Course of Geothermics

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Course Outline:

1. **Thermal conditions of the early Earth and present-day Earth's structure**
2. Thermal parameters of the rocks
3. Thermal structure of the lithospheric continental areas (steady state)
4. Thermal structure of the lithospheric oceanic areas
5. Thermal structure of the lithosphere for transient conditions in various tectonic settings
6. Heat balance of the Earth
7. Thermal structure of the sedimentary basins
8. Thermal maturity of sediments
9. Mantle convection and hot spots
10. Magmatic processes and volcanoes
11. Heat transfer in hydrogeological settings
12. Geothermal Systems

A practical definition

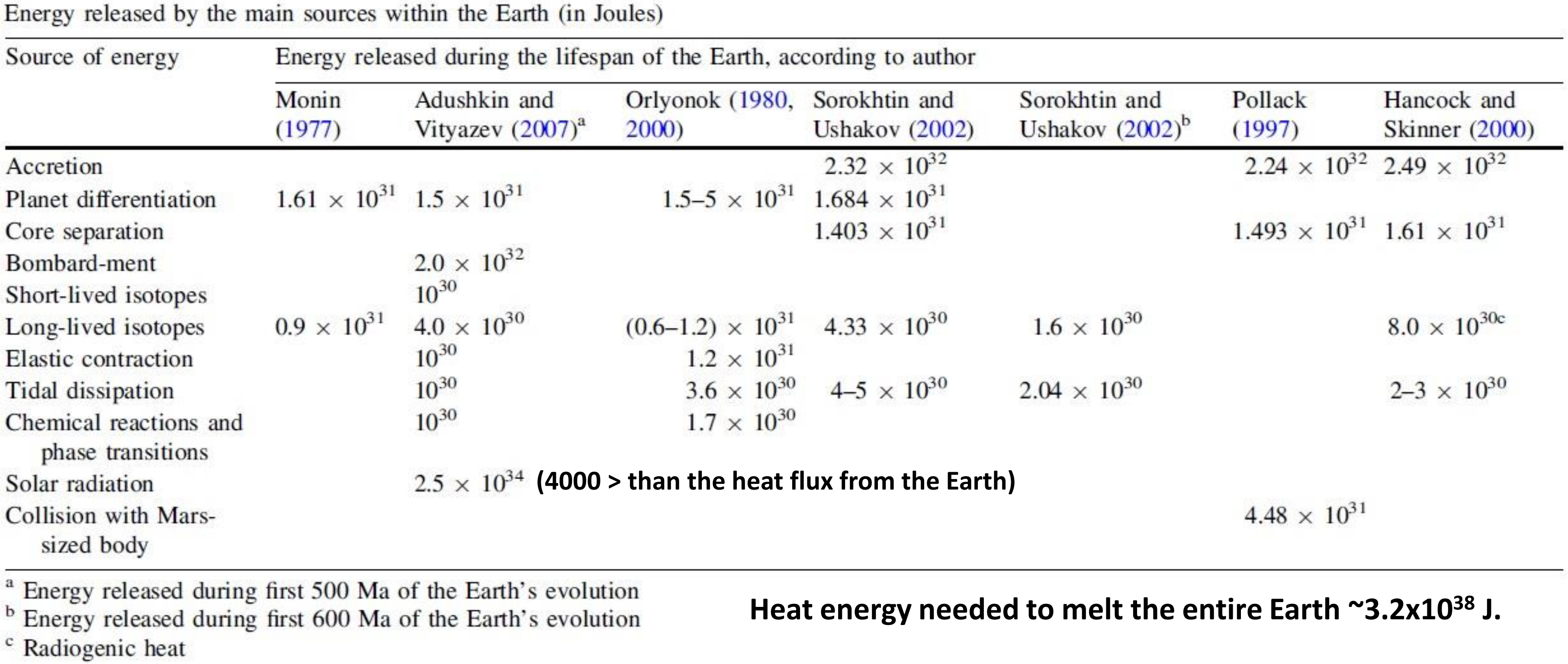
HEAT (Q, SI:[Joule]) is a form of Internal Energy of a material which is transferred through solid and fluid media by thermal interaction

TEMPERATURE (T, SI:[degC/K]) is a measurement of the Internal Energy of a system

Temperature and age of the Earth

- The initial temperature distribution within the Earth is unknown, but most of scientist agree that the Early Earth (< 50 Myr after the formation of the solar system) was in molten conditions, which facilitated the gravitational separation of dense metallic material from lighter silicates.
- Before the discover of the radioactivity in 1896 by Becquerel, Kelvin considered the Earth to be an inert body cooling in space, due to its positive thermal gradient with depth.
- By assuming that the Earth began as a sphere initially at constant temperature and using an average gradient of 36.5C/km, Kelvin calculate the age of the Earth to be ~ 200 Myr if the initial *T* was 5540°C or only 98 Myr if the initial *T* was 3870°C with an error margin of 20-400 Myr, due to the uncertainty of the thermal diffusivity of the crustal rocks.
- With the discover of the radioactivity, the age of the Earth was estimated to be ~4.55 Gyr and Turcotte (1980) attributed 83% of the present surface heat flow to the decay of radioactive isotopes and 17% to the cooling of the Earth and estimated a cooling rate of the mantle of 36C°Gyr^{-1}.

Sources of heat energy on Earth



Energy released by the main sources within the Earth (in Joules)

Source of energy	Energy released during the lifespan of the Earth, according to author						
	Monin (1977)	Adushkin and Vityazev (2007)[a]	Orlyonok (1980, 2000)	Sorokhtin and Ushakov (2002)	Sorokhtin and Ushakov (2002)[b]	Pollack (1997)	Hancock and Skinner (2000)
Accretion				2.32×10^{32}		2.24×10^{32}	2.49×10^{32}
Planet differentiation	1.61×10^{31}	1.5×10^{31}	$1.5–5 \times 10^{31}$	1.684×10^{31}			
Core separation				1.403×10^{31}		1.493×10^{31}	1.61×10^{31}
Bombard-ment		2.0×10^{32}					
Short-lived isotopes		10^{30}					
Long-lived isotopes	0.9×10^{31}	4.0×10^{30}	$(0.6–1.2) \times 10^{31}$	4.33×10^{30}	1.6×10^{30}		8.0×10^{30}[c]
Elastic contraction		10^{30}	1.2×10^{31}				
Tidal dissipation		10^{30}	3.6×10^{30}	$4–5 \times 10^{30}$	2.04×10^{30}		$2–3 \times 10^{30}$
Chemical reactions and phase transitions		10^{30}	1.7×10^{30}				
Solar radiation		2.5×10^{34} **(4000 > than the heat flux from the Earth)**					
Collision with Mars-sized body						4.48×10^{31}	

[a] Energy released during first 500 Ma of the Earth's evolution
[b] Energy released during first 600 Ma of the Earth's evolution
[c] Radiogenic heat

Heat energy needed to melt the entire Earth ~3.2×10^{38} J.

- The main sources of heat energy on Earth are (1) the energy of solar radiation, (2) the release of gravitational energy during planetary accretion, (3) gravitational energy during differentiation, (4) the contraction of the planet and its separation into its core, mantle and crust, (5) radioactive energy by short- (^{26}Al, ^{36}Cl, ^{60}Fe) and long-lived (^{235}U, ^{238}U, ^{232}Th, ^{40}K) radioactive isotopes.
- Some sources generated heat energy in the past (accretion, overall differentiation of the planet, decay of short-lived radioactive isotopes), while some others are sources of perpetual and ongoing generation of heat energy (radiogenic heat from the decay of long-lived radioactive isotopes, ensuing differentiation, tidal energy, chemical reactions, etc.).
- The sources of heat loss include the constant radiation of heat by the Earth into outer space and escaping matter, such as hydrogen (heat capacity ~ 14.3 kJ kg^{-1} K^{-1}), derived from decomposition of water from *UV* rays.

Accretion of the Earth

- The accretion process of Earth brought together matter which was originally dispersed in the proto-solar nebula, releasing gravitational energy (E_g ~$2.2\text{x}10^{32}$J).
- During accretion, the gravitational energy of impactors is first transformed into kinetic energy and then dissipated in the form of heat at the impact: (a) If all the energy is released at shallow depth and lost to space by radiation, accretion would raise the temperature of the Earth by <70 K. (b) If no energy is lost in the space and thus is used to heat the Earth, the temperature of the whole Earth is raised by:

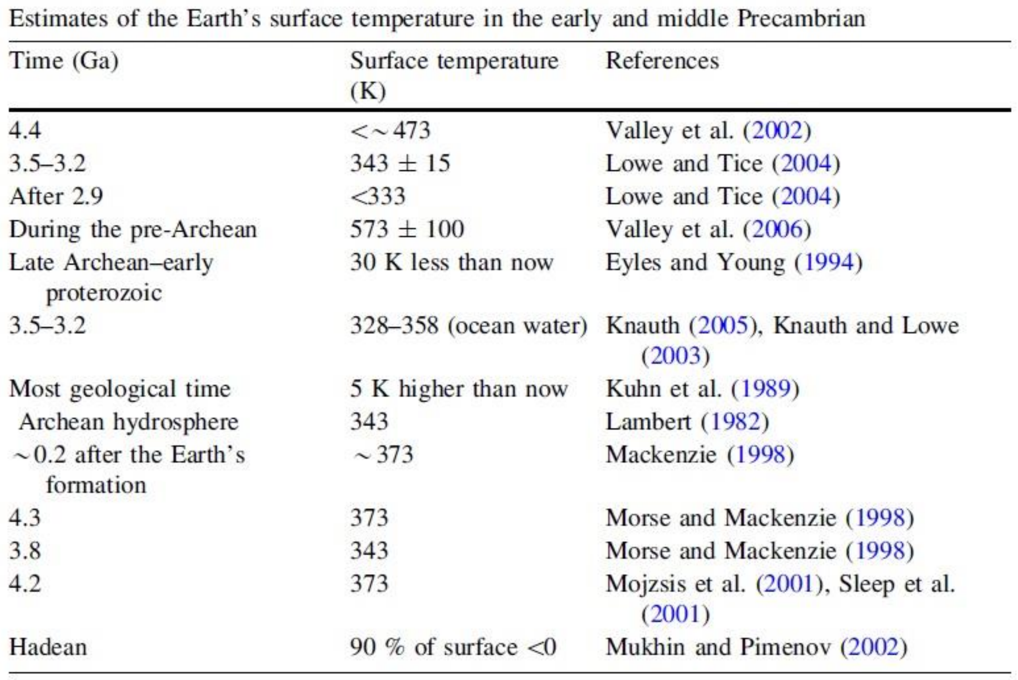
$$\Delta T = \frac{-E_g}{MC_P} \sim 3.75 \times 10^4 \text{ K}$$

M=mass of the Earth ($5.974\text{x}10^{24}$ kg) C_p= specific heat capacity (~ 10^3 J Kg^{-1} K^{-1})

- The actual evolution occurred between the two end-members cases, involving partial dissipation of the impact energy within the planet and radiative heat transfer througth the primordial atmosphere. The energy partition depended on the size of the impactors and time between the impacts: (1) with the increase of the sizes of impactors, the depth of energy release and time for energy transport to the surface increaesed as well (2) the increase of time between impacts enhanced the heat loss to the atmosphere.
- Almost all the gravitational energy in the formation of the core translated as heat energy and raised the Earth's temperature by about 3,000 K.
- During the Earth existence, the gravitational differentiation and decaying of long-lived radioactive elements has generated about 2.5 x 10^{31} J of heat energy accumulated inside the planet.

Thermal conditions during the early stages of the Earth's formation

- The uppermost layers of the Earth (up to a depth of ~1000 km) was heated to the melting point by the collision with large bodies between 4.5 and 3.8 Gyr.
- An impact initially deposits a large fraction of its kinetic energy in a region comparable in size to the projectile itself. The moving material generally opens a crater whose size and form depends on the relative importance of gravity and strength in the crater excavation.
- The major part of the kinetic energy of the impact is thus converted to heat in a roughly hemispherical region centered on the crater, which extends to a depth of only a few projectile diameters.
- The heat coming from short-lived isotopes could have created a surface temperature in the range 1200-1700 K.
- Due to these high temperatures, a magma-ocean covered the Earth from the Hadean to the beginning of the Archean (1-10 Myr to a few 100s Myr), with a surface temperature during its accretion likely in the range of 1,200–1,700 K.
- Felsic rocks were molten and could solidify when the temperature dropped below 1300 K.
- The survival time of the magma ocean depends on both the atmosphere and whether or not an insulating lid can develop. In the absence of these two effects, the lifetimes are very short, of order 10^3 years, while in presence of them the lifetime may be much longer, of order 10^8 years.

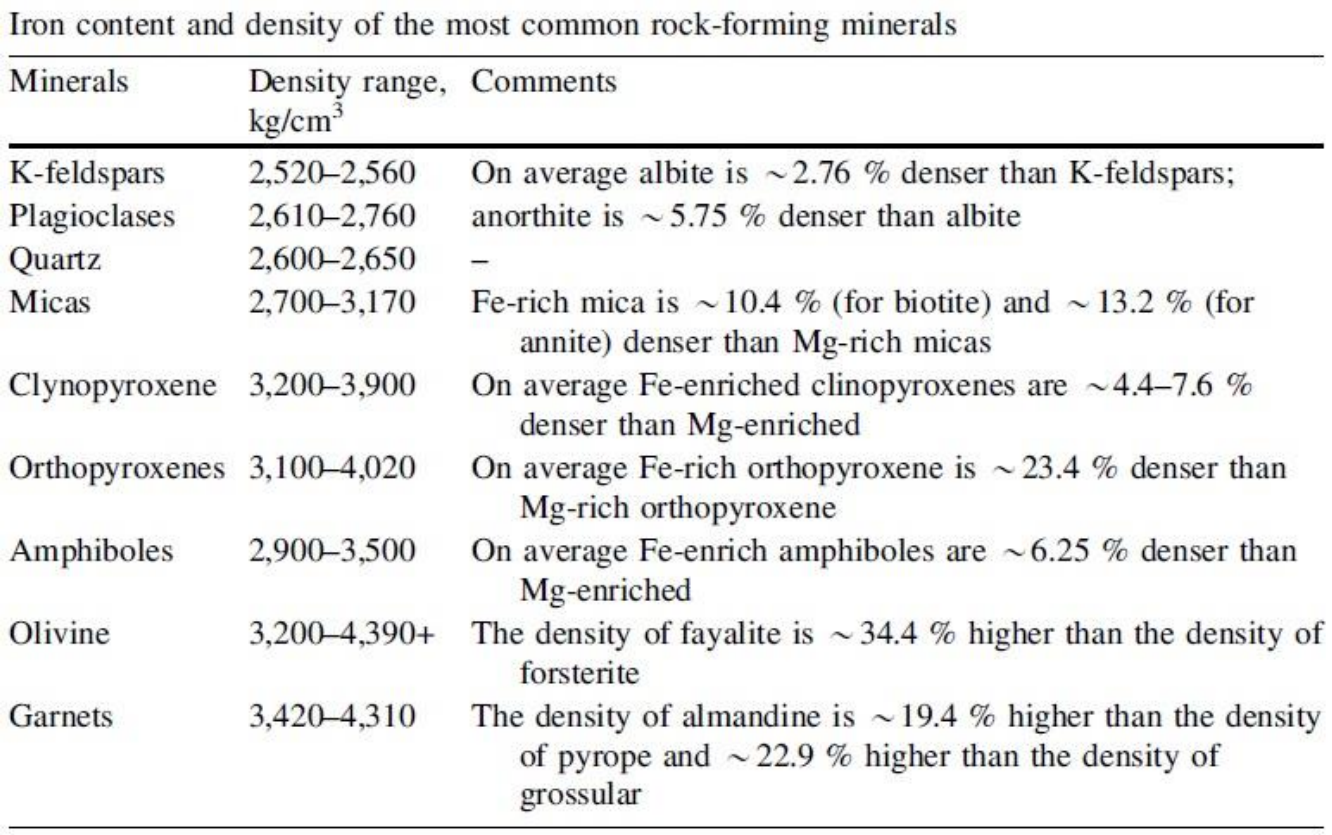


Estimates of the Earth's surface temperature in the early and middle Precambrian

Time (Ga)	Surface temperature (K)	References
4.4	<~473	Valley et al. (2002)
3.5–3.2	343 ± 15	Lowe and Tice (2004)
After 2.9	<333	Lowe and Tice (2004)
During the pre-Archean	573 ± 100	Valley et al. (2006)
Late Archean–early proterozoic	30 K less than now	Eyles and Young (1994)
3.5–3.2	328–358 (ocean water)	Knauth (2005), Knauth and Lowe (2003)
Most geological time	5 K higher than now	Kuhn et al. (1989)
Archean hydrosphere	343	Lambert (1982)
~0.2 after the Earth's formation	~373	Mackenzie (1998)
4.3	373	Morse and Mackenzie (1998)
3.8	343	Morse and Mackenzie (1998)
4.2	373	Mojzsis et al. (2001), Sleep et al. (2001)
Hadean	90 % of surface <0	Mukhin and Pimenov (2002)

Physical conditions during the early stages of the Earth's formation

- During formation of the magma-ocean, melts formed magmatic layers in ascending order of their density and iron content (uppermost layers were felsic and lowermost mafic).
- The density jumps between the different chemical layers was only about 30-100 kg/m^3.
- 2-3% of difference in density was sufficient to differentiate the magma-layers (no mixing, considering also the difference in composition).
- Difference in density between the current uppermost (3300 kgm^{-3}) and lowermost mantle (5500 kgm^{-3}) supports the hypothesis of the formation of different layers without interaction.

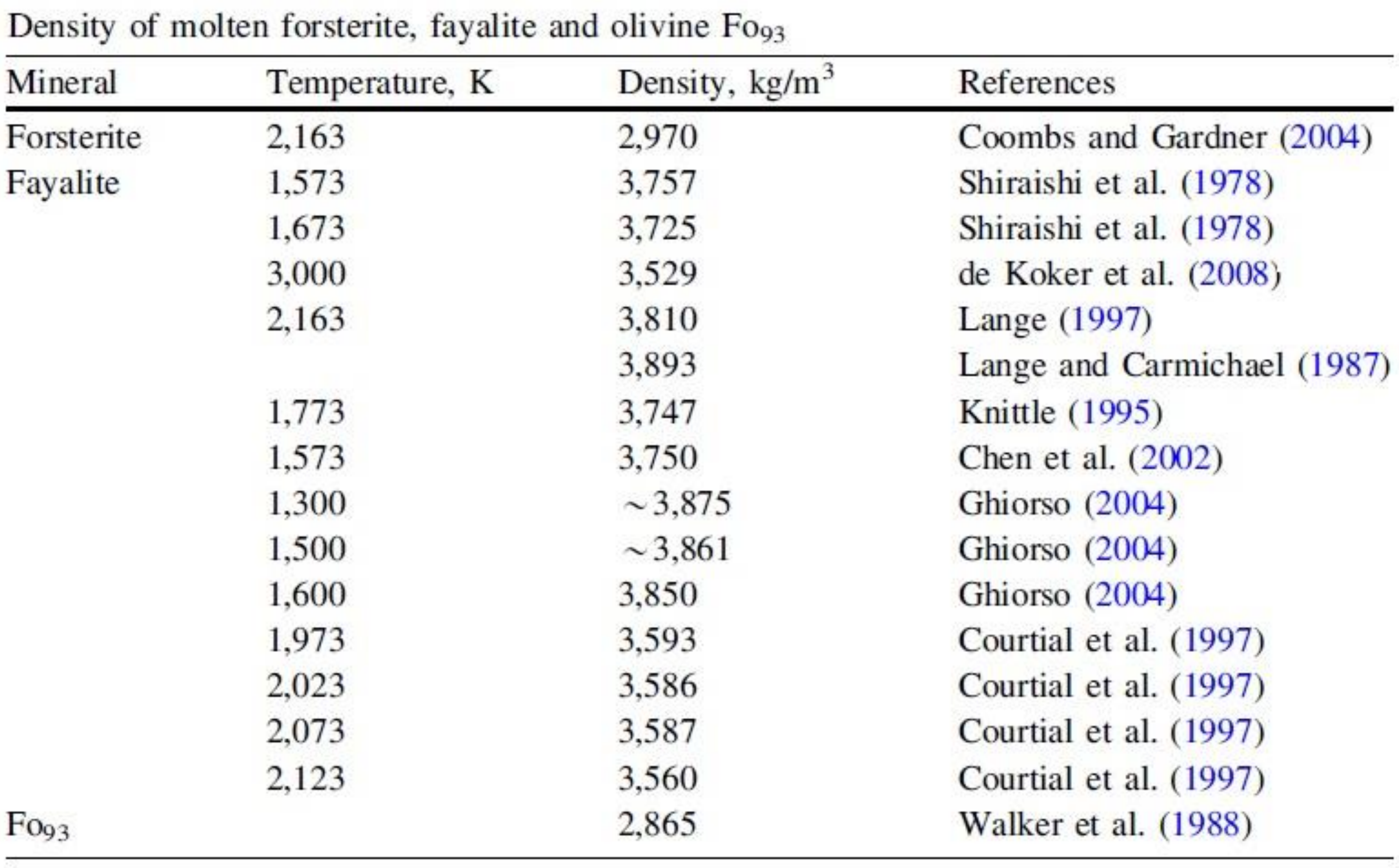


Iron content and density of the most common rock-forming minerals

Minerals	Density range, kg/cm^3	Comments
K-feldspars	2,520–2,560	On average albite is ~2.76 % denser than K-feldspars;
Plagioclases	2,610–2,760	anorthite is ~5.75 % denser than albite
Quartz	2,600–2,650	–
Micas	2,700–3,170	Fe-rich mica is ~10.4 % (for biotite) and ~13.2 % (for annite) denser than Mg-rich micas
Clynopyroxene	3,200–3,900	On average Fe-enriched clinopyroxenes are ~4.4–7.6 % denser than Mg-enriched
Orthopyroxenes	3,100–4,020	On average Fe-rich orthopyroxene is ~23.4 % denser than Mg-rich orthopyroxene
Amphiboles	2,900–3,500	On average Fe-enrich amphiboles are ~6.25 % denser than Mg-enriched
Olivine	3,200–4,390+	The density of fayalite is ~34.4 % higher than the density of forsterite
Garnets	3,420–4,310	The density of almandine is ~19.4 % higher than the density of pyrope and ~22.9 % higher than the density of grossular

Physical conditions during the early stages of the Earth’s formation

- Among the mafic and ultramafic magma-layers, the upper portions were formed by lither *Mg*-rich and the lower portions by *Fe*-rich magmas.
- When solidification of the magma ocean started the source of formation of magma increased.
- Nowadays, most of the basic and ultrabasic magmas should have formed under the lithosphere and the depth of the magma forming layers should increase permanently.

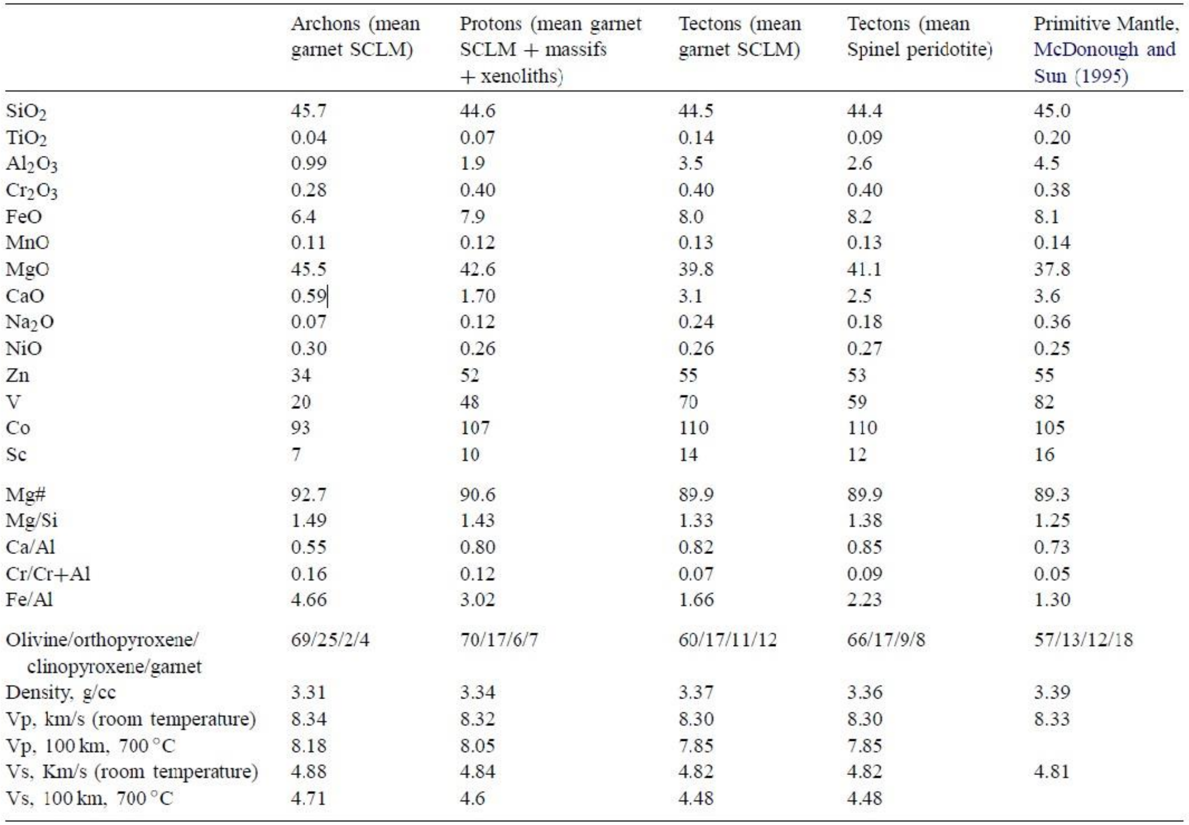


Density of molten forsterite, fayalite and olivine Fo_{93}

Mineral	Temperature, K	Density, kg/m^3	References
Forsterite	2,163	2,970	Coombs and Gardner (2004)
Fayalite	1,573	3,757	Shiraishi et al. (1978)
	1,673	3,725	Shiraishi et al. (1978)
	3,000	3,529	de Koker et al. (2008)
	2,163	3,810	Lange (1997)
		3,893	Lange and Carmichael (1987)
	1,773	3,747	Knittle (1995)
	1,573	3,750	Chen et al. (2002)
	1,300	~3,875	Ghiorso (2004)
	1,500	~3,861	Ghiorso (2004)
	1,600	3,850	Ghiorso (2004)
	1,973	3,593	Courtial et al. (1997)
	2,023	3,586	Courtial et al. (1997)
	2,073	3,587	Courtial et al. (1997)
	2,123	3,560	Courtial et al. (1997)
Fo_{93}		2,865	Walker et al. (1988)

Upper Mantle Composition *vs* Age

- The iron content in magmas increases systematically with decrease in their age and is minimal for the Archean rocks and maximal for rocks from the Cenozoic to the present.
- Cenozoic or younger continental basalts on average are almost all ferrobasalts (*Fe* = 12%).
- Oceanic basalts have a lower *Fe* content (8-12 %), indicating that these basalts formed at lower depth (shallower asthenosphere).
- Solidification of the magma-ocean very likely started with the formation of forsterite layer at a depth of about 100–120 km in areas of the current-day continental lithosphere.

