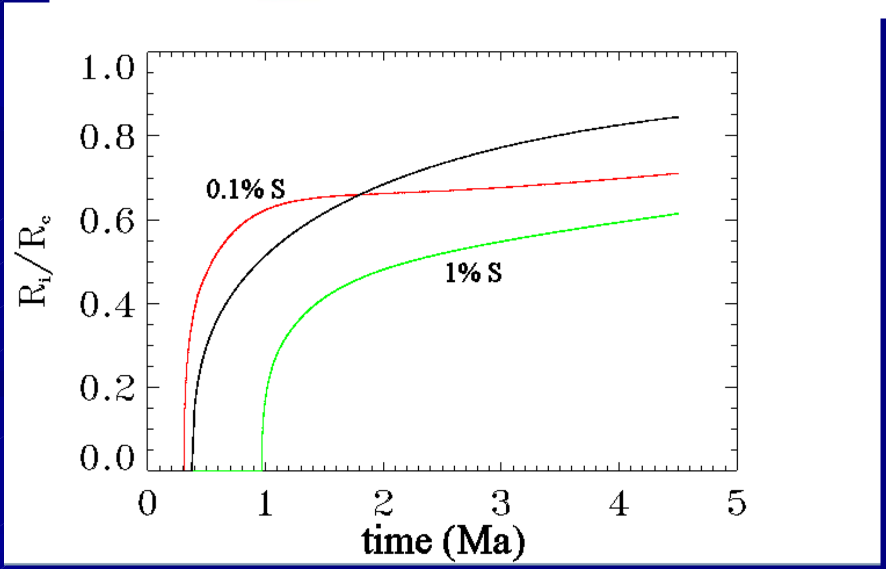
- Thermal history calculations using full convection codes
- Planet cools mostly by thickening its lithosphere



Dynamo Driven by Compositional Convection

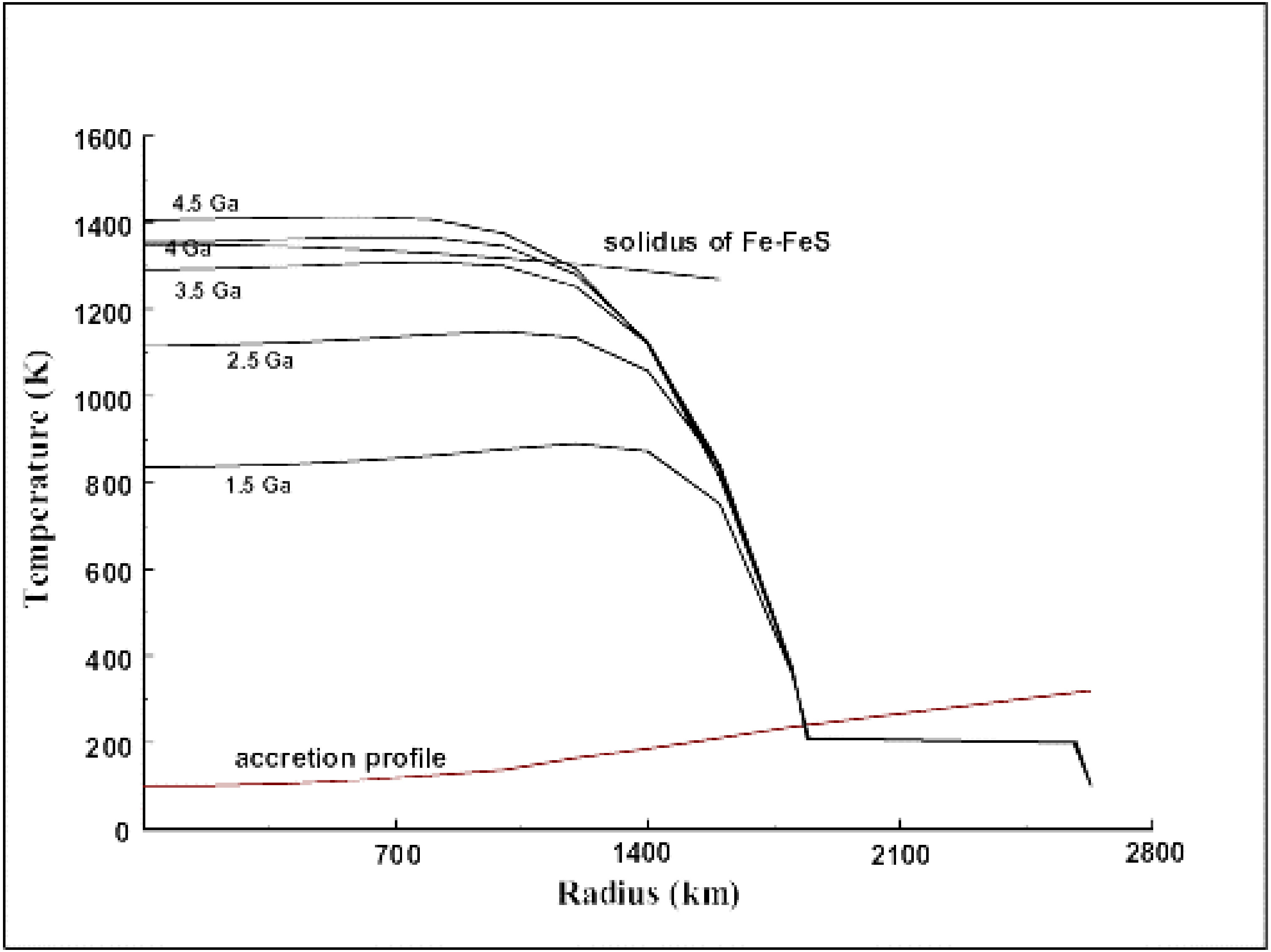
- Compositional bouyancy released by inner core growth after a 100 – 1000 Ma
- Small amount of S required; consistent with geochemical models