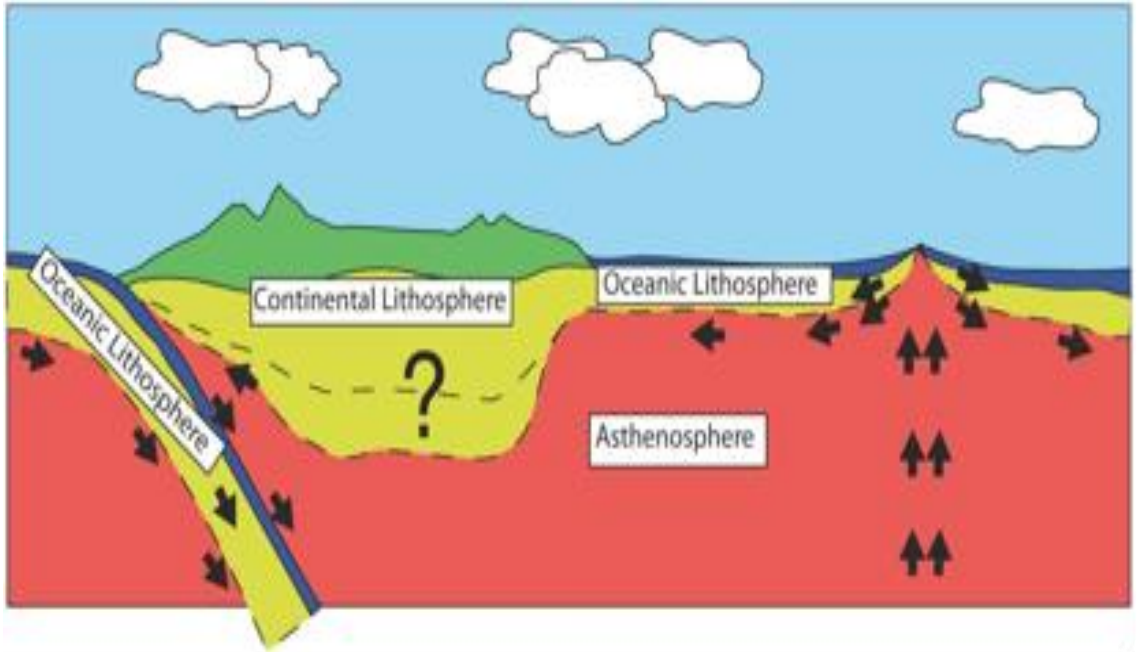


	Archons (mean garnet SCLM)	Protons (mean garnet SCLM + massifs + xenoliths)	Tectons (mean garnet SCLM)	Tectons (mean Spinel peridotite)	Primitive Mantle, McDonough and Sun (1995)
SiO_2	45.7	44.6	44.5	44.4	45.0
TiO_2	0.04	0.07	0.14	0.09	0.20
Al_2O_3	0.99	1.9	3.5	2.6	4.5
Cr_2O_3	0.28	0.40	0.40	0.40	0.38
FeO	6.4	7.9	8.0	8.2	8.1
MnO	0.11	0.12	0.13	0.13	0.14
MgO	45.5	42.6	39.8	41.1	37.8
CaO	0.59	1.70	3.1	2.5	3.6
Na_2O	0.07	0.12	0.24	0.18	0.36
NiO	0.30	0.26	0.26	0.27	0.25
Zn	34	52	55	53	55
V	20	48	70	59	82
Co	93	107	110	110	105
Sc	7	10	14	12	16
Mg#	92.7	90.6	89.9	89.9	89.3
Mg/Si	1.49	1.43	1.33	1.38	1.25
Ca/Al	0.55	0.80	0.82	0.85	0.73
Cr/Cr+Al	0.16	0.12	0.07	0.09	0.05
Fe/Al	4.66	3.02	1.66	2.23	1.30
Olivine/orthopyroxene/ clinopyroxene/garnet	69/25/2/4	70/17/6/7	60/17/11/12	66/17/9/8	57/13/12/18
Density, g/cc	3.31	3.34	3.37	3.36	3.39
Vp, km/s (room temperature)	8.34	8.32	8.30	8.30	8.33
Vp, 100 km, 700 °C	8.18	8.05	7.85	7.85	
Vs, Km/s (room temperature)	4.88	4.84	4.82	4.82	4.81
Vs, 100 km, 700 °C	4.71	4.6	4.48	4.48	

Griffin et al., 2003, Precambrian Research, 127

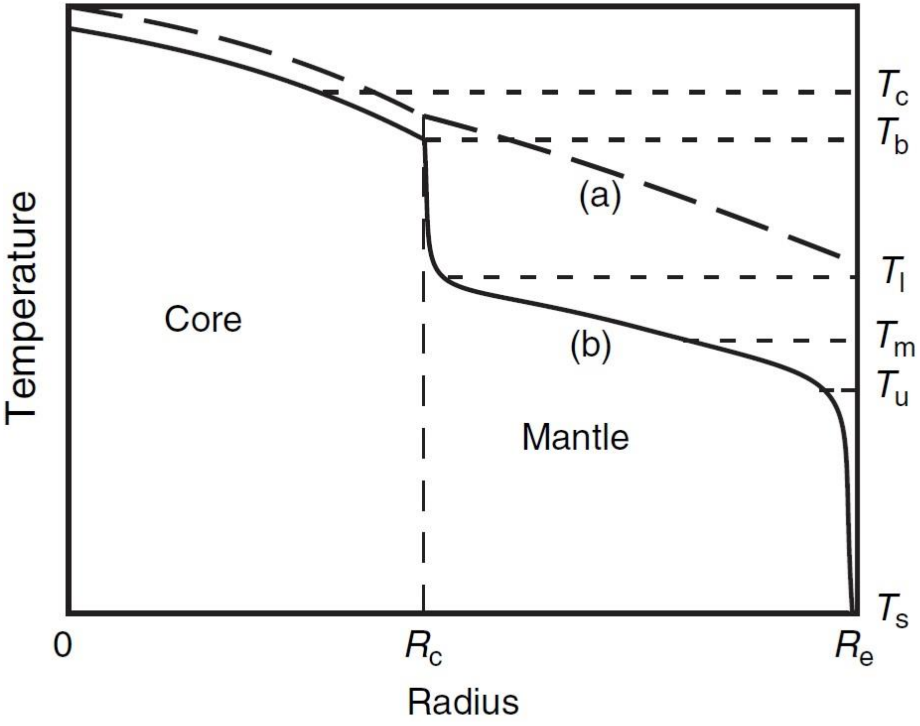
Thermal conditions of early lithosphere formation

- The cooling of the magma ocean was likely slow because numerous layers would only allow for separate convection within each layer.
- The cooling of the magma-ocean started with the formation of the fosterite layer (which has the highest melting point: 2163 and 2307 at 0.1 MPa and 3 GPa, respectively) at a depth of about 100 km (depth of xenoliths), still present under the cratons.
- The cooling effect of the thick early Earth's atmosphere caused solidification of the near surface magmatic layers, which would have increased the geothermal gradient, and consequently the surface heat flow, enhancing the rate of cooling of the near surface magmatic layers.
- Solid forsterite, having a low heat conduction coefficient, likely blocked the transfer of heat from the upper mantle below the forsterite layer to the surface, slowing down the cooling of the upper mantle below the forsterite layer and causing the bottom part of the present continental lithosphere to form with a significant delay.

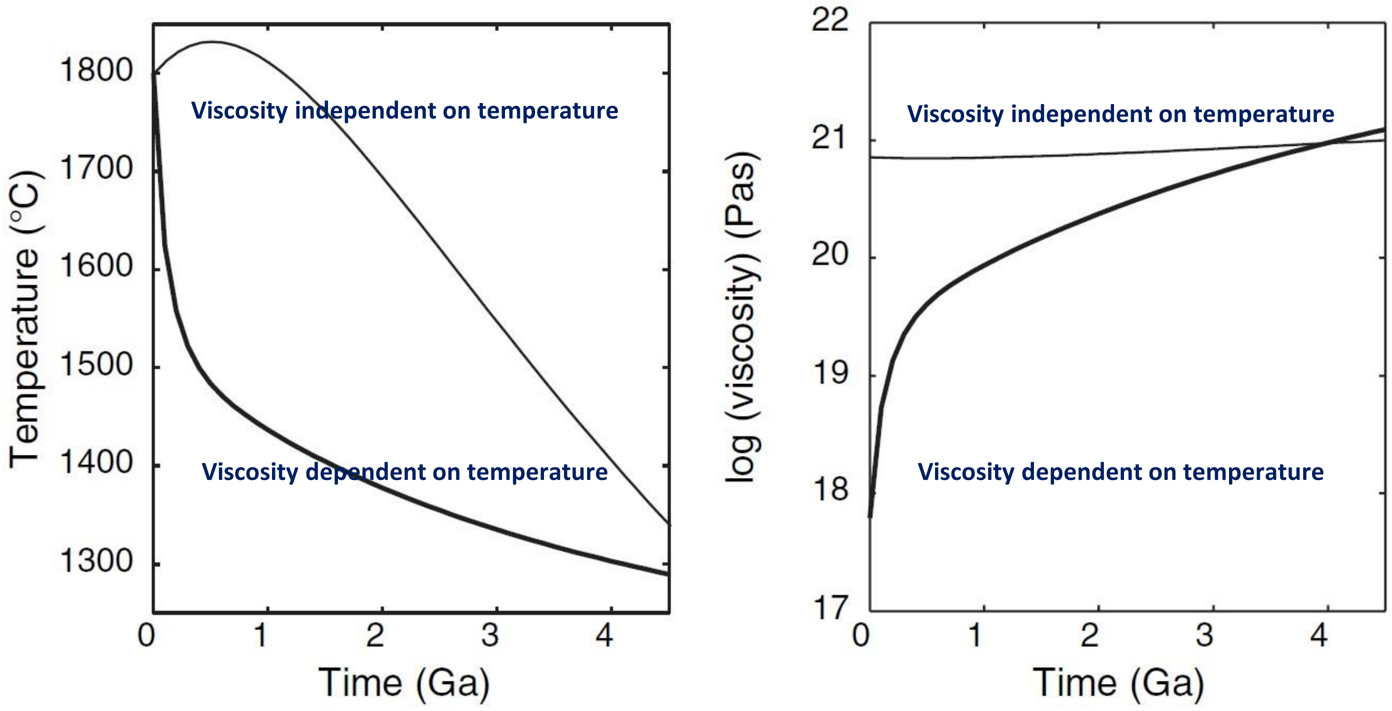


Present-day lithosphere conditions

Initial Earth’s thermal conditions



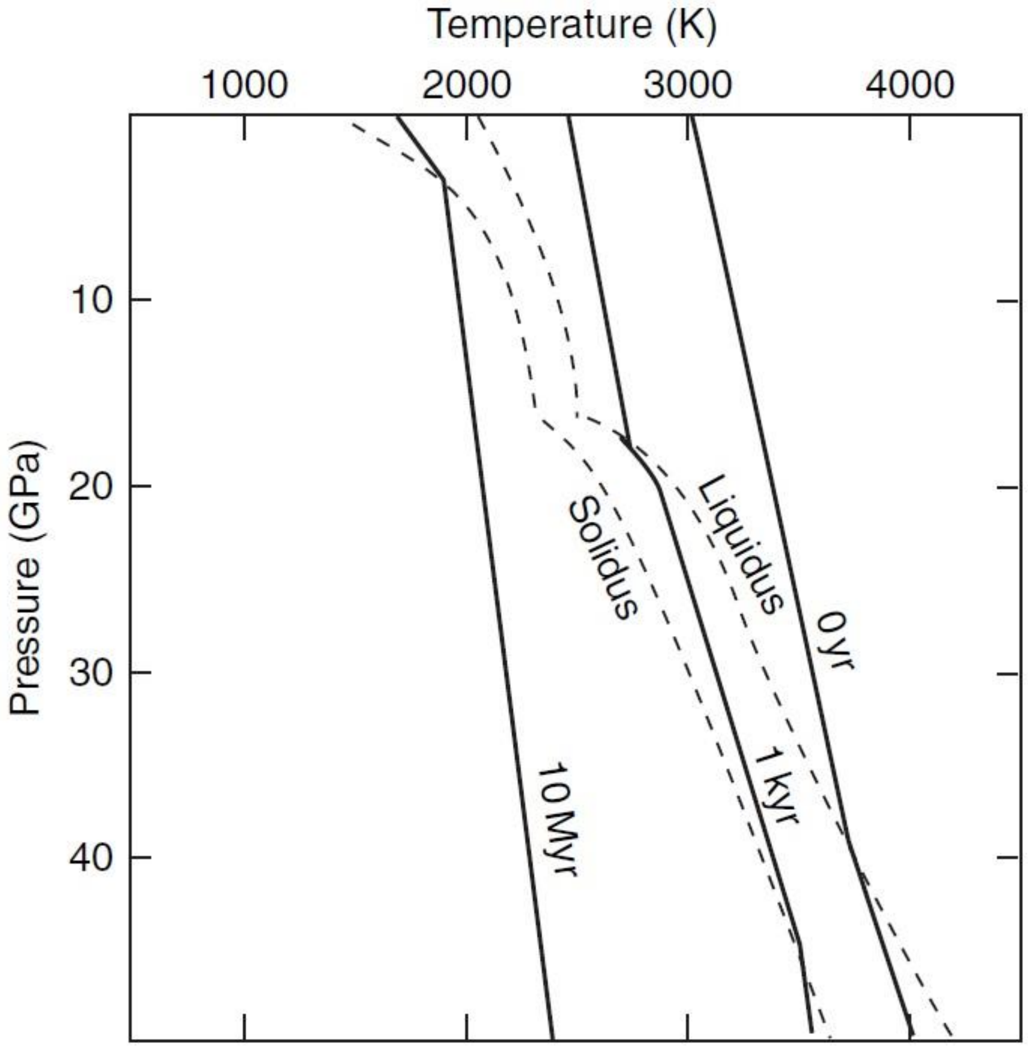
(a) Initial state. (b) Present state. Tc, mean core T; Tb, core–mantle boundary T; Tl, lower-mantle T; Tm, mean mantle T; Tu, upper-mantle temperature; Ts, surface T.



Davies, 2007, Treatise of Geophysics, vol. 9

- The peak temperature during the Earth’s formation is difficult to estimate, since it was presumably determined by a competition between heat deposition by large impacts and heat removal by the effect of further impacts, by conduction near the surface, by mantle convection, and by rapid cooling of surface melt (magma ocean).
- At the beginning the viscosity is much lower than at present, and this reduces the resistance to mantle convection, which can therefore remove heat very rapidly. This high heat loss causes the temperature to drop rapidly, but then the viscosity rises rapidly and reduces the heat loss.
- This early transient stage continues until the heat loss approaches the heat generation, at which point the initially large imbalance between them becomes small and thus the rate of decline of temperature slows.

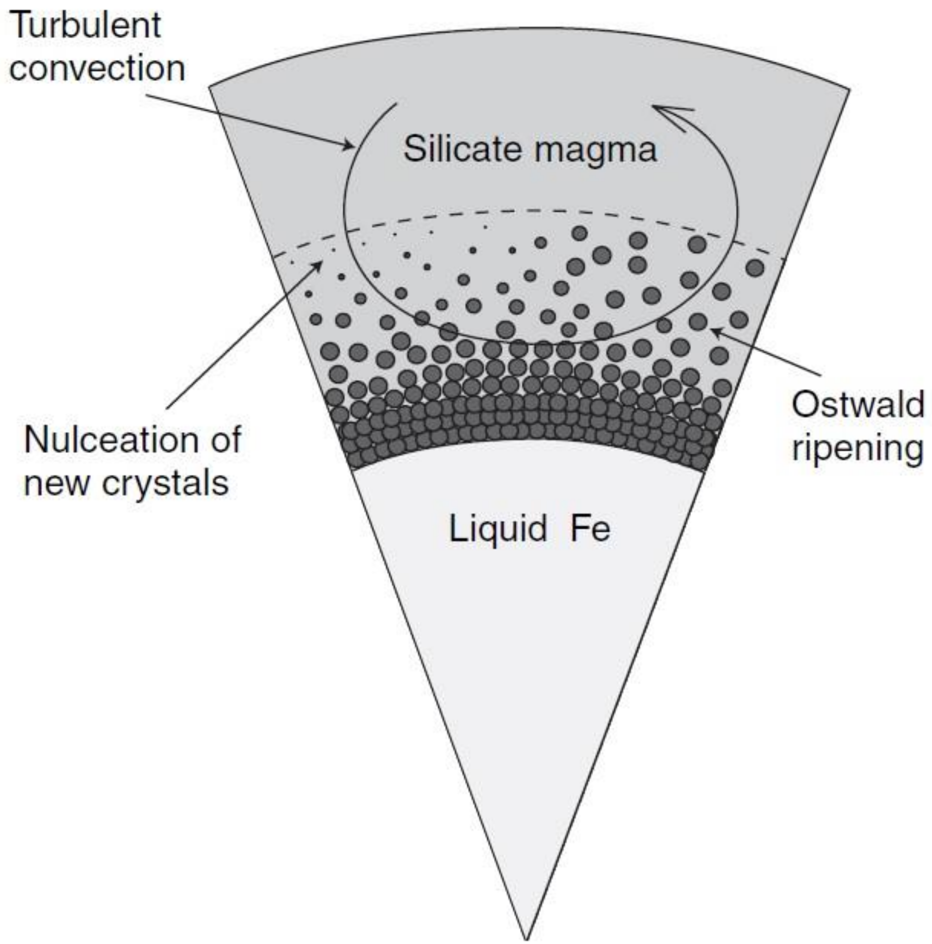
Initial Earth’s thermal conditions



Jaupart and Labrosse, 2007, Treatise of Geophysics, vol. 7

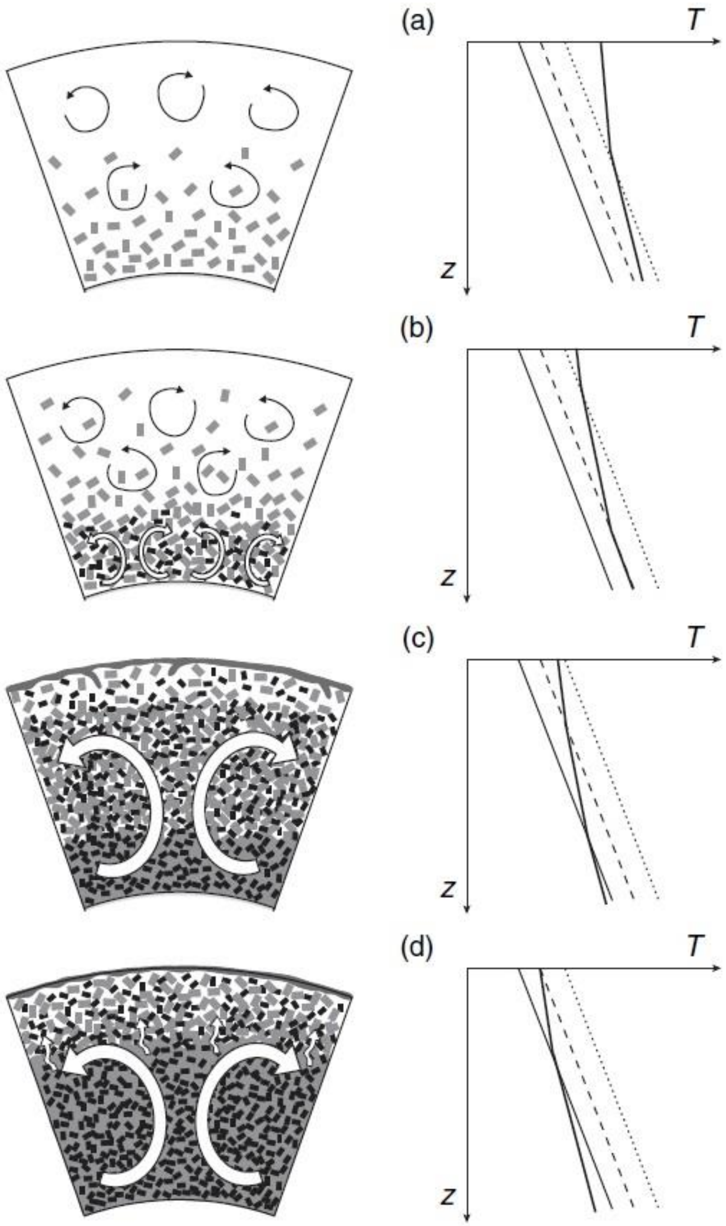
- At about 10 Myr, the solid content in the partially molten upper mantle layer reaches the threshold value of 60%, which marks the cessation of liquid behavior.
- After that time, convection is in the subsolidus regime controlled by solid behavior which still prevails today.

Crystallizzation of the magma ocean



Solomatov, 2007, Treatise of Geophysics, vol. 9

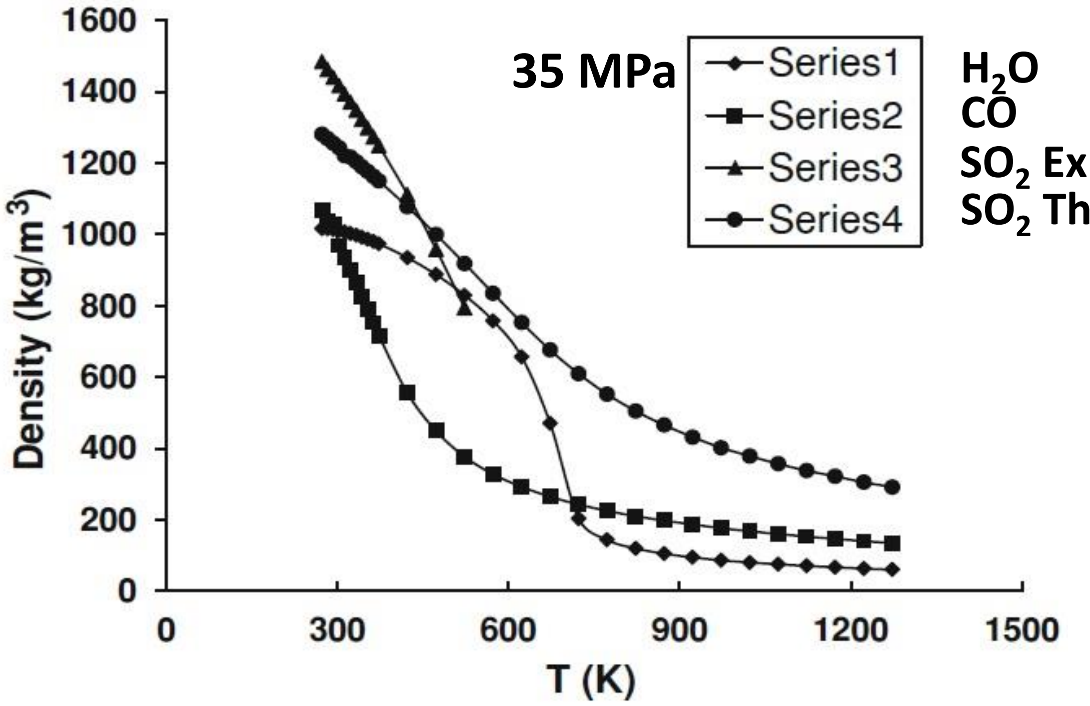
- The velocity controls the cooling rate of fluid parcels: the faster the cooling rate, the further the system is driven into the metastable state, the higher the nucleation rate and the more crystals are nucleated (faster cooling produces smaller crystals).



(a) The lower part of the magma ocean is below liquidus (dotted line), convection is controlled by melt viscosity and *T* (heavy solid line) is adiabatic;
(b) High viscosity, gravitationally unstable region with the maximum packing crystal fraction forms near the bottom of the magma ocean (dashed line);
(c) Cooling proceeds via solid state convection which is still fast at this stage. *T* in the high viscosity region below solidus (solid line) can be superadiabatic;
(d) The rate of cooling and crystallization of the remaining melt is controlled by solid-state convection and melt percolation. Convection becomes much slower, the heat flux drops, and a thick crust forms at the surface.

Early Earth’s Atmosphere

- During accretion, a magma-ocean with a surface temperature of >= 1300 K was formed, overlain by a thick, dense (>= 35 MPa) early terrestrial atmosphere composed of supercritical fluids with its main layers (sulfur-layer, carbon-layer, and water-layer) having different density.
- The early Earth’s atmosphere at a pressure of ~35 MPa was composed by the sulfur-layer (mostly SO_2) with a density of ~292 kg/m^3 and a thickness of 2.4 km, the carbon-layer (mostly CO) with a density of ~314.5 kg/m^3 and a thickness of 1.3 km, and the water layer with a density of 60.35 kg/m^3 and a thickness of 43.8 km.



H_2O
CO
SO_2 Experimental data
SO_2 Thermodynamic methods

Density of water decreases drastically to 14.4% at T=773 K and more slowly to a maximum of ~ 6% of the initial value.

Early Earth’s Atmosphere

- The solidified forsterite layer would have separated the magma-ocean into two parts with the lower *Fe*-rich magmas and the upper *Fe*-poor magmas below and above the forsterite layer, respectively.
- The felsic magma layer with the lowest melting point and cooled by the early Earth’s atmosphere started to solidify first within the upper part of the magma-ocean.
- At *T* between 1253 K and 673 K, $2CO=CO_2+C$ (graphite formation). With the cooling also transfer of SO_2 to SO_3 occurred, likely forming individual sub-layers: SO_3 10% at 1273 K and 100% at 673 K.
- Between 673 K and 550 K, H_2O and SO_3 layers came in contact forming H_2SO_4 (it starts the leaching of the Earth's surface).

T and P	$CO-CO_2$ (ρ and Th.)	SO_2-SO_3 (ρ and Th.)	H_2O (ρ and Th.)
1273 K 35 MPa	~314.5 kg/m^3 ~1.3 km (*B*)	~292 kg/m^3 2.4 km (*M*)	~60.35 kg/m^3 43.8 km (*T*)
1173 K 35 MPa	~341 kg/m^3 ~1.2 km (*B*)	~322 kg/m^3 ~2.2 km (*M*)	~75.3 kg/m^3 ~39.5 km (*T*)
823 K 35 MPa	~210 kg/m^3 ~1.0 km (*M*)	~505 kg/m^3 ~1.4 km (*B*)	~119.9 kg/m^3 ~22.1 km (*T*)
673 K 35 MPa	~265 kg/m^3 ~0.6 km (*T*)	~676 kg/m^3 ~1.1 km (*B*)	~470.1 kg/m^3 ~5.6 km (*M*)
573 K 35 MPa	~327.9 kg/m^3 ~0.8 km (*T*)	~835 kg/m^3 ~0.9 km (*B*)	~758.1 kg/m^3 ~3.5 km (*M*)
523 K 35 MPa	~377.3kg/m^3 ~0.5 km (*T*)	~ 794.4 kg/m^3 ~0.9 km (*M*)	~829.7 kg/m^3 ~3.2 km (*B*)

T=top, *M*=middle, *B*=bottom

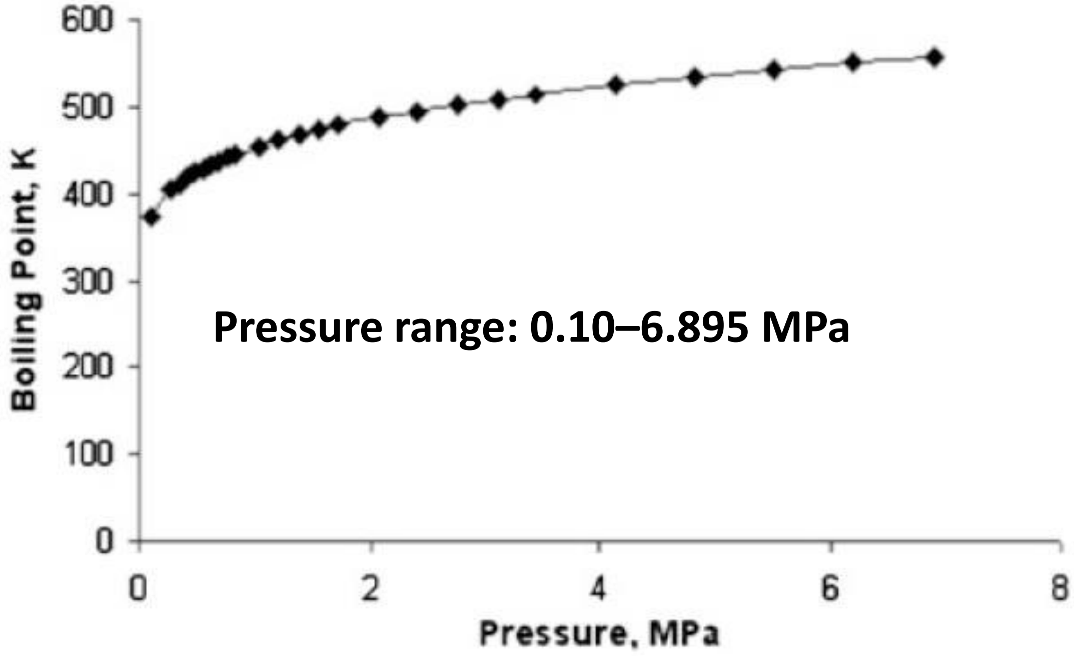
Age of crust formation and weathering

- The existence of ferric iron oxides formation (banded iron formation – BIF), leading to magnetite generation, starts when the temperature surface drops below 723-843 K (10 MPa-0.1 MPa), between 3.85 and 3.50 Gyr, and continues up to ~473-523 K at the corresponding age of 1.8 Gyr.

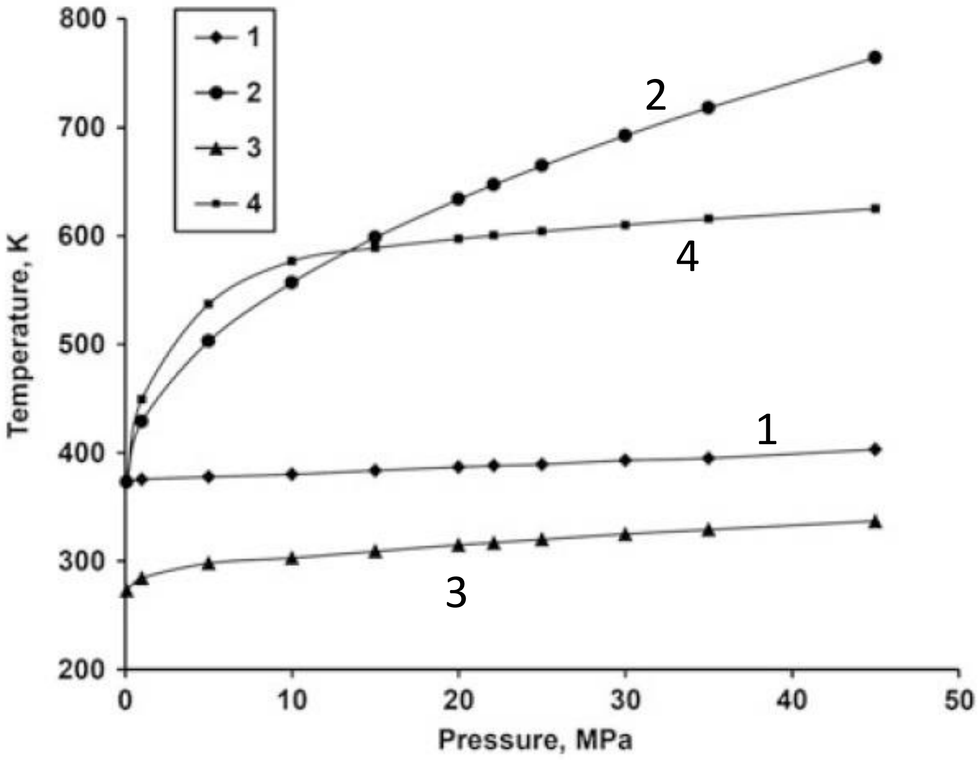
$$4FeO \longrightarrow Fe_3O_4 + Fe$$

- Oxidation processes started after 2.3 Gyr (when free oxygen was available) and thus didn't play a role in the deposition of BIFs during early Archean.
- Massive weathering of rocks started with the formation of H_2SO_4 at temperature < 700 K, when the H_2O and SO_3 layer became in contact and the sulfuric acid was liquid (its boiling point < 700 K at pressure > 35 MPa and at high concentration).
- Many elements are resistant to high concentration of sulfuric acid (e.g., Hf for concentrations up to 80%, Zr up to 70-75%, Pb up to 90%, Fe up to 75%, SiO_2 is insoluble).
- Dissolution of the oxides of albite (Na_2O, Al_2O_3, SiO_2) raises significantly (from 6.4 ppm to 318 ppm) with increase in *T* (373-623 K) and *P* (0.3-35 MPa).
- Hydrothermal alteration under acid conditions occurred ~3.4 Gyr and lasted till 3.2 Gyr: no crystallization of hydrothermal barite in veins, as a product of acid alteration between 3.2 and 1.8 Gyr (re-distribution of the sulfur-layer of the atmosphere with formation of sulfides and barites by 3.2 Gyr, before the water layer accesses to the surface).
- From 2.8 to 2.7 Gyr there was an intense magmatic activity (komatiite magmas), which may have caused the decomposition of sulfides which came in contact.
- The cooling of the Earth's surface became more efficient with the formation of a water-ocean, which depended on a *P* and *T* conditions.

Boiling Point of Water with P and T variations



Pressure range: 0.10–6.895 MPa



(1) P–T conditions for a constant density of water at 958.36 kgm^{-3}; **(2)** interpolation-extrapolation model of the variable boiling point of water with increases in pressure using as initial and boundary conditions the boiling point of water at 373.16 K at 0.1 MPa and the critical boiling point of water *T* = 647.36 K at *P* = 22.12 MPa; **(3)** P–T conditions of constant density of water at 1000 kgm^{-3}; **(4)** extrapolation model of the variable boiling point of water with increases in *P* using data from Speight (2005).

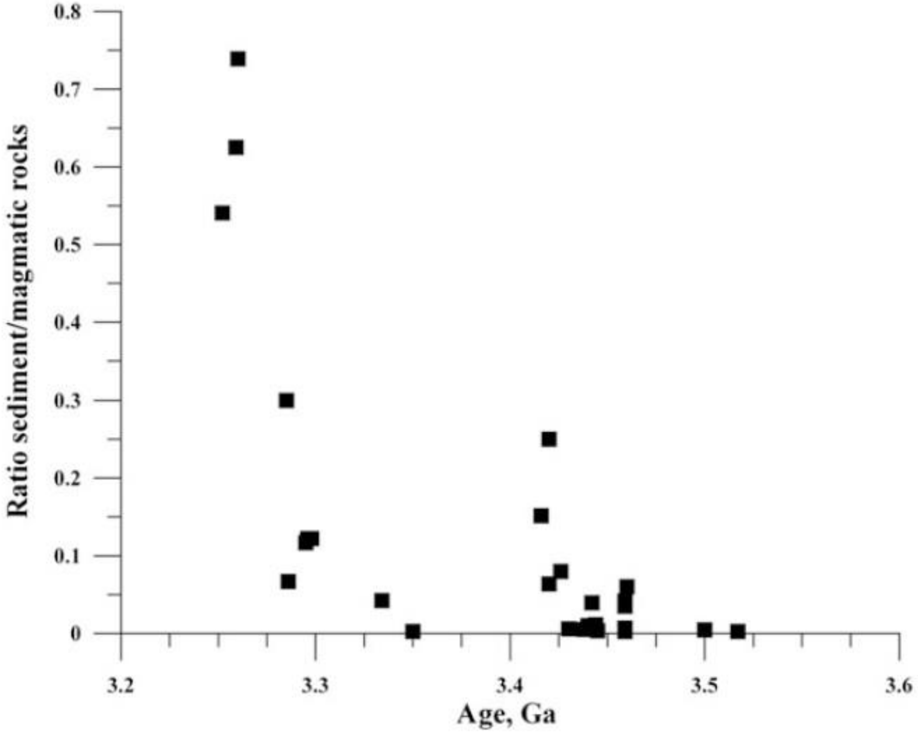
$$T_B - T_{B0} = A\sqrt{(P - P_0)}$$

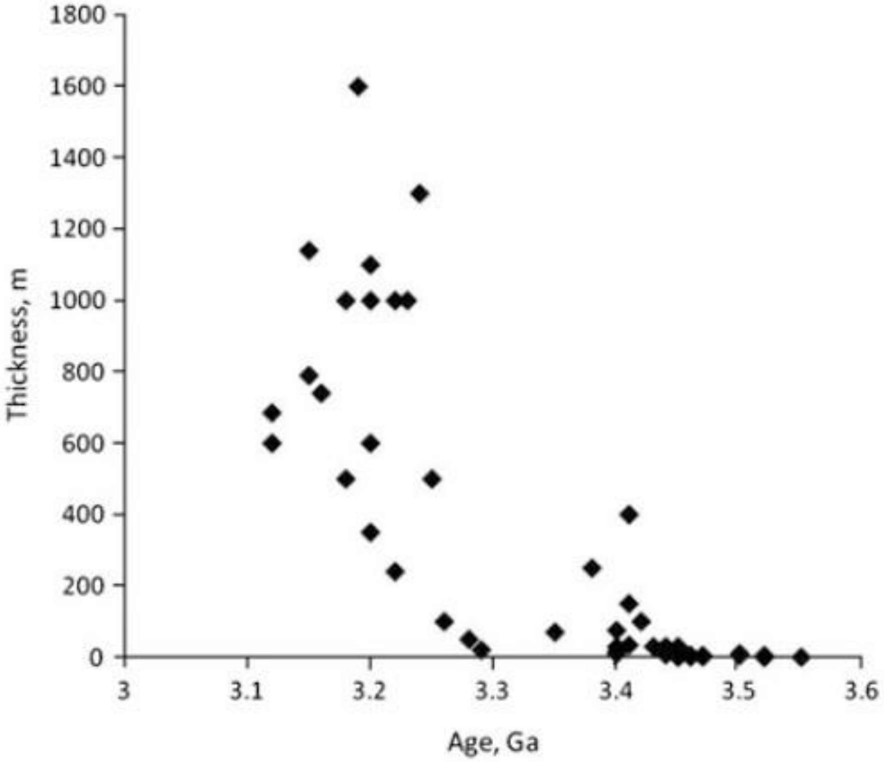
T_B and T_{B0} = boiling points of water at pressure P and at pressure P_0
A = constant defined by initial and boundary conditions

- **Formation of liquid water from vapor under a *P* of ~35 MPa, *T* of the Earth's surface and the atmosphere had to be below 615.5–717.9 K.**

Water-ocean layer: when did it form?

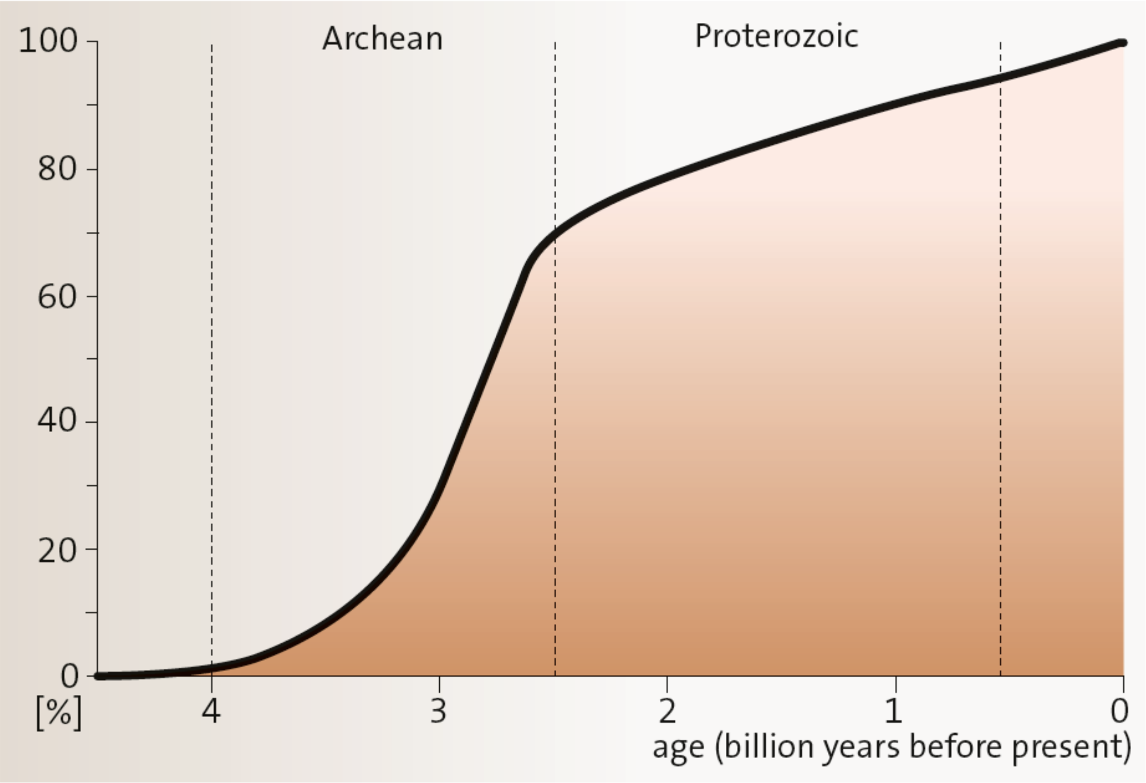
- Markers of the existence of a water-ocean are the thick deposits of sediments (carbonates and evaporates) and ophiolites.
- Carbonates are rare prior to ~2.9 Gyr and are not of sedimentary origin (metasomatic, 3.71-3.85 Gyr). They started to increase at the end of Early Archean, likely because of the absence of a water-ocean layer before this time or of the scarce solubility of CO_2 in the hot water.
- Carbonates form in a range of *T* of the ocean between 473-673 K, reached in the early Archean: Serpentinized komatiite of age ~3.4-3.25 Gyr have been found.
- Significant amount of sediments started to form at 3.26 Gyr (at the end of the Early Archean). Before this time, absence of carbonates and evaporites, but presence of barites formations (between 3.46-3.55 and 3.2 Gyr): water- and sulfur-layer in contact.

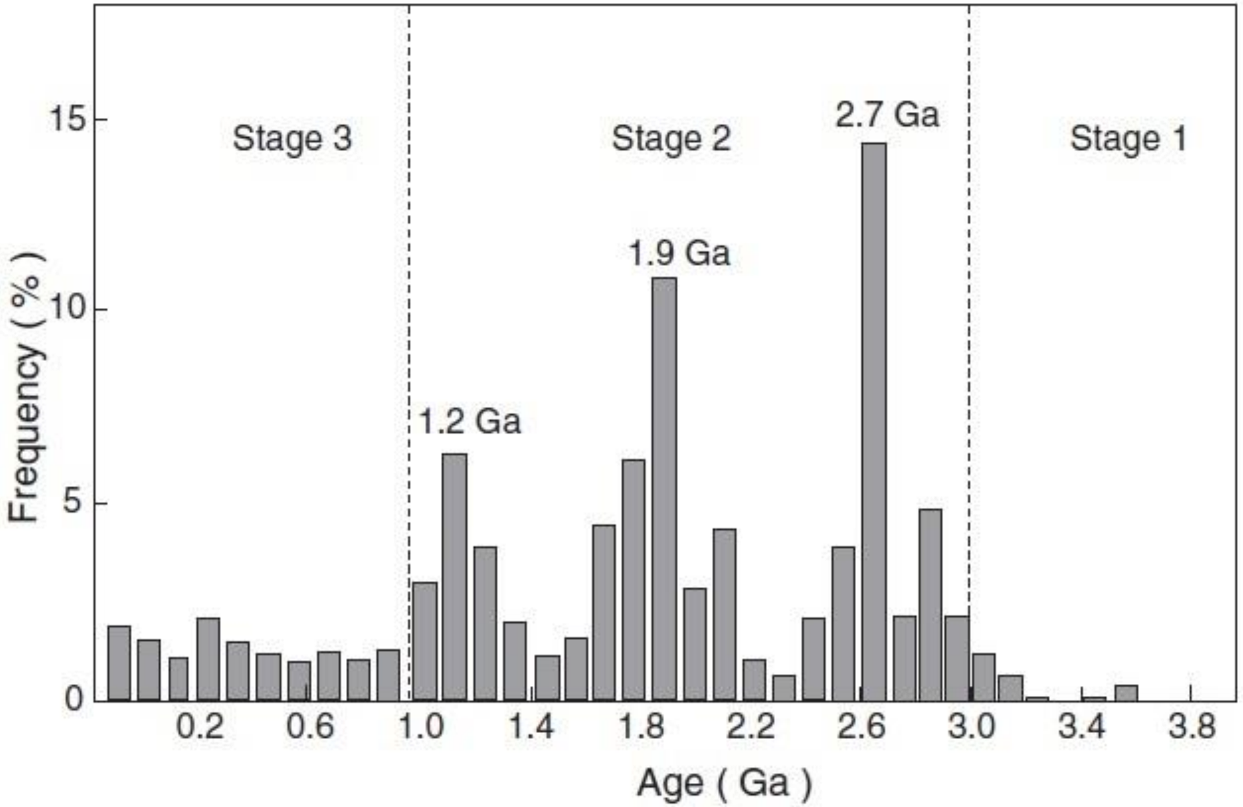




Crustal Growth

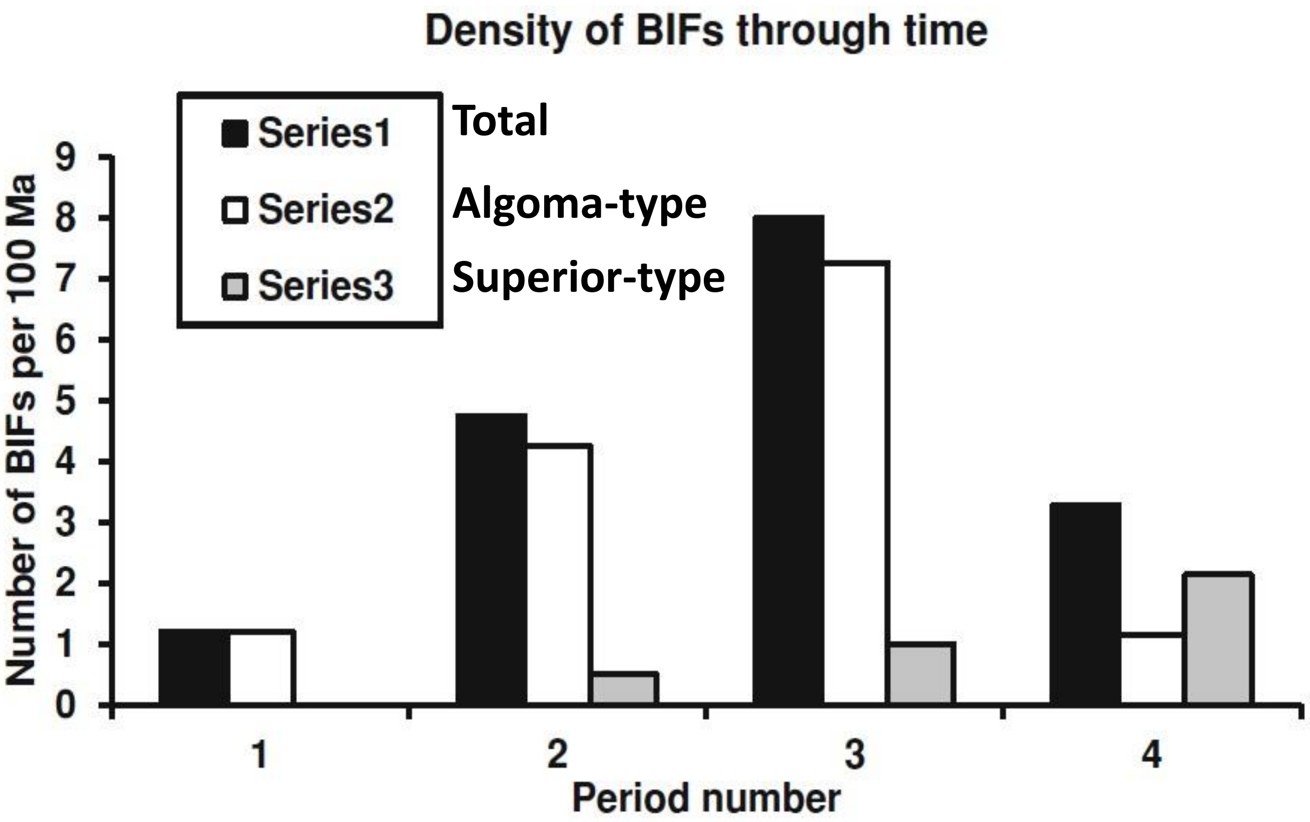
- Enormous crustal growth occurred in the second half of the Archean era (before only small, unstable continents), between 3.2 and 2.5 Gyr. This is due to the considerably more efficient production of rocks constituting continental crust above subduction zones since ca. 3.2 Gyr.
- Measurements of Nb/Th and Nb/U ratios could define the net production rate of continental crust since 3.8 Gyr (the different ratios potentially provide information on the extent of the chemical depletion and the amount of continental crust that was present on Earth at different times).
- These results and on those of isotopic age determinations suggest that crust production was episodic with rapid net growth at 2.7, 1.9, and 1.2 Gyr (with pulses of ≤ 100 Myr) and slower growth afterwards.
- Previous studies suggested that: 39% of the continental crust formed in the Archean, 31% in the Early Proterozoic, 12% in the Middle–Late Proterozoic, and 18% in the Phanerozoic.





Markers of lithospheric thermal conditions

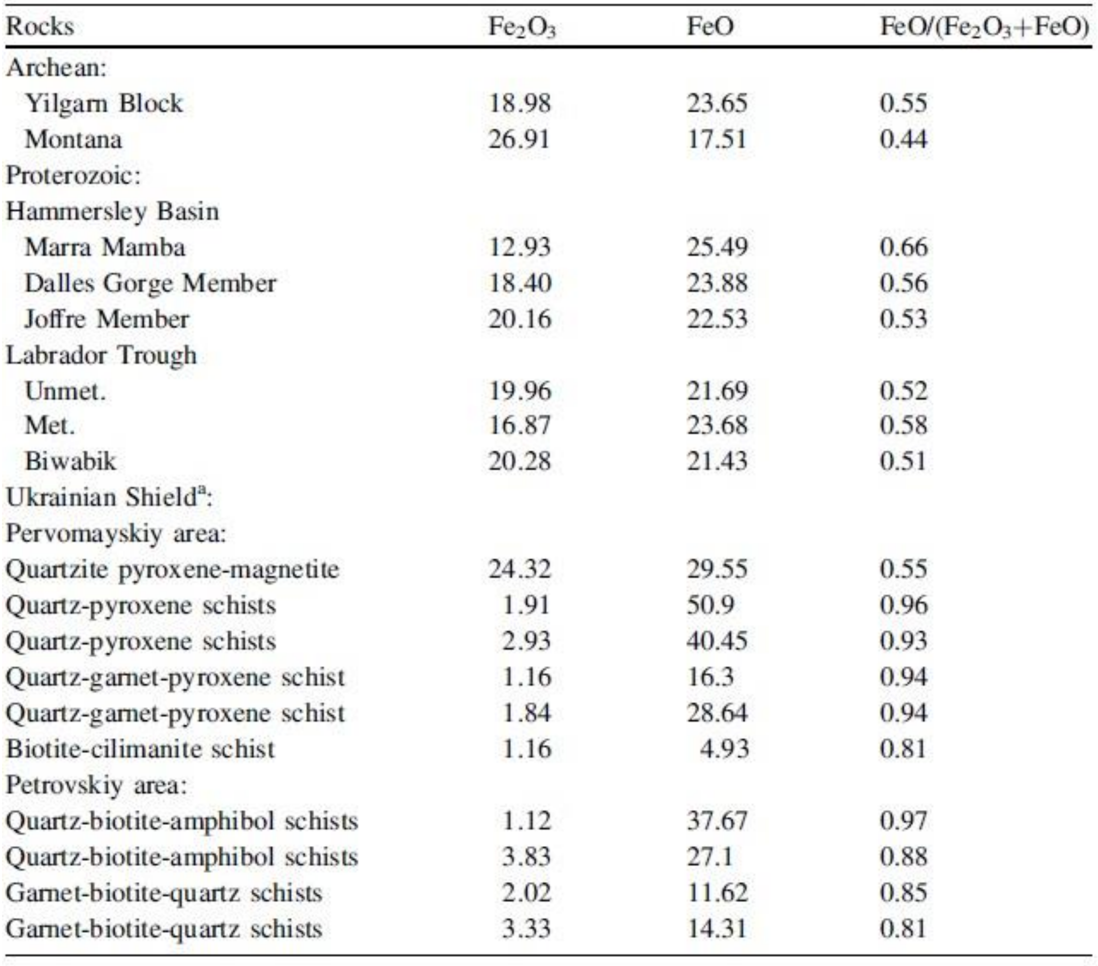
- Processes as the deposition of BIFs, and the formation of pyrite deposits are markers of anoxic atmospheric environments, while the formation of paleosols and red beds take place in oxygen-rich environments.
- Transformation of ferric iron to ferrous iron (TFFI) occurs with an increase of temperatures from low to high values and viceversa, since ferrous iron is stable at HP and HT and ferric iron is stable at LP and LT.



1 Early Archean, 2 Middle Archean, 3 Late Archean, and 4 Early Proterozoic

Ferrous and Ferric Iron distribution in Precambrian BIFs

- The coefficient of stability of iron(II) oxide ($\nu = FeO/(FeO+Fe_2O_3)$) is in the range of 0.44–0.55 and 0.51–0.66, for the Archean and Early Proterozoic BIFs, respectively, likely due to the reduction in temperatures to the zone of iron(II) stability or the rapid cooling of forming deposits.
- The excess of iron(II) oxide in the Precambrian metamorphic rocks may be related to the formation and stability of hematite, which can occur only in the presence of both water and oxygen.



Rocks	Fe_2O_3	FeO	$FeO/(Fe_2O_3+FeO)$
Archean:			
Yilgarn Block	18.98	23.65	0.55
Montana	26.91	17.51	0.44
Proterozoic:			
Hammersley Basin			
Marra Mamba	12.93	25.49	0.66
Dalles Gorge Member	18.40	23.88	0.56
Joffre Member	20.16	22.53	0.53
Labrador Trough			
Unmet.	19.96	21.69	0.52
Met.	16.87	23.68	0.58
Biwabik	20.28	21.43	0.51
Ukrainian Shield[a]:			
Pervomayskiy area:			
Quartzite pyroxene-magnetite	24.32	29.55	0.55
Quartz-pyroxene schists	1.91	50.9	0.96
Quartz-pyroxene schists	2.93	40.45	0.93
Quartz-garnet-pyroxene schist	1.16	16.3	0.94
Quartz-garnet-pyroxene schist	1.84	28.64	0.94
Biotite-cilimanite schist	1.16	4.93	0.81
Petrovskiy area:			
Quartz-biotite-amphibol schists	1.12	37.67	0.97
Quartz-biotite-amphibol schists	3.83	27.1	0.88
Garnet-biotite-quartz schists	2.02	11.62	0.85
Garnet-biotite-quartz schists	3.33	14.31	0.81

- **The interaction between the CO-layer and the SO_2-layer caused several chemical reactions resulting in the formation of such important reducing agents as elemental carbon and sulfur at *T* of 873-973K:**

$$2CO + SO_2 \rightarrow CO_2 + S^0$$

$$SO_2 + C \rightarrow CO_2 + S^0$$

- **C, CO_2, S, and SO_2 caused the reduction of hematite to magnetite at *T* of 723-843K and fast cooling during Earth's evolution could preserve iron(II):**

$$3Fe_2O_3 + C \rightarrow 2Fe_3O_4 + CO$$

$$3Fe_2O_3 + CO \rightarrow 2Fe_3O_4 + CO_2$$

$$6Fe_2O_3 + 2SO_2 \rightarrow 4Fe_3O_4 + 2SO_3$$

Paleosols and Red beds during Precambrian age

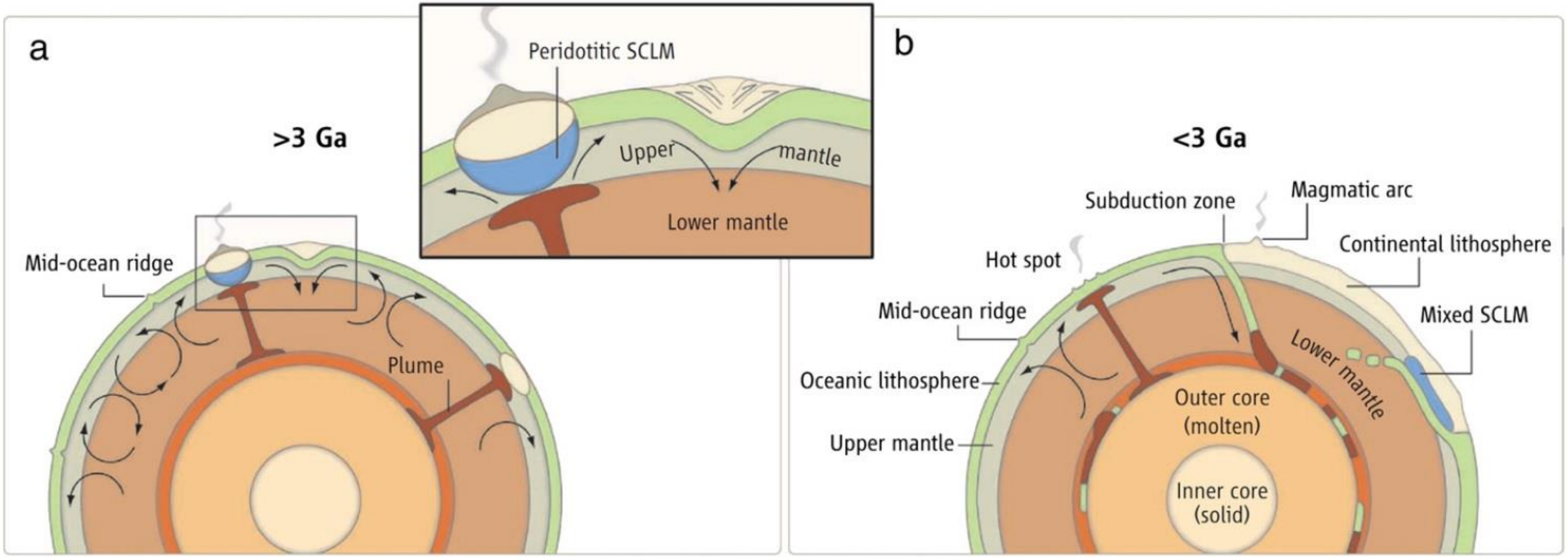
Chemical profiles of the paleosols are useful to mark the the transformation of the Earth's atmosphere from an anoxic to an oxic environment:

- If there was enough oxygen in the atmosphere, ferrous iron(II) would have oxidized to ferric iron(III) during weathering and the iron (III) within the paleosol would not be soluble or mobile, since only iron(II)-compounds are soluble in water.
- Pre-2.44 Ga paleosols suffered significant iron loss during weathering (anoxic environment), whereas the iron loss was negligible during their formation of the 2.2–2.0 Gyr and after (oxic environment).

Red beds, detrital sedimentary rocks rich in reddish-brown ferric oxide, were absent before 2.0-2.2 Gyr (anoxic environment, no direct contact between oxygen and surface).

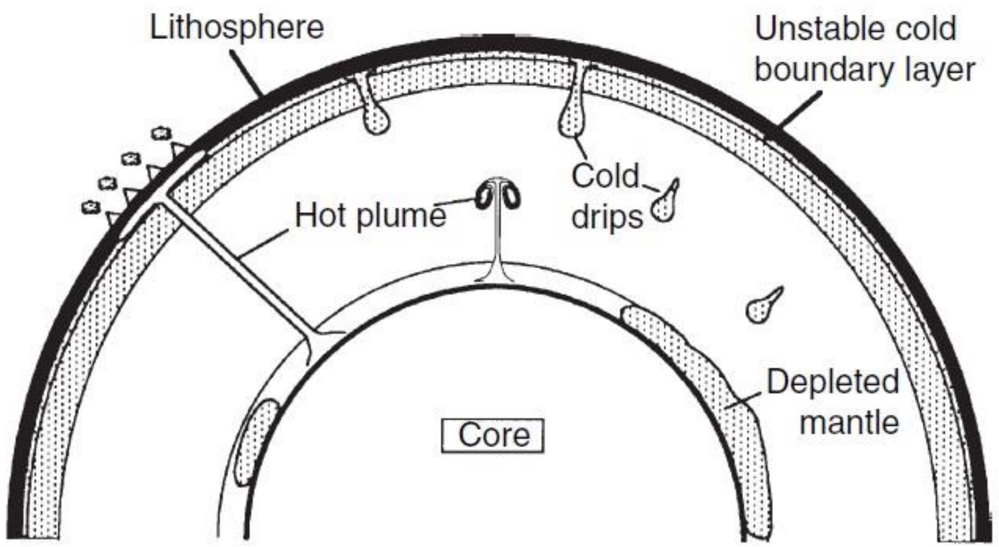
Precambrian Geodynamics

- Archean geodynamics was dominated by plume tectonics and the development of hot accretionary orogens with low topography.
- Due to the hot mantle temperature, slab break-off was more frequent in the Precambrian time and limited occurrence of ultrahigh-pressure (UHP) rocks. Mantle downwellings and slab break-off processes are likely to have played a key role in assembling and stabilizing the hot orogens.
- Both oceanic and continental lithospheres were rheologically weak due to the high temperatures.
- Numerical models suggest that the long-term stability of cratons sustaining multiple supercontinent cycles can be achieved if their viscosity and yield strength are sufficiently high and weak mobile belts are present along the boundaries of the cratons.
- Stable cratons facilitate subduction initiation of very young seafloor during continental growth and dispersal.
- Wide spread development of modern-style (cold) collision on Earth started during Neoproterozoic at 600–800 Myr. Cold collision created favorable conditions for the generation of UHP metamorphic complexes, which become widespread in Phanerozoic orogens.