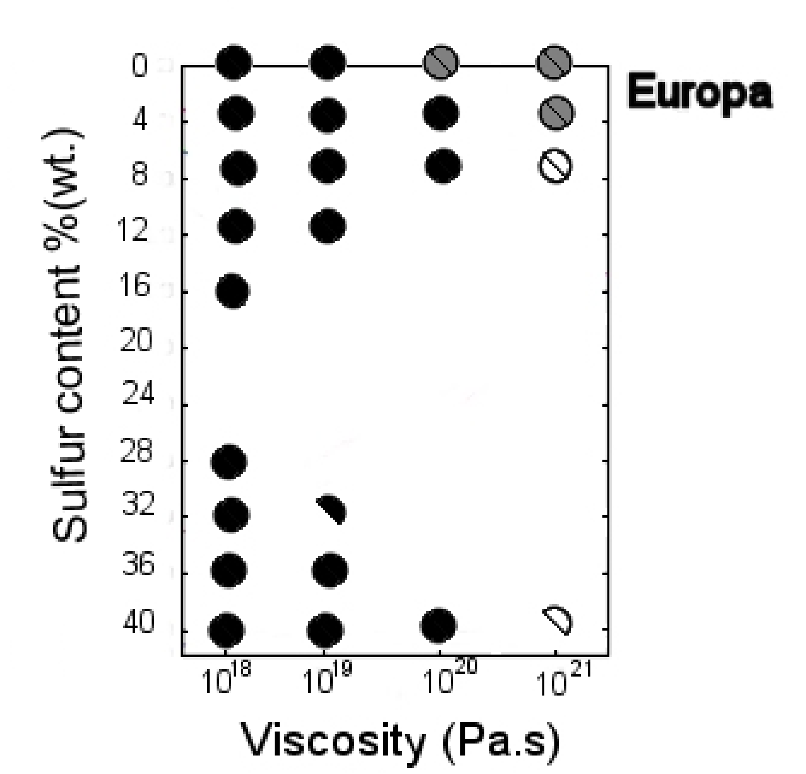


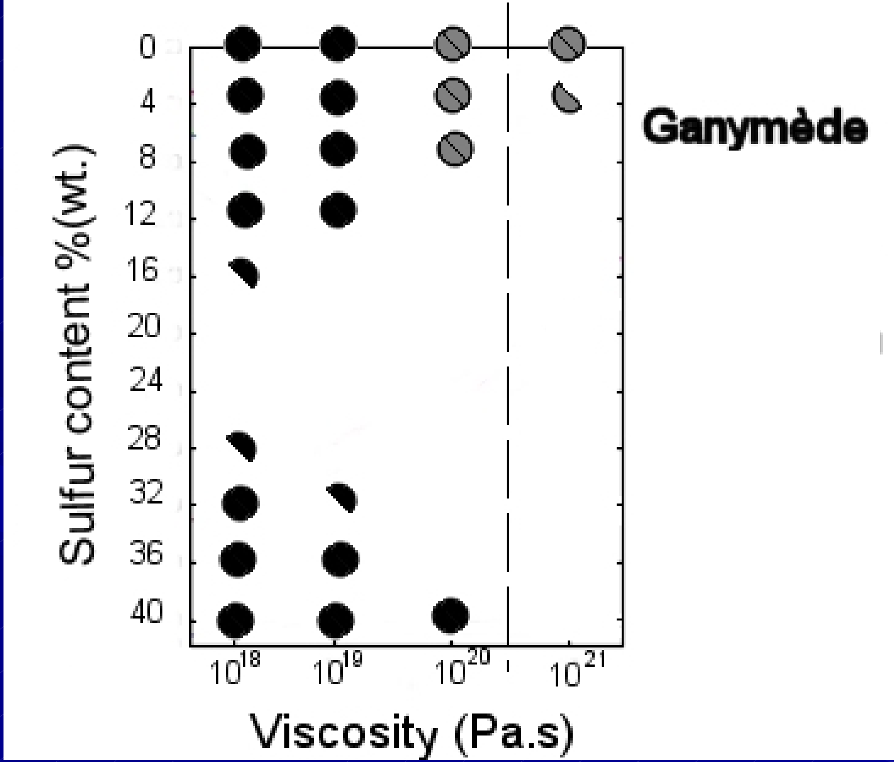
Europa and Ganymede

Temperature Profiles of Ganymede



Inner Core Growth

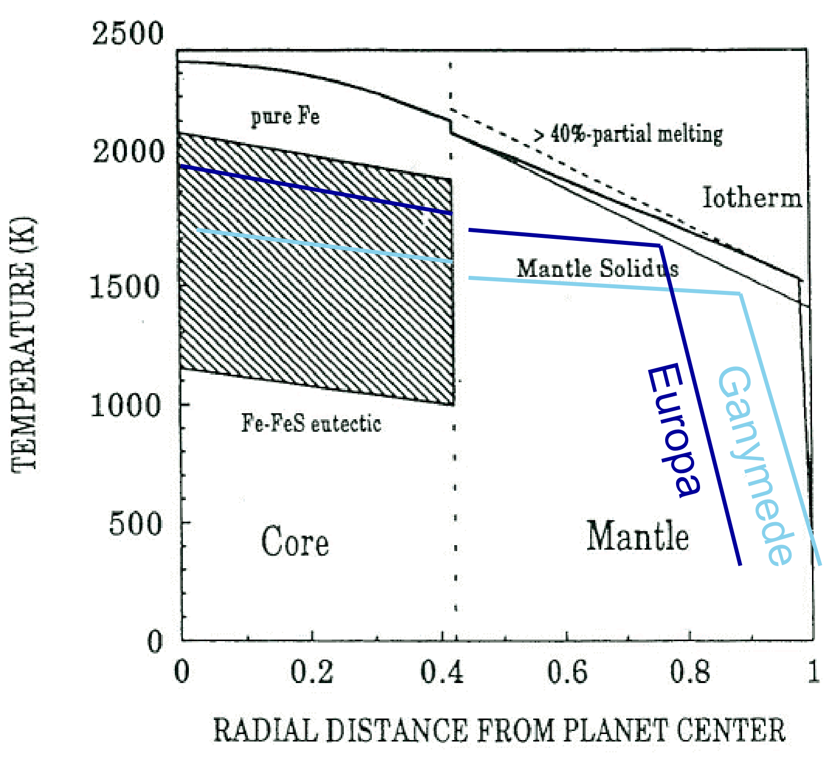




- ● Present-day inner core growth
- ○ Inner core growth early in the evolution, remanent magnetic field possib

Ganymede, Europa

- If Ganymede's core formed late
  - Slow differentiation is less favourable for a sufficient ΔT to drive a thermal dynamo.
- If core formed early
  - Dynamo action possible with chemical dynamo if low S content or weak rheology
- Tidal heating in Europa may frustrate present dynamo action



General Conclusions

- Early dynamos for One-plate planets are likely if there is a sufficiently large ΔT as a consequence of core formation
- Thermal dynamos for One-plate planets will last about 100 – 500 Ma
- Extended dynamo action requires efficient core cooling and an IC freeze-out (Earth, Mercury, and Ganymede)
  - Earth: plate tectonics
  - Mercury: Large core and low S content
  - Ganymede: weak rheology (plate tectonics?) due to water