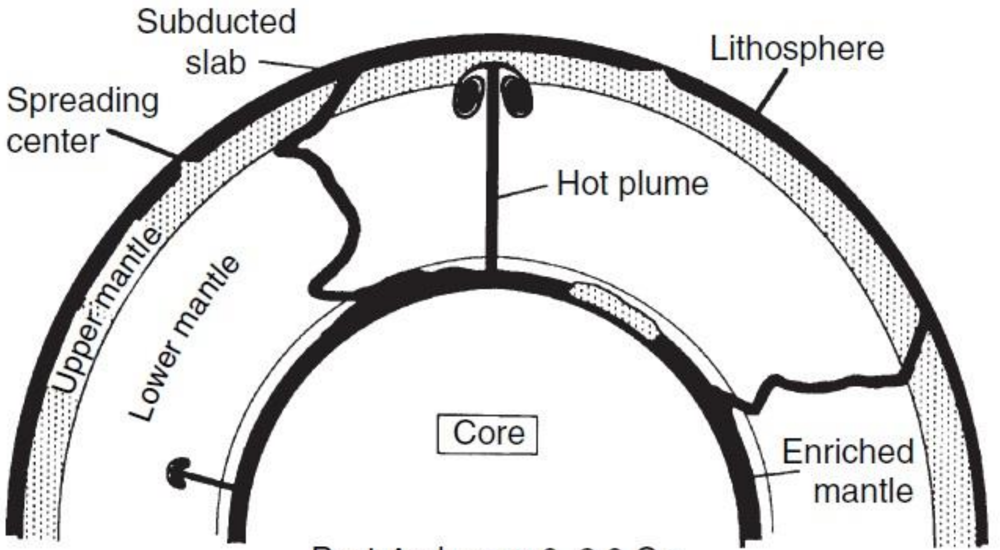


Gerya, 2014, Gondwana Research, 25

Precambrian Geodynamics



Pre-Archaean 4.0–4.5 Ga



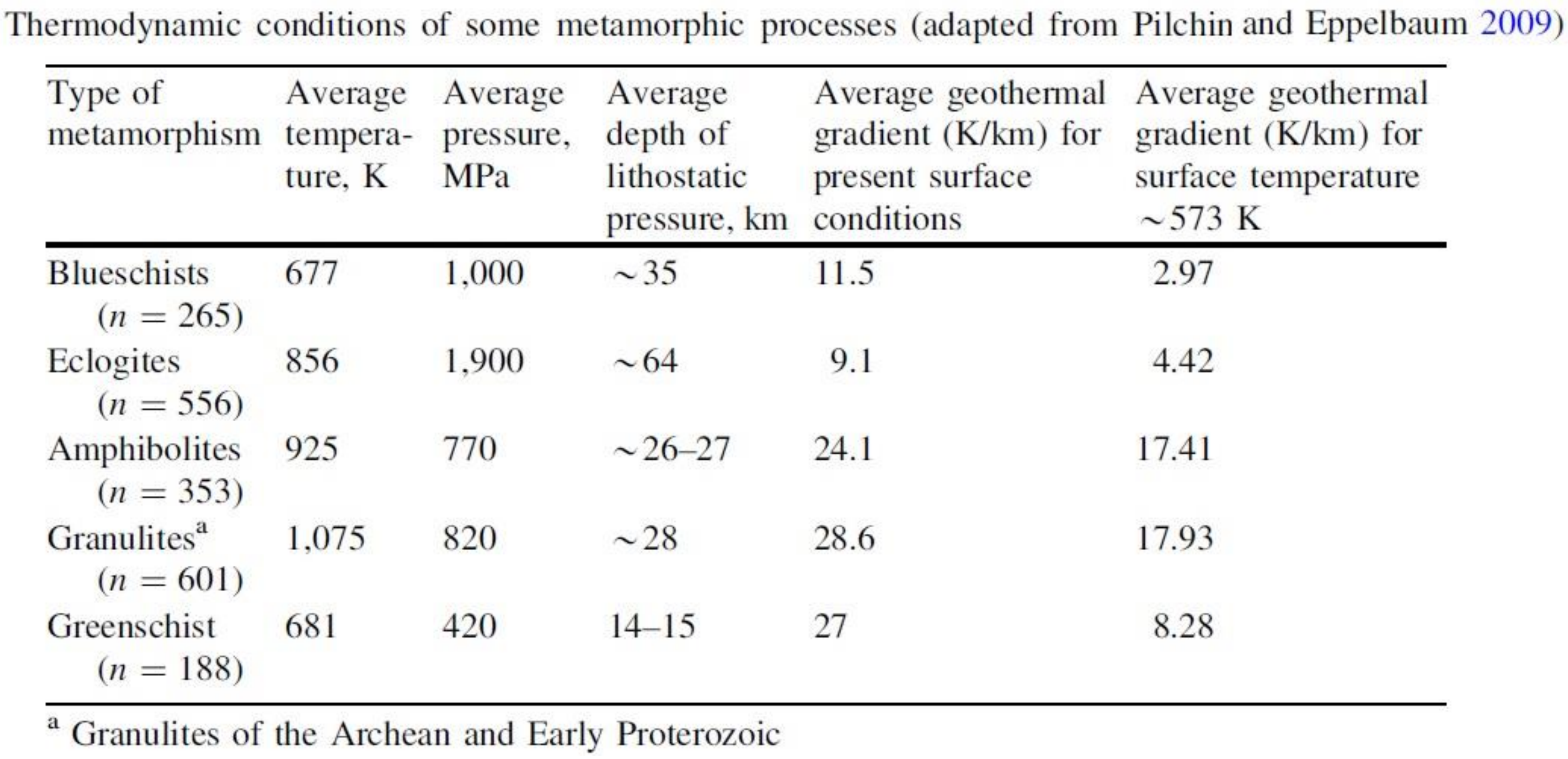
Post-Archaean 0–2.0 Ga

- During the pre-Archean or Hadean (4.5–4 Ga), the mafic crust was too buoyant to founder, and only the underlying mantle part of the thermal boundary layer (already cold) foundered.
- The post-Archean change in chemistry is attributed to the replacement of the early depleted D'' layer by enriched subducted mafic crust.
- The higher temperatures of the plume during the Archean, may be due to the fact that the earlier phase the D" layer covered only part of the core, leaving hot core directly in contact with mantle elsewhere and thus generating very hot plumes.

Davies, 2007, Treatise of geophysics, vol. 9

Metamorphic rocks of Precambrian

- The age of metamorphic rocks helps define the time period in the Earth's evolution when such thermal conditions took place at the specific depth in the regions where these rocks are located.
- Among metamorphic rocks, **greenschist rocks (MP and MT)** were the earliest to appear followed by **granulites (MP-HP and HT)** and **amphibolites (MP and MT-HT)**, while the oldest **eclogites (HP-UHP and HT)** appear only in the Early Proterozoic and **blueschists (MP-HP and LT)** only in the middle-late Proterozoic.



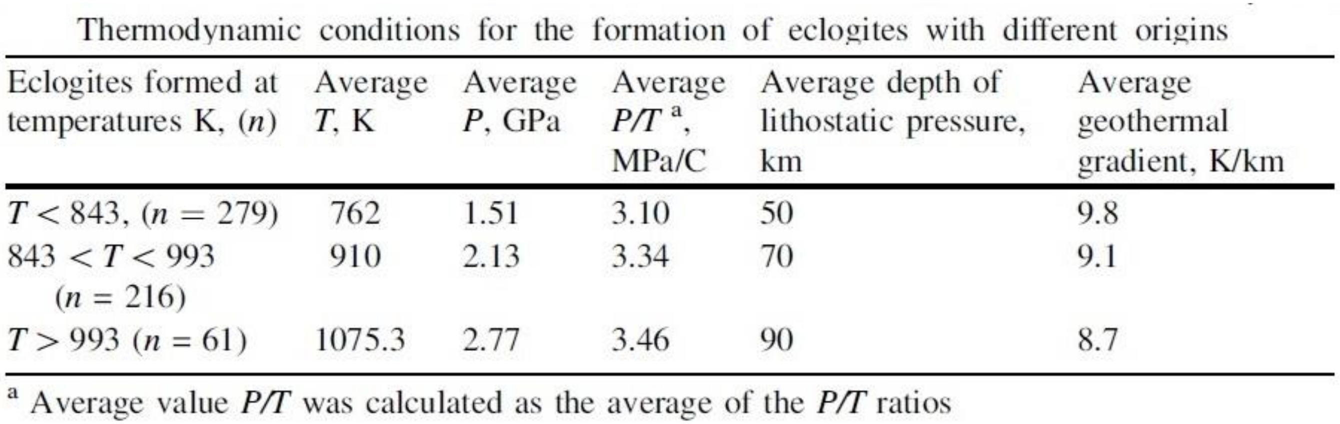
Thermodynamic conditions of some metamorphic processes (adapted from Pilchin and Eppelbaum 2009)

Type of metamorphism	Average temperature, K	Average pressure, MPa	Average depth of lithostatic pressure, km	Average geothermal gradient (K/km) for present surface conditions	Average geothermal gradient (K/km) for surface temperature ~573 K
Blueschists ($n = 265$)	677	1,000	~35	11.5	2.97
Eclogites ($n = 556$)	856	1,900	~64	9.1	4.42
Amphibolites ($n = 353$)	925	770	~26–27	24.1	17.41
Granulites[a] ($n = 601$)	1,075	820	~28	28.6	17.93
Greenschist ($n = 188$)	681	420	14–15	27	8.28

[a] Granulites of the Archean and Early Proterozoic

Metamorphic rocks of Precambrian

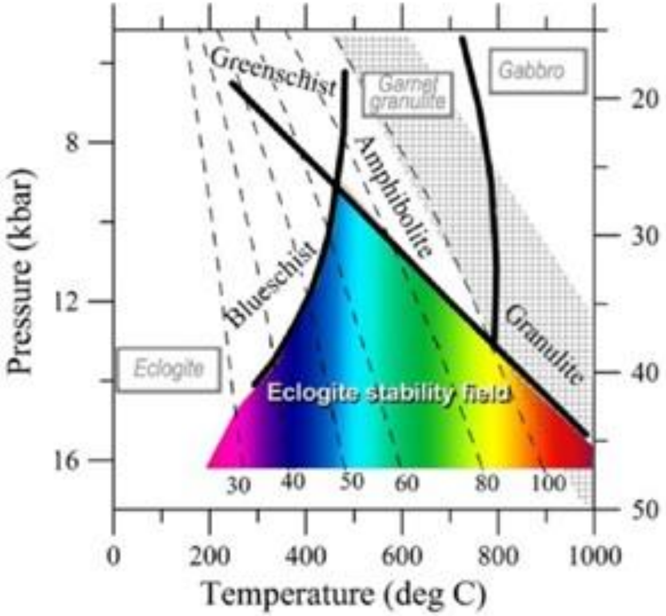
- Greenschists, granulites and amphibolites could be formed under lithostatic pressure, but metamorphic rocks such as eclogites and blueschists (characteristic of regions subjected to great horizontal displacement) require the presence of overpressure, not possible during the early stages of the Earth evolution.
- Amphibolites appeared after granulites, since their formation require lower temperatures and differently form granulites, require basic magmas, which did not form in the uppermost layers of the magma-ocean and water penetration through the very hot rocks of the young Archean crust at depths of 23-30 km.
- For eclogite metamorphism, the magma-ocean must have been solidified at a depth of 64 km, but for the formation of greenschists, granulites and amphibolites it only needed to be solidified at 14–15 km, 28 km and 26–27 km thick, respectively.



Thermodynamic conditions for the formation of eclogites with different origins

Eclogites formed at temperatures K, (*n*)	Average *T*, K	Average *P*, GPa	Average *P/T* [a], MPa/C	Average depth of lithostatic pressure, km	Average geothermal gradient, K/km
$T < 843$, ($n = 279$)	762	1.51	3.10	50	9.8
$843 < T < 993$ ($n = 216$)	910	2.13	3.34	70	9.1
$T > 993$ ($n = 61$)	1075.3	2.77	3.46	90	8.7

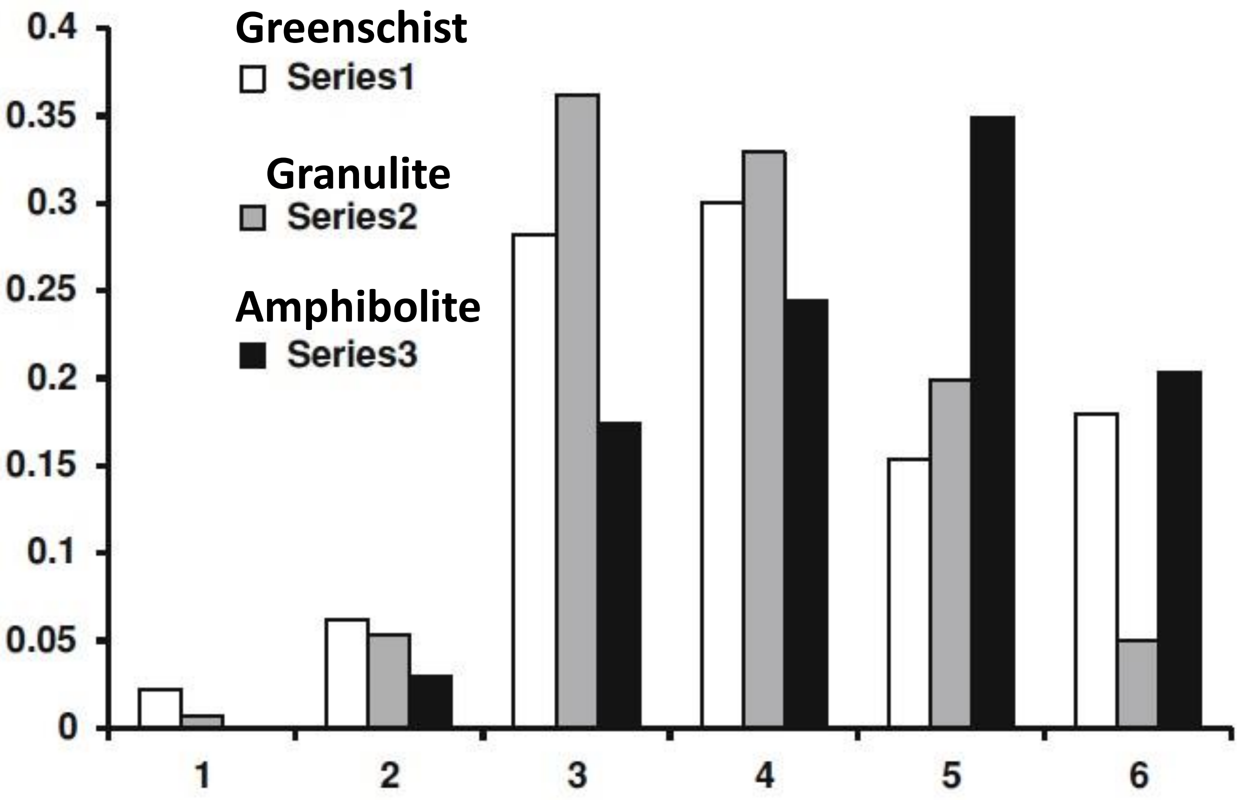
[a] Average value *P/T* was calculated as the average of the *P/T* ratios



Artemieva and Meissner, 2012, Tectonophysics 530-531

The eclogites are divided in: eclogites of low moderate temperatures (T<843 K), which require the transformation of ferrous iron oxide to ferric iron oxide, eclogites of moderate temperatures (843<T<993 K) and eclogites of high temperatures (T>993 K).

Metamorphic rocks of Precambrian



(1) Early Archean (2) Middle Archean (3) Late Archean (4) Early Proterozoic (5) Middle Proterozoic (6) Late Proterozoic

Greenschist metamorphic facies, requiring low to moderate *T* and *P*, always had favorable conditions within the crust, while the abundance of granulite facies rocks declined constantly with the cooling of the uppermost layers, since *T* ~1073 K at depths of ~28 km became increasingly unrealistic.

Thermodynamic conditions of metamorphic processes

Presence of external forces, which can control changes in volume (e.g., prevent it from expanding), stress within the volume increases to extremely high values.

α=thermal expansion coefficient
β= compressibility coeffecent

$$\alpha = \frac{(\partial V/\partial T)_P}{V_0} \qquad \beta = \frac{(\partial V/\partial P)_T}{V_0} \qquad P = P_0 + \frac{\alpha}{\beta}(T - T_0) - \frac{1}{\beta}\frac{\Delta V}{V_0}$$

The second term on the right side of the Eq. characterizes the dependence of pressure on the temperature of the rock. For $T > T_0$ it will be an increase of pressure compared to the lithostatic or hydrostatic value, whereas $T < T_0$ will lead to pressures below the lithostatic or hydrostatic value.
The third term on the right side of the equation represents pressure unloading (when $\Delta V > 0$).

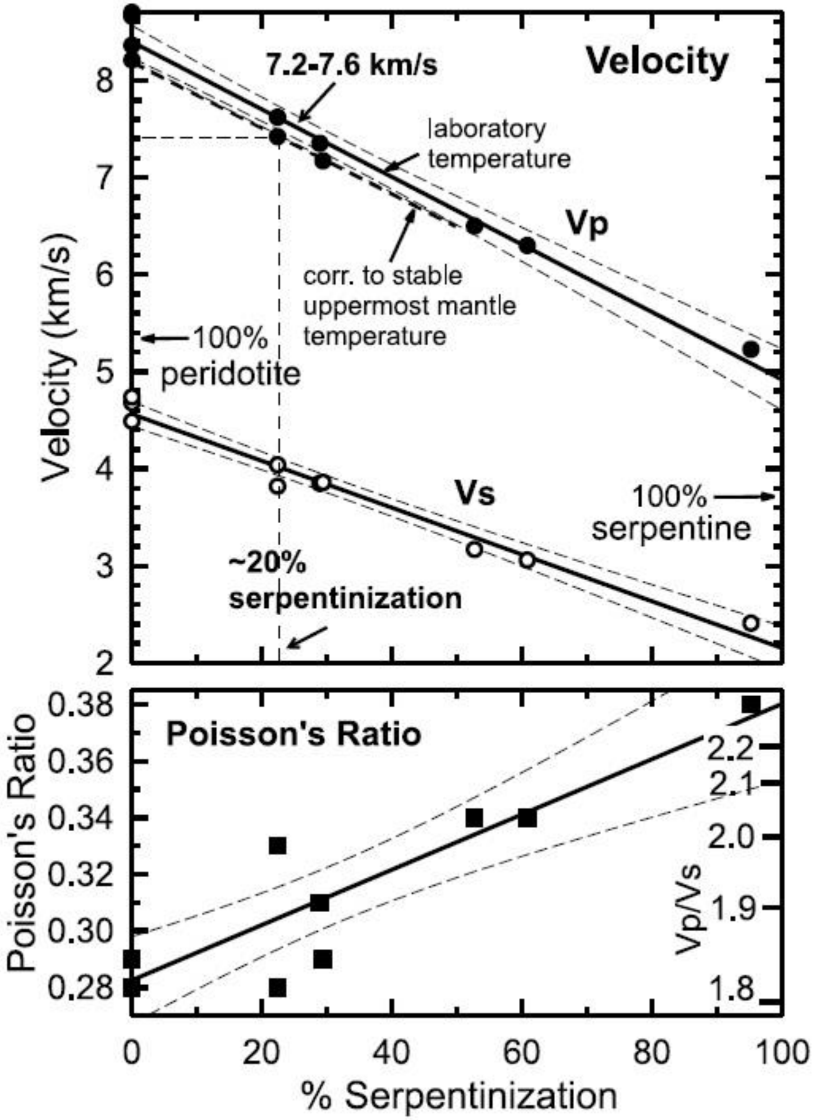
- We can notice that an increase of volume would lead to a decrease in pressure only in the case where the actual increase of volume was equal or greater than the change in volume caused by the process, but in all other cases it would lead to an increase in pressure.

- Since β for solid crustal rocks ~ 10^{-5} 1/MPa, an increase in volume of only 1 % can cause composite unloading of pressure of about 1000 MPa, but if there was not any change in volume, the pressure would increase by about 1000 MPa.

There are several processes that lead to the formation of significant overpressure capable of causing tectonic processes: (1) heating of underground water; (2) heating of great volumes of rocks; (3) melting of rocks; (4) rising of melt to the Earth's surface; (5) decompaction of crustal and upper mantle rocks.

Serpentinization

Serpentinite: low Vp and Vs velocity, low density, and high Vp/Vs ratio and Poisson's ratio



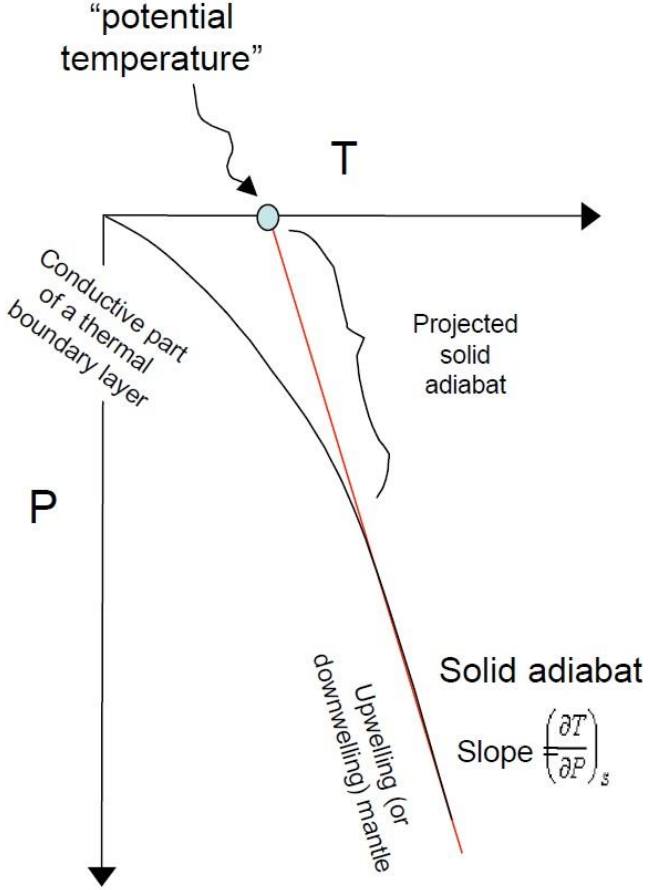
Hyndman and Peacock, 2013, EPSL, 212

- Serpentinization occurs in a *T* range of 473-773 K (shallow depths, during obduction processes or in the oceanic regions) and can cause an increase in volume up to 40 %.
- Serpentinites are the key component of ophiolite association and all peridotite plates obducted onto the surface are partially serpentinized.
- An increases of 15–25 % in the volume of the peridotite layer of oceanic lithosphere during its serpentinization can cause the horizontal expansion of the layer in any direction of at least 5–8 %.
- Serpentine minerals (lizardite, chrysotile, and antigorite) contain up to 13 wt% in water, have a low density of about 2600 kg/m^3, and a low viscosity of about 4 x 10^{19} Pa s.
- Lizardite is stable at T 358–458 K, chrysotile at T 403-458 and antigorite at 493–733 K. Since magnetite is a main mineral of serpentinite, a significant amounts of serpentinites should take place at *T* above the lower limit of TFFI (473–523 K).

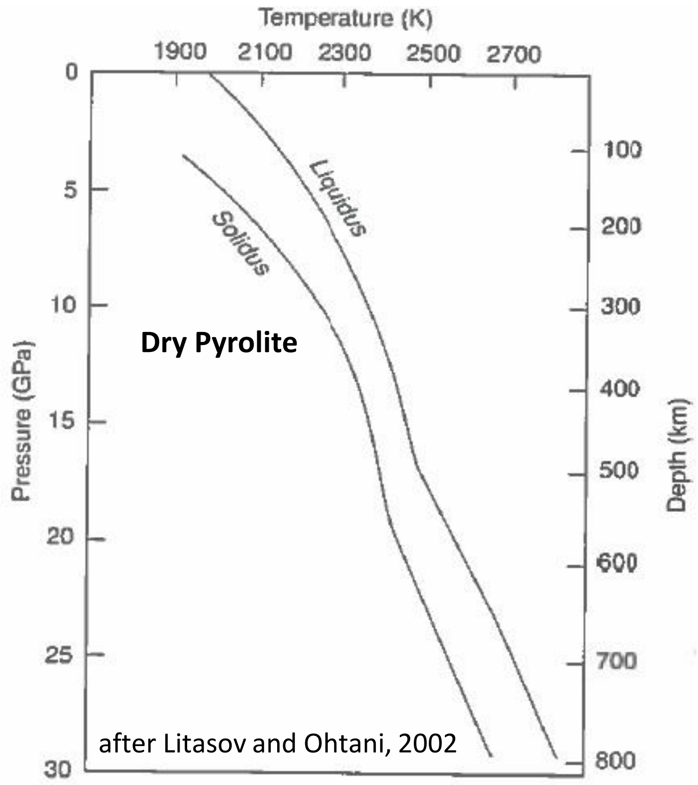
Average secular cooling rate

- After solidification of the magma ocean (at the end of the Early Archean) the potential temperature (T_p) of a pyrolitic mantle (peridotite + basalt) was ~1800 K.
- According to petrological data, T_P decreased from the Archean to 137 ± 8 - 187 ±42 K, which leads to a cooling rate of ~ 50 ± 20 K Gyr^{-1}.

The temperature of a given portion of the mantle if it was adiabatically decompressed to the surface of the Earth.

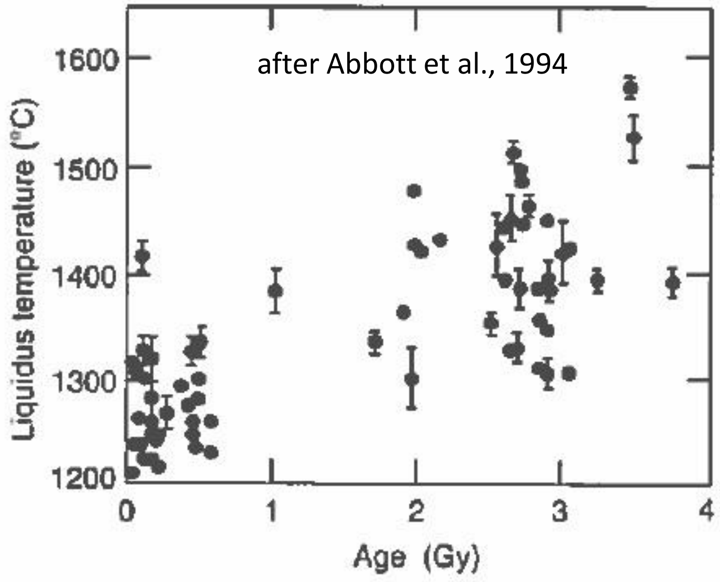


(after solidification of the magma ocean)



after Litasov and Ohtani, 2002

T_L **for basalts (ophoilites and greenstone belts)**



after Abbott et al., 1994

Heat transfer mechanisms

conduction

from Latin conductionere
“to bring together”

convection

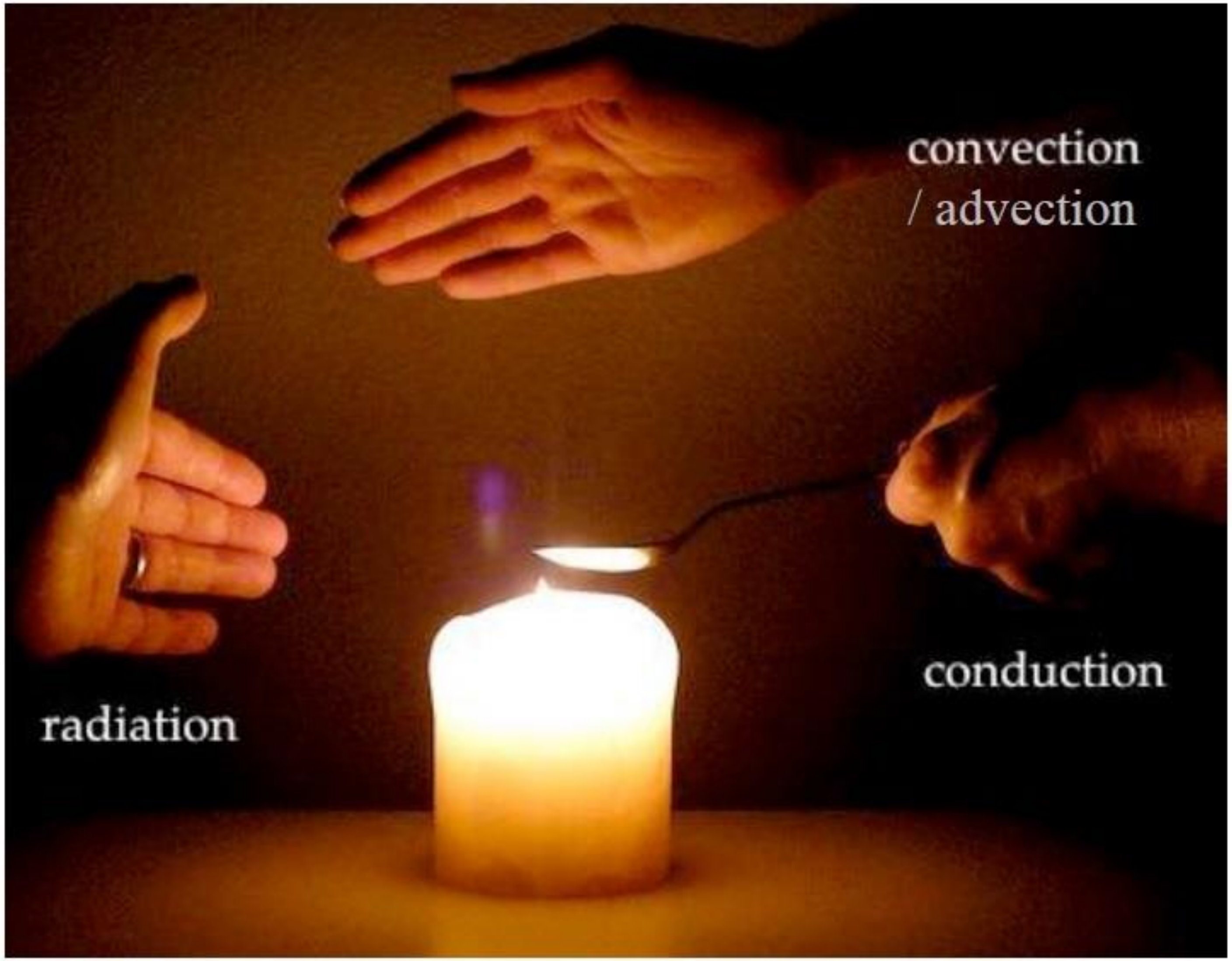
from Latin convehere
“to carry together”

advection

from Latin advehere
“to carry to”

radiation

from Latin radius
“ray, spoke”

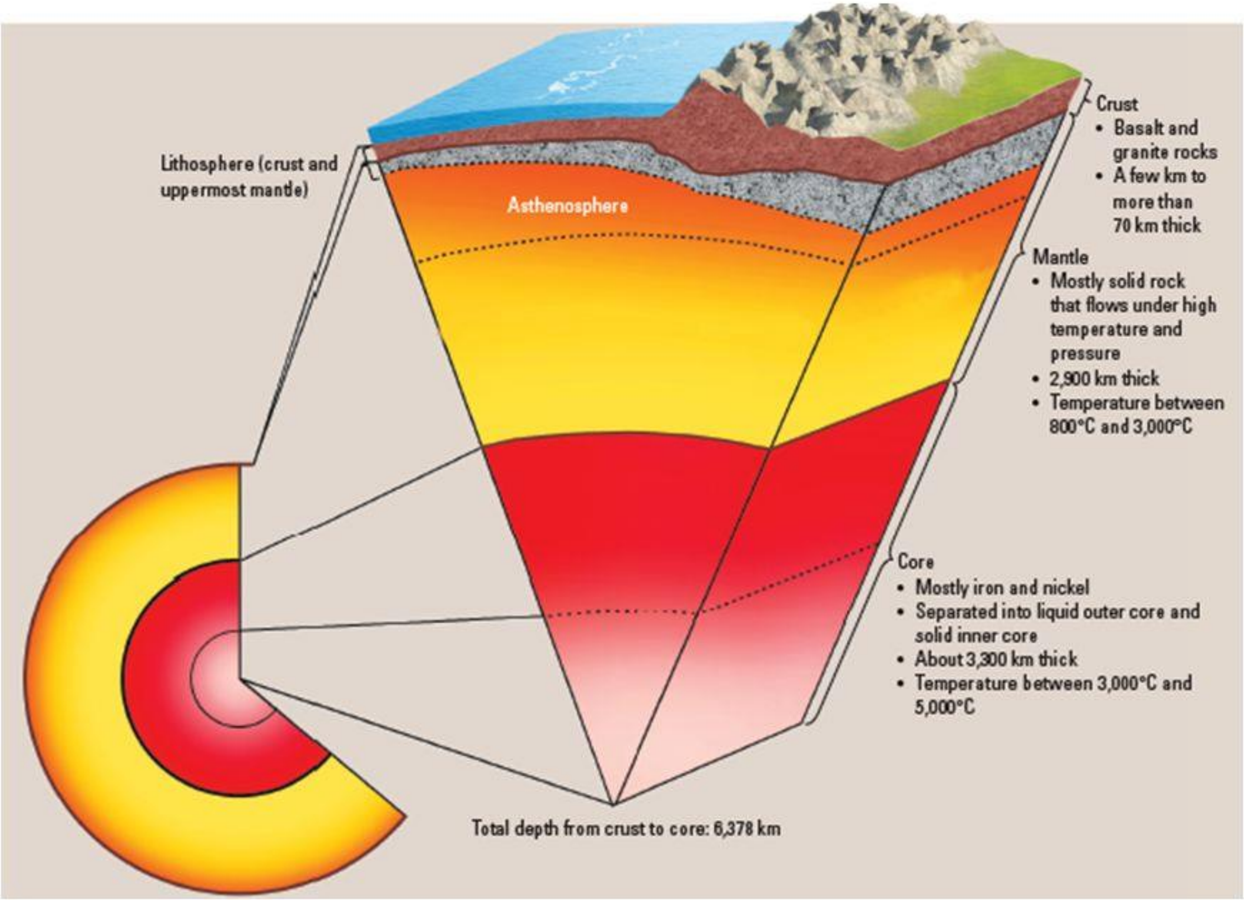


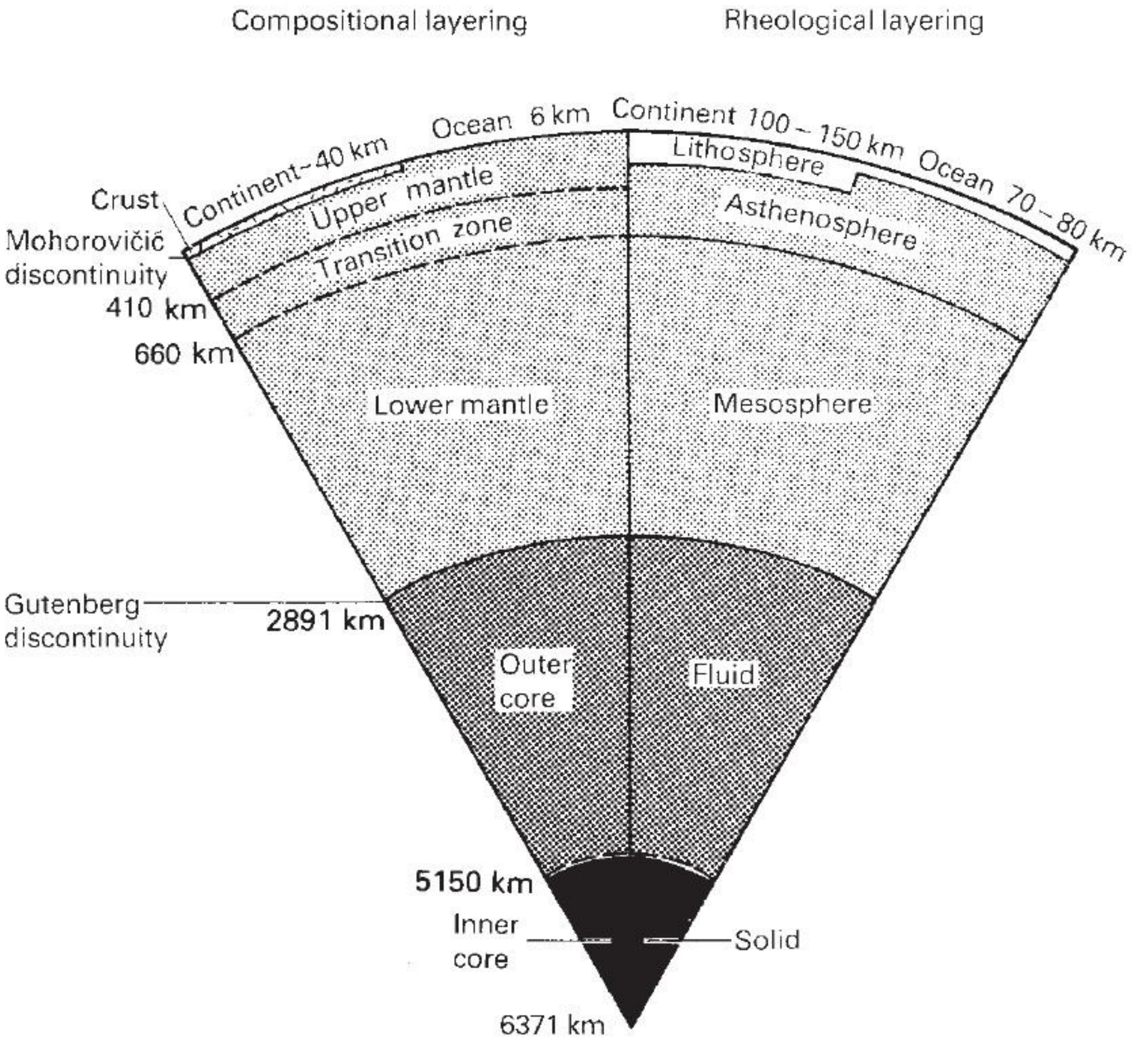
Heat transfer mechanisms

- **Conduction**
  - It is the slowest kind of heat transfer, which occurs through a material by atomic or molecular interaction with the material. It requires direct contact between medias transferring heat.
- **Radiation**
  - Direct transfer of heat as electromagnetic radiation It dominates the Earth‘s surface, since it requires an open surface
- **Convection**
  - Transfer of heat by movement of the molecules themselves. It is a very effective method of heat transfer by a fluid medium, based on circulation of cold fluid down and hot fluid up. It requires an open system conditions for the free movement of fluids.

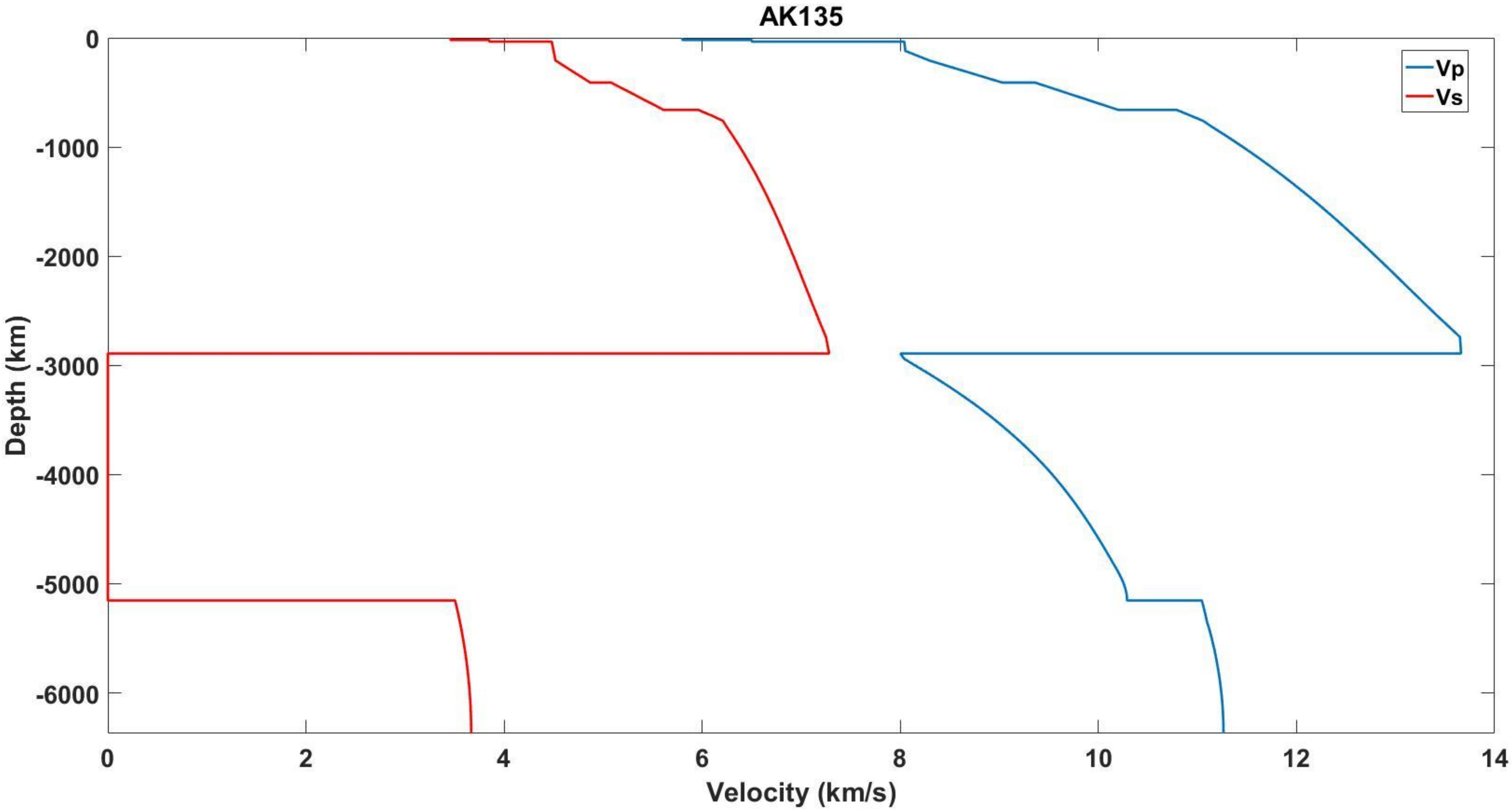
Earth’s Structure

What is the internal structure of the earth?

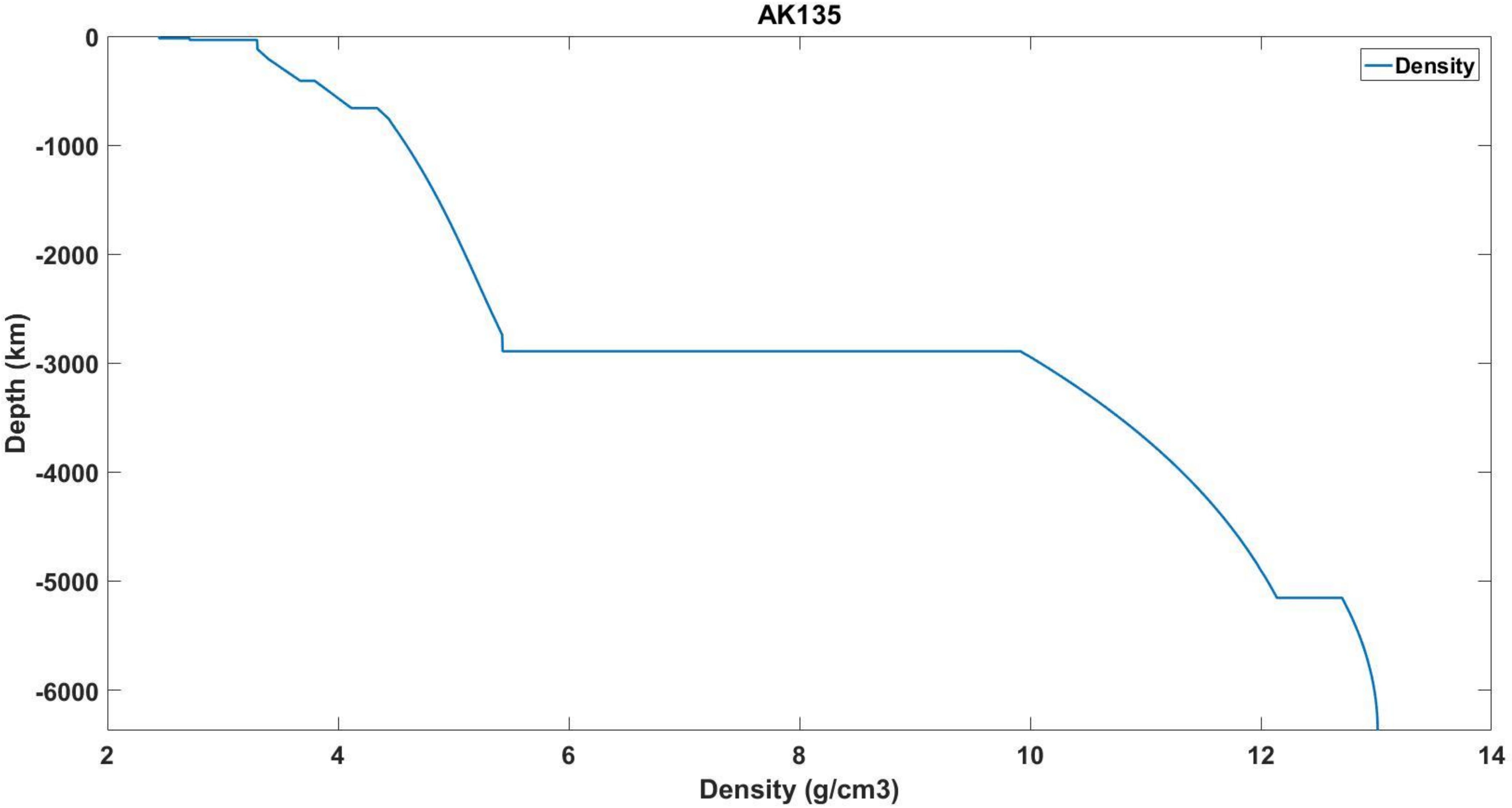




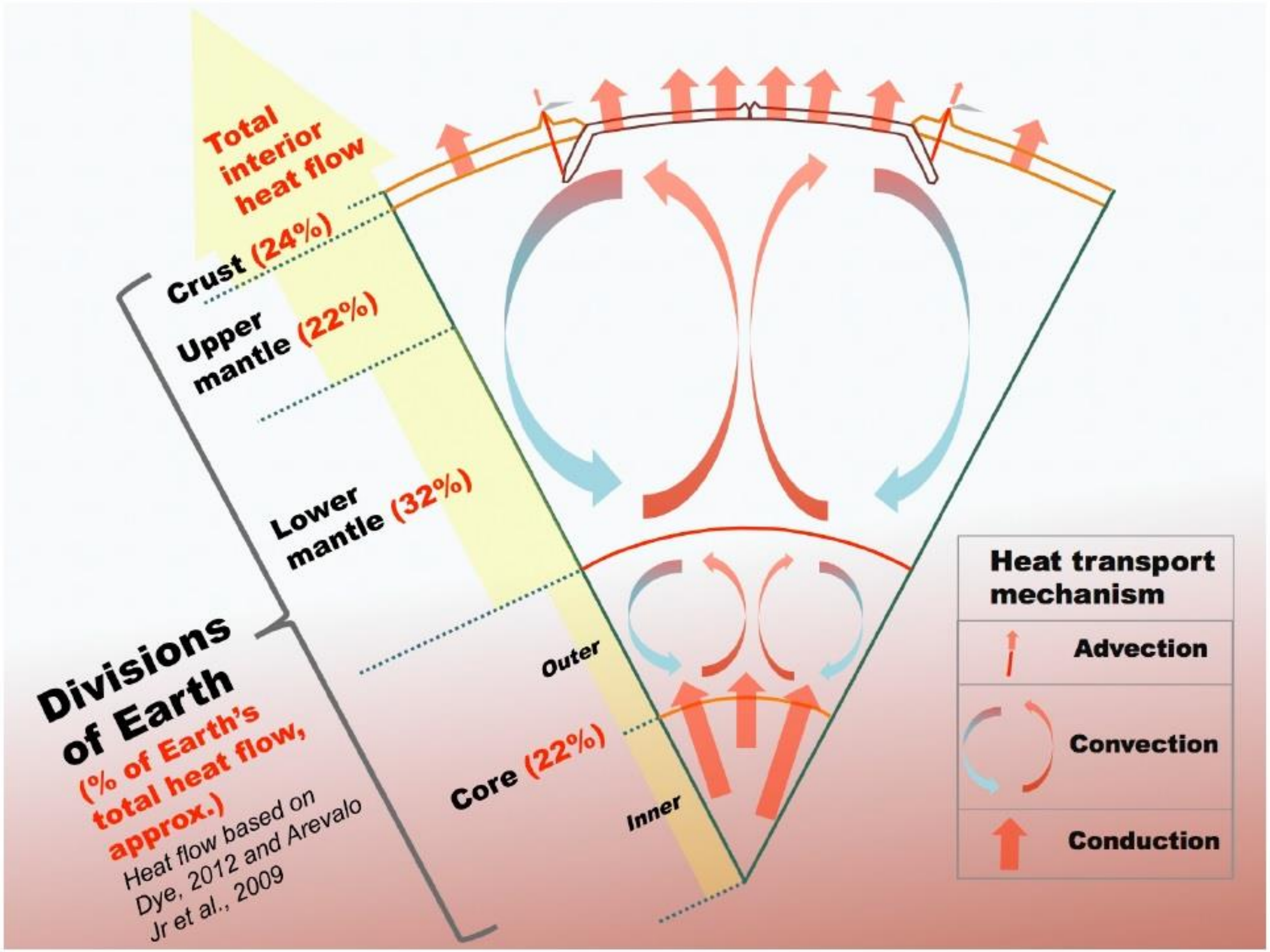
Earth’s Velocity



Earth's Density

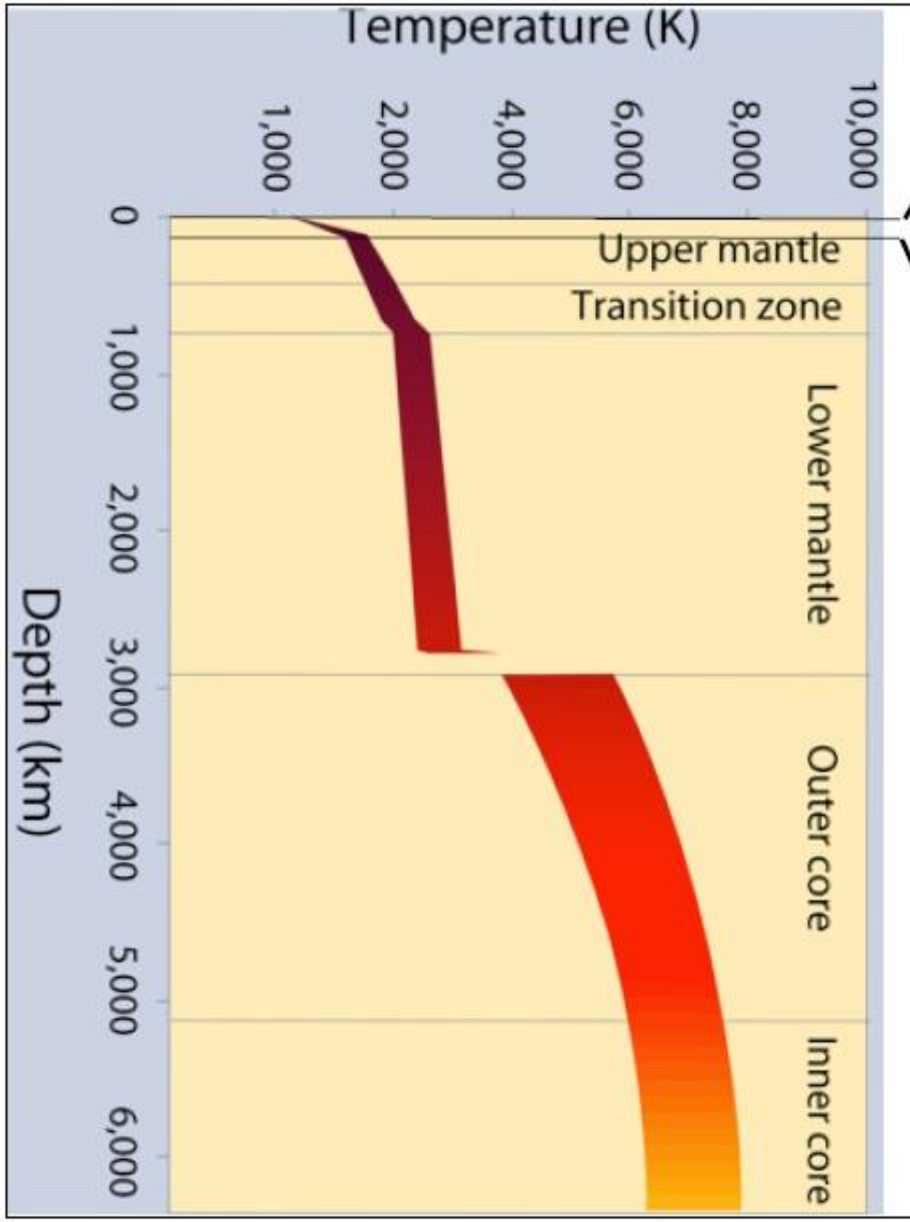


Heat Transport Mechanisms in the Earth



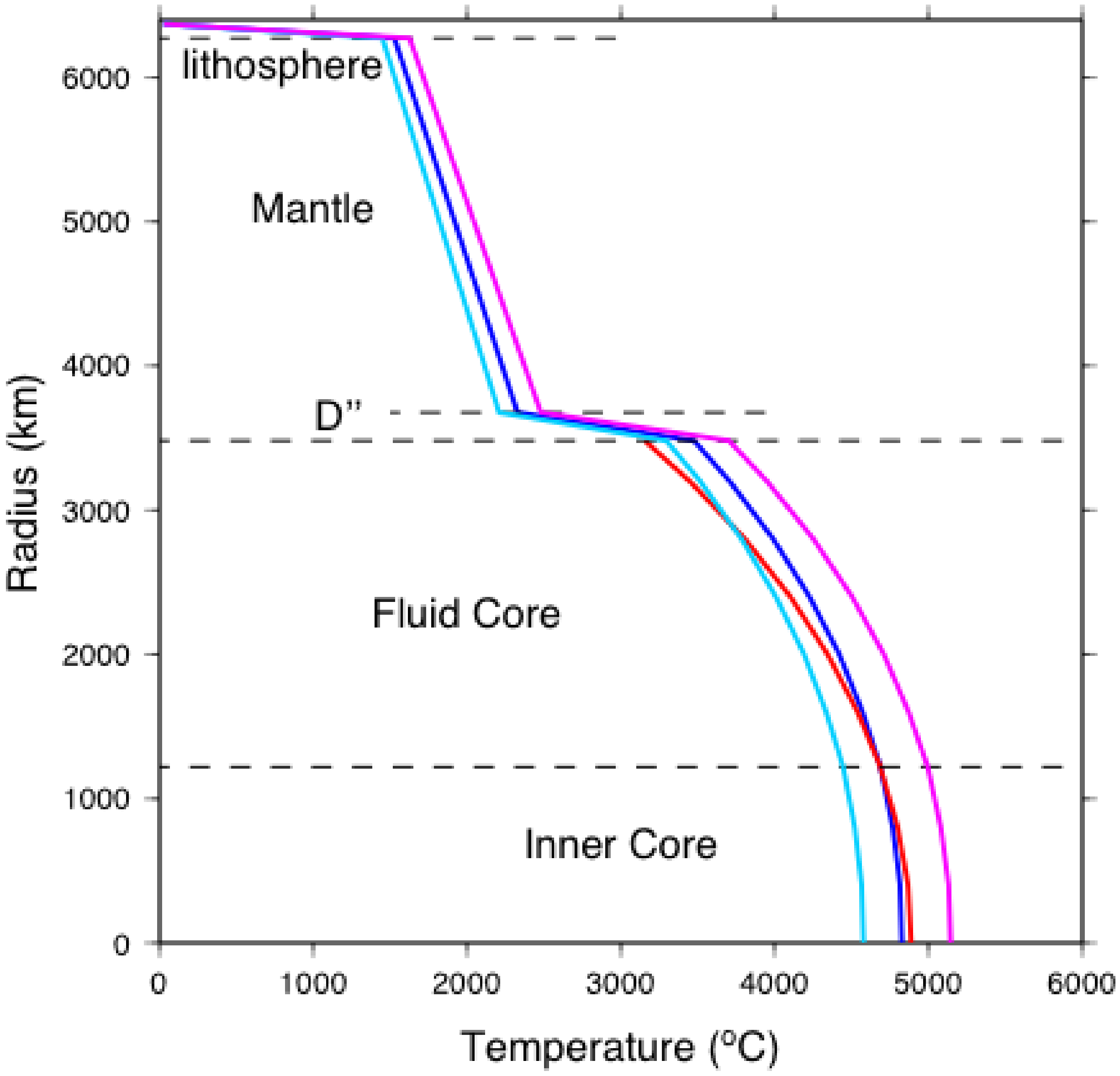
Heat flow based on Dye, 2012 and Arevalo Jr et al., 2009

Temperature profile inside Earth



The lithosphere corresponds roughly with the conductive boundary layer of the convective system

Thermal evolution of the Earth



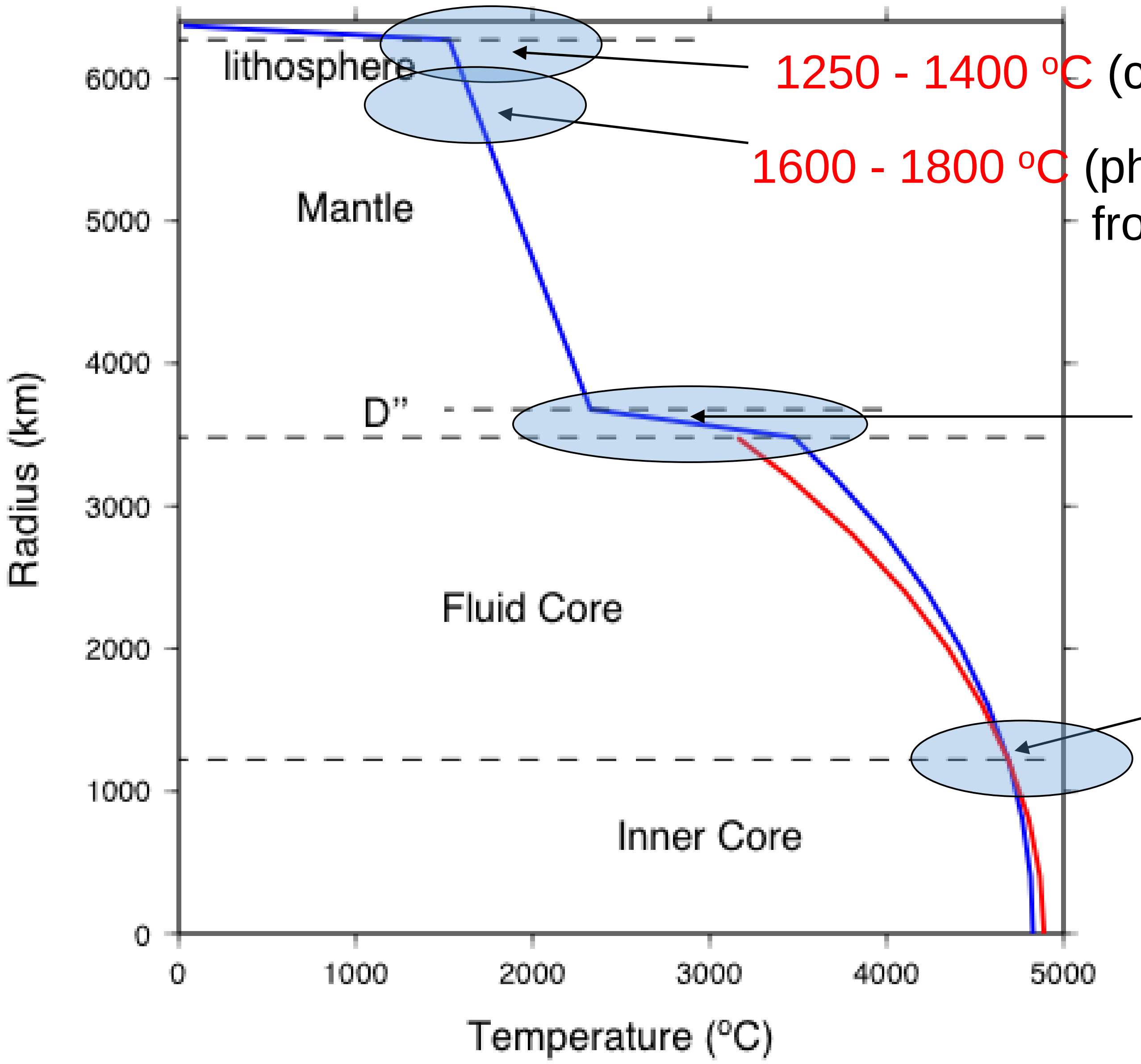
Present Temp.

Past Temp.

Future Temp.

Fe melting curve

Temperature profile inside Earth



1250 - 1400 °C (cooling of lithosphere)

1600 - 1800 °C (phase transition temp. from mineral physics)

dT ~ 800 - 1800 °C

5000 - 5300 °C (mineral physics)

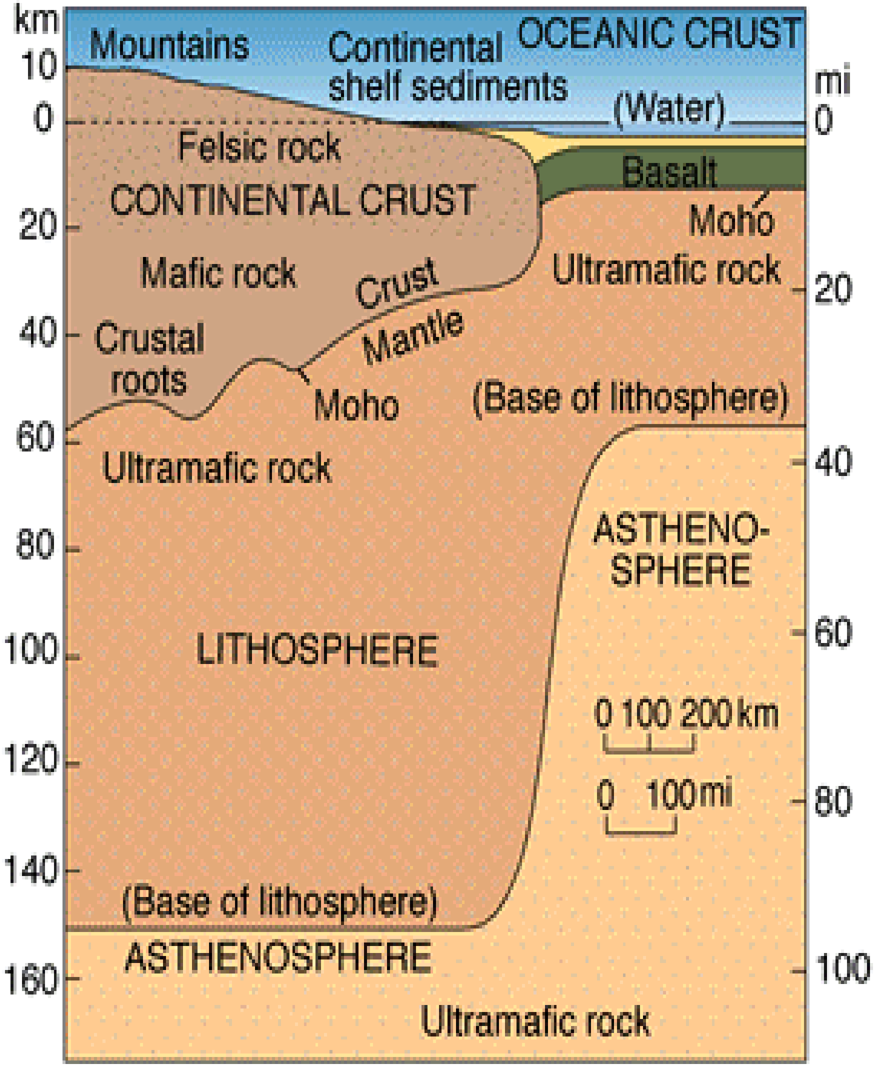
Oceanic and continental lithosphere

Oceanic lithosphere

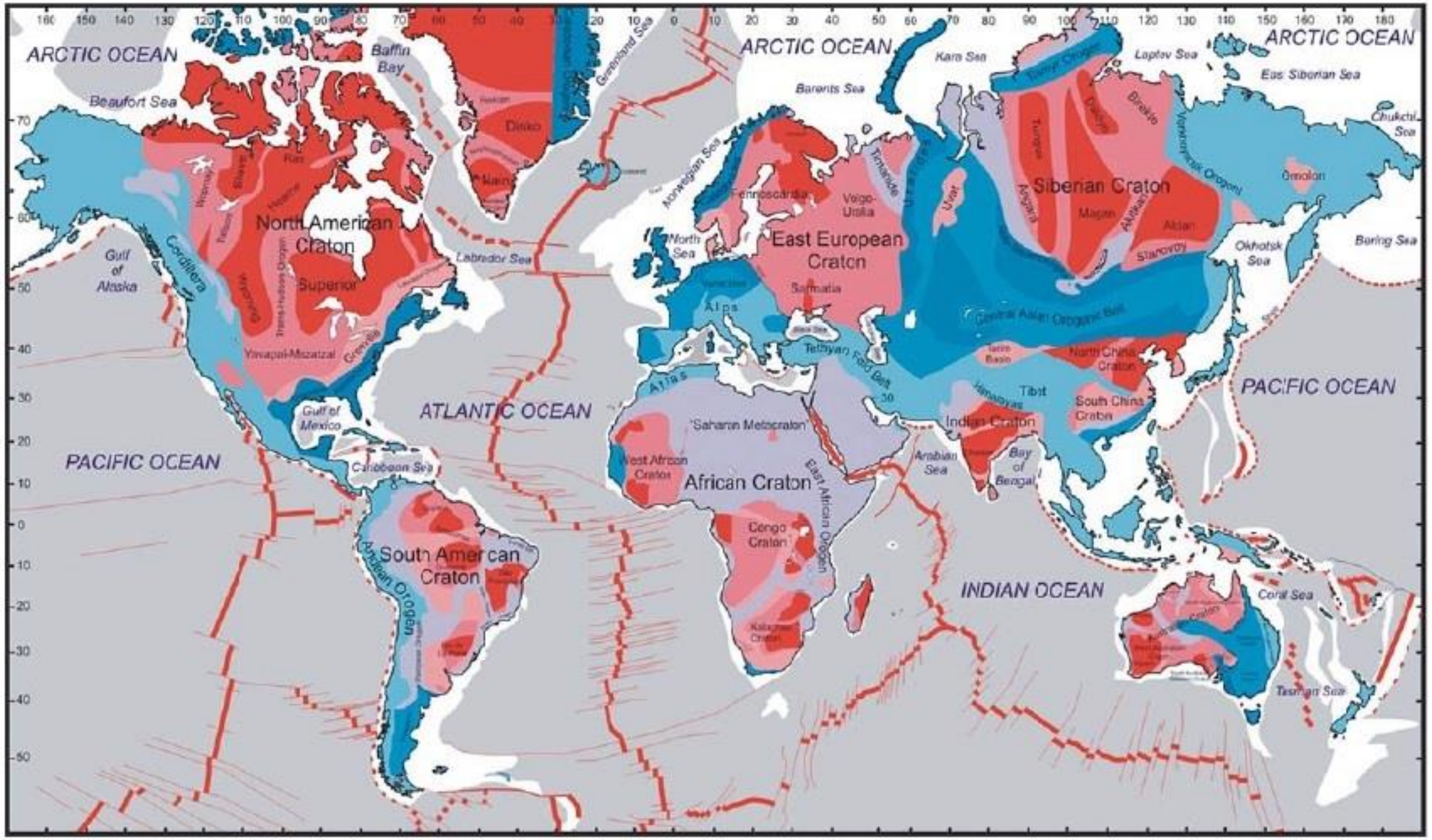
- Thin
    - Crust: 5 – 8 km
    - Lithospheric thickness increases with age (100-125 km)
- Young: less than 200 million years
- Heavy: ultimately always subduction
- Enriched in FeO and MgO
- Hardly any heat production

Continental lithosphere

- Thick
    - Crust: 20 – 60 km
    - Lithosphere: 25 – 250 km
- Old: More than 4 billion years
- Light: virtually never subduction
- Enriched in SiO2
- Substantial heat production



Basement age of continental crust



▭ Subduction zone

▭ Mid-oceanic ridges and transform faults

■ Oceanic crust

Continental crust

□ Shelf

■ Mesozoic–cenozoic (250–0 Ma)

■ Late Paleozoic (400–250 Ma)

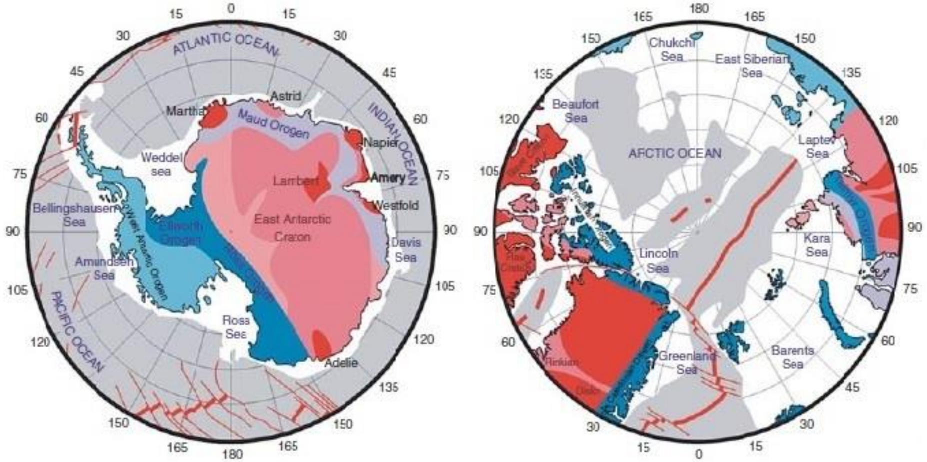
■ Early Paleozoic (543–400 Ma)

■ Neoproterozoic (1.0–0.54 Ga)

■ Mesoproterozoic (1.6–1.0 Ga)

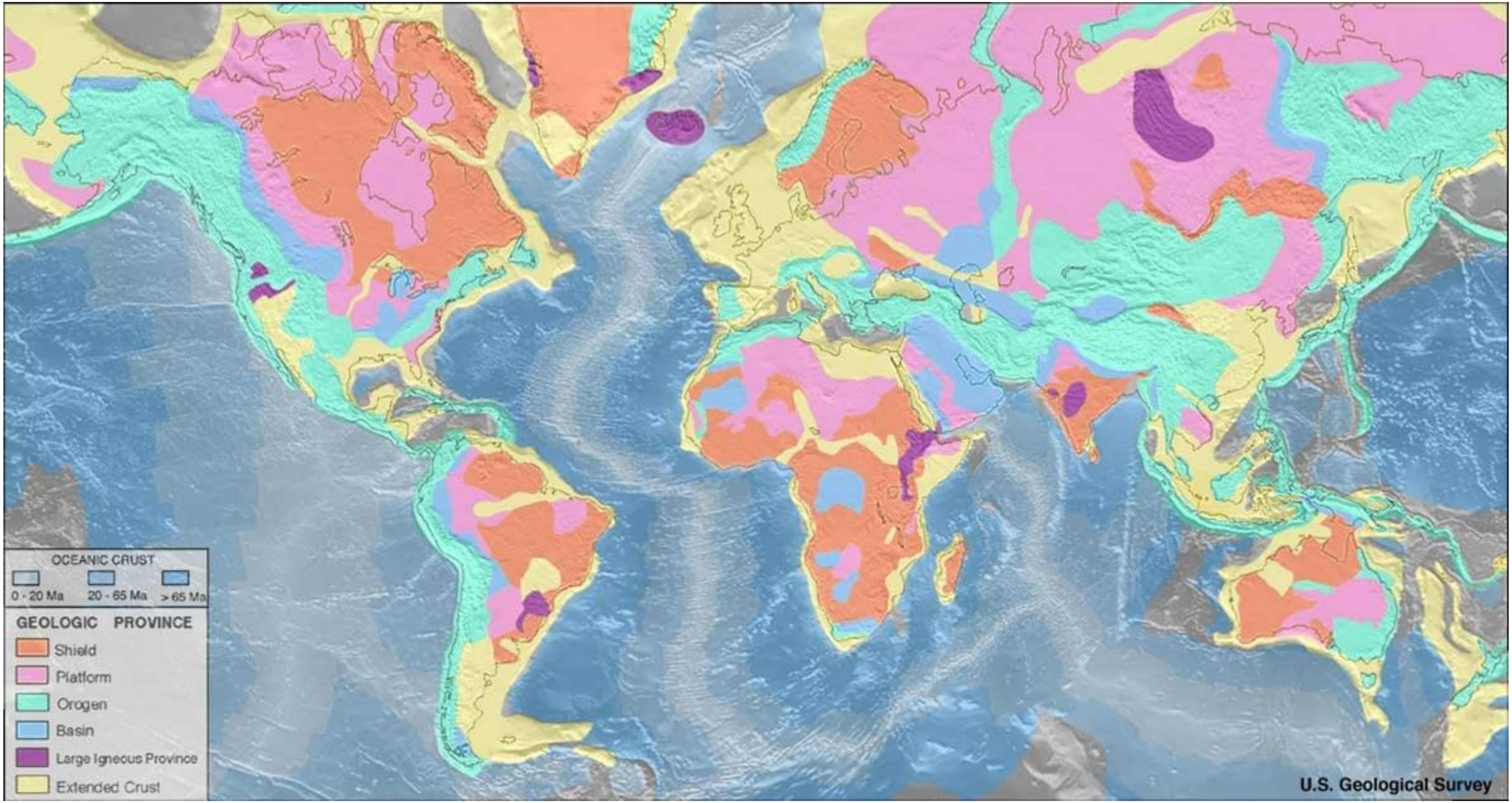
■ Paleoproterozoic (2.5–1.6 Ga)

■ Archaean (3.8–2.5 Ga)

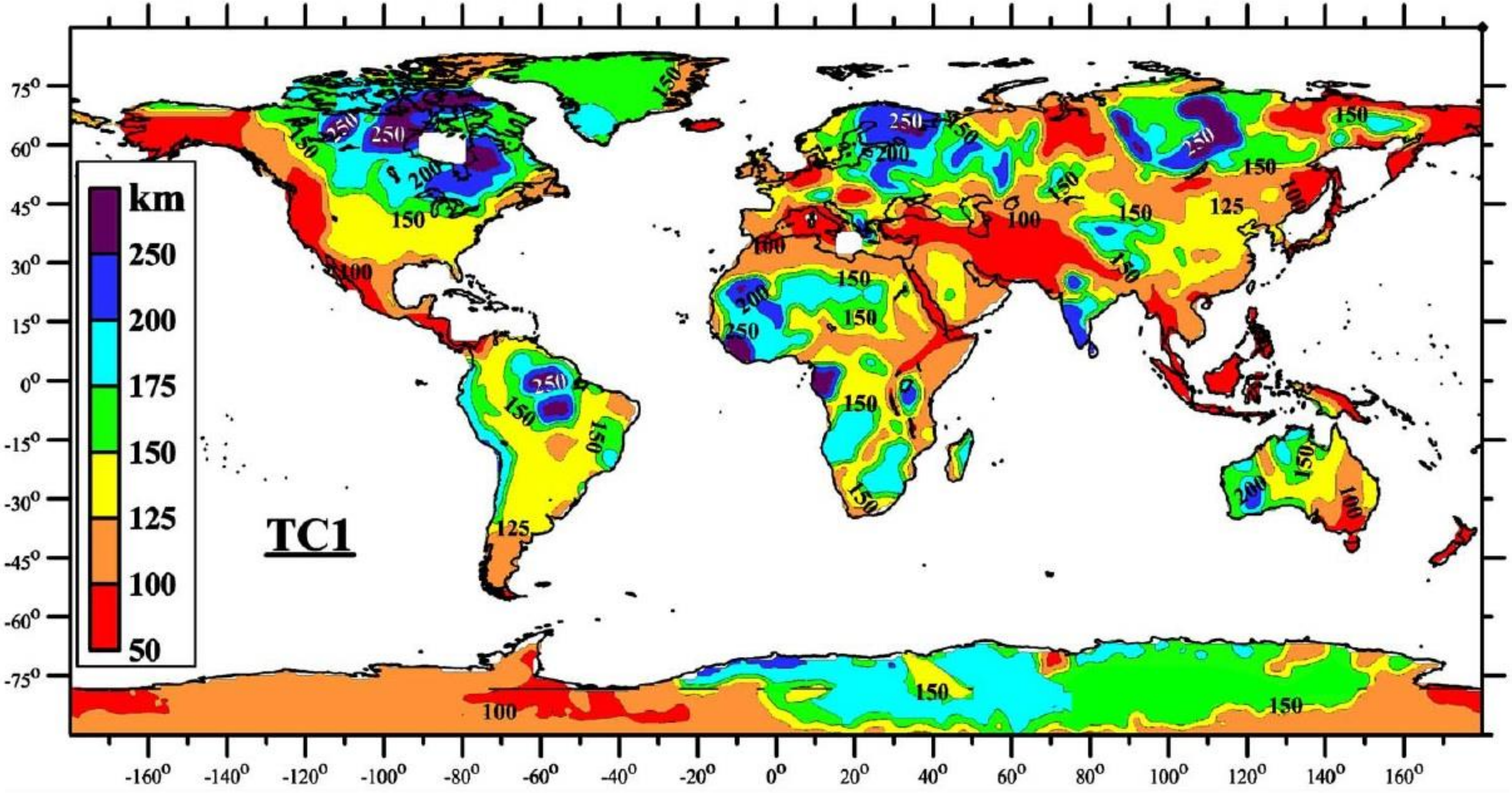


(Mooney, 2013, Treatise of Geophisics)

Continental crustal type



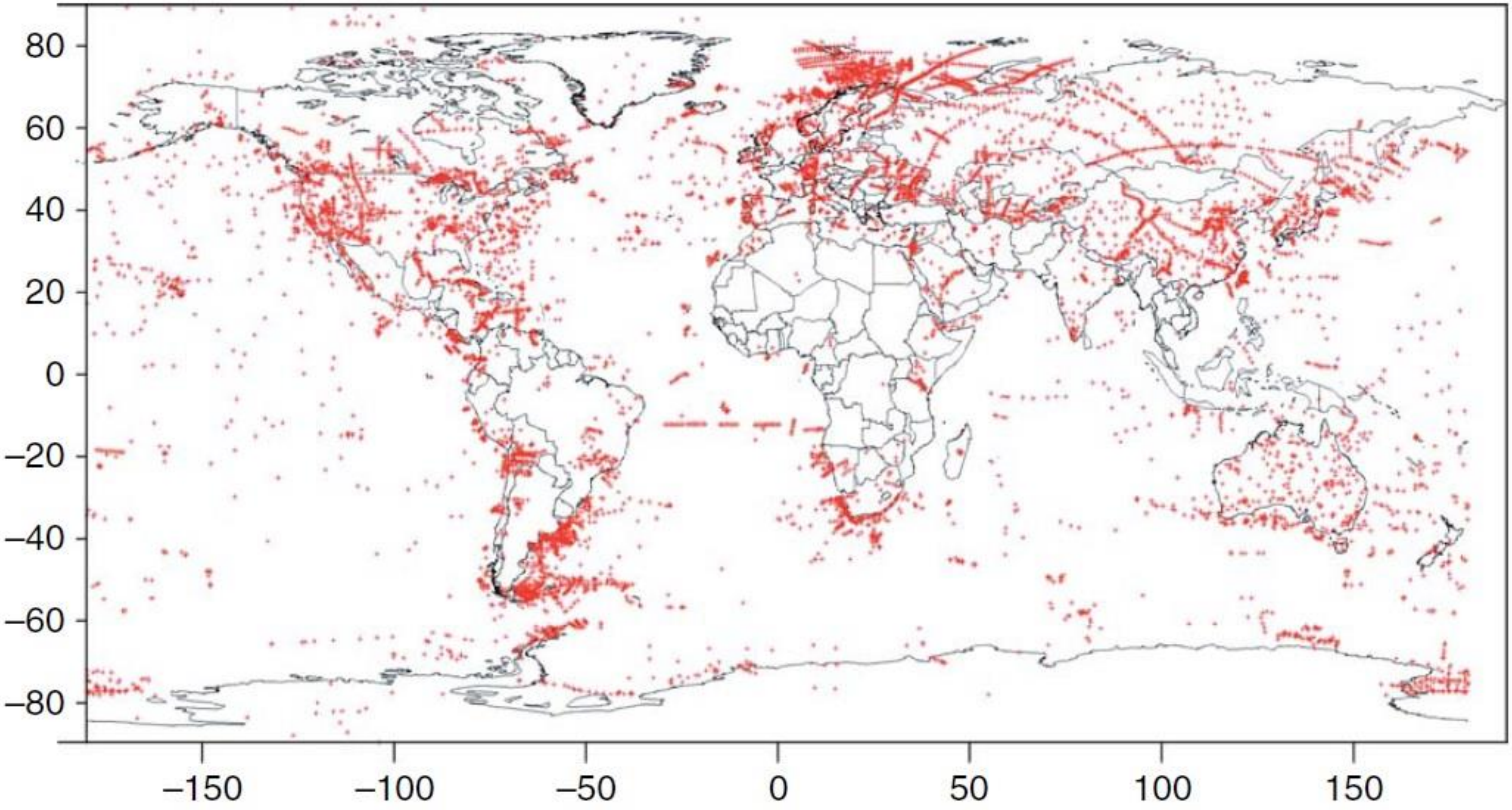
Thickness of the thermal lithosphere



Artemieva, 2006, EPSL

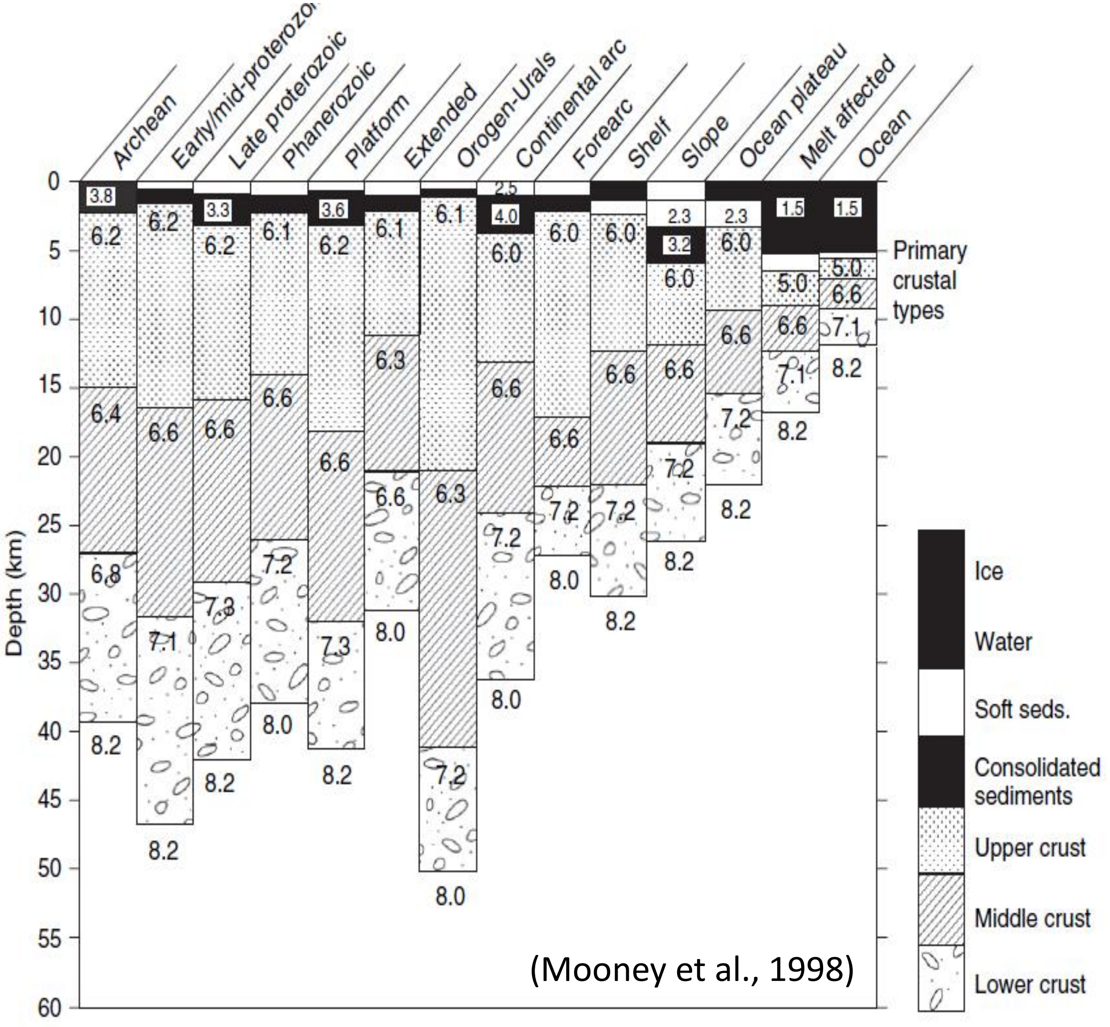
Seismic-depth measurements (1920-present)

(USGS Database)



Continental and Oceanic Crust

- The crust is the outer shell of the Earth (constitutes about 0.7% of the total mass of the crust–mantle system), involving the transfer of heat from the inner layers of the Earth and particularly from the mantle to the surface.
- Heat conduction occurring in the crust is extremely lithology dependent, since is due to (1) thermal conductivity λ and (2) heat generation *A*. λ and *A* decrease and increase from basic to acidic rocks, respectively.



(Mooney et al., 1998)

Continental crust (weighted average values):

Thickness = 41 km (SD=6.2 km)

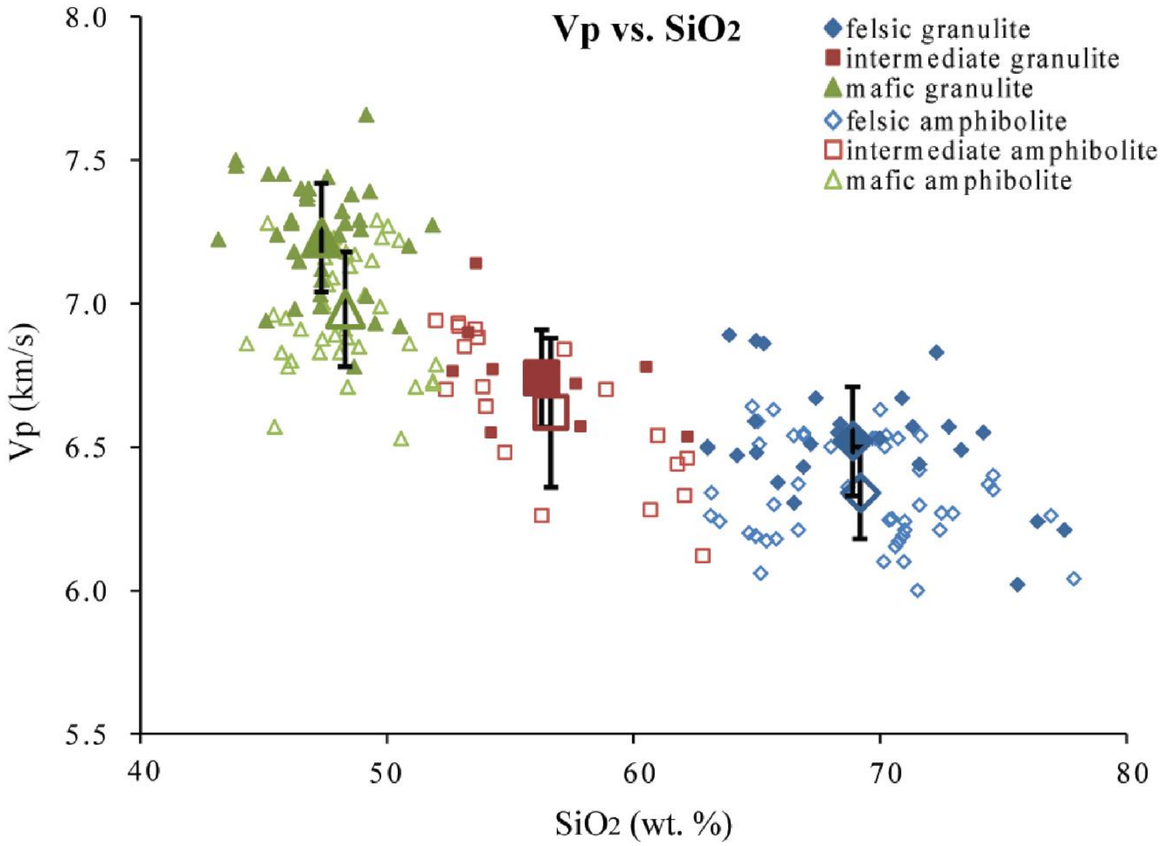
Vp=6.45 km/s (SD=0.21 km/s)

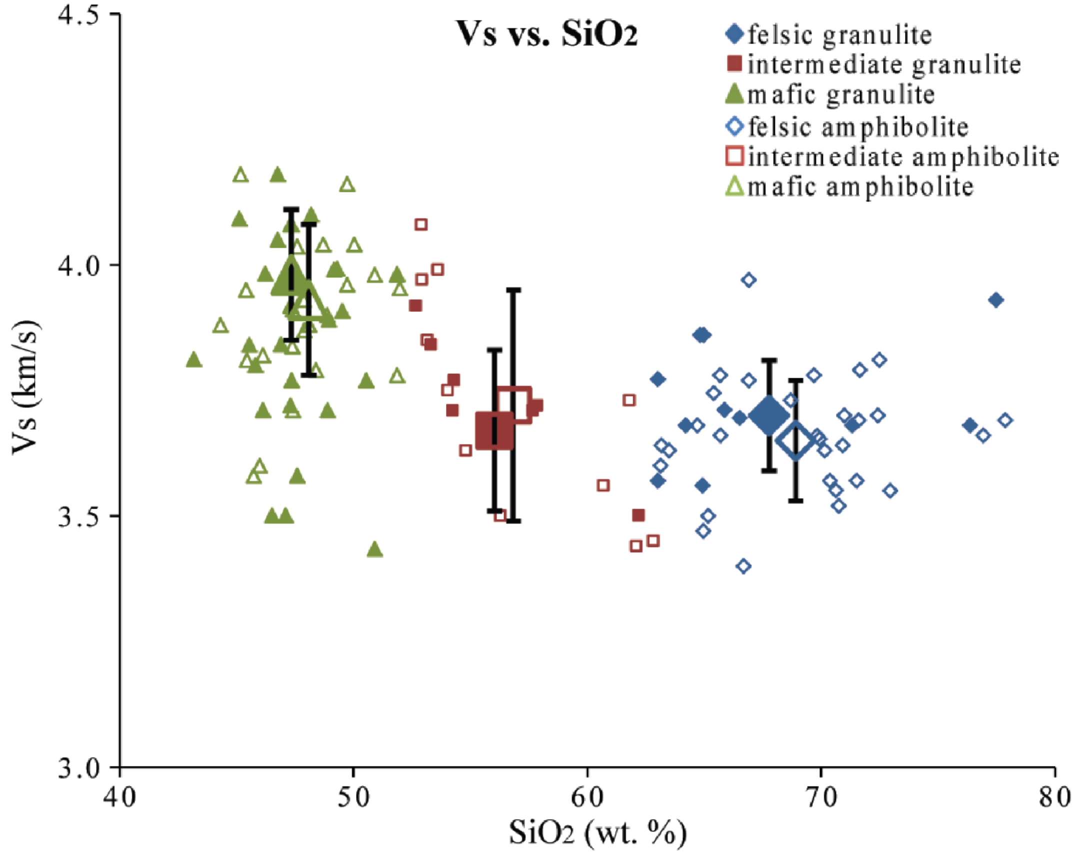
Oceanic crust (average values):

Thickness = 6 km, $\rho=3$ g/cm^3,

Age <=200 My

P-wave and *S-wave* velocity correlates with bulk composition

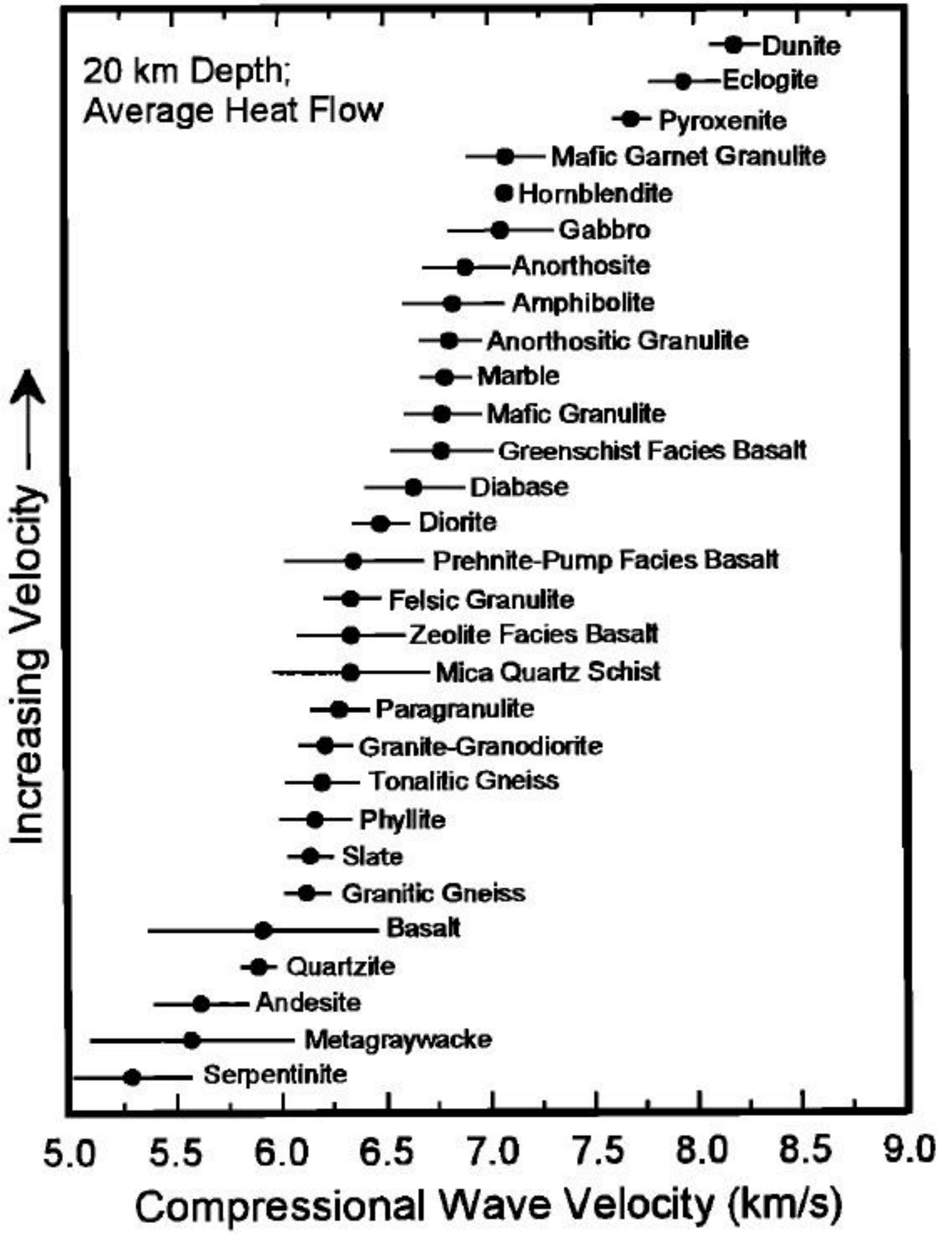




P=0.6 GPa, room *T*

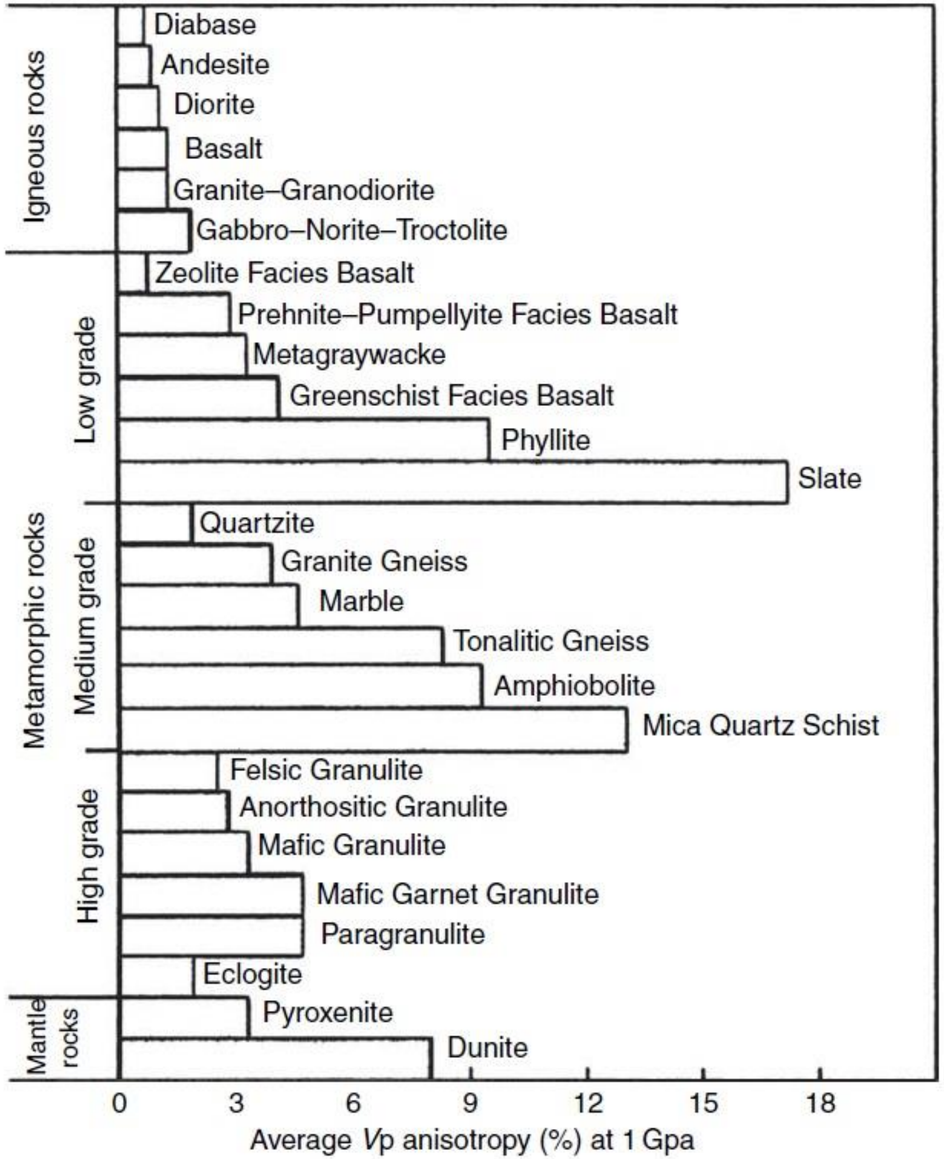
Huang et al., 2013

Rocks' P-wave velocity



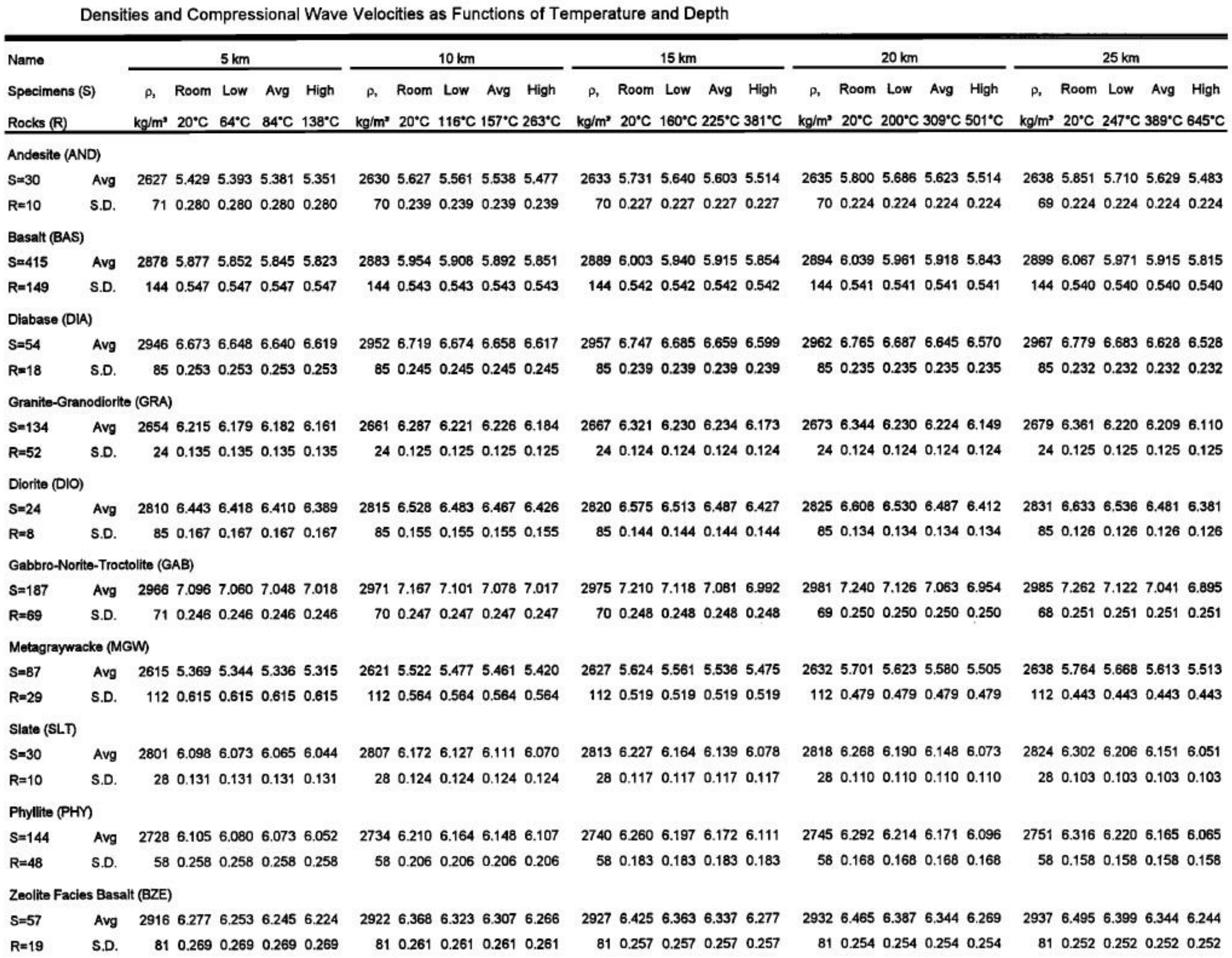
Average Anisotropy 100x (Vmax-Vmin)/Vavg

Elastic waves show a directional dependence in wave speed in many minerlas



Christensen and Mooney, 1995, JGR, 100

Rocks' Density *vs* *T* and *P*



Densities and Compressional Wave Velocities as Functions of Temperature and Depth

Name		5 km					10 km					15 km					20 km					25 km				
Specimens (S)		ρ,	Room	Low	Avg	High	ρ,	Room	Low	Avg	High	ρ,	Room	Low	Avg	High	ρ,	Room	Low	Avg	High	ρ,	Room	Low	Avg	High
Rocks (R)		kg/m^3	20°C	64°C	84°C	138°C	kg/m^3	20°C	116°C	157°C	263°C	kg/m^3	20°C	160°C	225°C	381°C	kg/m^3	20°C	200°C	309°C	501°C	kg/m^3	20°C	247°C	389°C	645°C
Andesite (AND)																										
S=30	Avg	2627	5.429	5.393	5.381	5.351	2630	5.627	5.561	5.538	5.477	2633	5.731	5.640	5.603	5.514	2635	5.800	5.686	5.623	5.514	2638	5.851	5.710	5.629	5.483
R=10	S.D.	71	0.280	0.280	0.280	0.280	70	0.239	0.239	0.239	0.239	70	0.227	0.227	0.227	0.227	70	0.224	0.224	0.224	0.224	69	0.224	0.224	0.224	0.224
Basalt (BAS)																										
S=415	Avg	2878	5.877	5.852	5.845	5.823	2883	5.954	5.908	5.892	5.851	2889	6.003	5.940	5.915	5.854	2894	6.039	5.961	5.918	5.843	2899	6.067	5.971	5.915	5.815
R=149	S.D.	144	0.547	0.547	0.547	0.547	144	0.543	0.543	0.543	0.543	144	0.542	0.542	0.542	0.542	144	0.541	0.541	0.541	0.541	144	0.540	0.540	0.540	0.540
Diabase (DIA)																										
S=54	Avg	2946	6.673	6.648	6.640	6.619	2952	6.719	6.674	6.658	6.617	2957	6.747	6.685	6.659	6.599	2962	6.765	6.687	6.645	6.570	2967	6.779	6.683	6.628	6.528
R=18	S.D.	85	0.253	0.253	0.253	0.253	85	0.245	0.245	0.245	0.245	85	0.239	0.239	0.239	0.239	85	0.235	0.235	0.235	0.235	85	0.232	0.232	0.232	0.232
Granite-Granodiorite (GRA)																										
S=134	Avg	2654	6.215	6.179	6.182	6.161	2661	6.287	6.221	6.226	6.184	2667	6.321	6.230	6.234	6.173	2673	6.344	6.230	6.224	6.149	2679	6.361	6.220	6.209	6.110
R=52	S.D.	24	0.135	0.135	0.135	0.135	24	0.125	0.125	0.125	0.125	24	0.124	0.124	0.124	0.124	24	0.124	0.124	0.124	0.124	24	0.125	0.125	0.125	0.125
Diorite (DIO)																										
S=24	Avg	2810	6.443	6.418	6.410	6.389	2815	6.528	6.483	6.467	6.426	2820	6.575	6.513	6.487	6.427	2825	6.608	6.530	6.487	6.412	2831	6.633	6.536	6.481	6.381
R=8	S.D.	85	0.167	0.167	0.167	0.167	85	0.155	0.155	0.155	0.155	85	0.144	0.144	0.144	0.144	85	0.134	0.134	0.134	0.134	85	0.126	0.126	0.126	0.126
Gabbro-Norite-Troctolite (GAB)																										
S=187	Avg	2966	7.096	7.060	7.048	7.018	2971	7.167	7.101	7.078	7.017	2975	7.210	7.118	7.081	6.992	2981	7.240	7.126	7.063	6.954	2985	7.262	7.122	7.041	6.895
R=69	S.D.	71	0.246	0.246	0.246	0.246	70	0.247	0.247	0.247	0.247	70	0.248	0.248	0.248	0.248	69	0.250	0.250	0.250	0.250	68	0.251	0.251	0.251	0.251
Metagraywacke (MGW)																										
S=87	Avg	2615	5.369	5.344	5.336	5.315	2621	5.522	5.477	5.461	5.420	2627	5.624	5.561	5.536	5.475	2632	5.701	5.623	5.580	5.505	2638	5.764	5.668	5.613	5.513
R=29	S.D.	112	0.615	0.615	0.615	0.615	112	0.564	0.564	0.564	0.564	112	0.519	0.519	0.519	0.519	112	0.479	0.479	0.479	0.479	112	0.443	0.443	0.443	0.443
Slate (SLT)																										
S=30	Avg	2801	6.098	6.073	6.065	6.044	2807	6.172	6.127	6.111	6.070	2813	6.227	6.164	6.139	6.078	2818	6.268	6.190	6.148	6.073	2824	6.302	6.206	6.151	6.051
R=10	S.D.	28	0.131	0.131	0.131	0.131	28	0.124	0.124	0.124	0.124	28	0.117	0.117	0.117	0.117	28	0.110	0.110	0.110	0.110	28	0.103	0.103	0.103	0.103
Phyllite (PHY)																										
S=144	Avg	2728	6.105	6.080	6.073	6.052	2734	6.210	6.164	6.148	6.107	2740	6.260	6.197	6.172	6.111	2745	6.292	6.214	6.171	6.096	2751	6.316	6.220	6.165	6.065
R=48	S.D.	58	0.258	0.258	0.258	0.258	58	0.206	0.206	0.206	0.206	58	0.183	0.183	0.183	0.183	58	0.168	0.168	0.168	0.168	58	0.158	0.158	0.158	0.158
Zeolite Facies Basalt (BZE)																										
S=57	Avg	2916	6.277	6.253	6.245	6.224	2922	6.368	6.323	6.307	6.266	2927	6.425	6.363	6.337	6.277	2932	6.465	6.387	6.344	6.269	2937	6.495	6.399	6.344	6.244
R=19	S.D.	81	0.269	0.269	0.269	0.269	81	0.261	0.261	0.261	0.261	81	0.257	0.257	0.257	0.257	81	0.254	0.254	0.254	0.254	81	0.252	0.252	0.252	0.252

Christensen and Mooney, 1995, JGR, 100

Rocks' Density *vs* *T* and *P*

Variations in the density of minerals and rocks with *T* and *P* depend on the thermal expansion (α) and compressibility (β):

$$\alpha = \frac{\left(\frac{\partial V}{\partial T}\right)_P}{V_0} \qquad \beta = -\frac{\left(\frac{\partial V}{\partial P}\right)_T}{V_0} \qquad \alpha = -\frac{1}{\rho}\frac{\partial \rho}{\partial T} \qquad \beta = \frac{1}{\rho}\frac{\partial \rho}{\partial P}$$

α~3.5 x 10^{-5} K^{-1}

β~10^{-11} Pa

$$\frac{1}{K} = \frac{1}{\rho}\left(\frac{\partial \rho}{\partial P}\right)$$

K=bulk modulus (1/β)

$$d\rho(T,P) = \left(\frac{\partial \rho}{\partial T}dT + \frac{\partial \rho}{\partial P}dP\right) = \rho\left(-\alpha dT + \beta dP\right) \Rightarrow d\ln\rho = \beta dP - \alpha dT$$

- Density decreases with increasing *T* and increases with increasing *P*.
- In cases of constant *α* and β, integration of equations versus *T* and *P* gives:

$$\int_{\rho_0}^{\rho} d\ln\rho = \int_{P_0}^{P} \beta\, dP - \int_{T_0}^{T} \alpha\, dT \Rightarrow \rho = \rho_0 e^{\beta(P-P_0)-\alpha(T-T_0)}$$

$$\alpha_V = \frac{1}{V}\frac{\partial(V)}{\partial T} = \frac{1}{\frac{1}{\rho}}\frac{\partial\left(\frac{1}{\rho}\right)}{\partial T} = \rho\frac{\partial\left(\rho^{-1}\right)}{\partial T} = -\frac{\rho}{\rho^2}\frac{\partial(\rho)}{\partial T} = -\frac{1}{\rho}\frac{\partial(\rho)}{\partial T} = -\frac{\partial\ln(\rho)}{\partial T}$$

ρ_0= density of a given material at *reference* pressure P_0 (10^5Pa = 1 bar) and temperature T_0 (typically 298.15 K = 25 ◦C).

Since both $\alpha(T - T_0)$ and $\beta(P - P_0)$ are typically very small, the equations can be simplified, using the rules that $e^a \approx 1 + a$ and $e^{-a} \approx 1 - a$ when $a << 1$):

$$\rho = \rho_0\left[1 + \beta\left(P - P_0\right)\right] \times \left[1 - \alpha\left(T - T_0\right)\right]$$

or

$$\rho(P,T) = \rho(P_0,T_0)\ [1- \alpha_0(T-T_0) + \frac{(P-P_0)}{K}]$$

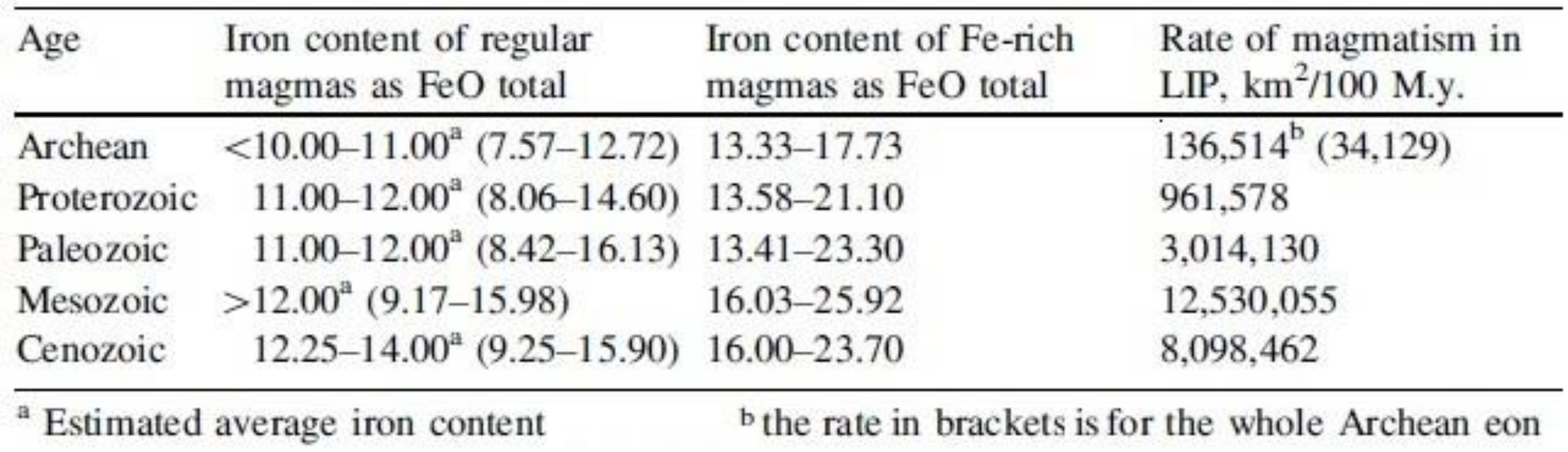
The equation is based on the assumption that *α* and β remain constant with *P* and *T*

Heat transfer in the crust

At the present-day the crust is the layer of the Earth that

(1) has an open surface for heat radiation and heat convection within the atmosphere and ocean.
(2) has sedimentary layer for heat convection by fluids within sediments.
(3) has direct contact with such cooling agents as the atmosphere and ocean.
(4) has the highest geothermal gradient, which gives importance to the heat transfer by conduction.
(5) has open faults and fracture systems for fluid and heat convection within the sediments.

- **During the Hadean and Early Archean the most effective kinds of heat transfer were: heat radiation from the surface of magma-ocean and convection within its layers, radiation by hot thick atmosphere and convection within it.**
- **The geothermal gradient was too low (about 0.3 $K\ km^{-1}$) to make effective heat transfer by conduction.**
- **The role of mass transfer by magmatic activity is increased with time (max in the Mesozoic), together with the iron content in magmas and average density of the lithosphere, which favored the conditions for the subduction process.**

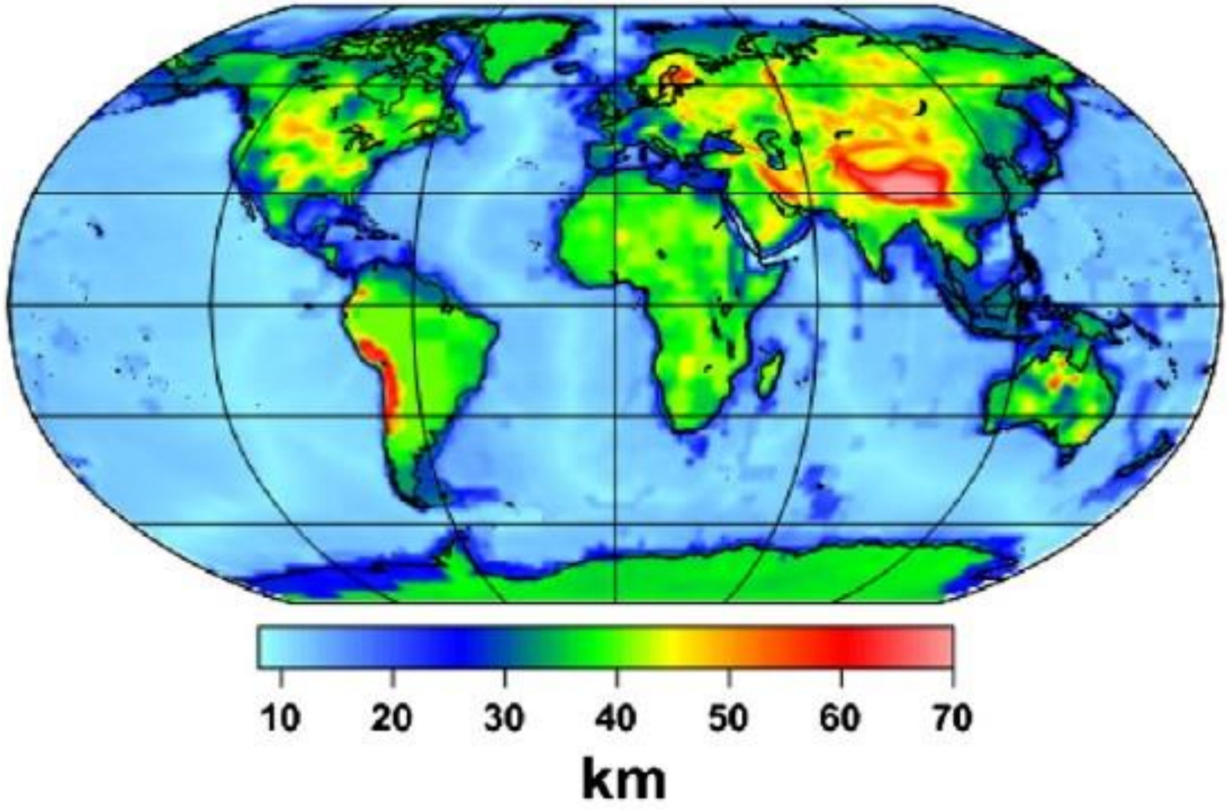


Age	Iron content of regular magmas as FeO total	Iron content of Fe-rich magmas as FeO total	Rate of magmatism in LIP, km^2/100 M.y.
Archean	<10.00–11.00[a] (7.57–12.72)	13.33–17.73	136,514[b] (34,129)
Proterozoic	11.00–12.00[a] (8.06–14.60)	13.58–21.10	961,578
Paleozoic	11.00–12.00[a] (8.42–16.13)	13.41–23.30	3,014,130
Mesozoic	>12.00[a] (9.17–15.98)	16.03–25.92	12,530,055
Cenozoic	12.25–14.00[a] (9.25–15.90)	16.00–23.70	8,098,462

[a] Estimated average iron content

[b] the rate in brackets is for the whole Archean eon

Moho Depth



Tesauro et al., 2012, Global and Planetary Change, 90-91

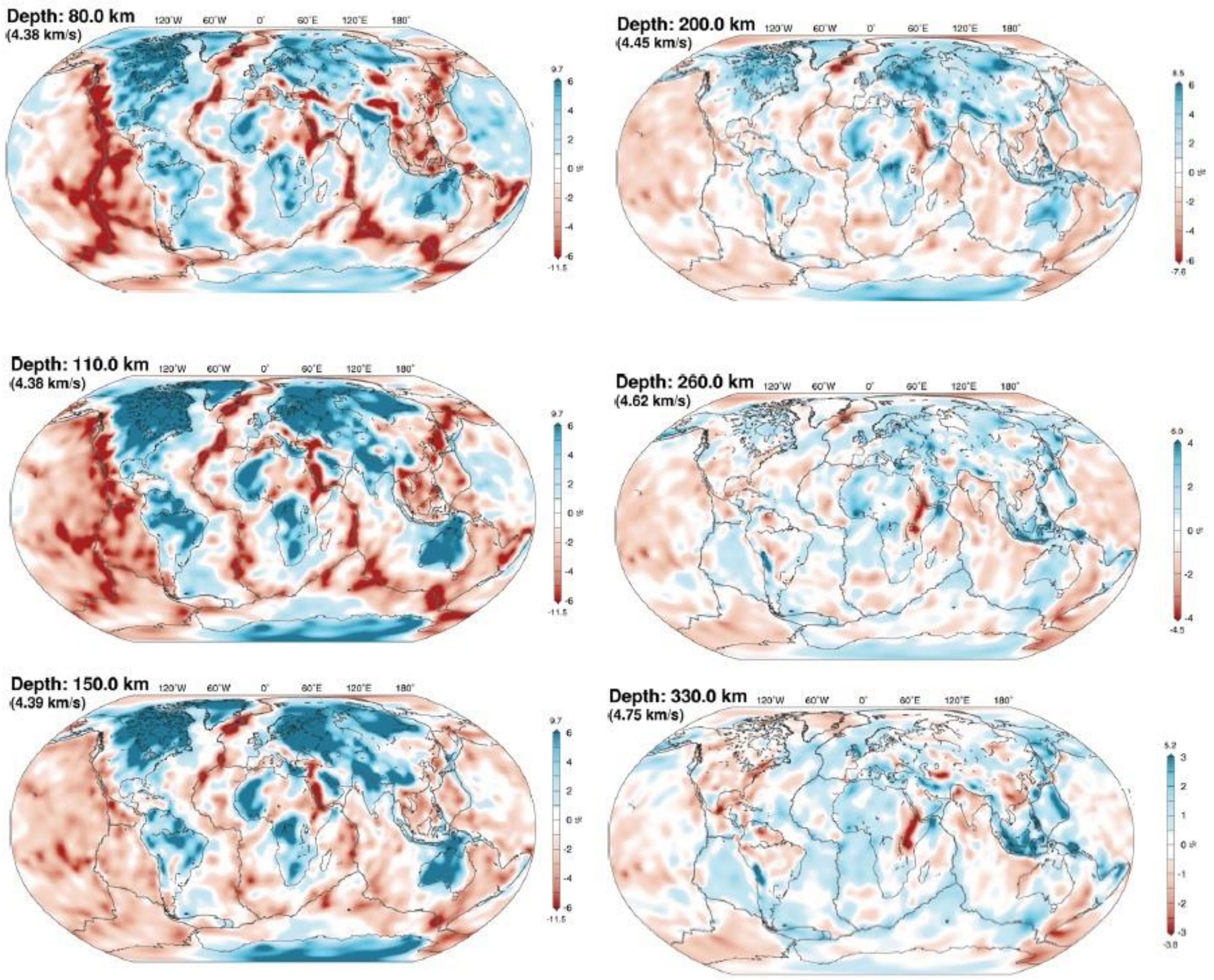
- Moho temperature in the continental areas depends on age, tectonic activity, thermal regime and the thickness of the crust.
- At a depth of ~40 km, Moho temperatures vary from 300-450 °C beneath the cratons, to about 800-1100 °C in thermally active areas (e.g., rift areas).

Upper mantle velocities

Seismic wave velocity is mainly a function of temperature:

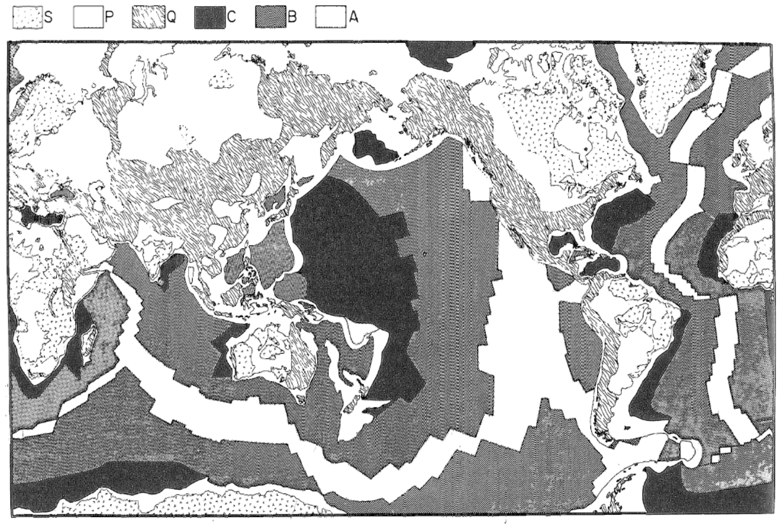
Warm → slower

Cold → faster



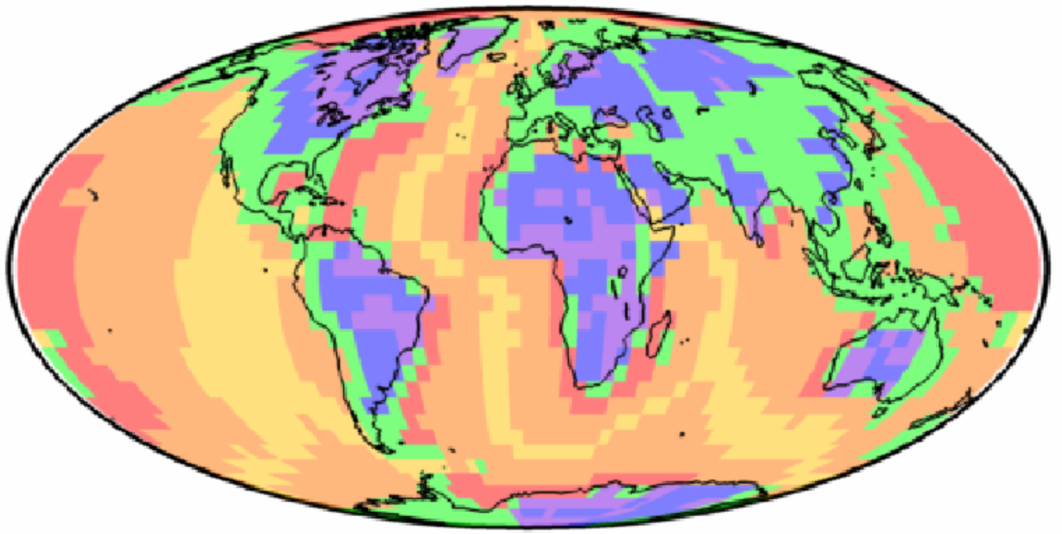
Scaheffer and Lebedev, 2013, Geophys. J. Int., 194

Global Tectonic Regionalization



GRT1 (Jordan, 1981)

- Three oceanic regions A, B, C
  - based on lithospheric age
- Three continental regions Q, P, S
  - based on generalized tectonic behavior during the Phanerozoic

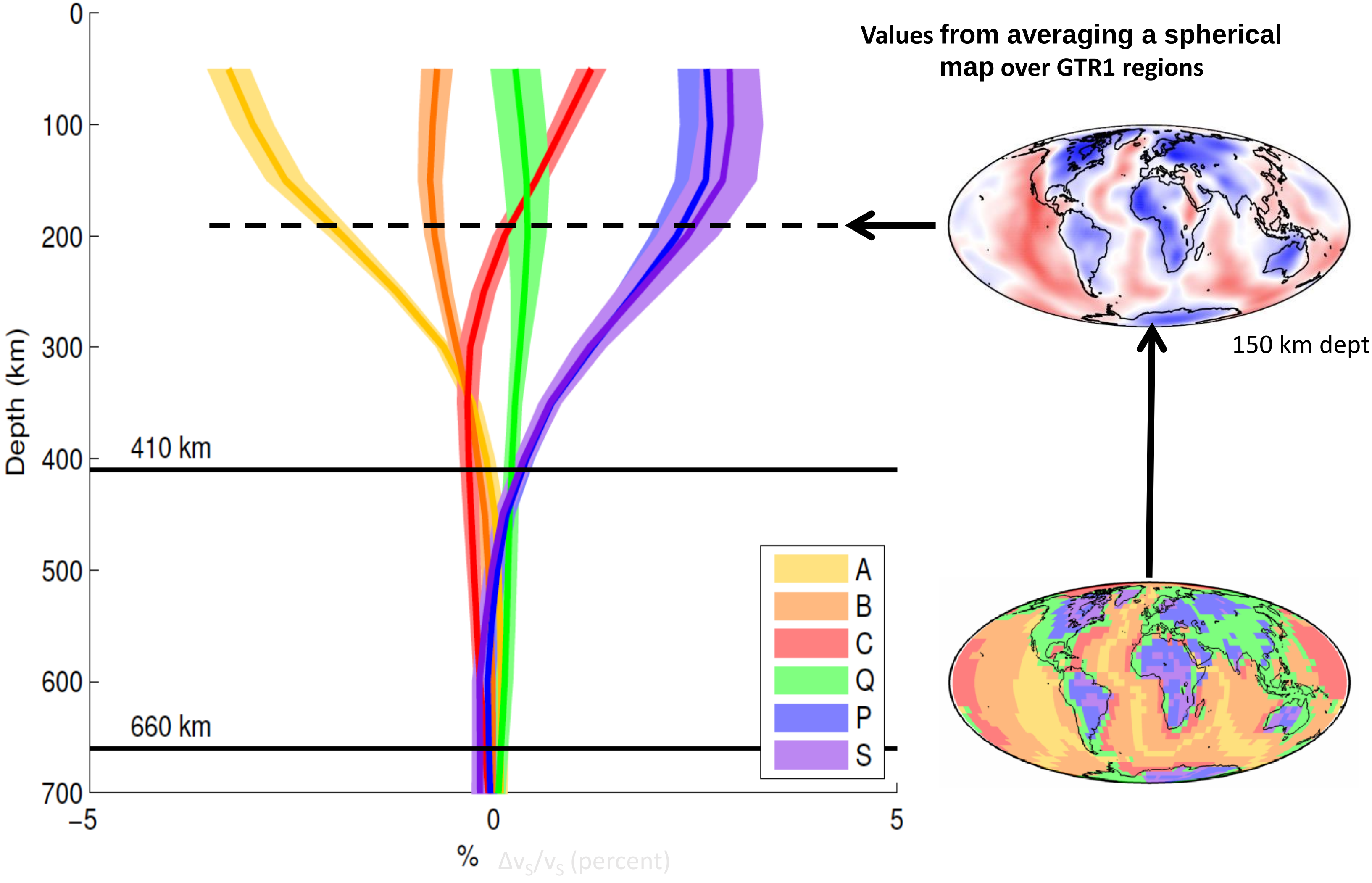


Precambrian Platform

Precambrian Shield

🟨 A- Young Oceans
🟧 B- Intermediate Oceans
🟥 C- Old Oceans
🟩 Q- Phanerozoic Orogenic Zones and Magmatic Belts
🟦 P - Precambrian Platform
🟪 S- Precambrian Shields and Platforms

Regional average seismic velocity

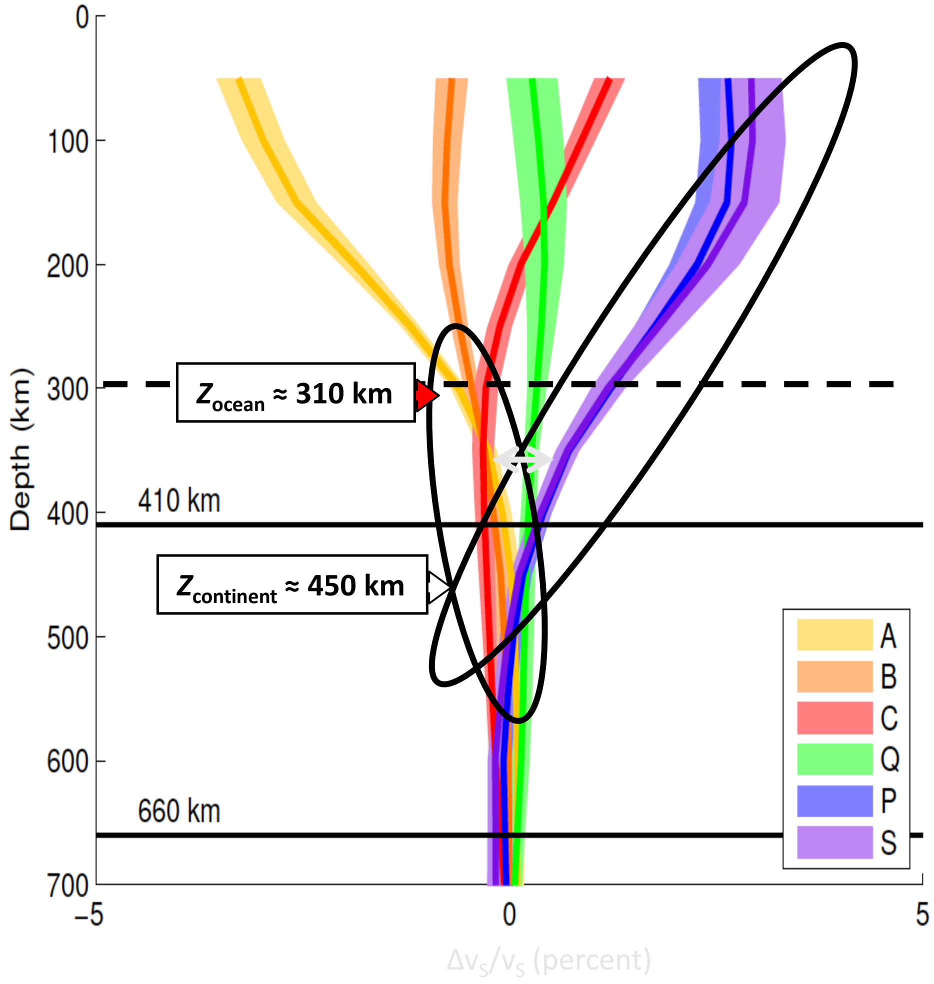


Values from averaging a spherical map over GTR1 regions

150 km depth

Model TX2008 (Simmons et al., 2009)

Regional average seismic velocity



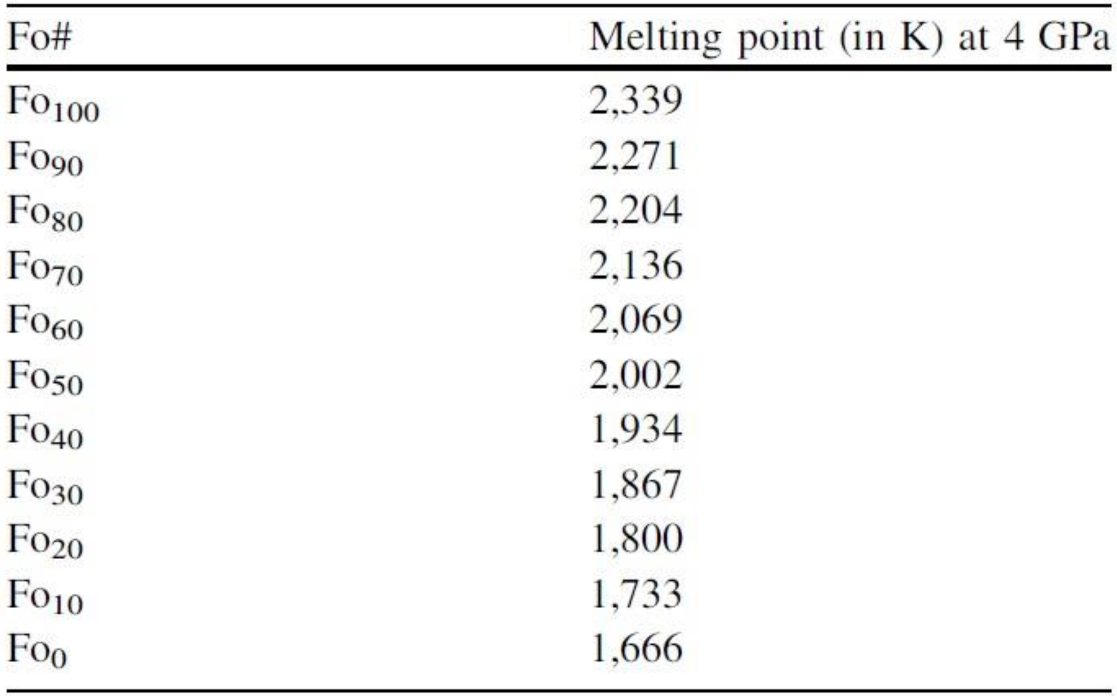
Model TX2008 (Simmons et al., 2009)

Common Features:

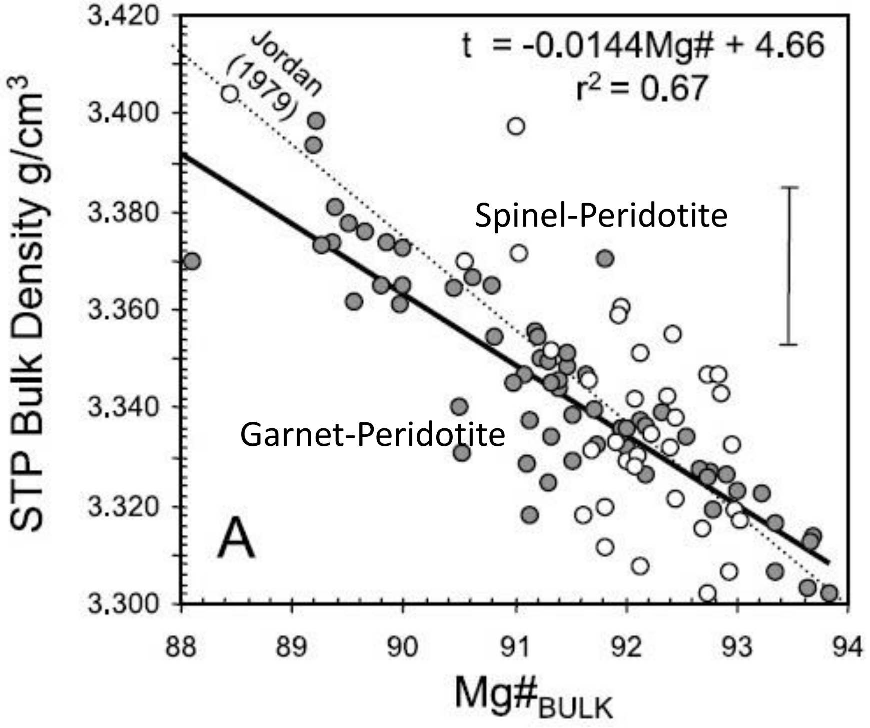
1. Platform and shield regions (P, S) show similar variations at all mantle depths
2. Ocean regions (A, B, C) show similar variations below 250-300 km
3. Differences between stable continents and oceans persist below 300 km

Thickness and composition of the cratonic lithosphere

- The thickness of the lithosphere increases in general with increases in age of the lithosphere and those younger magmatic rocks are usually richer in *Fe* than older rocks, because they erupted from greater depths.
- The melting point of rocks decreases with depth according to the increase of *Fe* content, as a result of the stratification of the magma-ocean. Then, the formation of partial melt should increase with depth below the forsterite layer.
- Peridotitic komatiites are restricted to the oldest cratons and were formed only in the Early Archean (~3.3 Ga) at temperatures ~2073 K, very close to the melting point of forsterite. These Mg-rich rocks, depleted in incompatible elements, had a very low density and were formed just below the forsterite layer, when the layers above had already solidified.



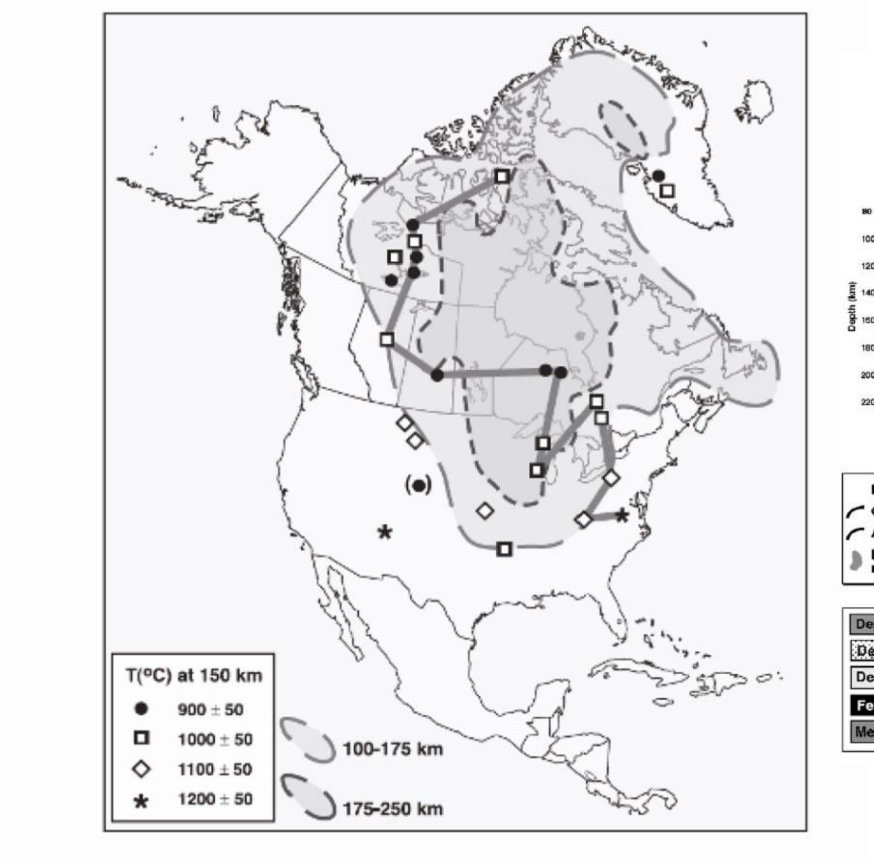
Fo#	Melting point (in K) at 4 GPa
Fo_{100}	2,339
Fo_{90}	2,271
Fo_{80}	2,204
Fo_{70}	2,136
Fo_{60}	2,069
Fo_{50}	2,002
Fo_{40}	1,934
Fo_{30}	1,867
Fo_{20}	1,800
Fo_{10}	1,733
Fo_{0}	1,666

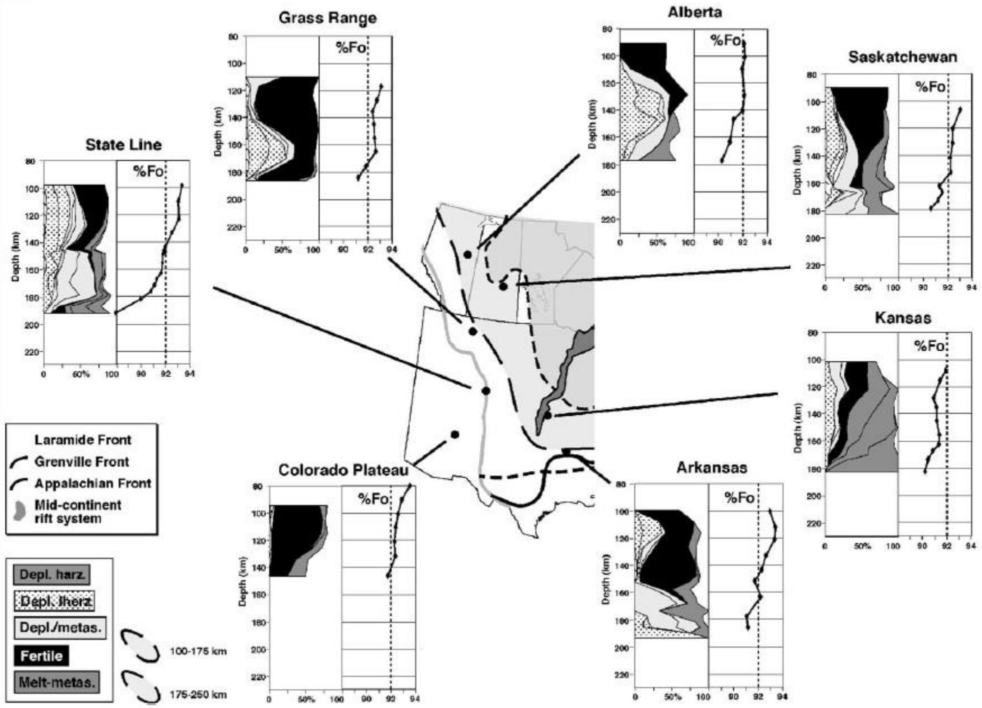


Lee, 2003, JGR, 108

Thickness and composition of the cratonic lithosphere

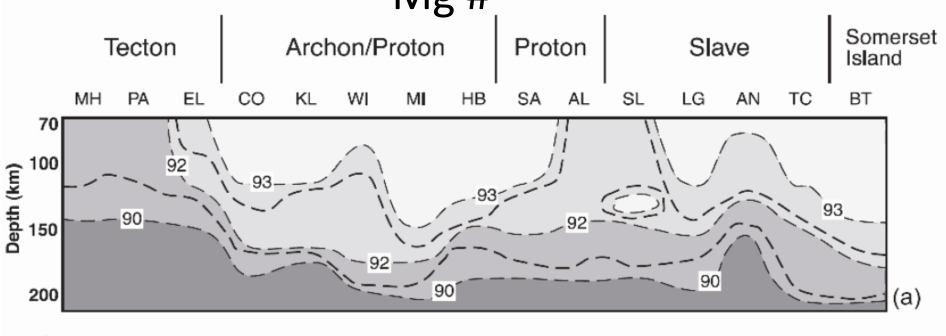
The lithosphere is *Fe* depleted up to a depth of 100-150 km





Griffin et al., 2003, Precambrian Research, 127

Mg #



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